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Clastic Sediment Partitioning in a Cretaceous Delta System, Western Canada: Responses to Tectonic and Sea-Level Controls

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Key words: Cretaceous, Cenomanian, Western Canada, Foreland basin, Deltas, Sea-level change, Tectonics, Eustasy, Clastic sedimentation, Sequence stratigraphy.

Abstract

The early–mid Cenomanian Dunvegan Formation represents a large delta complex that prograded at least 400 km from NW to SE. A regional stratigraphy based on marine transgressive surfaces and equivalent subaerial interfluves allows the formation to be subdivided into ten transgressive–regressive allomembers, labelled J to A in ascending order, each with an average duration of < 200 ky. Analysis of stacking patterns and facies distributions of parasequences within allomembers allows transgressive, highstand, falling stage and lowstand systems tracts to be identified. Extensive valley systems that average 1–2 km wide and 21 m deep can be traced for up to 320 km across the top surfaces of allomembers H to E. In their lower 20–40 km, valleys are filled with muddy heterolithic tidal facies but this changes to fluvial-dominated multi-storey channel-fills further up-valley. Interfluve surfaces are marked by palaeosols, the character of which indicate a protracted hiatus with extensive physical, chemical and biological modification of the parent material.

Changes in flexural subsidence rate are indicated by isopach patterns. Allomembers J–F have a sigmoidal prismatic geometry, successively offlapping to the SE. There is no evidence of thickening toward the orogen. In contrast, overlying allomembers E–A show progressive development of a depocentre along the western margin of the basin. The increasing accommodation rate on the updip coastal plain caused marine deltas to be starved of sediment, leading to progressive backstep of shorelines. Simultaneously, alluvial deposits within the depocentre show an upward increase in the proportion of subaqueous to subaerial facies, culminating in the incursion of brackish and finally marine waters. Thus tectonic subsidence rate had a first-order affect on both the volume of sediment available to build marine deltas and also on the local character of facies that accumulated on the coastal plain. The onset of flexural subsidence in allomember E appears to have resulted in subtle uplift of a forebulge, resulting in dramatic deflection of river systems.

Despite the clear tectonic signature, successive transgressions and regressions involved similar horizontal displacements of the shoreline, regardless of subsidence rate. This suggests that modest eustatic changes also influenced the accommodation available. Based on the measured horizontal excursions of the shoreline, the vertical thickness of alluvial strata, and realistic alluvial gradients, an average eustatic excursion of about 24 m is calculated. The incision of valley systems is attributed in part to periods of eustatic fall. However, valleys seem too long to be explained by eustasy alone, and hence secular changes in discharge are postulated as an additional forcing factor. Climatic cycles in the Milankovitch band may have been responsible for both eustatic and discharge variations.

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1. INTRODUCTION

Sedimentary successions record only three possible conditions: deposition, bypass or erosion. Various questions attend each option: if deposition took place, what were the processes, at what rate did they occur, and did the rates change over time? If non-deposition or bypass is indicated, then for how long, and why? If erosion occurred, what was the process, how much sediment was removed, and over what interval of time? These questions could be addressed in an 'ideal' stratigraphic succession that contained numerous, accurately-dated layers, and in which physical sequence-bounding surfaces and the main facies enclosed between those surfaces could everywhere be mapped. In reality, these requirements are rarely met in ancient sedimentary successions.

Perhaps one of the best opportunities to investigate the details of a basin's history is offered by the Mesozoic strata of the Western Canada Sedimentary Basin. This clastic-dominated basin extends northward into the NW Territories, and southward through the western United States to the Gulf of Mexico. The basin evolved from a divergent margin in the Palaeozoic to a retro-arc foreland basin in the Late Jurassic.

In Canada, the Palaeozoic and Mesozoic strata in this basin contain abundant oil and gas reserves which have been extensively investigated through several hundred thousand boreholes. Fair to excellent quality well logs are publicly available for every well, and abundant core has been cut from reservoir intervals. Although seismic sections are numerous, they are not publicly available. These rocks are also extensively exposed along the western margin of the basin in thrust sheets that form the Front Ranges and Foothills of the Rocky Mountains. Correlation of well logs in the basin to outcrop sections in the mountains not only allows logs to be calibrated against exposed lithologies, but also allows isolated outcrop sections to be placed in a regional stratigraphic and palaeogeographic framework. The main deficiency of the stratigraphic record is the relative paucity of well-dated stratigraphic marker horizons, although marine strata yield fossils that can, in the Cretaceous part of the section, provide relative age resolution of about 0.5 Ma.

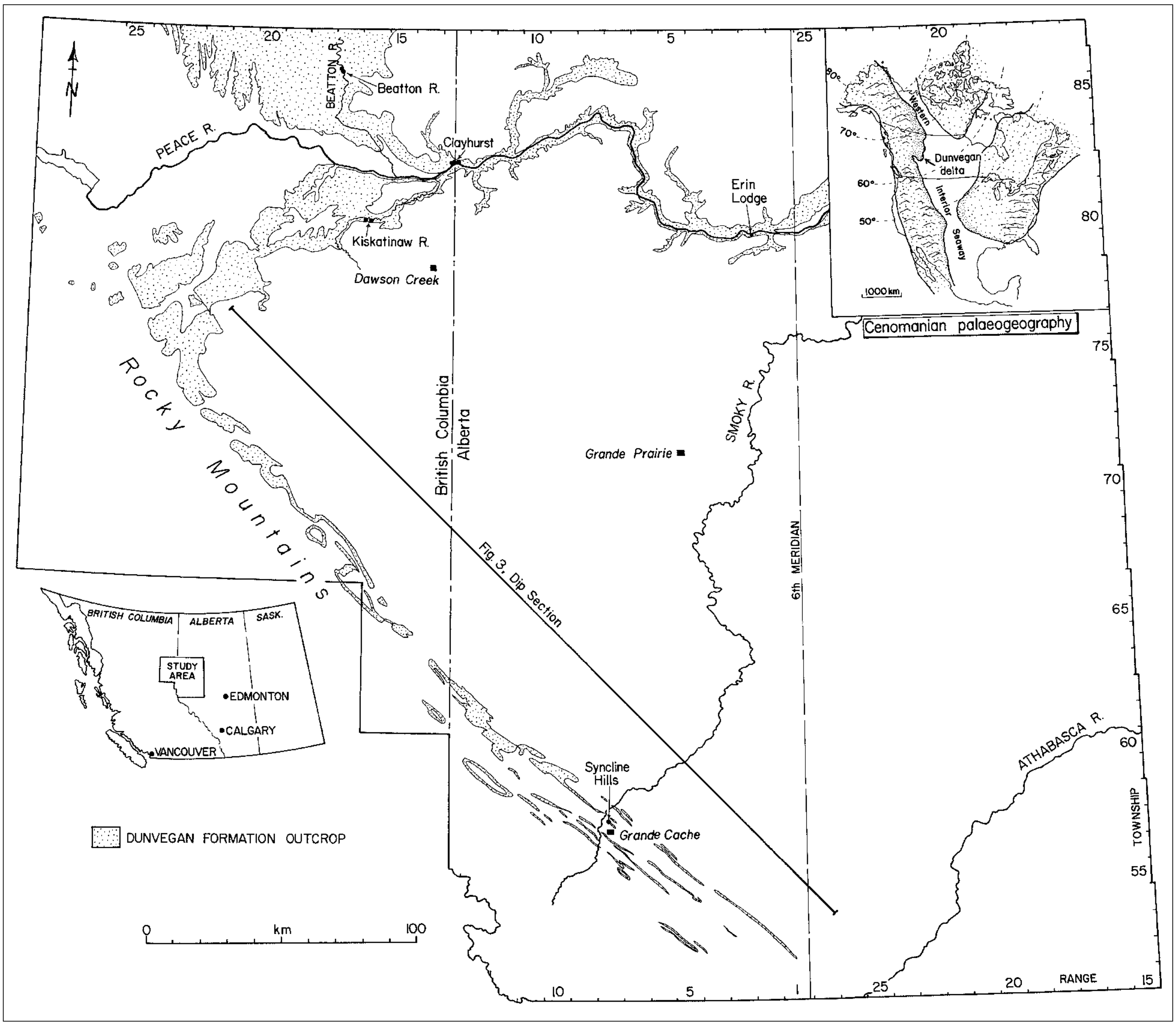


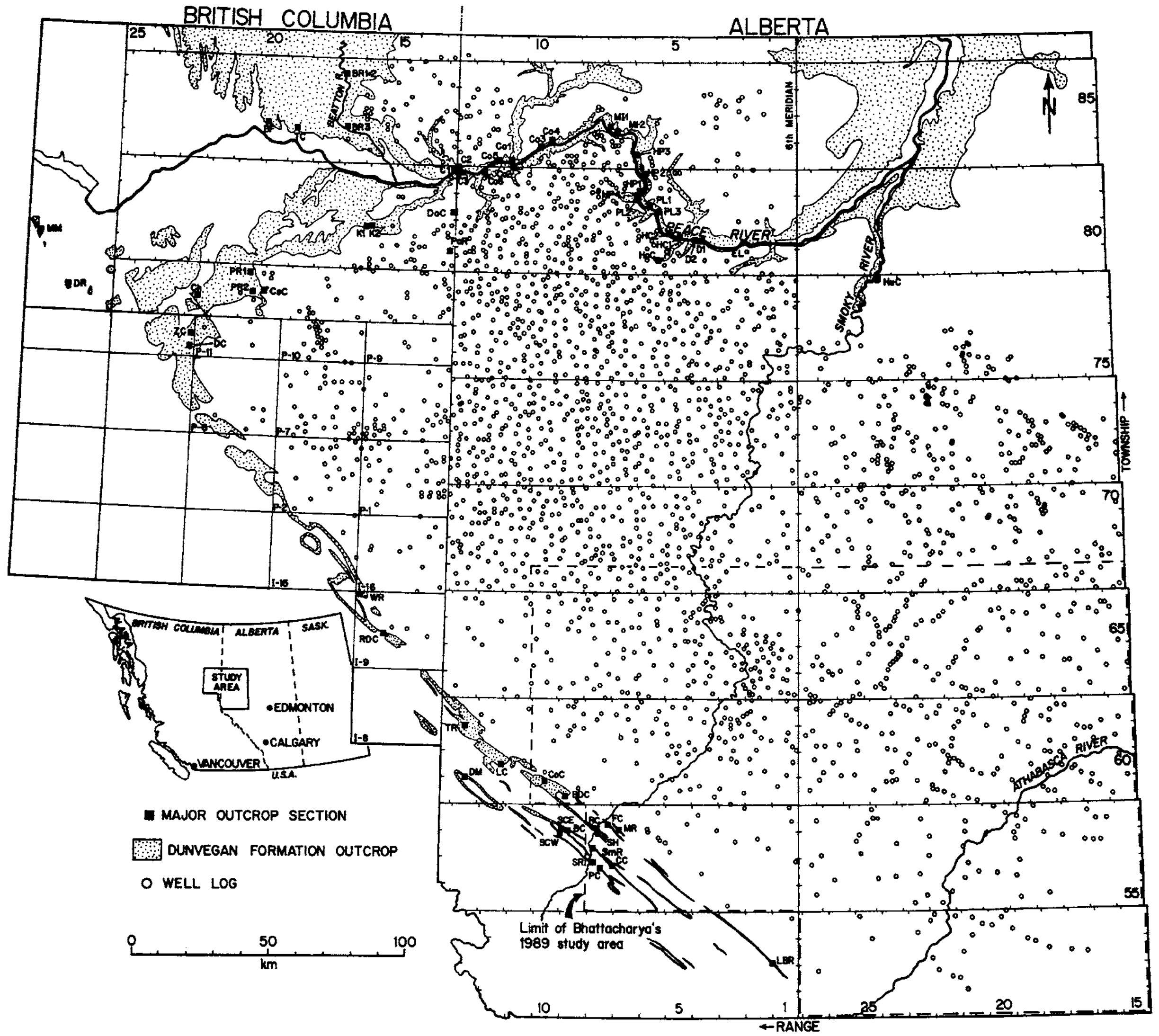
Fig. 1 Inset maps show location of study area in western Canada, and Cenomanian palaeogeography of the Dunvegan Formation within the Western Interior Seaway. The line of section shows the location of the dip section shown schematically in Fig. 3, and drawn to scale in Fig. 23 (from PLINT & WADSWORTH, in review).

This paper is concerned with the Dunvegan Formation, which is an early to mid-Cenomanian deltaic unit. This 90–270 m thick clastic wedge was deposited on the western margin of the Western Interior Seaway (Fig. 1), during a period of generally rapid subsidence of the North American plate resulting from dynamic and static loading mechanisms due to subduction and crustal shortening (MITROVICA et al., 1989). At least 400 km of NW to SE delta progradation can be demonstrated within the study area, which took place over about 2 My. The Dunvegan Fm. is extensively exposed in thrust sheets for several hundred km along the NW–SE trending Rocky Mountain Foothills, and also for 300 km in undeformed strata exposed in an E–W transect along the Peace River valley (Fig. 1). Over a study area of about 80,000 km^2, the stratigraphy of the Dunvegan Fm. was delineated using 2340 well logs and 60 major outcrop sections (PLINT, 2000; Fig. 2).

1.1. Aim of this paper

This paper summarizes the main results from previous studies concerned with various aspects of the Dunvegan Formation (PLINT, 1996, 2000, 2002; PLINT et al., 2001; McCARTHY & PLINT, 1998, 1999, in review; McCARTHY et al., 1999; McCARTHY, 2002; LUMSDON & PLINT, 2003 – in press; PLINT & WADSWORTH, in review). The analysis of the depositional history of the Dunvegan has involved several main steps.

1) Development of a robust allostratigraphy based on marine transgressive surfaces. This allows lateral facies relationships to be determined, and also permits isopach maps to be constructed. It is assumed that the bounding surfaces of allostratigraphic units approximate time lines. Unfortunately, there is no means available to determine absolute ages for the bounding surfaces.



OUTCROP LOCALITY ABBREVIATIONS

BC	Bergamot Creek	DC	Dickebusch Creek	K	Kiskatinaw River	RC	Railway Cutting
BDC	Beaverdam Creek	DM	Deadhorse Meadows	LBR	Little Berland River	RDC	Red Deer Creek
BF	Bear Flat	DoC	Doe Creek	LC	Lynx Creek	SCE	Sheep Creek, east
BR	Beatton River	DR	Dokie Ridge	MI	Many Islands	SCW	Sheep Creek, west
C	Clayhurst	EL	Erin Lodge	MM	Mount McAllister	SH	Syncline Hills
CC	Cowlick Creek	FC	Flood Creek	MR	Muskeg River	SmR	Smoky River
Ch	Chetwynd road cut	HC	Hines Creek	PC	Pearl Creek	SR	Sulphur River
Co	Cotillion	HoC	Howard Creek	PcR	Pouce Coupe River	TC	Tea Creek
CoC	Copton Creek	HP	Highland Park	PL	Pratts Landing	TR	Torrens River
CsC	Coldstream Creek	HuC	Hunting Creek	PR	Pine River	WR	Wapiti River
D	Dunvegan					ZC	Zonnebeke Creek

Fig. 2 Map of study area showing stratigraphic data base, including location of 2340 well logs and principal outcrop sections in the Foothills and Peace River valley (from PLINT, 2000).

2) Construction of palaeogeographic maps for each allomember showing maximum marine transgressive and regressive limits. This provides a picture of the regional shoreline trends and also permits quantification of horizontal movements of the shoreline.
3) Construction of isopach maps for each allomember. These show the geometry of the accommodation available for each allomember. Because each allomember aggraded essentially to sea level, isopach patterns can be interpreted in terms of subsidence patterns.
4) The analysis of sedimentary facies within a palaeogeographic and accommodation framework allows vertical and lateral facies changes to be interpreted in terms of proximity to the shoreline and to spatial variation in accommodation rate.
5) Mapping of palaeo-valley systems in both subsurface and outcrop. The resulting maps reveal the influence of various tectonic features (faults, forebulge) on drainage systems.
6) Analysis of palaeosols mantling interfluve surfaces between valley systems. This reveals both vertical and lateral variations in the character and evolution of palaeosols that reflect palaeodrainage conditions and rates of clastic flux related to distance from the margins of valleys.

7) Sequence stratigraphic interpretation, including definition of systems tracts on the basis of facies associations, stratal lap-out patterns, palaeogeography and palaeogeomorphology. The main aim is to interpret the principal controls (tectonic, eustatic, climate) on changing patterns of deposition and erosion, utilizing elements 1–6 above.

1.2. Collaboration

This paper presents a compilation of results from the research of the author on the regional stratigraphy and sedimentology of the Dunvegan Fm. However, very significant contributions to this understanding have been made by my post-doctoral fellows and graduate students whom I wish to acknowledge here. Dr. Jennifer WADSWORTH investigated the sedimentology of the valley-fill deposits exposed along the Peace River, was the first to recognize tidal features, and made preliminary maps of their subsurface distribution. Dr. Paul McCARTHY undertook palaeopedological investigations of the interfluve surfaces between the valleys, and confirmed the existence of major hiatal surfaces. Bira FACCINI spent part of his PhD project investigating the sedimentology and stratigraphy of the Dunvegan strata in the valley of the Kiskatinaw River, and documented the vertical changes in alluvial facies in response to changes in accommodation rate. Matthew LUMSDON conducted detailed facies analysis of alluvial strata near the orogenic margin of the basin and showed how drainage conditions and facies successions on the alluvial plain were strongly linked to regional subsidence rates and patterns. The work of these colleagues is indicated at appropriate points by references to our publications.

2. REGIONAL GEOLOGICAL SETTING

During the Cretaceous, North America was progressively flooded from the south by warm saline waters of the ancestral Gulf of Mexico – Tethys, and from the north by cooler, less saline water of the Boreal Ocean. These marine embayments merged just before the Albian–Cenomanian boundary (96 Ma) to form the Western Interior Seaway, which continued to widen until mid-Turonian time (90 Ma) (WILLIAMS & STELCK, 1975; KAUFFMAN & CALDWELL, 1993; HAY et al., 1993; Fig. 1, inset). Static and dynamic loading of the lithosphere above the subducting Farallon Plate produced a wide sedimentary basin comprising two main parts. Static loading by the orogenic wedge generated a relatively narrow (<300 km), rapidly-subsiding flexural foredeep, to the east of which lay a broader region, 1000–1500 km wide, of moderate subsidence that reflected dynamic loading by large-scale mantle flow (MITROVICA et al., 1989; PANG & NUMMEDAL, 1995; CATUNEANU et al., 1997). Thus, the Interior Seaway can be attributed, in part, to static and dynamic loading of the North American Plate. However, eustatic rise, broadly ascribed to accelerated plate spreading rates (KAUFFMAN & CALDWELL, 1993), was also a factor with Turonian sea level estimated to have peaked at 180 m (SAHAGIAN & JONES, 1993) to 300 m (McDONOUGH & CROSS, 1991), above present level.

Relative sea level rise at the Albian–Cenomanian boundary led to widespread deposition of marine shales, marking the onset of the Greenhorn Cycle (KAUFFMAN & CALDWELL, 1993). In Alberta and British Columbia, this transgressive phase was accompanied by rapid flexural subsidence, recorded by a westward-thickening wedge of laminated, organic-rich marine shale of the Shaftesbury Formation (equivalent to the Hasler, Goodrich and Cruiser formations in the NW part of the basin in British Columbia; STOTT, 1982; Fig. 3). The Shaftesbury Formation coarsens upward into deltaic strata of the Early to Middle Cenomanian Dunvegan Formation which represents a major progradation of the shoreline. However, unlike the fluvio-deltaic Mountain Park Fm. strata below the Shaftesbury Formation, which prograded towards the north and NW, progradation of the Dunvegan Fm. was in the opposite direction, towards the SE, parallel to the Cordillera (Fig. 3). This change in regional sediment dispersal direction probably records the northward migration of the centre of maximum Cordilleran uplift as a result of the lateral migration of accreted terranes along the continental margin during the Cretaceous (EISBACHER, 1981; POULTON, 1994). The principal source of sediment for the Dunvegan Formation is considered to have lain in northern British Columbia and southern Yukon, where the formation is dominated by alluvial sandstones and conglomerates (STOTT, 1982). Palaeocurrent observations (PLINT, unpubl. data) and mapping of valley systems and shorelines (PLINT, 1996, 2000, 2002; PLINT & WADSWORTH, in review) show that the principal fluvial palaeoflow direction was toward the SE, although several subordinate sources of sediment were scattered along the Cordilleran margin of the basin further to the south.

Isopach maps of successive allomembers of the Dunvegan Fm. (discussed in more detail below), show that flexural subsidence of the foredeep was minimal during deposition of the older five allomembers, J–F, each of which constitute broadly prismatic packages of sediment that show no significant thickening towards the orogen. In contrast, the younger five allomembers, E–A show major thickening towards the orogen, indicating the onset of renewed subsidence during and following allomember E time (PLINT et al., 2001, LUMSDON & PLINT, 2003 – in press). This phase of subsidence led to major backstep of the Dunvegan delta complex in allomember A time as accommodation rate gradually exceeded sedimentation rate, culminating in widespread marine transgression at the base of the overlying Kaskapau Fm. (Fig. 3).

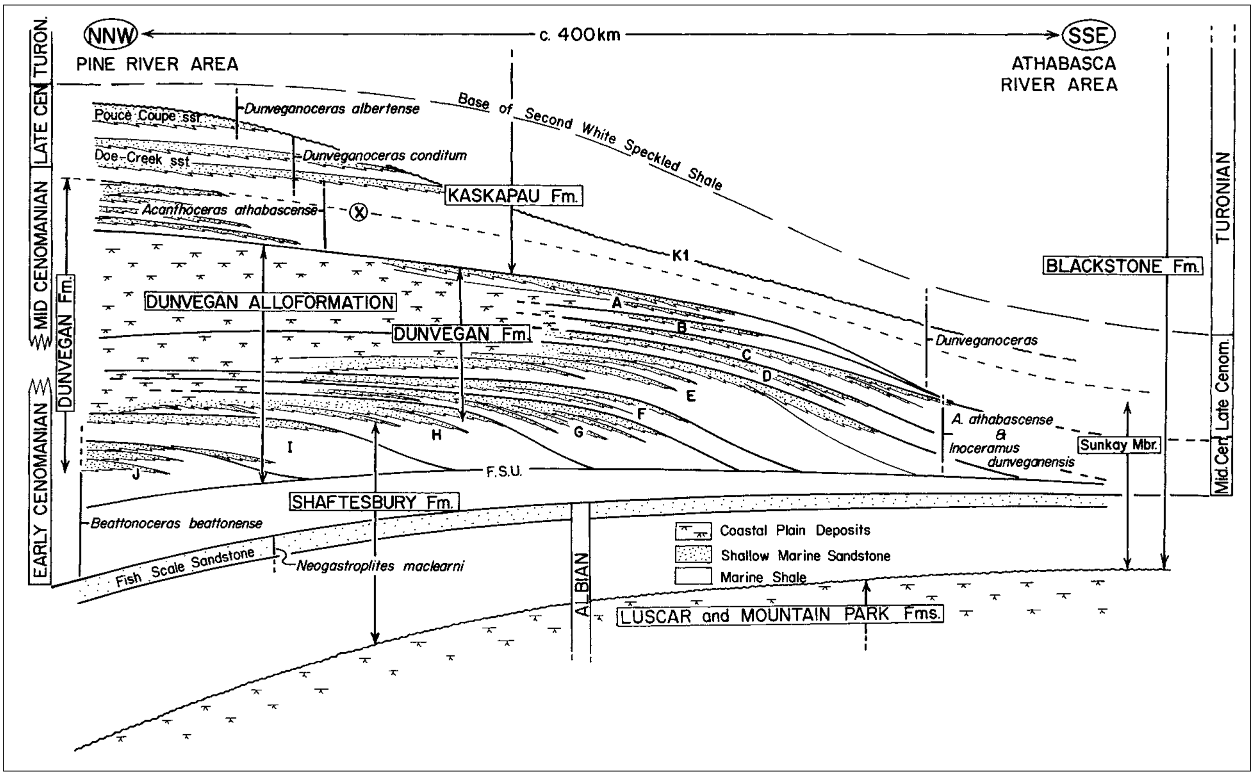


Fig. 3 Schematic NW–SE dip section showing the Dunvegan Fm., the stratigraphic terminology applied to the under- and overlying units, and the stratigraphic range of important Zone fossils. The boundaries of the lithostratigraphic Dunvegan Fm., and the allostratigraphic Dunvegan alloformation are also indicated. The former is very diachronous (from PLINT et al., 2001).

3. STRATIGRAPHY

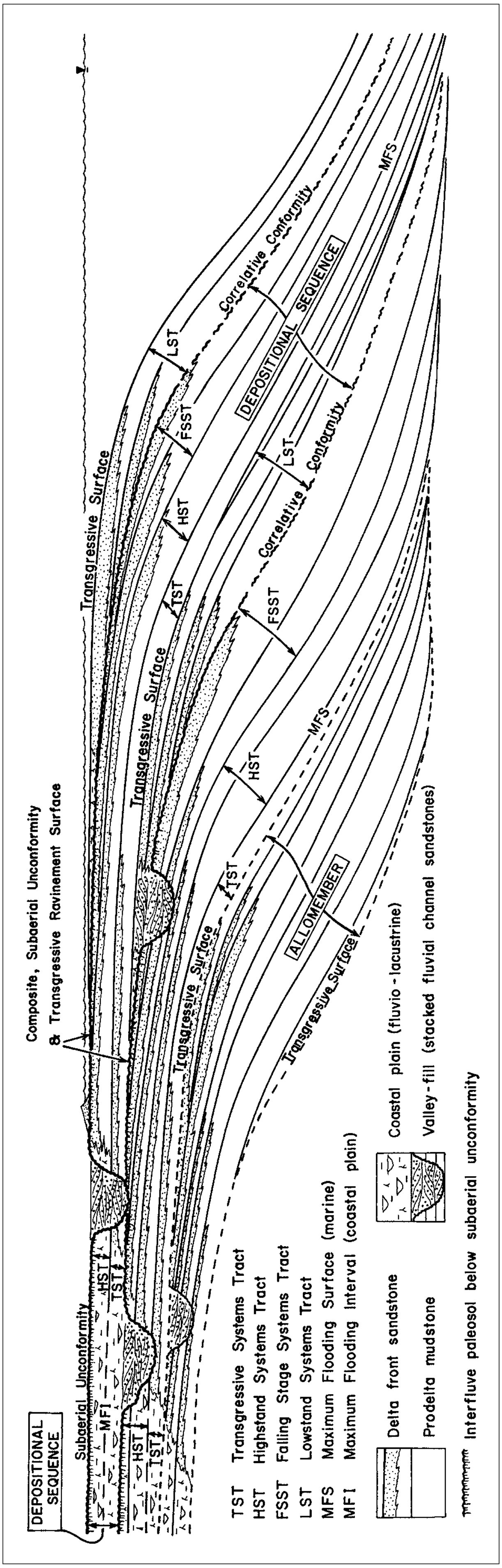
3.1. Lithostratigraphy

The Dunvegan Formation ranges from 90–270 m thick and crops out extensively on thrust sheets in the Alberta and British Columbia Foothills, and also in the Peace River Valley, where undeformed strata are exposed in the vicinity of Fort St. John, and extend for over 300 km to the east (STOTT, 1982; Fig. 2). The Dunvegan Formation comprises a succession of alluvial and shallow marine sandstones, siltstones and shales that separate underlying marine shales of the Shaftesbury Formation from overlying marine shales and minor sandstones of the Kaskapau Formation (Fig. 3). It has long been recognised that the Dunvegan Formation has an interfingering, diachronous relationship with the underlying Shaftesbury, and overlying Kaskapau formations (STELCK & WALL, 1955; STELCK et al., 1958; ALBERTA SOCIETY of PETROLEUM GEOLOGISTS, 1960; SINGH, 1983), and the boundaries of the Dunvegan Formation are placed at the first and last appearance of sandstone in an otherwise shale-dominated succession. For mapping purposes, this definition serves well. However, in order to analyze in more detail the depositional history and palaeogeography of the Shaftesbury, Dunvegan and Kaskapau formations, it is useful to subdivide the rocks into smaller, genetically-related packages on the basis of mappable surfaces. Such an *allostratigraphic* approach was introduced by BHATTACHARYA (1989).

3.2. Allostratigraphy

Allostratigraphy involves the subdivision of the stratigraphic record into packages "defined and identified on the basis of their bounding discontinuities" (NORTH AMERICAN COMMISSION ON STRATIGRAPHIC NOMENCLATURE, 1983). Bounding discontinuities can take several forms, including subaerial erosion surfaces, the upper surfaces of palaeosols, erosional marine ravinement surfaces, and weakly-erosional to non-erosional marine flooding surfaces (Fig. 4).

In a subsurface study, BHATTACHARYA (1989) demonstrated that the Dunvegan Formation could be subdivided into seven transgressive–regressive packages, bounded by regional marine flooding surfaces. The seven packages, or allomembers, were designated G to A in ascending order, and each was shown to consist of several component, sandier-upward successions, termed 'shingles' (broadly equivalent to parasequences; BHATTACHARYA, 1989; BHATTACHARYA & WALKER, 1991a). Analysis of facies successions in core led BHATTACHARYA & WALKER (1991b) to interpret the Dunvegan Formation in terms of a series of prograding and aggrading delta complexes. Because of their regional extent, allomembers were interpreted to record an allogenic, tectonic control on subsidence rate.



In contrast, the more localised distribution of individual shingles suggested an autogenic origin, controlled by delta distributary switching (BHATTACHARYA & WALKER, 1991a).

PLINT (1996, 2000) extended the allostratigraphic scheme of BHATTACHARYA & WALKER (1991a), tracing the Dunvegan northward from township 67 to correlate with outcrop sections on the Peace and Beatton rivers. An additional three allomembers, J–H were added to the original scheme of BHATTACHARYA (1989), and it is likely that further north, yet more allomembers could be defined, but which downlap and pinch out before reaching the study area.

3.3. Recognition of depositional sequences

It can be shown that the marine ravinement and flooding surfaces that define allomembers in more downdip areas can be traced updip into surfaces bounding well-developed interfluve palaeosols (McCARTHY & PLINT, 1998; McCARTHY et al., 1999; PLINT et al., 2001). These palaeosol-bounding surfaces can in turn be traced into erosional surfaces that define a network of valleys incised into the upper surfaces of most of the Dunvegan allomembers (Fig. 4). Therefore, each component allomember of the Dunvegan Formation preserves evidence for relative sea level *rise,* expressed by sandstones filling valleys, by the abrupt juxtaposition of offshore marine or lagoonal mudstone above sandier, shallower-water deposits separated by a flooding or ravinement surface, and by changes in fluvial style in coastal plain areas (McCARTHY et al., 1999). Evidence for relative sea level *fall* and fluvial incision is provided by extensive valley systems (PLINT, 2002; PLINT & WADSWORTH, in review), interfluve palaeosols (McCARTHY et al., 1999), and, in shallow marine areas, by erosive-based delta-front sandstones (PLINT, 1988, 1996; PLINT & NUMMEDAL, 2000).

Of the ten Dunvegan allomembers studied, only allomembers J and E show a distinct downward shift of facies and onlap in the delta front area, as is required for the definition of a classical Exxon type 1 sequence (VAN WAGONER et al., 1988). The remaining allomembers show well-developed offlap to subtle onlap of

Fig. 4 Diagram showing how allomembers and depositional sequences are defined. Allomembers are bounded by regional marine transgressive surfaces (heavy dashed lines) and their lacustrine equivalents on the coastal plain. Valley fills are included with the underlying allomember. Depositional sequences are bounded on the coastal plain by a subaerial unconformity that defines valleys and interfluves. In downdip areas, this surface continues as the upper surface of the offlapping, falling stage systems tract (which may be a palaeosol, or a ravinement surface), and, basinward of the shoreline, as the correlative conformity. This surface is onlapped by valley-filling and marine deposits of the lowstand and transgressive systems tracts. Recognition of the four marine systems tracts relies heavily on regional mapping of shingle stacking patterns (from PLINT et al., 2001).

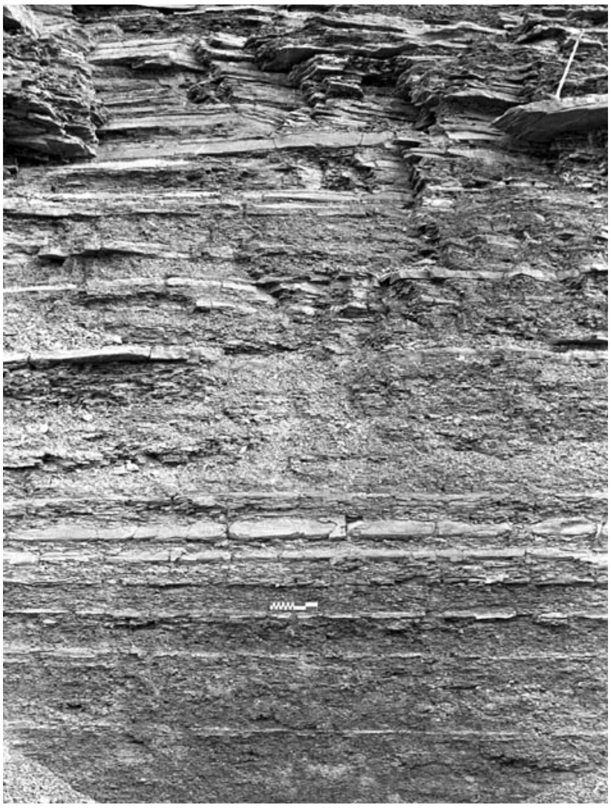


Fig. 5 Detail of prodelta facies consisting of cm-scale interbeds of very fine sandstone and mudstone. Sandstone beds are sharp-based with small grooves and flutes, and upper surfaces are typically ornamented with combined-flow ripples. Beatton River section 1; scale bar = 20 cm.

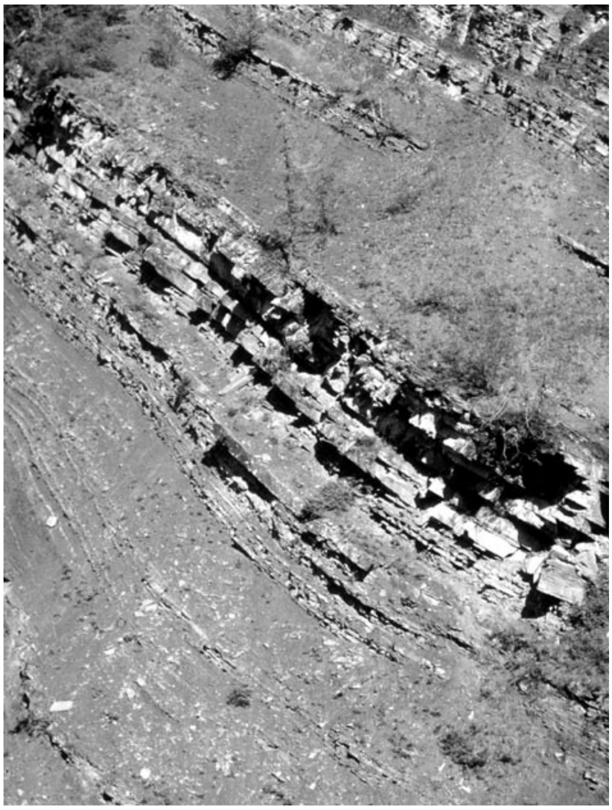


Fig. 6 Three successions of prodelta storm beds grading up into thicker sandstones of the delta front. The main package of sandstone is 4 m thick and is capped by a coal. Flood Creek, Alberta.

parasequences in the delta front area and represent both falling stage and lowstand systems tracts, *sensu* PLINT & NUMMEDAL (2000), and PLINT et al. (2001).

4. SEDIMENTARY ENVIRONMENTS AND FACIES

The Dunvegan Fm. can be subdivided into three main depositional environments: prodelta, delta-front and coastal plain (including valley-fills). The main sedimentary characteristics of each of these environments are summarized below.

4.1. Prodelta

Prodelta deposits consist mainly of cm to mm-bedded dark grey siltstones and mudstones. Interbedded very fine sandstones are mm to cm thick and become thicker and more abundant upward (Fig. 5). Thicker sandstone interbeds are parallel-laminated or show wave or combined-flow ripples. Rare, cm- to dm-scale gutter casts may be present on the base of sandstone beds. In the more offshore facies, gutter casts are oriented parallel to shore, but in more nearshore deposits, they rotate to shore-perpendicular. Bioturbation is generally sparse, and includes *Chondrites*, *Planolites*, *Teichichnus*, and *Zoophycos*. These sediments are organized in 5–20 m thick, sandier-upward successions interpreted to record progradation of individual delta lobes (Fig. 6).

4.2. Delta-front

Delta-front deposits are dominantly fine- to locally medium-grained sandstone and form units from 1 to about 10 m thick. The bases of sandstone bodies may be either gradational with underlying prodelta facies (Fig. 7), or sharp and erosive, commonly with spectacular gutter casts (Figs. 8, 9). Swaley cross-stratification and parallel lamination typify fine sandstone whereas dm-scale trough crossbedding is present in the upper fine- to lower medium-grained sandstones. The delta-front sandstone bodies are commonly capped by 1–2 m of planar-laminated to massive rooted sandstone interpreted as beach and backshore deposits. Sharp-based delta front sandstones are widely-recognized throughout the formation and are interpreted to record shoreface progradation during relative sea level fall (i.e. during the falling stage systems tract; PLINT & NUMMEDAL, 2000).

Delta-front sandstones can be mapped in well-logs, and their maximum progradational extent in each allomember can be traced. For each allomember, shorelines face consistently towards the E to SE and

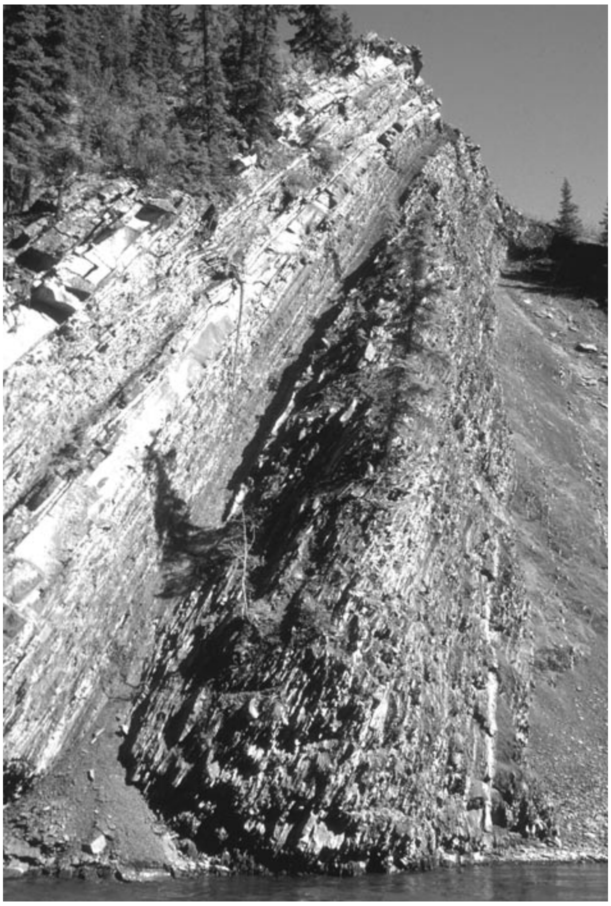


Fig. 7 Stacked prodelta to delta-front facies successions. The middle succession is 4 m thick; Sulphur River, Alberta.

Fig. 8 Delta front sandstone unit resting erosively on dark laminated prodelta mudstone. Sandstone is 7 m thick and is sharply overlain by laminated dark marine mudstone; person for scale. Railway cutting section, Grande Cache, Alberta.

can be traced continuously for several hundred km. The palaeogeographic maps of allomembers J to A (Figs. 10–16) show that the delta shoreline was broadly cuspate to lobate, suggestive of moderate to strong wave reworking of individual delta lobes. Large bulbous delta lobes up to 80 km wide and 100 km long are recognized in allomembers I, D, E, and F, and are interpreted as lowstand deltas deposited at the end of periods of slow relative sea level fall.

Fig. 9 Detail of large canoe-shaped gutter casts on the base of the delta front sandstone shown in Fig. 8. Gutter cast is 80 cm. wide. Railway Cutting section, Grande Cache, Alberta.

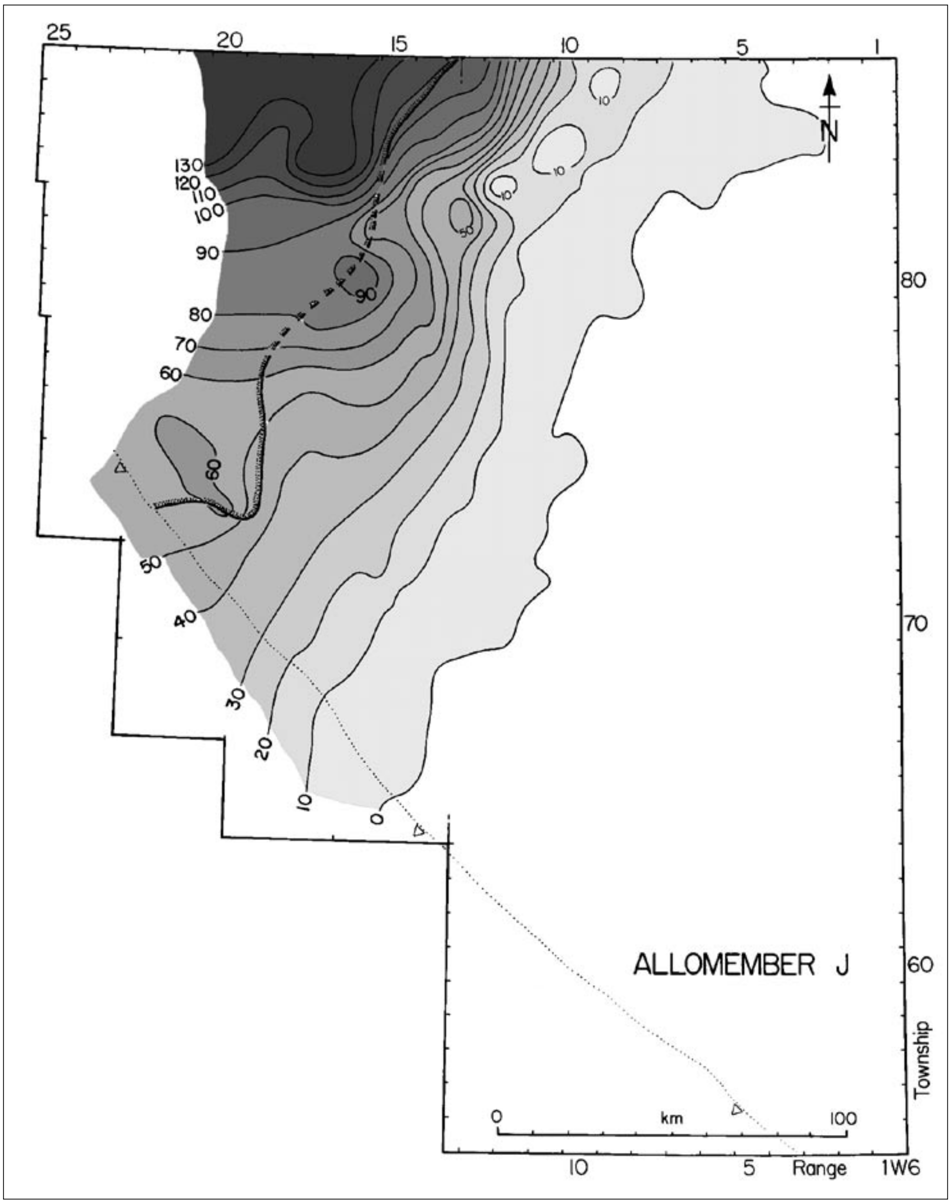


Fig. 10 Palaeogeographic and isopach map of Dunvegan allomember J. Heavy stippled line in this and figures 11–16 indicate the maximum progradational extent of the delta-front sandstone, mapped from gamma ray log signatures. Note how isopach lines trend perpendicular to the orogen in this map and in Figs. 11–14. This pattern is interpreted to reflect deposition at a time when the rate of flexural subsidence was negligible. Contour interval in figures 10–16 = 10 m.

4.3. Coastal Plain

Coastal plain deposits dominate the NW portion of the delta complex and comprise a complex array of facies broadly representative of a generally poorly-drained environment, traversed by numerous non-migrating river channels that formed ribbon sandstones in a matrix of lacustrine, marsh and poorly-drained floodplain mudstone, interstratified with crevasse splay and thin lacustrine delta sandstones. The abundance of tracks shows that the area was inhabited by numerous dinosaurs, primarily *Hadrosaurs*, but with *Ankylosaur*, *Theropod*, crocodilian, large avian and possibly pterosaur tracks also recognized (M. LOCKLEY, pers. comm., 1997; R. McCREA, pers. comm., 1999; SCOTT, 2000). Coastal plain facies have been grouped into three *facies associations* that form distinct *nonmarine systems tracts* (PLINT et al., 2001) that reflect differing sedimentation : accommodation ratios. The facies characteristics of each systems tract are discussed in more detail below.

4.3.1. Channel-dominated, low-accommodation systems tract

This systems tract consists of multi-storey, laterally-accreted meandering channel sandstones, typically 5–8 m thick, that were only deposited within valleys, the base of which forms the lower boundary of the systems tract. There is very little floodplain mudstone preserved within the valley-fills. Within a valley-fill, mud-free, cross-bedded sandstone generally dominates the lower storeys, whereas the uppermost, and most complete storey is sometimes much more mud-rich, with well-developed heterolithic stratification (Figs. 17, 18). This suggests that the latest stage of valley-filling may have taken place under tidally-influenced conditions (cf. SHANLEY et al. 1992; SMITH, 1987; WILLIS, 1997). The uppermost

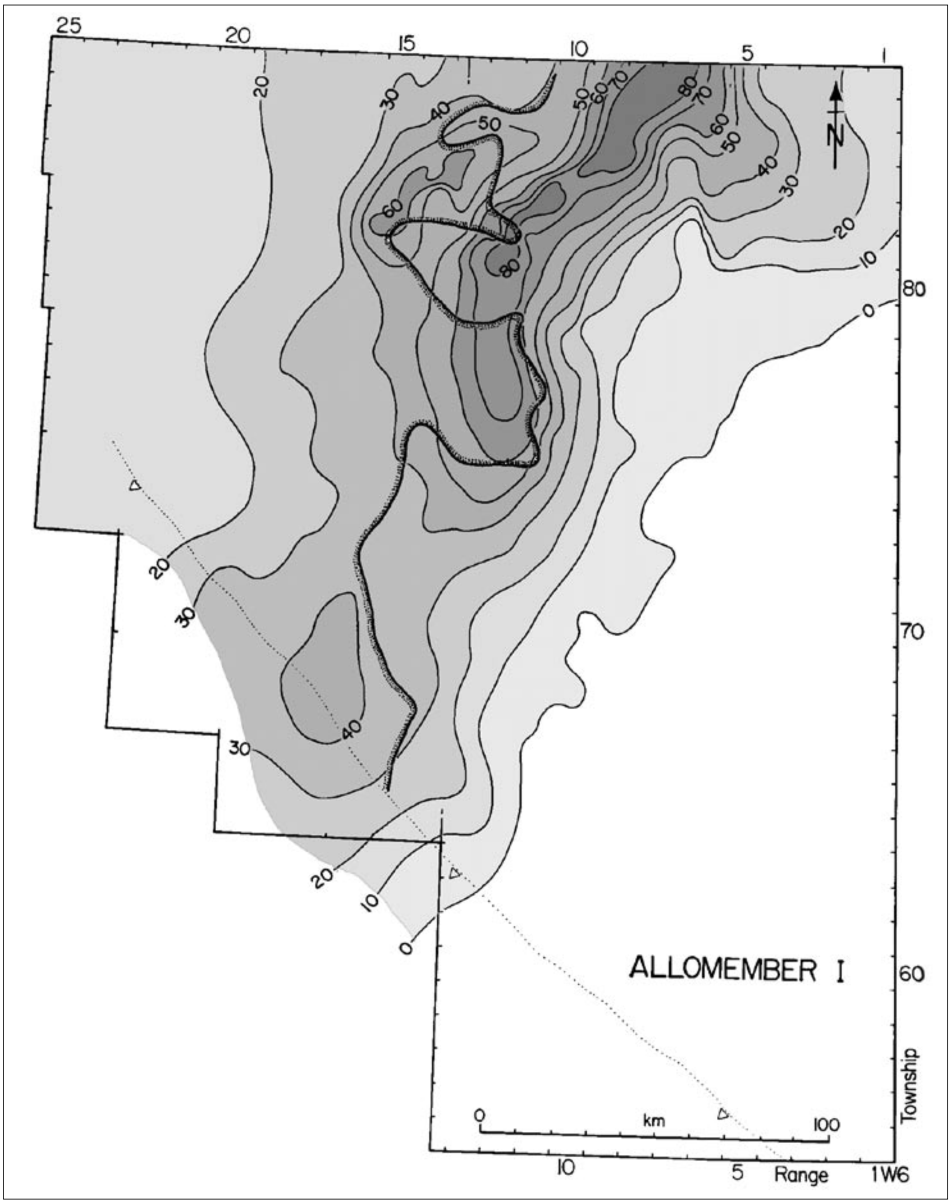


Fig. 11 Palaeogeographic and isopach map of Dunvegan allomember I.

1–5 m of the valley-fill is sometimes occupied by a sandier-upward succession, commonly highly bioturbated, interpreted to record final filling of the valley by a small bay-head delta (Figs. 18, 19). The bioturbation suggests incursion of brackish water, and this is consistent with the progressive drowning of the valleys and updip migration of marine influence (cf. McLAURIN & STEEL, 2000). The upper boundary of the systems tract is placed at the generally abrupt transition (an expansion surface of MARTINSEN et al., 1999), to lacustrine-dominated facies, that record a major change in the accommodation/supply ratio. The prevalence of multi-storey sandbodies and scarcity of preserved floodplain mudstone suggests that these deposits reflect a relatively low accommodation rate that permitted protracted reworking of alluvial deposits within the valley (cf. ASLAN & BLUM, 1999; MARTINSEN et al., 1999). Available evidence does not permit us to differentiate sediments deposited during regional base-level rise from possible terraces deposited during the falling stage (cf. BLUM & PRICE, 1998).

4.3.2. Lacustrine-dominated, high-accommodation systems tract

This systems tract is characterized by widespread, well-laminated to poorly-laminated and carbonaceous mudstone and coal facies that represent a range of lake and wetland environments. These facies enclose, and are interstratified with ribbon channel sandbodies (Fig. 20), crevasse splays and sandier-upward, lake-fill successions (Fig. 21). Palaeosols are generally poorly developed and hydromorphic in character (palaeosol types 1–3, fig. 8 of McCARTHY et al., 1999). The base of this systems tract is defined in two ways. It may abruptly overlie multistorey channel sandstones filling valleys, where the facies transition records the abrupt expansion of deposition beyond the confines of the val-

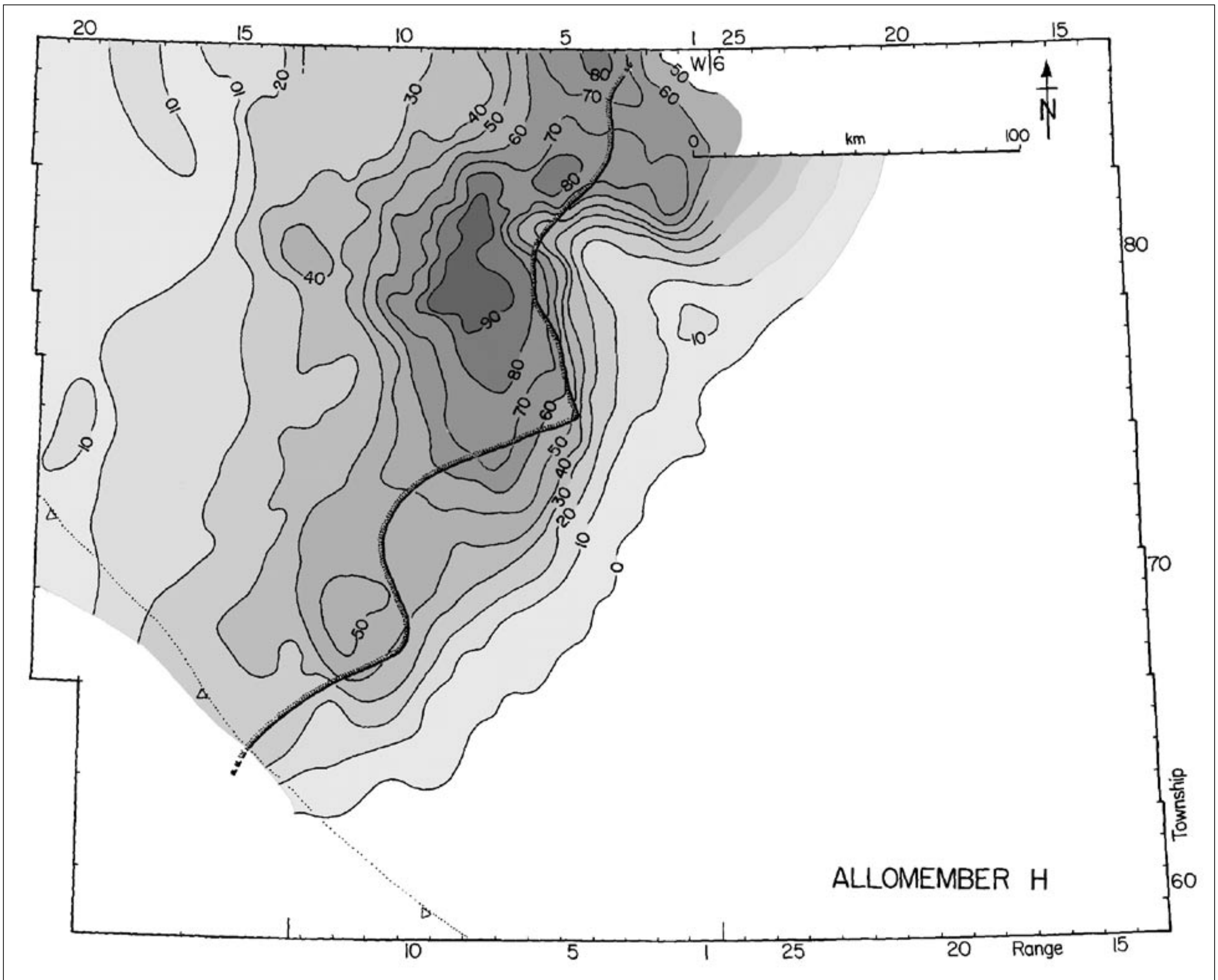


Fig. 12 Palaeogeographic and isopach map of Dunvegan allomember H (from PLINT et al., 2001).

leys (cf. MARTINSEN et al., 1999). The prevalence of lacustrine facies indicates that accommodation was generated faster than it could be filled, resulting in standing bodies of water (cf. HAMPSON et al., 1999). Alternatively, it may abruptly overlie a sharp, micro-erosional surface above a well-developed palaeosol that defines an interfluve surface (McCARTHY et al., 1999).

It is significant that *bioturbated* lacustrine deposits are consistently observed for a few metres, *immediately* above sequence-bounding interfluve palaeosols. Further up section, bioturbation disappears until the next interfluve surface is crossed. This observation suggests that interfluves were initially flooded by brackish-water lagoons, with an attendant burrowing fauna. This interpretation is consistent with the observation (McCARTHY & PLINT, 1998, in review) of barite crystals in mudstones immediately above the top H and top G interfluve surfaces. This mineral is known to precipitate in modern hydromorphic soils associated with saline groundwater (STOOPS & ZAVALETA, 1978)

The lacustrine-dominated, high-accommodation systems tract is the principal component of the nonmarine portion of sequences. It is here emphasized that the maximum flooding interval which is stratigraphically equivalent to the time of maximum marine transgression, consistently lies *within*, and not at the top of the lacustrine-dominated systems tract. This observation suggests that, for a significant interval of time after maximum transgression, accommodation was generated faster than it could be filled.

4.3.3. Palaeosol-dominated, low-accommodation systems tract

This systems tract is dominated by blocky mudstones that represent moderately- to well-developed palaeosols (palaeosol types 4–5, fig. 8 of McCARTHY et al., 1999). Commonly, several well-developed palaeosol profiles are found in close vertical juxtaposition. These palaeosols are generally interstratified with crevasse-splay sandstones and thin units of lacustrine and wetland facies. The base of this systems tract is defined where laminated lacustrine mudstones pass upward, typically over 1–2 metres, into a succession consisting of relatively well-developed complex palaeosols that are welded together (McCARTHY et al., 1999). The top of the palaeosol-dominated systems tract is defined by a

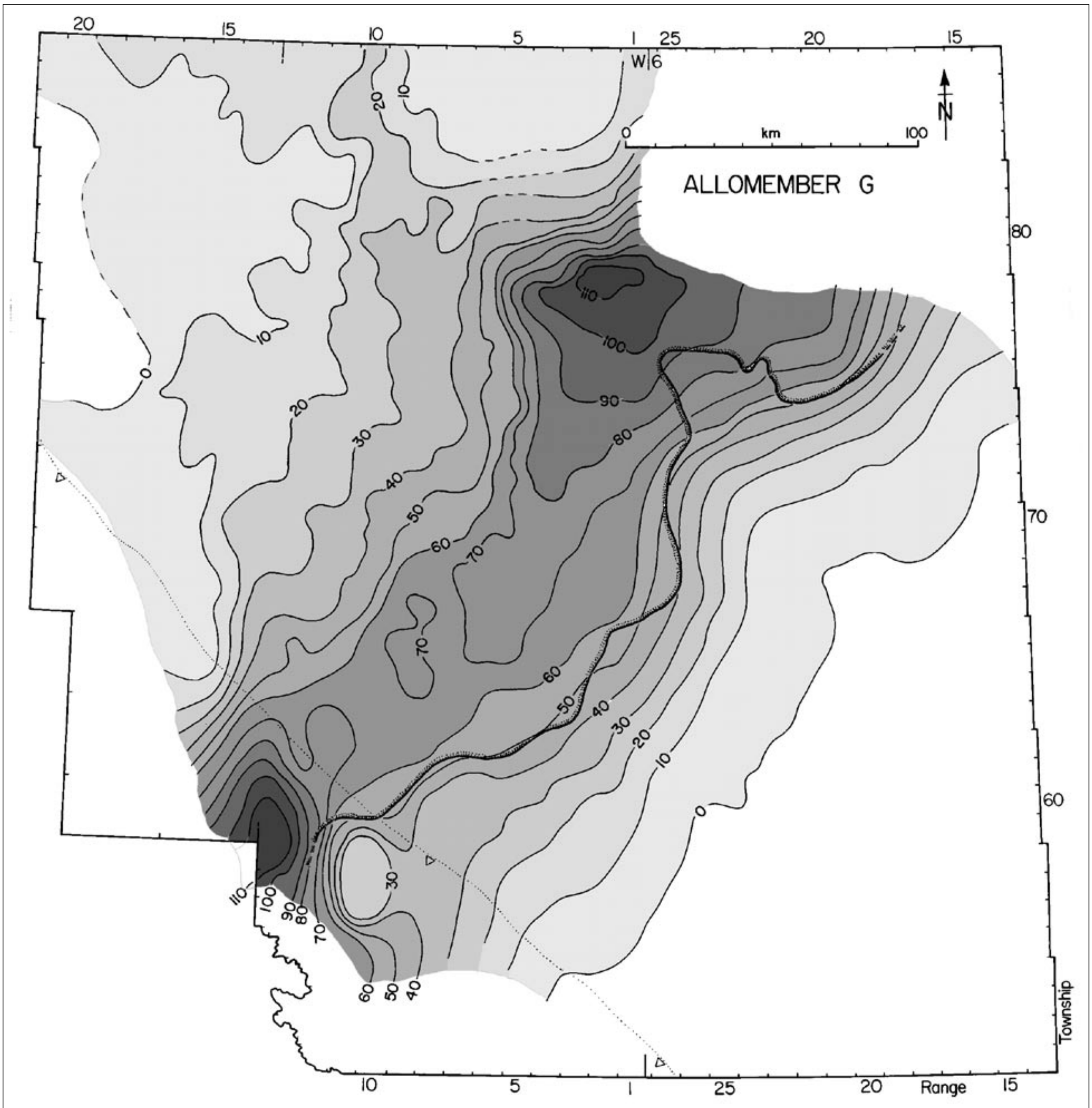


Fig. 13 Palaeogeographic and isopach map of Dunvegan allomember G (from PLINT et al., 2001).

subaerial erosion surface marking an interfluve (McCARTHY & PLINT, 1998; in review; McCARTHY et al., 1999; McCARTHY, 2002) that is expressed as a sharp, but commonly subtle stratigraphic discontinuity having a micro-erosional relief of centimeters to decimeters.

This systems tract is interpreted to represent a floodplain environment in which autogenic processes continued to build and fill minor depositional topography, but where the overall rate of vertical aggradation gradually decreased to nil, permitting progressively more time for pedogenesis. The upper part of each sequence-bounding palaeosol succession represents several tens of thousands of years of sediment starvation and *in-situ* polygenetic pedogenesis (*sensu* MARRIOT & WRIGHT, 1993). These palaeosols represent a period of nil to negative accommodation during which pedogenic overprinting is recorded in the upper part of the interfluve palaeosol complex (McCARTHY & PLINT, 1998; McCARTHY et al., 1999; Fig. 22). The significance and spatial variability of interfluve palaeosols is discussed in more detail below.

5. PALAEOVALLEY SYSTEMS

Extensive valley systems exist at the tops of Dunvegan allomembers H, G, F and E (Fig. 23). Valley-filling deposits are well-exposed along the Peace River valley between the Alberta–British Columbia border and Erin Lodge, some 140 km downstream, and also on

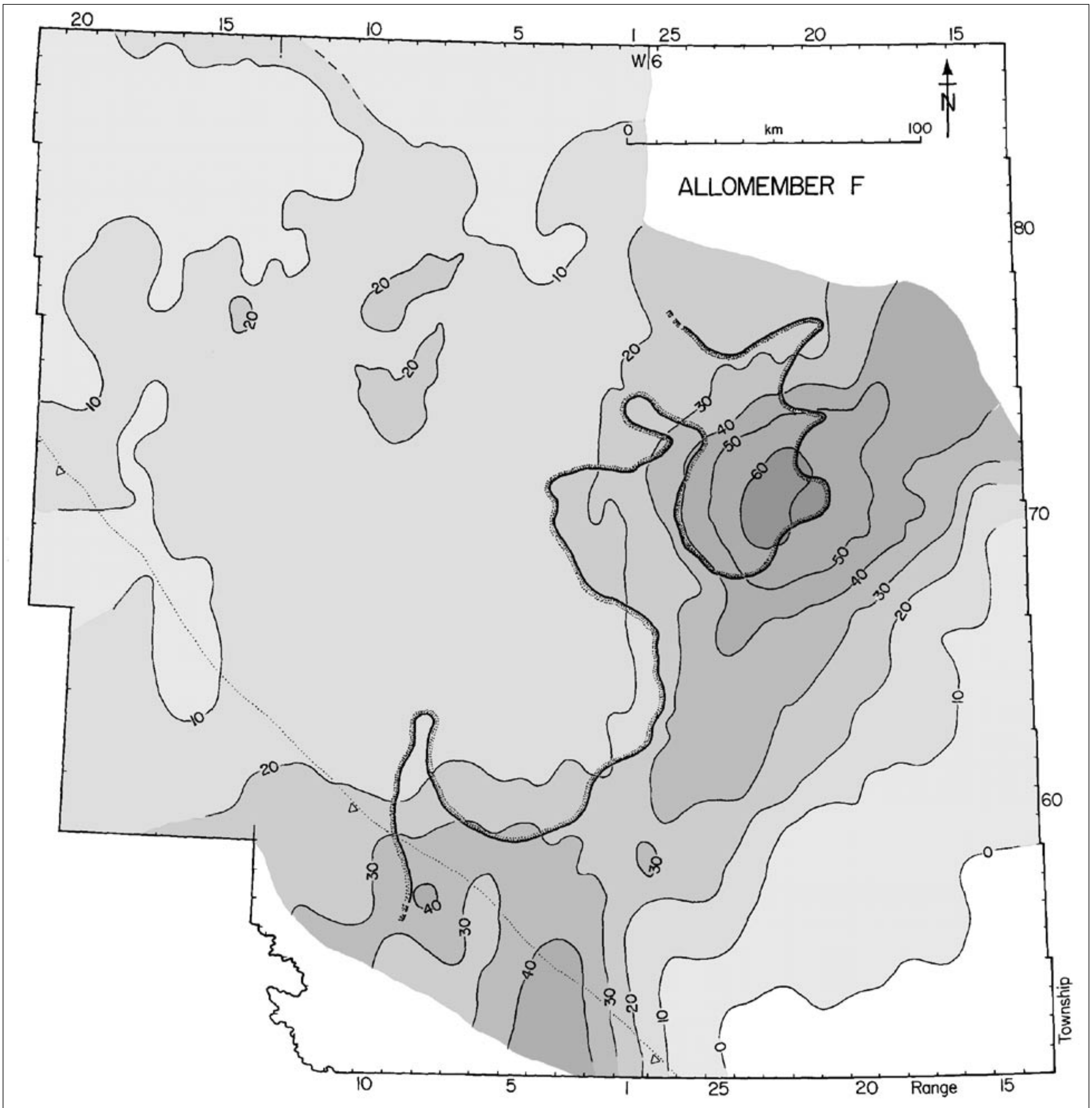


Fig. 14 Palaeogeographic and isopach map of Dunvegan allomember F (from PLINT et al., 2001).

the Beatton and Kiskatinaw rivers (Fig. 1; PLINT, 2002). Valley-filling sandstones up to 30 m thick are also present at the top of allomembers D, C and A. However, these younger sandbodies are typically <1 km wide and are generally impossible to map in subsurface with the available well density: they will not be discussed further in this paper. Up to 4800 wells and 40 outcrops were used to map valley systems incised into the top surfaces of allomembers H to E. Individual valleys can confidently be mapped for distances of 110 km in allomember H, 250 km in allomember G, 200 km in allomember F, and 320 km in allomember E. Valleys maintain a remarkably consistent depth over their entire length, with maximum depths ranging from 31–40 m, with an overall average of about 21 m. Valley width varies from a few hundred metres to a maximum of about 8 km, with most valleys being about 1–2 km wide. Cross-valley depth varies abruptly suggesting the development of a terraced morphology.

Limited core in downdip valley reaches and abundant outcrop in up-valley reaches permits five main facies to be identified. Up-valley reaches are dominated by fine to medium grained, trough crossbedded sandstone. Widely-distributed but uncommon tidal bundles defined by carbonaceous drapes, bundle sequences and sigmoidal cross-stratification are present up to 30 km landward of the marine transgressive limit, and indicate tidal backwater effects, probably under a microtidal regime. Rippled and planar laminated sandstone is an accessory facies. Locally-preserved lenses of rooted,

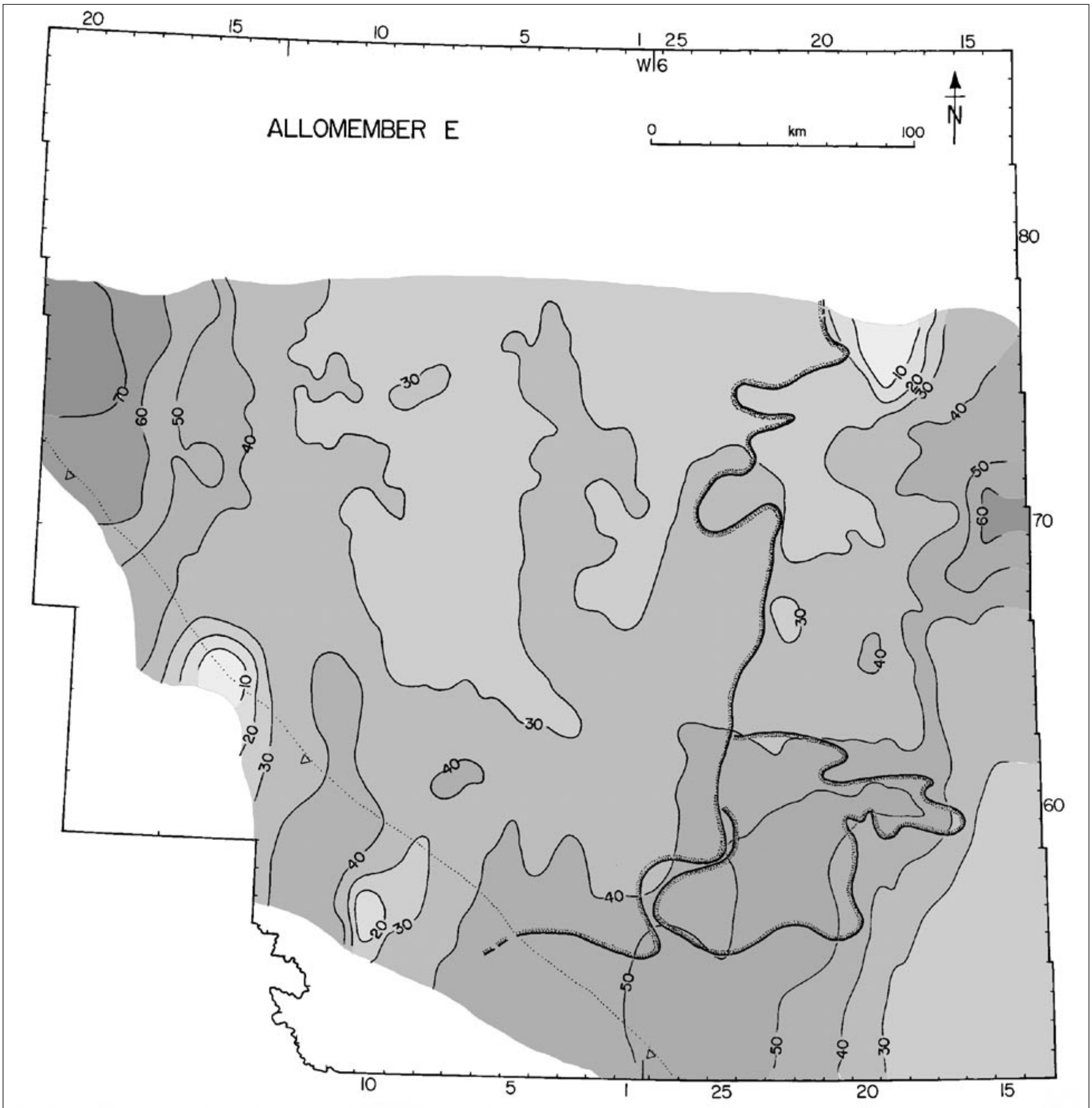


Fig. 15 Palaeogeographic and isopach map of Dunvegan allomember E. Note the appearance of a semi-circular depocentre in the NW corner of the study area, interpreted to record the onset of renewed flexural subsidence (from PLINT et al., 2001).

blocky to laminated mudstones and coals are interpreted to represent remnants of floodplain deposits preserved within the valleys. Inclined heterolithic stratification on a dm- to m-scale is common at the top of valley-fills up to 110 km landward of the marine transgressive limit; this facies probably records a complex interaction between tidal processes and ?seasonal river discharge fluctuations. Down-valley reaches, typically within about 50 km of the lowstand shoreline, have a sandstone-dominated lower portion and a mudstone-rich upper portion in which a variety of thinly-bedded heterolithic strata are encountered. Heterolithic facies may be organized in sandier-upward successions that might represent upward-shallowing bay-head deltas, or muddier-upward successions that might represent tidal flats or muddy IHS filling a channel. Ubiquitous sub-mm scale mud drapes and bidirectional ripple cross lamination provide strong evidence for tidal processes (Fig. 19). Soft-sediment deformation and micro-faulting are abundant and might reflect dewatering, slumping, earthquake shock and dinoturbation (PLINT & WADSWORTH, in review).

Within about 20–40 km of the maximum regressive shoreline, valleys in each allomember become unrecognizable, with the exception of one major channel in each of allomembers E and F, which obviously continues as a distributary that feeds a major lowstand delta lobe. The most seaward part of the delta plain lacking valleys

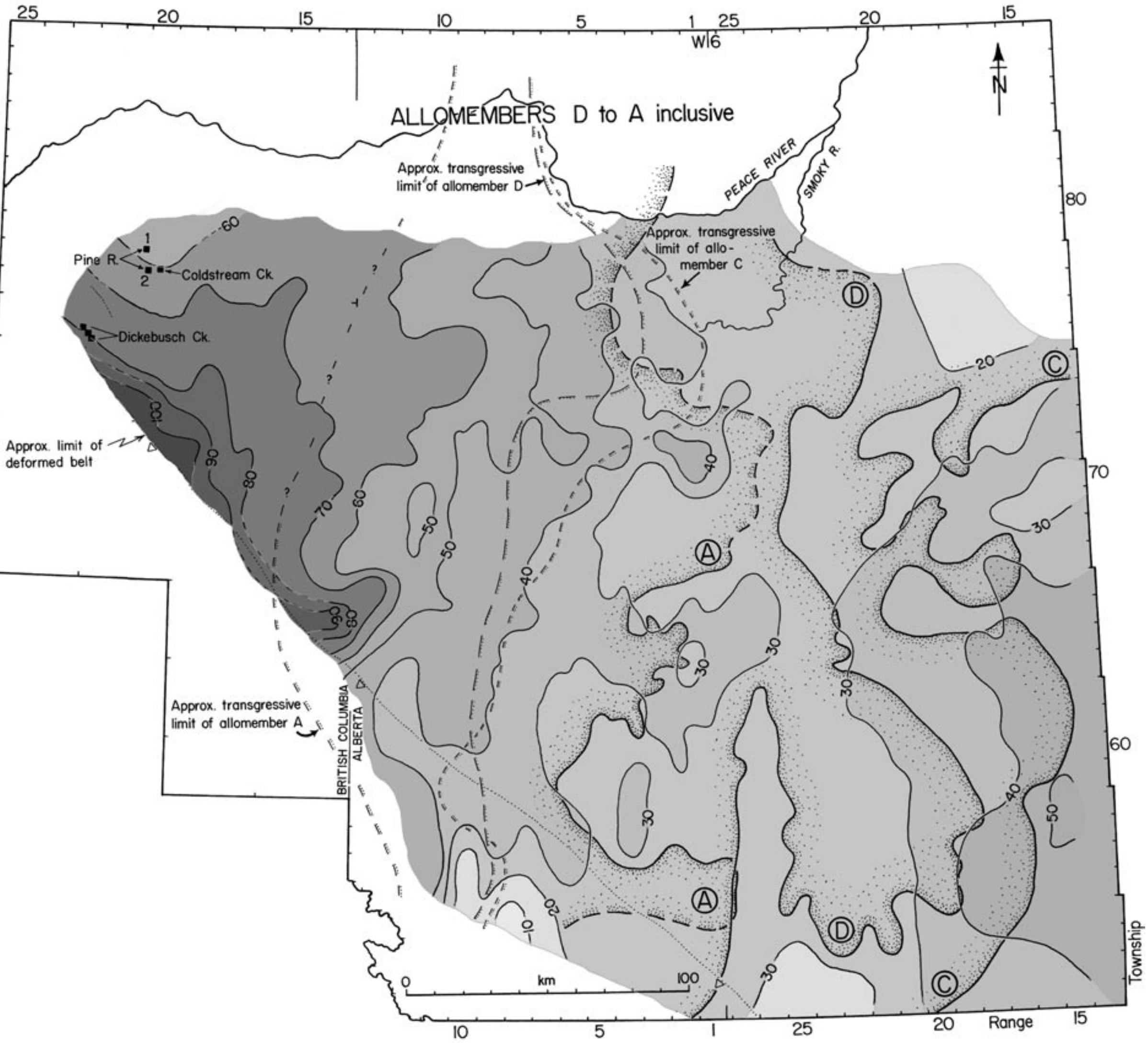


Fig. 16 Composite map showing the maximum progradational shorelines of Dunvegan allomembers D, C and A, and also isopachs for the entire D–A interval. Note the prominent depocentre along the western margin of the study area, and progressive rotation of isopach lines from N–S to NW–SE (from LUMSDON & PLINT, 2003).

is interpreted as the LST, deposited when sea level was slowly rising and rivers started to backfill their valleys. Valley filling took place primarily during the early TST. In downdip areas, the top of the valley fill is overlain, at a *Glossifungites* transgressive surface, by a few dm of intensely bioturbated muddy sandstone that constitutes a very thin marine TST. Further landward, a thin coal commonly separates the valley-fill from overlying marine deposits; the coal typically extends about 30 km seaward of the maximum marine transgressive limit and records the development of sediment-starved mires behind the transgressing shoreline (Fig. 19). Further landward, in areas not inundated by the sea, valley-fills are typically overlain by laminated lacustrine mudstones, thin coals or pedogenic mudstones.

6. INTERFLUVE PALAEOSOLS: SPATIAL VARIABILITY AND PALAEOCLIMATIC SIGNIFICANCE

The palaeosols that mantle interfluve surfaces are similar to modern Alfisols and each example records (i) aggradation on an alluvial/coastal plain; (ii) a subsequent static and/or degradational phase related to valley incision, non-deposition and soil thickening; and (iii) a final aggradational phase related to valley-filling and renewed sedimentation across the coastal plain (McCARTHY & PLINT, in review; Fig. 22). Within this overall template, however, variations in palaeosol thickness, redoximorphic features, illuvial clay content, and geochemistry suggest developmental control by hydrological characteristics that were influenced both by the nature of the underlying alluvial sediments, and

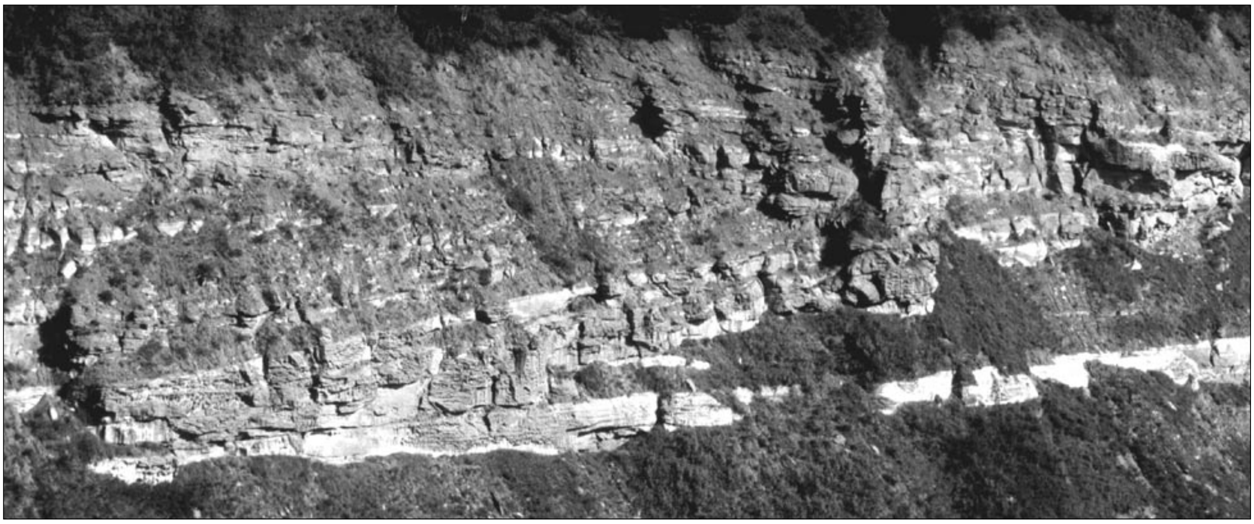


Fig. 17 Multi-storey valley-fill at Beatton River, B.C. Note how base of valley-fill cuts downward from right to left, eventually intersecting the top of a sheet-like delta-front sandstone. Lower storeys are crossbedded sandstone but the uppermost storey shows well-developed inclined heterolithic stratification, possibly deposited under the influence of seasonal discharge fluctuations. Maximum thickness of valley-fill = 23 m (from PLINT & WADSWORTH, in review).

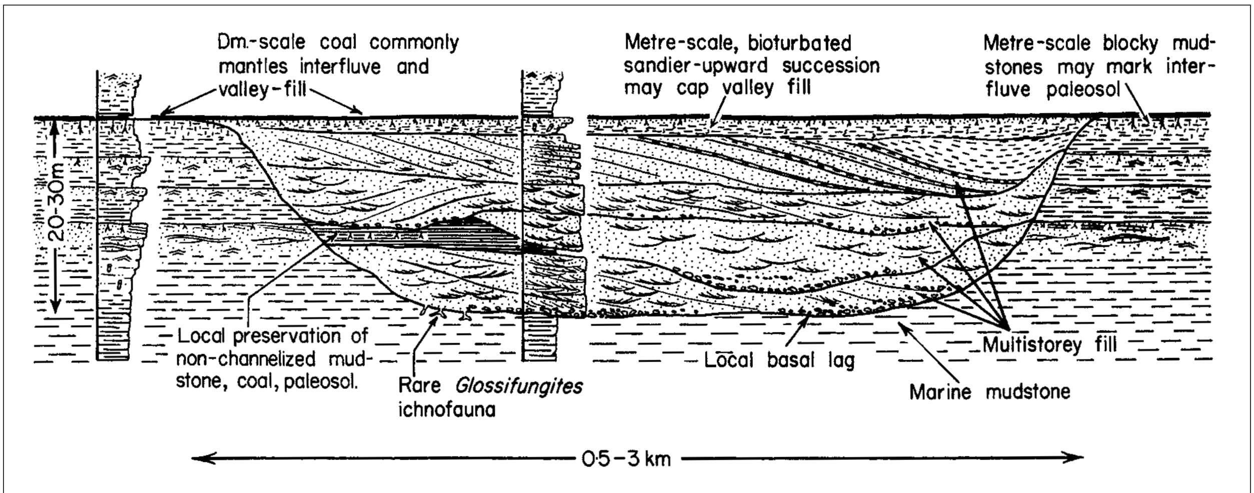


Fig. 18 Transverse section summarizing the main characteristics of valley-fills in an up-valley, fluvial-dominated reach (from PLINT & WADSWORTH, in review).

by distance from the dissected valley edge. Palaeosols that developed closer (few hundred m to few km) to valley margins are thicker, have greater quantities of illuvial clay, and display characteristics suggestive of well-drained conditions relative to those palaeosols that developed further (>5–10 km) from valley margins. Variations in drainage and palaeotopography during landscape dissection resulted in different palaeosol development styles on interfluve surfaces that can be shown, on the basis of physical correlation, to have the same geomorphic age.

Renewed coastal plain aggradation and syndepositional pedogenesis is marked, at sites a few km from valleys, by an obvious increase in grain size, multiple organic horizons, sedimentary microlamination, and the presence of sphaerosiderite. These features suggest the resumption of periodic flooding of the coastal plain as fluvial channels once again migrated beyond the valley walls. The presence of barite coupled with elevated levels of Ca, Mg, Sr, and P suggest that interfluve palaeosols adjacent to valley margins were initially flooded with brackish water (McCARTHY & PLINT, 1998; DRIESE et al., 1992). In contrast, an interfluve palaeosol located 15 km from the nearest valley is directly capped by a coal, indicating water-table rise, increased accommodation rate, but negligible clastic supply (Fig. 22). The interfluve palaeosols contain abundant pedogenic clay minerals (interstratified illite–vermiculite) in the <0.2 mm fraction that yield $\delta^{18}O$ values of -12.9‰ to -11.6‰. These isotopic ratios are considered to record the composition of the Cretaceous rainwater at this palaeo-latitude of about 65°N, and suggest soil temperatures in the range 9–16°C (VITALI, et al., 2002). This temperature estimate is consistent with

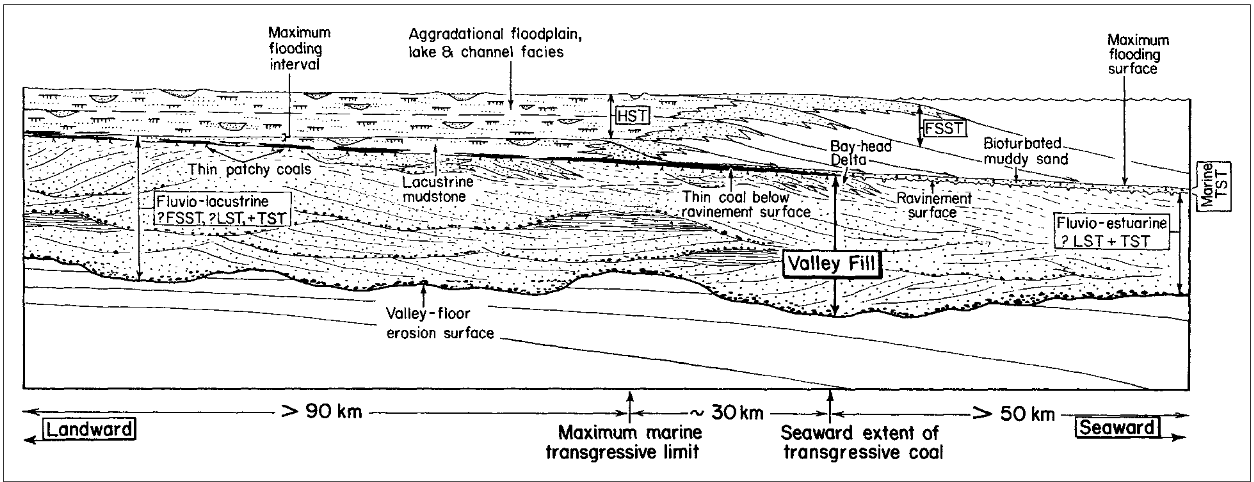


Fig. 19 Longitudinal section summarizing the main characteristics of valley-fills from the valley mouth to a point about 180 km up-valley (from PLINT & WADSWORTH, in review).

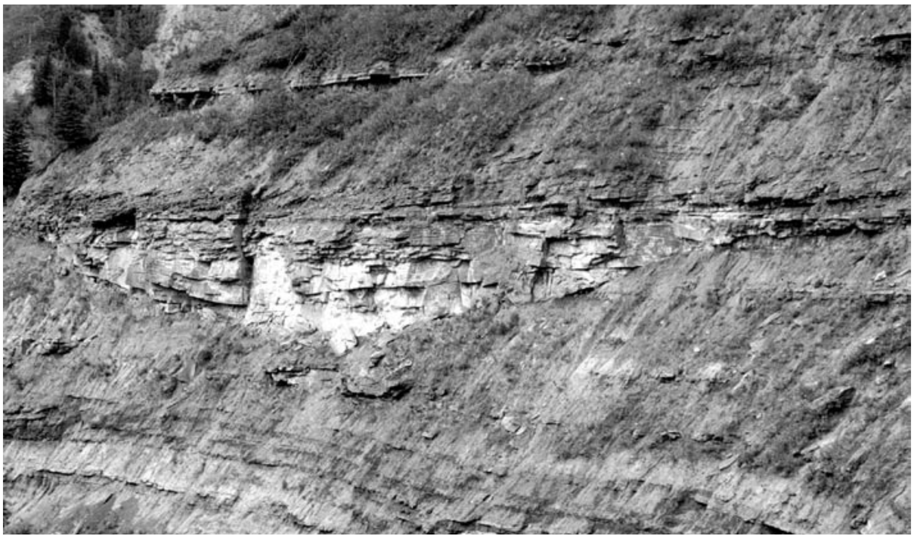


Fig. 20 Example of lenticular fluvial sandstone (up to 8 m thick), interpreted to represent an anabranch of an anastomosed river, enclosed in grey pedogenic and lacustrine mudstones. Coldstream Creek, B.C.

estimates based on pedological characteristics (McCARTHY & PLINT, 1999) and on palaeobotanical information (7–13°C; UPCHURCH & WOLFE, 1993).

7. RECOGNIZING ACCOMMODATION CHANGE

Evidence of accommodation changes, both positive and negative, can be inferred from various sedimentological and stratigraphic features of the Dunvegan Fm. Seven main lines of evidence will be summarized: 1) Marine transgressive surfaces; 2) Isopach patterns; 3) Distribution and character of nonmarine facies; 4) Valley systems; 5) Interfluve palaeosols; 6) Sharp-based delta-front sandstones, and 7) Palaeodrainage patterns.

7.1. Regional marine transgressive surfaces

Regionally-mappable marine transgressive surfaces are among the most recognizable features of the Dunvegan Fm., allowing the formation to be divided into allomembers (Fig. 23). Importantly, the well log response that corresponds to these transgressive surfaces can be traced inland for up to about 150 km across the coastal plain that lay landward of the transgressive limit of the marine shoreline. Tracing these non-marine (lacustrine) flooding surfaces allows the complete geometry of the allomember to be determined, including both the marine and non-marine portions. Transgressive surfaces provide clear evidence of an increase in the accommodation : sedimentation ratio which led to shoreline transgression. Comparison of maps in Figs. 10–16 shows how the transgressive distance varied within and between each allomember. Figure 24 summarizes the 'average' distance that the shoreline moved landward and seaward for each allomember.

7.2. Regional isopach maps and accommodation changes

Figures 10–14 illustrate the regional palaeogeography and geometry of the lower 5 Dunvegan allomembers,

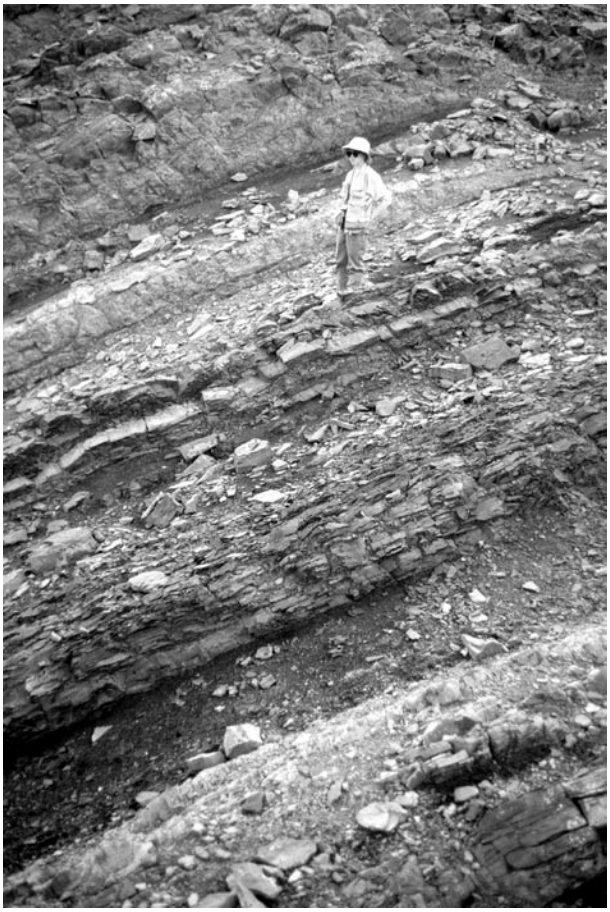


Fig. 21 Examples of stacked sandier-upward lake-fill successions typical of much of the Dunvegan coastal plain. Lynx Creek, Alberta; person for scale.

J to F. Allomembers J to F are markedly sigmoidal in profile, thinning updip onto the coastal plain, and also seaward through downlap onto the condensed section (FSU marker, Fig. 23). The maps for allomembers J to F very clearly illustrate the progressive basinward (i.e. southeastward) shift of the depocenter. Bathymetric deeps (i.e. water-filled space) remaining between delta lobes were filled by particularly thick intervals of the succeeding allomember. This pattern may reflect differential compaction over older lobes, leading to diversion of rivers into the intervening, topographically low areas (cf. BLUM & PRICE, 1998). Isopach lines in Figs. 10–14 are essentially parallel, and trend NE–SW, perpendicular to the orogen. *The isopachs show no evidence of thickening towards the Foothills deformed belt*. This critical observation shows that during deposition of allomembers J to F, this part of the basin experienced essentially no differential tilting toward the orogen.

Figures 15 and 16 illustrate the palaeogeography and geometry of allomember E and allomembers D–A combined (because the bounding surfaces of allomembers D, C and B cannot be mapped across the entire study area). The isopach pattern for allomembers E to A shows a dramatic change relative to older allomembers. Allomember E shows a pronounced updip thickening towards the NW with the development of an isopach 'moat' that extends about 70 km from the present deformation front. All of this semi-circular depocentre is filled with nonmarine deposits which reach 80 m in thickness in outcrop in the Pine River – Coldstream Creek area (LUMSDON & PLINT, 2003 – in press; Fig. 2). In the SE however, isopach patterns for allomember E resemble those of older allomembers in preserving the broad NE–SW trend, with the thickest parts of allomember E filling the bathymetric deep fronting allomember F.

The semi-circular sedimentary 'moat', first recognized in the NW part of allomember E, expands in allomembers D–A to a radius of about 250 km, across which allomembers D–A thicken from about 30 to 100 m (Fig. 16). Note how isopach lines progressively rotate to parallel the orogen, reflecting the increasing importance of flexural subsidence towards the SW.

The thickening of allomembers E to A in the NW corner of the study area is interpreted to record the onset of a new phase of flexural subsidence adjacent to the active margin of the basin (cf. ROGERS, 1998). The generation of new alluvial accommodation in the updip portion of the delta system had a pronounced affect on deposition in the marine realm downdip. Allomembers J–E are relatively thick and the delta-front areas are relatively sand-rich. In contrast, the deltas in allomembers D–A (located in the SE corner of the study area; Fig. 16), are relatively thin and sand-starved. This is interpreted to reflect sediment partitioning between alluvial and marine depocentres, with the balance between the two progressively shifting in favour of the alluvial realm, as the orogen-proximal subsidence rate increased through allomembers E to A.

7.3. Distribution and character of nonmarine facies

In each allomember, only the lower 2–4 parasequences, which comprise the transgressive and highstand systems tracts, include both marine delta front / prodelta *and* correlative coastal plain deposits (Fig. 25). Younger parasequences offlap seaward and consist of only marine delta front / prodelta facies (PLINT et al., 2001). The offlapping deltaic parasequences are interpreted to have been deposited during the falling stage systems tract (PLINT & NUMMEDAL, 2000; PLINT et al., 2001). The top of the HST became a non-depositional surface during the FSST. During this time, rivers began to incise valleys, starting at the seaward margin of the highstand systems tract (PLINT, 2002). Rivers incised progressively headward during the FSST, and also continued to erode valleys across the upper surface of the FSST as sea-level continued to fall. Interfluves between valleys were starved of sediment for many tens of ky, during which time they developed thick and mature soil profiles (McCARTHY et al., 1999, McCARTHY & PLINT, in review).

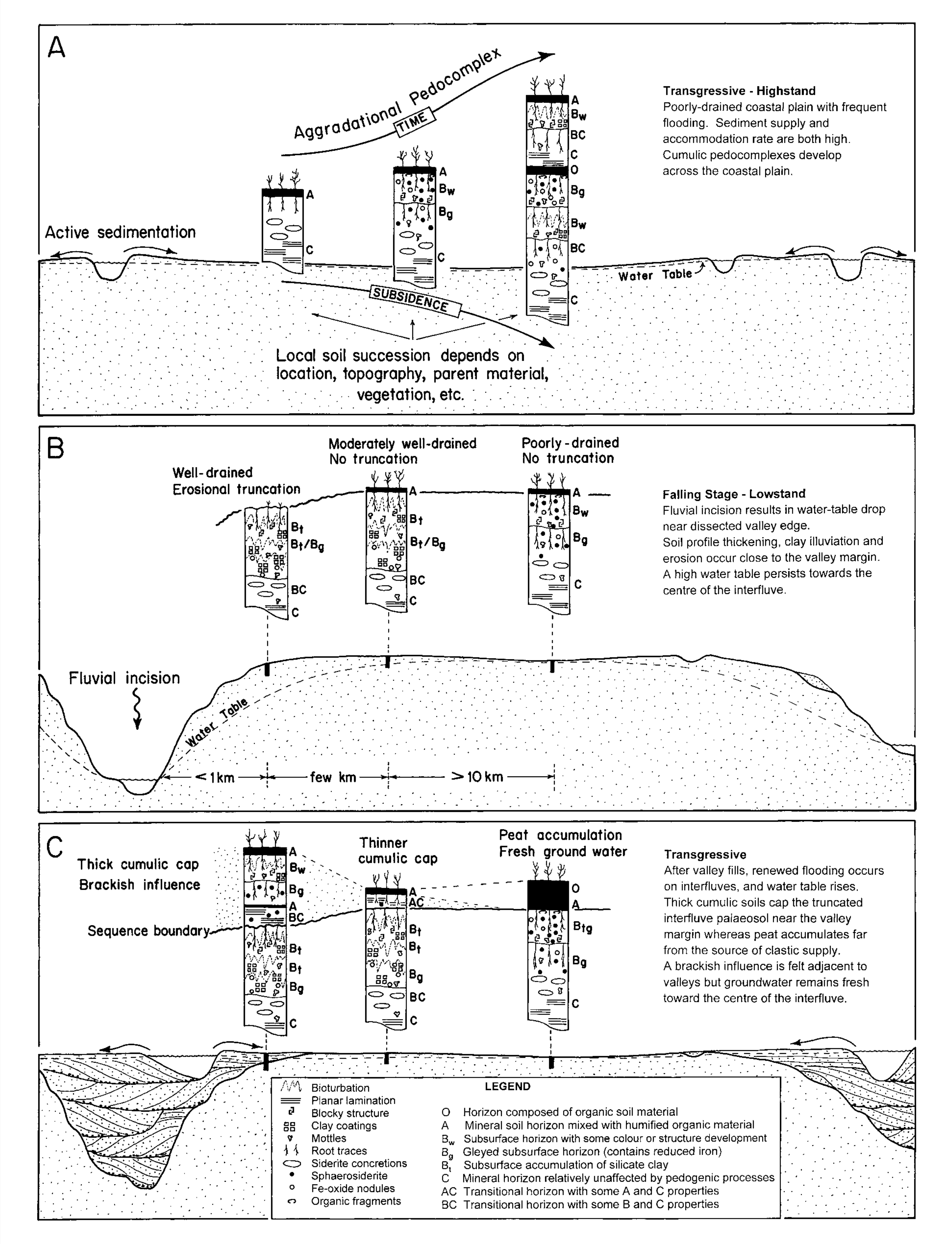


Fig. 22 Diagram summarizing the developmental history of interfluve palaeosols, and also showing how the history and character of the palaeosol varies with distance from the margins of palaeovalleys. The distance from the valley margin strongly influenced the drainage of the interfluve, and also controlled the rate of sediment flux once the valley had been filled (from McCARTHY & PLINT, in review).

The restriction of coastal plain deposits to the TST and HST, which have an aggradational to progradational stacking pattern, show that alluvial accommodation was only generated during the rising half of the relative sea-level cycle. The fact that the alluvial deposits in allomembers I–F form a gradually updip-thinning sheet suggests that accommodation was generated by eustatic sea-level rise rather than by flexural subsidence. The

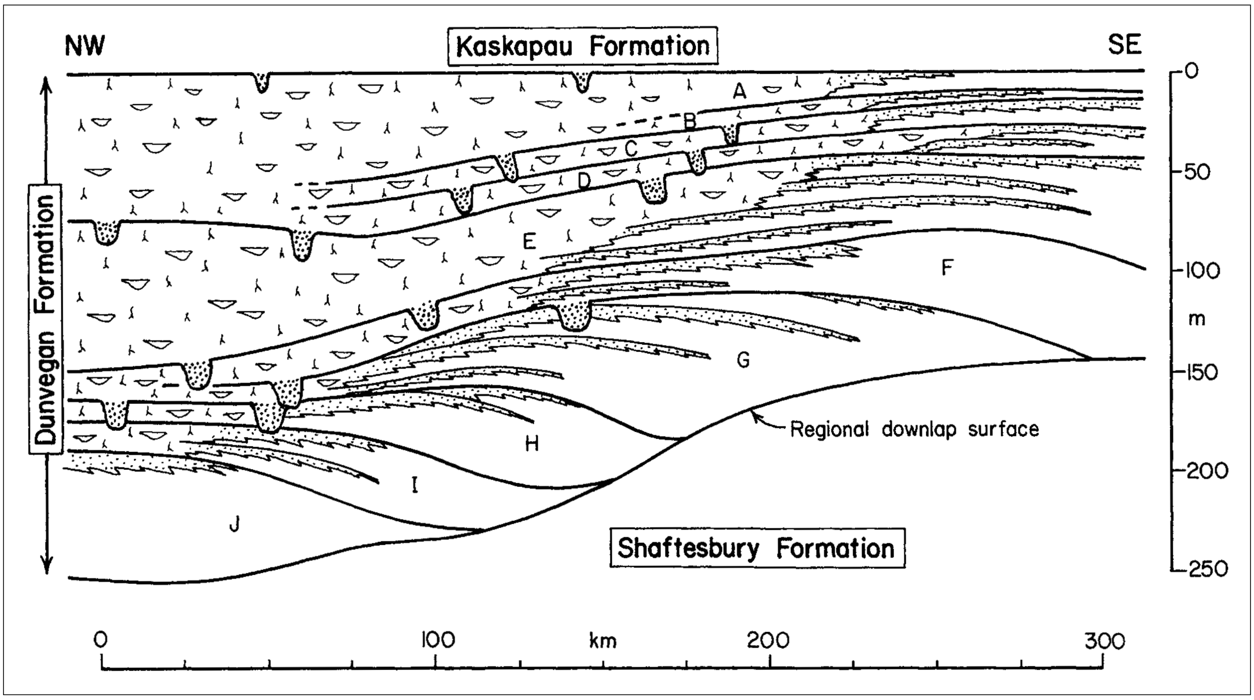


Fig. 23 Summary dip section (located in Fig. 1) drafted to scale, with the top of allomember A as a datum, showing the stratigraphic distribution of valley-fills. Only valleys on allomembers H–E can be mapped with the available well density (from PLINT & WADSWORTH, in review).

updip and westward thickening of alluvial deposits in allomembers E–A however, reveals an additional component of flexural subsidence along the basin margin. Thus, the overall geometry of allomembers ('sigmoidal-prismatic' in allomembers J–F, versus wedge-shaped in allomembers E–A) reveals a fundamental difference in the geometry of accommodation generation. For allomembers J–F, tectonic tilting was negligible and eustatic change appears to provide the simplest explanation for the parasequence stacking pattern and for the restriction of alluvial deposits to the rising half of the relative sea level cycle. In contrast, updip thickening of alluvial deposits in allomembers E–A records the onset of flexural subsidence which partially overprinted the effects of eustatic changes in the more distal (SE) part of the basin.

Allomembers E to A contain semi-circular depocentres in the NW portion of the study area (Figs. 15, 16). Alluvial rocks filling those depocentres are well-exposed in NE British Columbia. Non-channelised alluvial facies represent lake, mouth-bar, levee and crevasse splay environments, and organic histosol (coal), poorly-drained, intermediate redox and better-drained palaeosols. Siderite, organic debris and dinosaur tracks are

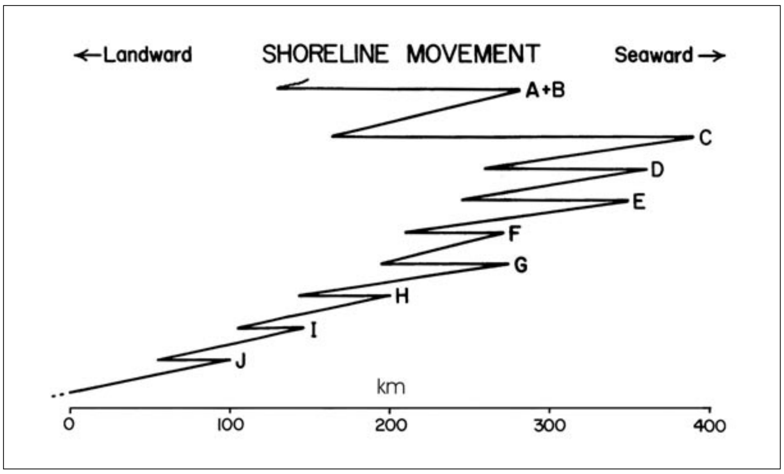


Fig. 24 Scale drawing summarizing the average shoreline displacement for each transgressive–regressive cycle of the Dunvegan Fm. Note the similarity in shoreline displacement for allomembers I–C. Only the transgressions at the tops of allomembers C and A show a transgressive distance that greatly exceeds the regressive distance, probably reflecting an accelerating subsidence rate and increasing accommodation:sedimentation ratio (see also Fig. 16).

ubiquitous elements of these facies. Channelized facies are dominated by non-migrating (probably anastomosed) sandstone channel-fills (LUMSDON & PLINT, in press).

The non channelized facies form two associations which are interpreted on the basis of intrinsic facies characteristics, to represent different accommodation rates. Association A is dominated by crevasse splays,

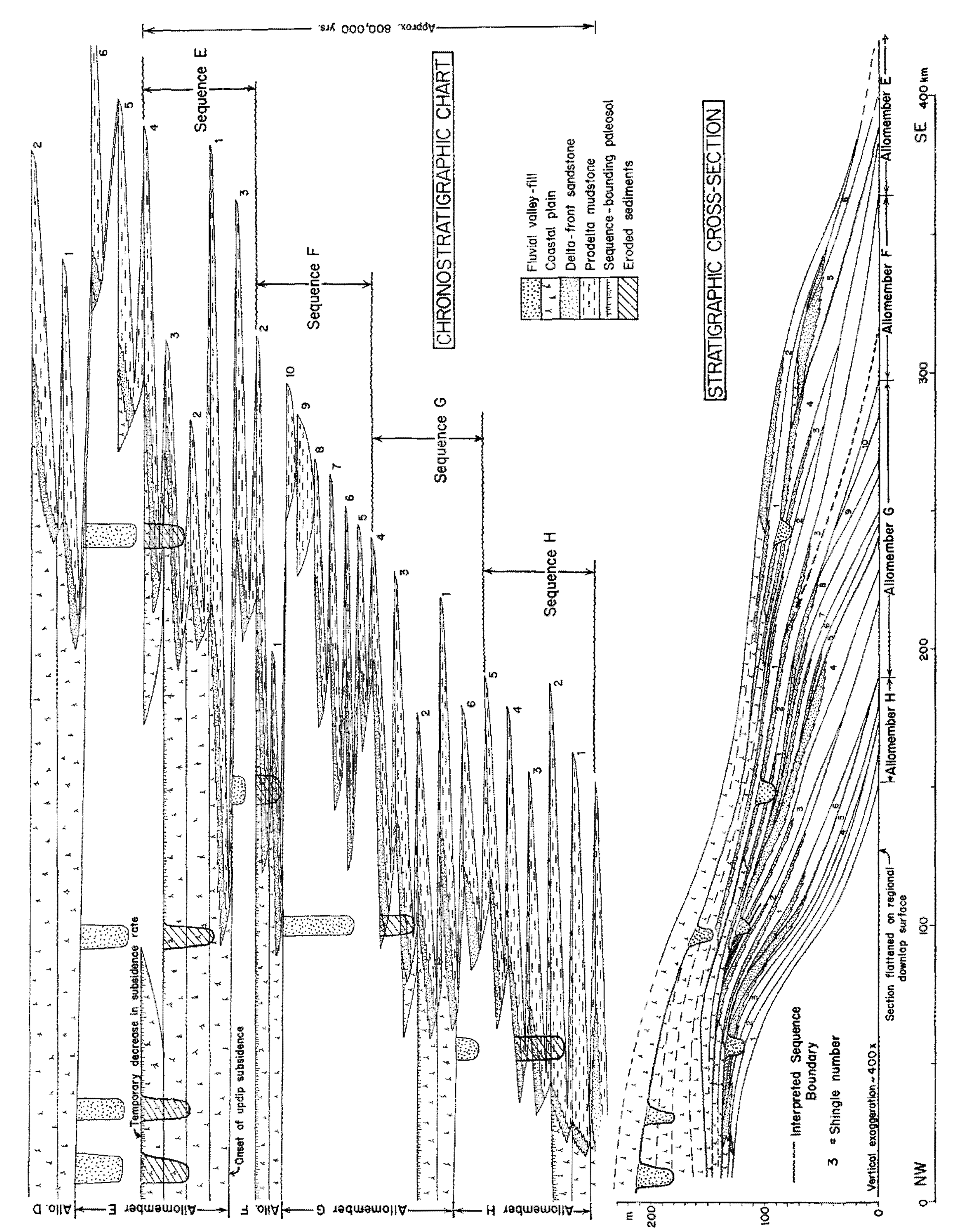


Fig. 25 Dip section, drawn to scale and datumed on the downlap surface, showing (below) the organization of parasequences within allomembers H–D. The upper diagram is a chronostratigraphic representation of the lower cross-section, based on the assumption that each parasequence represents an equal interval of time. The chart emphasizes the depositional hiatus on the upper coastal plain that develops during the falling stage and lowstand systems tracts (from PLINT et al., 2001).

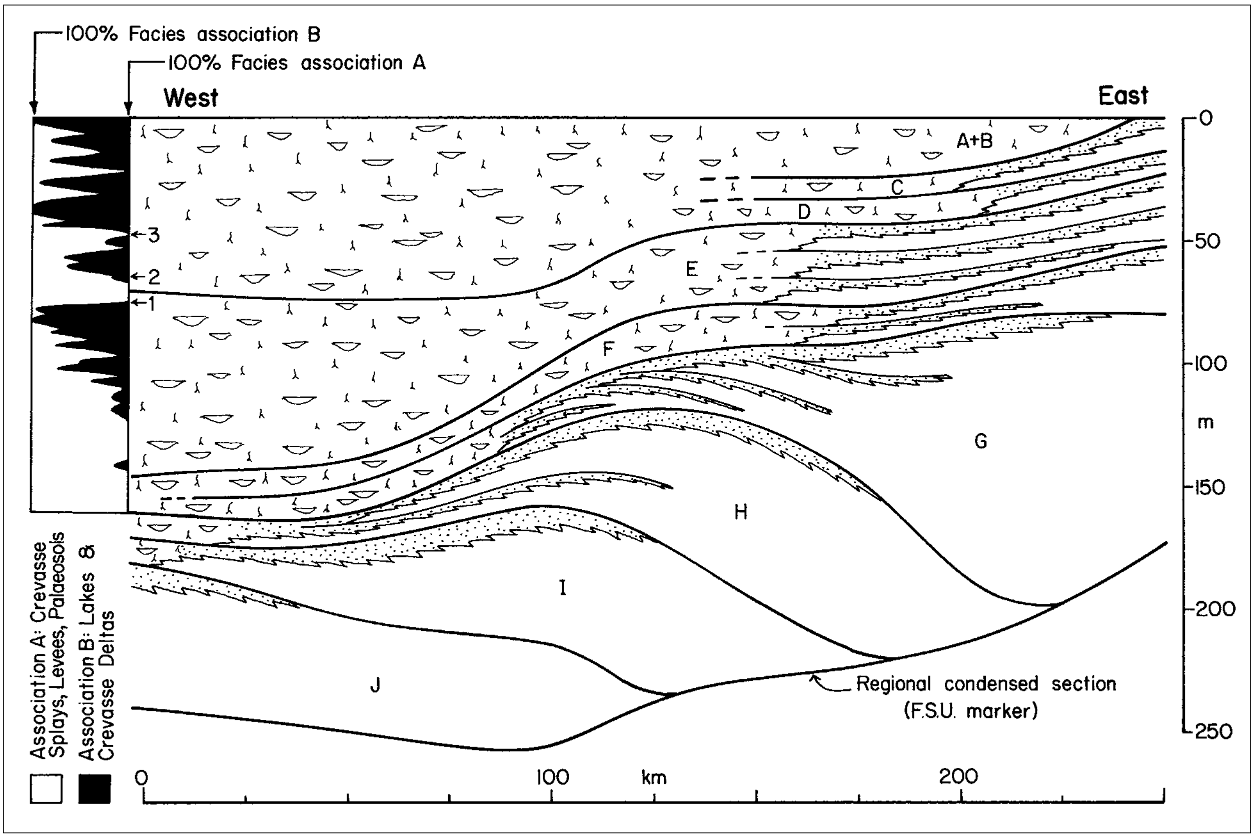


Fig. 26 Summary stratigraphic section oriented E–W across the northern part of the study area (Fig. 1) showing, to scale, the thickness variations of the Dunvegan allomembers. Flexural subsidence was negligible in allomembers J–F, but became increasingly rapid in allomembers E–A as shown by the great expansion of the section towards the west. The increased accommodation rate is reflected in the vertical changes in alluvial facies in outcrop, with an overall upward increase in the proportion of waterlogged and subaqueous facies (association B) relative to better-drained facies (association A; from LUMSDON & PLINT, 2003).

levees, intermediate redox and better-drained palaeosols. Association B is dominated by lakes, mouth-bars, organic and poorly-drained palaeosols. Association A, indicating relatively well-drained conditions, dominates allomembers G to the lower part of E. From near the middle to near the top, allomember E is progressively dominated by association B, but the uppermost 5 m shows an abrupt return to better-drained palaeosols (Fig. 26). Allomembers D to A show alternations between associations A and B, although overall, association B predominates upward. The basal units of the overlying Kaskapau Fm. comprise a 40–50 m thick transitional succession dominated by facies of association B, but also including brackish-water lagoonal deposits, overlain by marine deposits (Fig. 26).

When the vertical changes in alluvial facies associations are considered in the context of the *geometry* of their host allomembers, the relationship between facies and accommodation rate becomes clear. Allomembers H to F thin towards the NW suggesting minimal accommodation in that direction, and this is reflected in their constituent alluvial facies (association A), which are dominated by relatively well-drained environments. The upward change within allomember E, from association A to relatively poorly-drained association B is interpreted to record an upward increase in accommodation rate. This change is independently suggested by the isopach ‘moat’ in allomember E, attributed to renewed flexural subsidence (Figs. 15, 26). A 5 m thick pedocomplex at the top of allomember E suggests a final phase of very low accommodation rate. Allomembers D–A are again dominated by poorly-drained freshwater environments, which pass upward into rocks of the basal Kaskapau Fm. which show progressively increasing marine influence. This facies succession reflects an increasing accommodation rate, which is also suggested by regional isopach patterns (Fig. 16).

7.4. Valley systems

Valley systems can be mapped for up to 320 km, and certainly extend to the NW beyond the study area (PLINT, 2002; PLINT & WADSWORTH, in review). Representative maps of valley systems on allomembers F and E are shown in Figs. 27 & 28. A crucial observation is that, at least over the study area, valleys show no systematic change in depth with distance upvalley. The considerable length, modest depth (average 21 m) and lack of longitudinal shallowing suggests that the valleys might reflect both eustatic and discharge-related forcing mechanisms. The incision of valleys into offlapping deltaic deposits suggests that there was some relation-

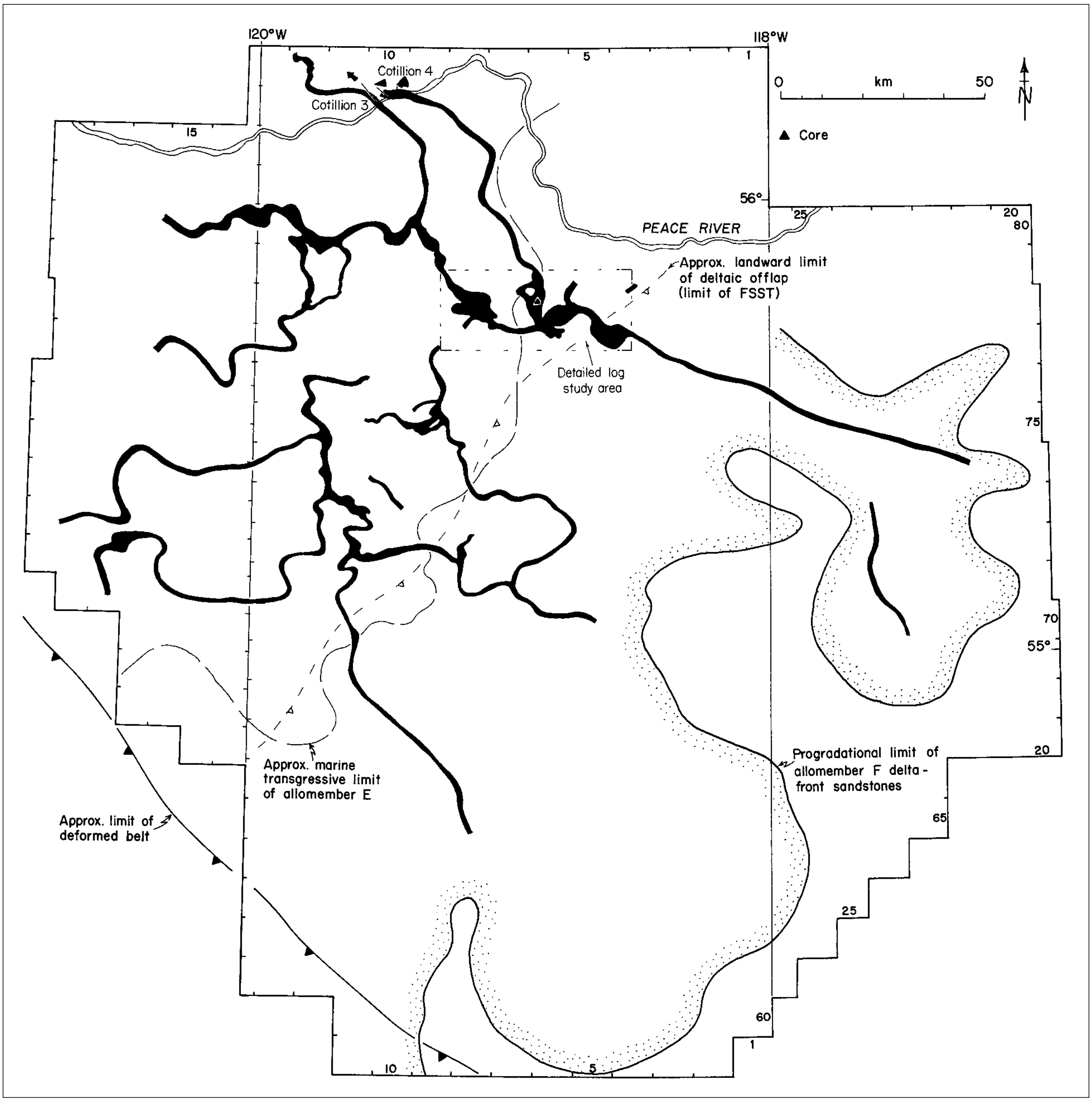


Fig. 27 Palaeogeographic map of allomember F showing the maximum progradational limit of the marine shoreline, the boundary between aggradational (HST) and offlapping (FSST) parasequences, the distribution of valleys, and the maximum transgressive limit of the succeeding allomember E shoreline (from PLINT & WADSWORTH, in review).

ship to relative sea-level fall which triggered erosion, starting at the offlap break fronting the HST. However, valleys seem ‘too long’ for the depth of incision, and the lack of headward shallowing suggests that an additional mechanism operated from upstream to effect incision. It is possible that cyclical changes in the ratio of sediment load to discharge might have been responsible for river incision far upstream (PLINT, 2002). It is not unlikely that changes in river discharge and small eustatic changes were both linked to global climatic oscillations in the Milankovitch band, and in combination, were responsible for most of the changes in accommodation pattern. Intermittently-active flexural subsidence was the third recognizable influence on the system.

7.5. Interfluve palaeosols

Interfluve palaeosols are recognized on the basis of abundant clay coatings, Fe oxide nodules and mottles which indicate the presence of Bt horizons. The prolonged stability of these surfaces is also indicated by the fragmentation, ageing and assimilation of clay coatings, the intense biological reworking of the matrix, concentration of residual Zr and TiO_2, and increased kaolinite content. Minor erosion of the interfluve surface is suggested by the presence in the palaeosols of papules, pedorelicts and embedded grain argillans, all indicative of minor remobilization of the palaeosol. The sharp irregular upper boundary of the palaeosol

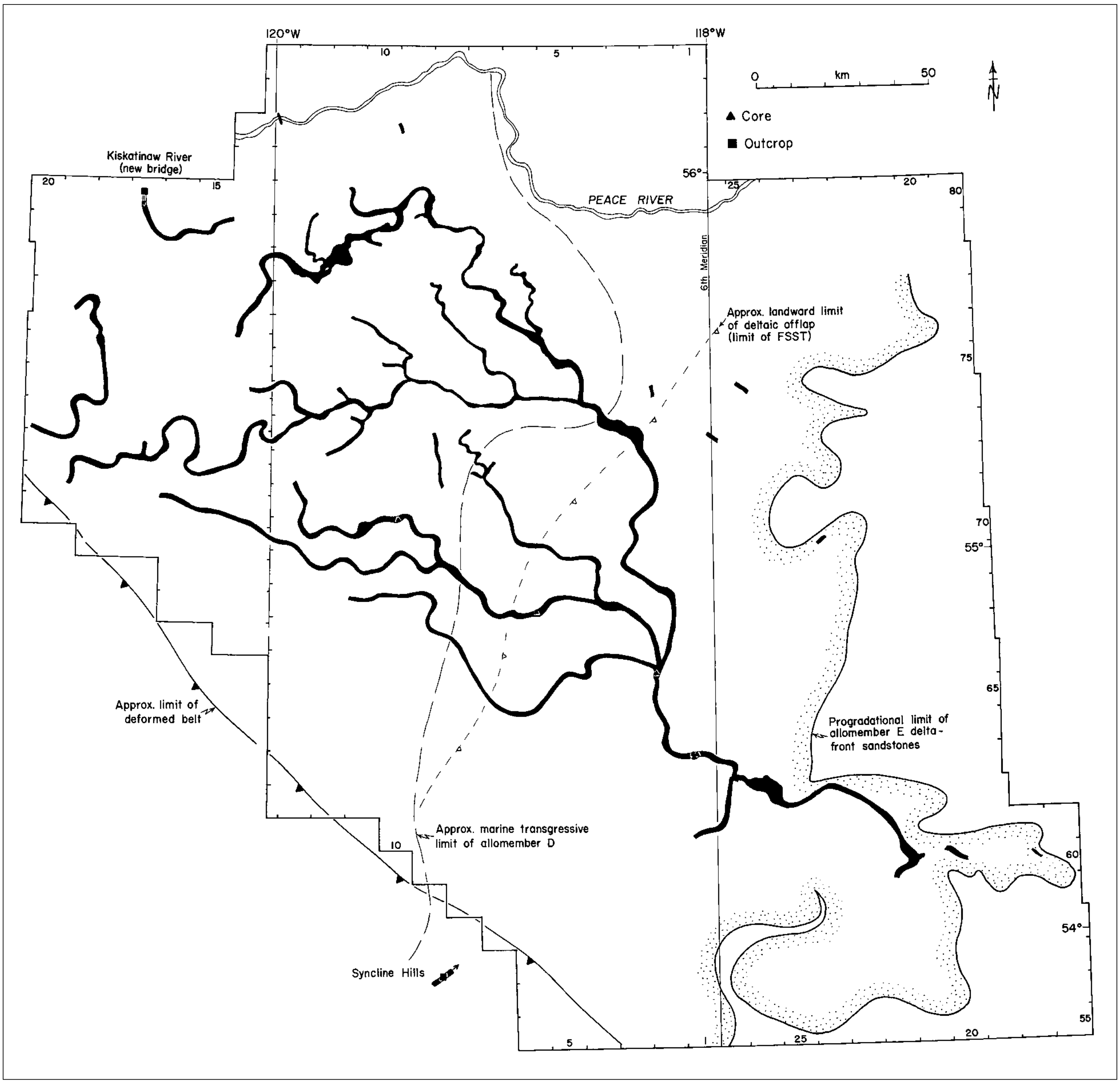


Fig. 28 Palaeogeographic map of allomember E showing the maximum progradational limit of the marine shoreline, the boundary between aggradational (HST) and offlapping (FSST) parasequences, the distribution of valleys, and the maximum transgressive limit of the succeeding allomember D shoreline (from PLINT & WADSWORTH, in review).

(i.e. the sequence boundary) also indicates subtle erosion (McCARTHY et al., 1999). Immediately above the sequence boundary, the sediment tends to be well-laminated and grain size increases, containing abundant silt and very fine sand. This indicates renewed supply of sediment to the surface and low levels of physical reworking. The presence of abundant, partially-oxidized siderite and multiple surface organic layers indicates increased water-logging and flooding in response to a rising, but fluctuating water table.

The stratigraphic position and vertical spacing of these very distinctive palaeosols is predictable to within 1–2 m on the basis of the projection of sequence boundaries from nearby well logs. This gives us confidence in our interpretation of sequence-bounding unconformities on the basis of well log correlations. The characteristics of the interfluve palaeosols indicate very significant depositional hiati, of the order of tens of ky, which corroborates the evidence provided by valleys for lengthy periods of negative accommodation across the coastal plain.

7.6. Sharp-based delta-front sandstones

Progradation of a delta typically produces a succession in which prodelta mudstone grades up into interbedded sandstone and mudstone and finally clean delta front sandstone. This Walthurian succession of facies indicates a gradual change in environmental conditions from deeper to shallower. In contrast to this ‘norm’, erosive-based delta front sandstones are widespread in the Dunvegan Fm. Swaley cross-stratified or cross-bedded

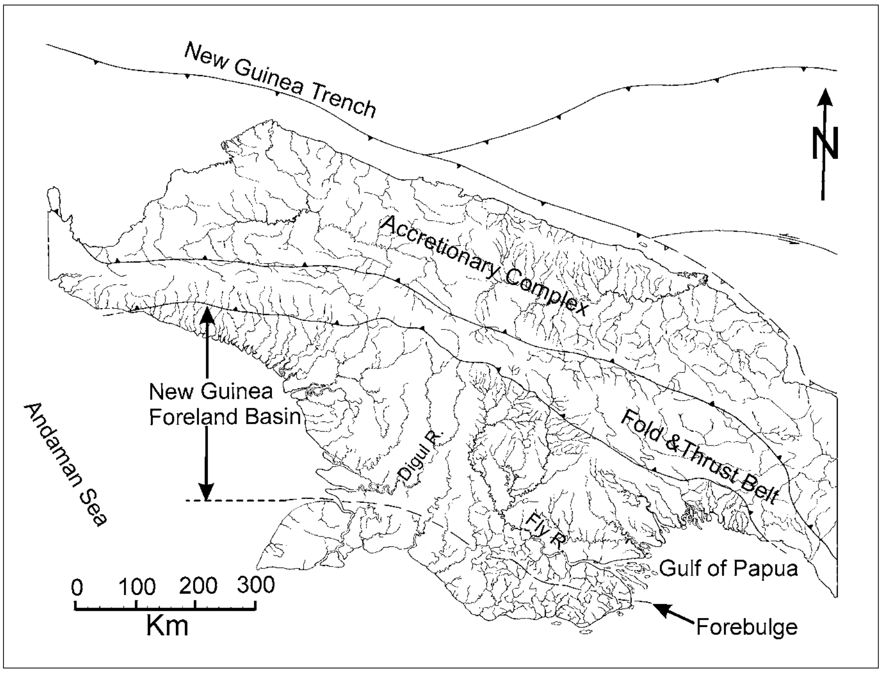


Fig. 29 Map of rivers in the central portion of the New Guinea foreland basin. The distinct drainage divide close to the southern coast is interpreted to reflect the position of the forebulge. Note how the Fly and Digul river systems flow initially perpendicular to the fold and thrust belt, but distally, are diverted by and flow parallel to the forebulge.

sandstone of the delta-front rests sharply on laminated mudstone or siltstone of the prodelta (Fig. 8). The contact is generally ornamented with large gutter-casts oriented shore-normal. This succession implies removal of accommodation causing shallow water processes to impinge directly on a muddy, low-energy prodelta sea floor. Eustatic fall or tectonic uplift seem to provide the only reasonable explanation for this facies succession (PLINT, 1988, 1996; PLINT & NUMMEDAL, 2000). It cannot be satisfactorily explained simply by changes in sedimentation rate.

Isopach maps provide no evidence for regional tilting, particularly in allomembers J–F, and therefore it is difficult to attribute accommodation loss and relative sea level fall to flexural uplift. In view of this geometric constraint, the pronounced offlap of deltaic parasequences has been attributed to eustatic fall (PLINT et al., 2001). One consequence of eustatic fall would be accommodation loss at the shoreline, resulting in the development of sharp-based delta-front sandstones.

7.7. Palaeodrainage patterns

Observations and computer simulations (GARCIA-CASTELLANOS, 2002) reveals two consistent and characteristic features of the drainage system in foreland basins. Smaller tributary streams drain perpendicular to the orogen but merge into a large river that flows parallel to the orogen. Secondly, the large axial river lies on the orogen-distal side of the basin, close to the forebulge. This pattern is exemplified by the New Guinea foreland basin (Fig. 29) in which the tributaries of the two main rivers, the Fly and Digul flow perpendicular to the orogen before turning abruptly to east and west, draining into the Gulf of Papua, and the Arafura Sea respectively. In the south a low but distinct drainage divide is revealed by the small rivers that traverse the southern margin of the island (Fig. 29) and it seems likely that this divide marks the crest of the forebulge. The interpreted forebulge lies 250–300 km from the deformation front in the New Guinea Highlands; this distance is consistent with the flexural wavelength of the lithosphere, and with observations in other foreland basins.

Drilling data are sufficiently dense that the course of the Dunvegan palaeovalleys can be reconstructed with some confidence. Valleys on allomembers H, G and F have a broadly rectilinear pattern that is interpreted to reflect control of the precursor rivers by faults in the Palaeozoic basement. A large SE-prograding lowstand delta is evident in allomember F (Fig. 27). Overall, valleys are oriented perpendicular to the coast. The valleys on allomember E show a different pattern (Fig. 28) in which all the tributary valleys flowing from the W and SW merge eastward into a single trunk valley that flows SE, parallel to the coast for over 150 km before terminating in a large lowstand delta lobe. This pattern bears a remarkable similarity to that of the Digul and Fly rivers in New Guinea. It is postulated that a subtle forebulge deflected the main valley on allomember E from a shore-perpendicular to a shore-parallel trend. The drainage pattern and main tectonic elements of New Guinea and of allomember E are compared, at the same scale, in Fig. 30. Valleys on allomembers H to F are not deflected shore-parallel. This might be evidence that a forebulge was not a significant topographic element on the coastal plain. This in turn might reflect a phase of negligible flexural deformation, which is also indicated by the isopach maps. In contrast, the dramatic shore-parallel re-orientation of valleys in allomember

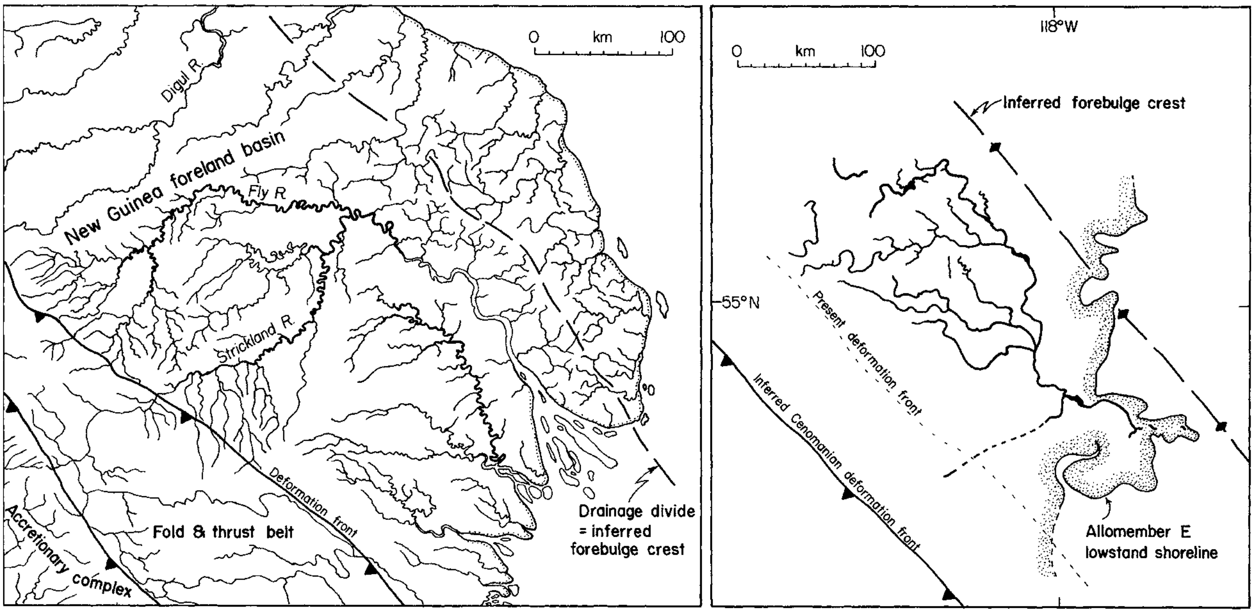


Fig. 30 Maps comparing the pattern, scale and palaeogeographic relationships of the modern Fly River (flipped and rotated) so as to mimic the orientation of the valley system in Dunvegan allomember E.

E suggests that rivers were at this time influenced by a forebulge. It may be no coincidence that the apparent uplift of a forebulge is linked to the onset of renewed flexural subsidence, as revealed by the isopach maps of allomembers E and A–D.

7.8. Quantification of Eustatic Change

Isopach maps (Figs. 10–14) show that in allomembers J–F, flexural subsidence was unimportant. Nevertheless, transgressions and regressions, and the cutting and filling of valleys continued to occur. Because the geographic limits of transgressive and regressive shorelines, and the seaward limits of valleys can be mapped with some accuracy, it is possible to determine the *horizontal* extent of transgressive and regressive events in each allomember. By applying geologically realistic gradients to the surfaces over which the shoreline migrated, and including the thickness of vertically-accreted coastal plain deposits, it is possible to make some first-order estimation of the vertical change of sea-level. This approach is summarized in Fig. 31.

The maximum transgressive limit of marine shorelines commonly lies within 10–20 km of the previous HST/FSST offlap break, and rarely inundates the HST coastal plain to landward. This implies a consistent balance between accommodation and sediment supply for each T/R cycle. The width and inferred slope of the FSST (60 km x 1:2:500) implies a slope height of about 24 m (Fig. 31). An additional average of 11 m of coastal plain sediment accumulated during the TST and HST, suggesting a minimum vertical accommodation of about 35 m. Calculation of water and sediment loads on the basis of simple Airy isostasy suggests that a eustatic rise of about 24 m would result, after sedimentation to sea level, in about 35 m of vertical accommodation (PLINT & WADSWORTH, in review). This estimate of eustatic change does not take account of the elastic strength of the lithosphere which would tend to decrease the amount of subsidence for any given sediment and water load. It is therefore probable that eustatic excursions were somewhat less than 24 m. It is perhaps no coincidence that the average depth of valleys (21 m) is close to the eustatic change inferred on geometric criteria. Modest eustatic fall at the coastline drove FSST delta progradation and also probably effected river incision across the HST and older FSST surfaces. However, the small (<24 m) and slow (?50–100 ky) eustatic change that is indicated by this analysis seems unlikely to have been capable of driving valley incision for >300 km upstream, as is observed (PLINT & WADSWORTH, in review). We are led to infer an additional, upstream forcing mechanism, perhaps linked to changes in discharge:sediment load ratio, controlled by climatic cycles in the Milankovitch-band.

8. CONCLUSIONS

A regional stratigraphy based on marine transgressive surfaces and approximately equivalent subaerial interfluve surfaces allows the deltaic Dunvegan Formation to be subdivided into ten transgressive–regressive allomembers with an average duration of <200 ky. Allomembers prove to be practical and robust stratigraphic units, well-suited to a data-set that consists mostly of well logs. More sophisticated interpretation of changing accommodation patterns is based on analysis of the stacking pattern of component parasequences within allomembers. This approach allows transgressive and highstand systems tracts to be recognized. These contain

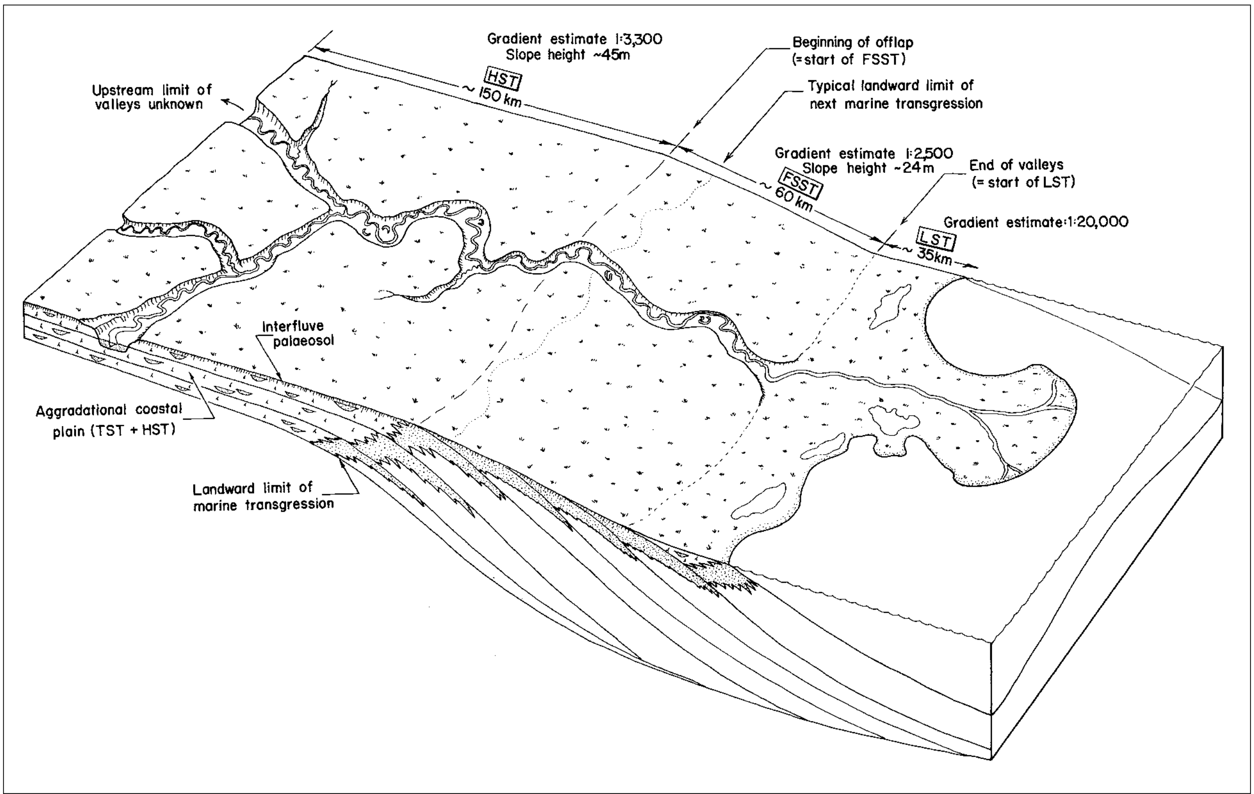


Fig. 31 Block diagram summarizing the stratigraphic, geographic and topographic relationships between valleys and various systems tracts of a delta complex. Strata assigned to the TST and HST include aggradational coastal plain deposits. Offlapping deltaic strata of the FSST lack coeval coastal plain deposits and prograde an average of 60 km during slow relative sea level fall. Fluvial incision is inferred to have occurred during this time, and was accompanied by development of very mature interfluve palaeosols. Valley systems generally terminate about 35 km from the maximum progradtional limit of the delta front, and this limit is interpreted to indicate the FSST/LST boundary. Aggradational and onlapping LST strata average 35 km wide, but include major lowstand deltas constructed at the termini of valley systems. On the basis of the assumed slopes and measured geographic dimensions, it is possible to infer cyclic eustatic changes of about 24 m (see text for discussion) (from PLINT & WADSWORTH, in review).

both marine and aggradational coastal plain deposits. Offlapping deltaic parasequences that lack an updip alluvial component are assigned to the falling stage systems tract. Aggradational to onlapping parasequences at the seaward limit of the allomember are assigned to the lowstand systems tract. The seaward termini of valleys are considered to mark the geographic boundary between FSST and LST. Valley systems that are typically 1–2 km wide, average 21 m deep, and > 320 km long are incised into the top surfaces of most allomembers; valleys have been mapped in detail for allomembers H to E. Valleys are back-filled with muddy heterolithic tidal facies within a few tens of km of the valley mouth. Further updip, fluvial processes dominated, producing multi-storey sandstone-dominated channel-fills. A weak tidal influence is inferred locally, and the effects of semi-diurnal tides extended about 30 km inland from the marine shoreline. Major palaeosols mantle interfluves. A wide range of pedological features in the interfluve palaeosols indicate a protracted hiatus with extensive physical, chemical and biological modification of the parent material. The drainage of interfluve palaeosols progressively deteriorated with increasing distance from valley margins.

Allomembers can be grouped on the basis of isopach patterns. Allomembers J–F have a sigmoidal prismatic geometry, successively offlapping to the SE. There is no evidence of thickening toward the orogen. In contrast, allomembers E–A show progressive development of a depocentre along the western margin of the basin, presumably in response to renewed flexural subsidence. The increasing accommodation rate on the coastal plain had a profound effect on marine deltas further downdip, which were starved of sediment, leading to progressive backstep of shorelines. Simultaneously, alluvial deposits within the depocentre show an upward increase in the proportion of subaqueous to subaerial facies, culminating in the incursion of brackish and finally marine waters. Thus tectonic subsidence rate had a first-order affect on both the volume of sediment available to build marine deltas and also on the local character of facies that accumulated on the coastal plain. Renewed flexural subsidence appears to have resulted in subtle uplift of a forebulge, resulting in dramatic deflection of river systems in allomember E, and also probably in allomember D.

Despite the clear tectonic signature, transgressions and regressions of the shoreline took place over similar distances, regardless of whether subsidence was negligible (allomembers J–F) or accelerating (allomembers E–A). This suggests that modest eustatic changes also influenced the accommodation rate. On the basis of the measured horizontal excursions of the shoreline, the vertical thickness of alluvial strata, and realistic alluvial gradients, an average eustatic excursion of about 24 m is calculated. The incision of valley systems is attributed in part to periods of eustatic fall. However, valleys seem too long to be explained by eustasy alone, and hence secular changes in river discharge:sediment load ratio are postulated as an additional forcing factor. Climatic cycles in the Milankovitch band may have been responsible for both eustatic and discharge variations.

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