ESTIMATION OF SOURCE-TO-SINK MASS BALANCE BY A FULCRUM APPROACH USING CHANNEL PALEOHYDROLOGIC PARAMETERS OF THE CRETACEOUS DUNVEGAN FORMATION, CANADA

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ABSTRACT: Trunk rivers transport the bulk of the sediment in a source-to-sink (S2S) system, and total mass passing through any cross section (i.e., fulcrum) of a trunk-river over geologic time should allow matching of source-area sediment delivery budgets to downstream sediment volumes deposited in the basin. We analyze the paleohydrology of ancient trunk channels and link downstream deltaic strata of Alloformation E of the Cretaceous Dunvegan Alloformation in the Western Canadian Sedimentary Basin to test the total mass-balance fulcrum approach. Bankfull channel depth and width, grain size, paleoslope, velocity, and discharge are derived from outcrop, core, and well logs. Some parameter estimates use multiple methods, providing a range of values and serve as a cross check of independent methods. Estimates of annual flood frequency and paleodischarge, associated with long-term geologic-time estimates, are derived from chronostratigraphic analysis and allow calculation of cumulative sediment discharge. Isopach maps are used to measure sink-area sediment volumes. The results indicate that the trunk river of Alloformation E was 10–15 m deep and 150–250 m wide, carried fine- to medium-grained (<200 microns) sand, and flowed over a low-gradient paleoslope of 4.1–6.1x10^-5. Annual total sediment discharge is estimated to have ranged from 5.4 to 12x10^6 m^3. Within 25,000 years, the river is estimated to have transported 135–307 km^3 of sediment into the basin. This is consistent with the 130 km^3 of sediment mapped in the study area. However, the upper-range estimate of sediment delivered into the sink is 2.5 times the measured sediment volume in the map area, which, if accurate, suggests significant sediment escape. This supports the hypothesis that in Dunvegan time, mud was widely dispersed southward, along the Alberta Foreland Basin by geostrophic currents associated with storm processes and counterclockwise oceanic gyres in the Cretaceous Seaway.

INTRODUCTION

The source-to-sink (hereinafter referred to S2S) concept encompasses source sediment demudation from the initial catchment area, transportation or transient storage through the system, and ultimate deposition in the basinal sink (Castelltort and Van Den Driessche 2003; Somme et al. 2011; Romans and Graham 2013; Jaeger and Koppes 2015; Leithold et al. 2015; Bhattacharya et al. 2016). The S2S concept has been increasingly applied to assess sediment budgets of both modern and ancient systems (Allen 2008b; Somme et al. 2009; Carvajal and Steel 2012; Somme and Jackson 2013). Given an ideally closed S2S system, the total sediment volume produced from the catchment should match the volume accumulated in downstream sinks (Paola 2000; Strong et al. 2005; Covault et al. 2011; Somme et al. 2011; Hajek and Wolinsky 2012; Paola and Martin 2012; Allen et al. 2013; Matenco et al. 2013; Michael et al. 2013; Petter et al. 2013; Holbrook and Wanans 2014; Sadler and Jerolmack 2015).

In estimating total mass balance, it is assumed that the total sediment volume passing through any cross section of a longitudinal trunk fluvial system in the transfer zone during a certain time interval should match both the sediment volume delivered from the source area and the sediment volume passing through that cross section. This point in a cross section in the sediment routing system serves as a “fulcrum” that can be used to balance the total mass between source and sink (Holbrook and Wanans 2014). The largest-scale incised trunk rivers transport the bulk of the sediment in a S2S system, and best serve as key fulcrum points (Blum et al. 2013; Romans and Graham 2013; Romans et al. 2015). The fulcrum approach requires calculating the instantaneous paleodischarge and other paleohydraulic parameters of trunk rivers and integrating this over the duration of the associated stratigraphic record of a given S2S system. The fulcrum approach can be applied without knowing source area properties, although knowledge of the catchment climate regime can be very useful, and does not require a closed downstream basin (Asselman 1999; Bhattacharya and Tye 2004; Hutton and Syvitski 2008; Parker et al. 2008; Whittaker et al. 2010; Blum et al. 2013; Holbrook and Wanans 2014; Bhattacharya et al. 2016). The fulcrum approach enables estimates of catchment areas and relief, amount of sediment transported in the systems, and the volume of sediment accumulation in the basins. These estimates also can be compared to those made from other approaches, such as the BQART method (see Syvitski and Milliman 2007), the use of regional climate curves (see Davidson and North 2009), other numerical models (Paola and Mohrig 1996; Paola 2000; Lee et al. 2002; Fedele and Paola 2007; Duller et al. 2010; Whittaker et al. 2011; Hajek and Wolinsky 2012; Allen et al. 2013; Liang et al. 2015), volumetric calculations from the stratigraphic record (Carroll et al. 2006; Carvajal and Steel 2012; Petter et al. 2013), and thermochronometric data (Michael et al. 2014).

The purpose of this paper is to estimate total mass balance by calculating long-term sediment flux through fulcrum analysis of trunk-river
deposits, with mapped downstream sediment accumulations in the Upper Cretaceous Dunvegan Alloformation, in the Alberta Foreland Basin, Canada, using a combination of outcrop and subsurface data (Fig. 1). The Dunvegan Formation has been extensively studied, incorporating extensive outcrop and subsurface data to develop a high-resolution allostratigraphic and chronostratigraphic framework (Bhattacharya and Walker 1991a, 1991b; Bhattacharya 1992; Bhattacharya 1993; Gingras et al. 1998; McCarthy et al. 1999; Plint 2000, 2002; Plint et al. 2001; McCarthy and Plint 2003; Plint and Wadsworth 2003, 2006; Plint and Kreitner 2007; Plint et al. 2009). Bhattacharya and Walker (1991) and Plint (2000) subdivided the Dunvegan Alloformation into 10 allomembers, seven of which are labelled G to A in ascending order, and these allomembers have been correlated and mapped on the basis of regional transgressive surfaces (Fig. 2). In sequence stratigraphic terms, each allomember represents a flooding-surface-defined genetic stratigraphic sequence (cf. Galloway 1989), each of which contains several parasequences (cf. Van Wagoner et al. 1990) that prograded towards the SE into the Cretaceous Interior Seaway. Eight of the allomembers contain recognizable paleovalley systems that can be linked to their downdip shoreline deposits (Bhattacharya 1993; Plint and Wadsworth 2003). The Dunvegan is thus an ideal candidate for fulcrum analysis owing to the ability to identify and characterize the trunk rivers within a specific incised-valley system and correlate these to the downdip lowstand deltaic and offshore shelf systems tracts (Bhattacharya and Walker 1991a; Plint 2002; Plint and Wadsworth 2006) (Fig. 3). More importantly, these Dunvegan lowstand system tracts are smaller-scale features than modern continental-scale rivers, and are entirely confined to the Western Interior Seaway. Therefore, the Dunvegan represents a potentially closed system (Bhattacharya et al. 2016). In this study, Allomember E was selected for more detailed fulcrum analysis because it has the clearest linkage between the tributary drainage system and its lowstand delta deposit (Bhattacharya 1993; Plint and Wadsworth 2006). Allomember E also comprises a river-dominated delta system, with minimal tidal influence and limited wave reworking (Bhattacharya and Walker 1991a; Bhattacharya 1993) (Fig. 4). Previous mapping of the lowstand river-dominated delta lobes (Bhattacharya 1993) enables volumetric estimation of the downstream sediment accumulation (Bhattacharya and MacEachern 2009) (Fig. 3).

METHODS

The workflow for testing this fulcrum approach includes: 1) identifying the trunk river and estimating trunk-river dimensions and attributes for...
Fig. 2.—Regional cross section across the Alberta Foreland Basin, illustrating the allostratigraphic interpretation of the Upper Cretaceous Dunvegan Formation. The Dunvegan comprises several stacked allomembers (A to G). Each allomember is bounded by a regional transgressive flooding surface. Each allomember internally consists of several smaller-scale, offlapping, shingled parasequences that map as sandy delta lobes and their associated prodelta mudstones. Many delta lobes can be correlated into their updip feeder valleys, such as in Parasequence E1 (Bhattacharya and MacEachern 2009).

Fig. 3.—Paleogeographic map of valleys and lowstand deltas in Allomember E of the Dunvegan Formation (modified after Bhattacharya and MacEachern 2009).
further paleohydrologic estimates, 2) calculating instantaneous bankfull paleodischarge and sediment load, 3) estimating annual sediment discharge using paleoclimate proxies (e.g., floodplain paleosol and paleobotanical data) and appropriate modern climatic analogs, 4) determining total duration of the river and its valley from stratigraphic analysis, 5) evaluating total mass delivered to and through the trunk river over this geological time period, 6) measuring sediment volumes in the sink, and 7) reconciling sediment volumes delivered to the fulcrum point from the source area to measured sediment accumulations in the sink area (Fig. 5).

**Fig. 4.**—Allomember E, with a well identified trunk river, is the most river-dominated delta with the least tidal influence. This allows accurate paleohydrological parameter estimates without tidal landward reworking (modified after Bhattacharya 2006, which was modified based on Galloway 1975).

**Fig. 5.**—Work flow for fulcrum analysis.
Dimension Analysis of Trunk Rivers

The criteria for identification of trunk rivers in ancient deposits have been reviewed by Olariu and Bhattacharya (2006) and Bhattacharya et al. (2016). In this study, regional outcrop and subsurface mapping (e.g., Fig. 3) has identified both tributary and trunk valleys in the Dunvegan (Plint 2002; Plint and Wadsworth 2003). To estimate trunk-river paleohydraulics, dimensional information regarding the trunk rivers is required, including average bankfull channel depth \( d_m \) and bankfull channel width \( W_c \). Grain size \( D \) and channel depth can be used to estimate paleoslope \( S \), and average flow velocity \( U \).

**Trunk-River Bankfull Depth.**—Bankfull channel depth can be estimated on the basis of the thickness of fully preserved fining-upward facies successions (stories) as measured in outcrops, cores, or well logs with a small correction for compaction (Bridge and Tye 2000; Bhattacharya and Tye 2004; Adams and Bhattacharya 2005; Miall 2006; Allen 2008b; Davidson and Hartley 2010, 2014; Whittaker et al. 2015; Hajek and Heller 2012; Hajek and Wolinsky 2012; Blum et al. 2013; Hampson et al. 2013; Petter et al. 2013; Bhattacharya et al. 2016). Sedimentary structures, such as dune-scale cross-bedding, and bar accretion deposits, can also be used to estimate channel depth, independently of story thickness (LeClair and Bridge 2001; Bhattacharya and Tye 2004; Davidson and Hartley 2010; Foreman et al. 2012; Hajek and Wolinsky 2012; Holbrook and Wanas 2014). Dune height can be estimated from thickness of dune-scale cross-set using certain empirical equations, which have been shown to scale to channel depth (see LeClair and Bridge 2001; Reesink et al. 2015). Thickness measurements have an error of about 10% (i.e., 10 cm per meter) and estimates of channel depths can have uncertainties of up to 60% using the LeClair and Bridge (2001) method, and about 25% using measurements of story thickness.

**Trunk-River Width.**—Channel and channel-belt width were measured from the previously published outcrop studies (e.g., Plint 2002; Plint and Wadsworth 2003) oriented perpendicular to the flow direction. In subsurface, or areas of limited exposure, channel width was estimated based on various empirical equations depending on the fluvial style (e.g., single-thread meandering channel vs. braided channels; as presented in Leopold and Maddock 1953; Bridge and Mackey 1993; Hampson et al. 2013; and Holbrook and Wanas 2014). Large-scale lateral-accretion sets in point-bar deposits in Dunvegan outcrops show that the trunk river of the Allomember E valley was a single-thread meandering channel (Plint and Wadsworth 2003) (Fig. 6). Uncertainties of length measurements of accretion beds are about 10%, and uncertainties in converting point-bar accretion length to channel width have an uncertainty of about 25%. As a consequence, width estimates based on outcrop exposures of meander belts have an uncertainty of about 35%.

Other scaling relationships of channel dimension are also applied to estimate channel width and constrain the results. For example, based on Gibling’s (2006) channel width/thickness ratio classification, trunk rivers in valley fills typically have a width/thickness ratio ranging from 5 to 50 (see Blum et al. 2013), depending on tectonic uplift and climate. The *entrenchment ratio*, defined as valley width over bankfull channel width (Rosgen 1997), was also used to evaluate estimates of channel width.
Grain Size.—Grain size can be used to estimate shear stress on the bed, which is determined by flow depth and relief, and controls the sediment transport mode (bedload vs. suspended load) and capacity, coupled with velocities of water flow and sediment transport (Hajek and Wolinsky 2012). The median grain size ($D_{50}$) value was used to estimate water discharge and to estimate paleoslope (Holbrook and Wanans 2014), and the suite of grain size distribution values ($D_{20}$, $D_{50}$, $D_{100}$, and $D_{16}$) are used to estimate total instantaneous sediment discharge. Grain size was derived directly from core and thin-section data, and used to generate cumulative distribution plots. The grain-size measurements, made with a standard grain-size card, have an error of about 1/2 phi (about 30%).

Paleoslope.—Paleoslope is represented by the longitudinal profile of a sedimentary system (Posamentier and Vail 1988; Blum and Tornqvist 2000) and can be estimated using multiple approaches, such as the use of empirical equations, stratigraphic stacking patterns, and lap-out relationships, including measurement of the onlap distance of coastal prisms versus the thickness mapped on regional dip-oriented cross sections (e.g., Blum and Tornqvist 2000), elevation drop of channel bases along a valley long profile (e.g., Bhattacharya et al. 2016), and paleohydrologic data (i.e., backwater and bayline length) (Blum et al. 2013).

We used the empirical equation of Holbrook and Wanans (2014): 

$$\tau_{w50} = (d_m S)/(PD_{50}) = \text{constant} \tag{1}$$

where $S$ is slope, $P$ is submerged dimensionless density of sand–gravel sediment ($\rho - \rho_w$), and $D_{50}$ is grain size. An assumed sediment density of 2.65 g/cm$^3$ yields a dimensionless submerged density ($P$) of 1.65. The bankfull Shields number for dimensionless shear stress ($\tau_{w50}$) is 1.86 (also see Parker 1978; Dade and Friend 1998; and Parker et al. 1998).

Backwater length is defined as the landward limit that downstream effects can propagate upstream (Paola and Mohrig 1996; Hajek and Wolinsky 2012; Blum et al. 2013) (Fig. 7), which can be used to estimate slope associated with channel depth, and may determine the gravel–sand boundary in the upstream reaches of fluvial systems (Bhattacharya et al. 2016). The Backwater length is expressed as

$$L_b = d_m S \tag{2}$$

where $L_b$ is backwater length. Slope estimates are considered to be uncertain by a factor of 2.

Flow Velocity.—Flow velocity ($\bar{U}$) was estimated using 3D bedform-phase diagrams of Rubin and McCulloch (1980) (Bhattacharya and Tye 2004; Bhattacharya and Maceachern 2009; Bhattacharya et al. 2016), as bedforms are a function of flow velocity, depth, and grain size. Sedimentary structures and grain-size data are directly observed from outcrop and core. It is assumed that the dominant bedforms record bedload transport during flood events. In the Dunvegan, dune-scale cross beds are the dominant sedimentary structure observed in the cores and outcrop, and higher-flow-regime bedforms, such as upper plane bed, were not observed. Lower flow regimes are reflected by ripple cross lamination in upper parts of channel stories and likely reflect waning, lower flow stages of the river hydrograph. Flow depth is estimated from channel-story thickness. Flow velocity was also estimated independently, based on hydraulic equations. Here we used the Chézy coefficient, hydraulic radius, and estimated slope to calculate the bankfull flow velocity (Holbrook and Wanans 2014), and compare this to the value estimated using the bedform-phase-diagram method. The bankfull flow velocity is a function of basal shear stress, and is determined by gradient and flow depth. The mean flow velocity of the trunk river of Allomember E was estimated using the equation:

$$\bar{U} = C_f (RS)^{1/2} \tag{3}$$

where $C_f$ is the Chézy Coefficient ($C_f = (g)^{0.581} (d_m/k_s)^{1/6}$, where $d_m$ is mean bankfull water depth, and $k_s$ is roughness element ($k_s = 3D_{50} + 1.1(1 - e^{-2.5})$), where $A =$ bedform height = $d_m / 8$ (after LeClair and Bridge 2001); $\psi = A/\lambda$, where $\lambda =$ bedform wavelength = 7.3 $d_m$ (see Equation 4b), $R$ is hydraulic radius, and $S$ is paleoslope.

Paleodischarge Estimation

Paleodischarge was calculated using the flow velocity and channel dimensions estimated using the methods described above. Paleodischarge of rivers was also estimated by integrating drainage-basin area with paleoclimatic parameters (e.g., Davidson and North 2009), and these are in
given depth to the full water column, \( \bar{u} \) represent the duration of trunk-river deposition are required. Mean formation, mean annual sediment discharge rate and long-term time scales longer geological period of time. In order to estimate the total sediment dune bedforms, which are interpreted to reflect the flood stage of the river. characterizes flow conditions at the time of migration of the dominant volume of sediment passing through any river cross section per second, from the equations and methods discussed above were used to evaluate the calculation method. The suspended-load discharge per unit width (\( \frac{Q_s}{w_c} \)) is obtained: 

For estimation of bankfull suspended-load discharge, the Van Rijn (1984) equation is used, which employs entrainment thresholds derived from grain-size distribution of bed load to determine suspended-load concentrations averaged over depth across a unit of channel width (see Holbrook and Wanas (2014)). The method comprises 13 steps with 12 equations, which constrain variables in the final Equation 8 (Appendix 2, see Supplemental Material). The reader is referred to Holbrook and Wanas (2014) and Van Rijn (1984) for a more detailed description of the calculation method. The suspended-load discharge per unit width (\( q_s \)) is multiplied by estimated channel width (\( w_c \)) to estimate total suspended-sediment discharge (\( Q_s \)) (Equation 8).

\[
Q_s = \frac{F w_c d_m c_s}{9} \quad (7)
\]

\[
Q_a = q_s (w_c) \quad (8)
\]

where \( F \) is the suspension factor, correcting for the concentration at the given depth to the full water column, \( \bar{u} \) is mean flow velocity, \( d_m \) is mean bankfull flow depth, and \( c_s \) is the concentration of suspended sediment at the reference depth. Two alternative methods were introduced to estimate suspended-load discharge (see Garcia and Parker 1991; Wright and Parker 2004); however, with the condition of low gradient and fine-grained sediment, the Van Rijn (1984) method is considered the most applicable in the Dunvegan (Holbrook and Wanas 2014).

Estimation of Annual Sediment Discharge

The values of bedload and suspended-load sediment discharge derived from the equations and methods discussed above were used to evaluate the volume of sediment passing through any river cross section per second, defined as an instantaneous discharge rate. This rate effectively characterizes flow conditions at the time of migration of the dominant dune bedforms, which are interpreted to reflect the flood stage of the river. Nevertheless, the fulcrum approach aims to quantify mass balance over a longer geological period of time. In order to estimate the total sediment volume delivered by the river during the given time period of stratal formation, mean annual sediment discharge rate and long-term time scales that represent the duration of trunk-river deposition are required. Mean annual sediment discharge can be estimated from rating curves of sediment discharge in analog modern systems; this is, however, not typically possible for stratigraphic systems. Mean bankfull discharge for various climate regions can be estimated as a function of bankfull depth and drainage-basin area (Simon et al. 2004; Davidson and North 2009). Nevertheless, few attempts have been made to estimate bankfull discharge on a yearly basis for deep-time systems.

In this study, we use modern analogs which show that the main proportion of the total annual sediment load is transported by rivers typically in a short-term period. These periods may reflect rainfall or seasonal climate events, resulting in flood-intensified peak flow periods, which typically occur over hours to days (Sadler 1981; Dott 1996; Clark 1998; Alexander et al. 2001; Meybeck et al. 2003; Vandenberghe 2003; Jerolmack and Sadler 2007; Kettner and Syvitski 2008; Fatorić and Chelleri 2012; Miall 2014; Cramer et al. 2015). The hydrological term “annual flood” refers to the period of maximum discharge for each year, regardless of whether the river has actually over topped its banks (Arora and Boer 2001). Compilations of annual flood durations in modern rivers show that they typically vary from a few days to a few weeks, but normally are on the order of one week (Wolman and Miller 1960; Andrews 1980; Nolan et al. 1987; Meybeck et al. 2003; Powell et al. 2006). From such statistical compilations, the following empirical equation has been obtained:

\[
Q_{\text{max}} = Q_{\text{base}} \times D/b \quad (9)
\]

where \( Q_{\text{max}} \) is annual discharge, \( Q_{\text{base}} \) is bankfull discharge, \( D \) is peak flood duration (days), and \( b \) represents the percentage of total sediment transported during peak flood. The flood duration and the proportion of sediment volume transported during flood events were estimated by reconstructing paleoclimatic conditions and classifying paleoclimatic regimes on the basis of modern analogs. The reconstructed paleoclimatic conditions are compared to the Köppen-Geiger climate classification for modern systems, which will allow for application of a modern climatic-zone analog (Kottek et al. 2006).

Evaluation of Cumulative Sediment Load

In order to estimate total sediment budget in the system, the total depositional duration must be integrated with estimate of annual discharge. The total depositional duration of individual parasequences of Allomember E was estimated by previous sequence stratigraphic analysis. The number of parasequences identified in each allomember is variable but ranges from \( 2 \) to \( 4 \), and the extent and maturity of paleosols related to valley incision cycles can be used to estimate durations of individual valleys (Bhattacharya 1992; McCarthy and Plint 1999). The total sediment load passing through the trunk-river fulcrum point in the Dunvegan Allomember E can be calculated by integrating the paleohydrologic parameters over the total duration.

Sediment Accumulation in the Sink

The total sediment volume deposited downstream of the trunk river was estimated using isochron maps of Allomember E (Bhattacharya 1993). The total isopach map of Allomember E was digitized in ArcGIS and converted to a vector grid comprising geographic coordinates, which enabled accurate estimates of sediment volume. Similarly, digitizing a sandstone isochron map of a single Parasequence E1 of Allomember E, in ArcGIS allowed sand sediment volume to be estimated. The sandstone isochron map of Parasequence E1 shows a lobate geometry, interpreted as a discrete delta lobe (Bhattacharya 1992). The map shows closure of the \( 4 \) m contour line, but the \( 0 \) and \( 2 \) m contour lines are not closed to the southwest due to loss of data in the thrust-fold-belt uplift. Isochon contours of the entire Allomember E, which include the shaly facies, thin uniformly basinward
from a maximum of 65 m, in the delta lobe center, down to about 20 m. The zero edge of downlapping shales was mapped by Plint et al. (2009). Where appropriate, we manually closed contour lines by assuming uniform geometric trends. These maps provide values of area and thickness that are used to calculate sediment volumes.

RESULTS

Allomember E of the Dunvegan Formation was deposited during a lowstand and includes an extensive tributary incised-valley system and a trunk valley that converged to feed a river-dominated progradational delta deposit (Figs. 2, 3) (Bhattacharya and MacEachern 2009). In order to reflect uncertainties, we estimated a range of reasonable parameters instead of a single value.

Trunk-River Dimensions

Trunk-River Bankfull Depth.—Plint and Wadsworth (2003) showed that individual channel-fill thickness in the valley systems range from 3 m to 10 m, with an average of 8–10 m. The isolith map of sandy valley fill associated with Parasequence E1 does not exceed 16 m thick, although some postdepositional truncation by the overlying transgressive surface is observed. Sand-filled valleys range from 0.5 km to 2 km wide (Fig. 8). Maximum thalweg channel depth in the Allomember E valley was estimated to be about 16 m, as measured from the thickness of fining-upwards channel stories in core in the valley fill (Fig. 9). In outcrop, inclined heterolithic stratification (IHS) of point-bar deposits in the valleys indicates a 12 m deep channel that is about 100–150 m wide (Fig. 6). Similarly, based on the data of Plint and Wadsworth (2003), Bhattacharya et al. (2016) estimated a large tributary channel that feeds the Allomember E trunk river, to be about 16 m deep and 150 m wide, based on heights of lateral-accretion bars from outcrop (Fig. 6). Statistics from well-log data, shown by Plint (2002), indicate that tributary channels of Allomember E are 10 m deep on average. Nonetheless, theoretically, trunk-river depth should increase downstream following a confluence of tributary channels.

Plint and Wadsworth (2003) documented the average cross-set thickness in the Dunvegan valley systems to be about 0.5 m (Fig. 9). This suggests bankfull channel depths ranging from 7 m to 18 m. Compilation of these channel-depth values, estimated from multiple methods, suggests mean bankfull channel depth values of about 10–15 m.
Trunk-River Width.—Trunk-river width should be a small fraction of valley width (Gibling 2006; Blum et al. 2013; Hampson et al. 2013); thus the channel widths should be less than 500 m, given the valley widths of 0.5–2 km. Single-thread meandering characteristics and a low gradient suggest that the entrenchment ratio should be greater than 2.2 (see Rosgen 1997), and accordingly the channel width should be less than 230 m. Plint and Wadsworth (2003) inferred that upstream tributary channels are about 100–150 m wide, and Bhattacharya et al. (2016) suggested that the trunk river of Allomember E must be somewhat wider than this. Therefore, we estimated trunk-river channel width to range from 150 to 230 m.

Grain Size.—Compilation of core data from nine wells (Appendix 1, see Supplemental Material) shows the overall sediment grain-size distribution of the Allomember E valley, from which key grain-size values (\(D_{50}\), \(D_{90}\), \(D_{84}\), and \(D_{16}\)) were derived (Fig. 10). Grain sizes of the trunk river varies from very fine to medium grained, with a \(D_{50}\) grain-size value of 0.20 mm (~200 microns), categorized as fine-grained sandstone. The relatively small and well-sorted grains suggest a low paleoslope.

Paleoslope.—Paleoslopes, estimated from Equation 1, are \(4.1-6.1 \times 10^{-5}\). In contrast, Plint et al. (2009) used qualitative reasoning to estimate paleoslopes of the Dunvegan alluvial and coastal plain to be in the range of \(2-3 \times 10^{-5}\), which are an order of magnitude steeper than the slopes calculated using the empirical equation herein.

Regional mapping of the Dunvegan Formation shows that conglomeratic facies are restricted to about 600 km inland (Stott 1982), which supports a low-gradient system. In addition, the bayline limit may also help characterize the slope. The bayline is defined as the boundary between the upper and lower delta plain, and its limit equals the tidal range divided by slope (Bhattacharya et al. 2016). Rare tidal deposits in the Dunvegan suggested a microtidal to low mesotidal regime (Plint 2003), suggesting a tidal height of \(<2\ m\). The two different paleoslopes yield the estimated bayline limits as high as 36–54 km, for the lower slope, or 4–10 km for the higher slope. Paleogeographic maps illustrated that the bayline limit (from the lower delta plain to river mouth) of Allomember E should be on the order of tens of kilometers. Additionally, Plint and Wadsworth (2003) suggested a 30 km tidal backwater length for the Dunvegan. The paleoslope estimated using this longer tidal backwater length would be on the order of \(10^{-5}\), which is more consistent with the lower paleoslopes calculated above using the empirical equation. Combining the estimates from various approaches suggests that the paleoslope estimated from the empirical equation is more reasonable than the steeper gradient originally hypothesized by Plint and Wadsworth (2003). The estimated lower slope is also in agreement with the dominant fine sand grain size observed in the valley fills.

Flow Velocity.—The flow velocity, estimated using the Chézy equation, is about 1.0 m/s. Using the estimated bankfull channel depth of 10–15 m, observation of predominantly upper-fine sandstone from cores, and the abundance dune-scale cross-bedding observed from both core and outcrops (Plint and Wadsworth 2003) (Figs. 8–10), the velocity estimated from the bedform phase diagram is in the range of 0.9–1.4 m/s (modified after Rubin and McCulloch 1980).

Discharge Estimation

The bankfull discharge \((Q_{bf})\) of the E1 trunk river is estimated to be \(1.6–3.7 \times 10^{3}\ m^3/s\); the bedload sediment discharge \((Q_{bd})\), estimated by the

\[ Q_{bf} = \frac{1}{2} \rho_{water} g A_{s} V_{bf} \]

\[ Q_{bd} = \rho_{sand} A_{bd} V_{bd} \]

where \(\rho_{water}\) is the density of water, \(g\) is the acceleration due to gravity, \(A_{s}\) is the cross-sectional area of the channel, \(V_{bf}\) is the bankfull velocity, \(\rho_{sand}\) is the density of sand, and \(A_{bd}\) is the cross-sectional area of the bedload material. The \(Q_{bf}\) and \(Q_{bd}\) estimates are crucial for understanding the sediment transport capacity and the potential for sedimentation in the Allomember E valley.
Chezy Coefficient method, ranges from $0.08 \, \text{m}^3/\text{s}$ to $0.12 \, \text{m}^3/\text{s}$; and the suspended-load sediment discharge ($Q_{ss}$) estimated by the Van Rijn equations ranges from $2.1 \, \text{m}^3/\text{s}$ to $4.8 \, \text{m}^3/\text{s}$ (Table 1).

Estimation of Annual Sediment Discharge

In the Cenomanian, the Dunvegan deltaic complex was located in the northern mid-latitude warm humid climate zone (NMW) (Fig. 12; Hay and Floegel 2012), which is equivalent to the warm temperate fully humid warm summer (Cfb) climate zone in the Köppen-Geiger climate classification. Geographically, the climate should be similar to northwest North America or western Europe in modern times (Köppen-Geiger climate classification map). This is in agreement with the climate classification of the Dunvegan Formation, according to Davidson and North (2009).

Meybeck et al. (2003) discussed the sediment-flux variability and flood duration by synthesizing global river data in a variety of basin areas and climate regimes. The percentage of sediment transported within 2% of the total time is suggested to be correlated to drainage-basin area; thus, based on the estimated drainage-basin area of the Dunvegan Formation (on the order of $10^{-5} \, \text{km}^2$, according to Plint and Wadsworth 2006), the trunk river may transport $10^{-4}$–$0.4 \%$ of the total suspended sediment within 2% of the discharge time (Fig. 13). Additionally, research on the Rhine River, which is considered as a modern counterpart of the Dunvegan river based on Davidson and North (2009), also suggested that peak flow transports about 25% of the sediment. In this study, we similarly assume that 25% of the total sediment is transported within 2% of one year (7.3 days) as the bankfull discharge duration. Therefore, Equation 10 can be written as

$$Q_{max} = Q_{bf} \times 7.3/25\%$$

Based on these parameters, the calculated annual bedload volume ranges from $1.9 \times 10^5$ to $3.0 \times 10^5 \, \text{m}^3$, and the calculated annual suspended-load volume ranges from $5.2 \times 10^6$ to $1.2 \times 10^7 \, \text{m}^3$ (Table 1).

Duration of Deposition

Previous work shows that the entire Dunvegan Alloformation was deposited in about 2 million years (Plint and Wadsworth 2006; Bhattacharya et al. 2016). Each of the ten allomembers should thus be deposited in no more than 200,000 years. Plint and Wadsworth (2006) documented four paleovalley systems, and suggested that each paleovalley system may be formed within 100,000–200,000 years. McCarthy et al. (1999) and Plint et al. (2001) also estimated that each Dunvegan allomember represented 150,000 to 200,000 years. Moreover, Plint et al. (2001) inferred that the offlap pattern of marine parasequences, and the maturity of interfluve paleosols, might represent as much as half of that time (Kraus 1999). They suggested that river entrenchment began early and

<table>
<thead>
<tr>
<th>Channel</th>
<th>Bankfull Channel Width (m)</th>
<th>Average Bankfull Depth (m)</th>
<th>Slope</th>
<th>Grain Size ($D_{16}$, $D_{50}$, $D_{84}$, $D_{90}$) (mm)</th>
<th>$Q_{bf}$ Water (m$^3$/s)</th>
<th>$Q_{bf}$ Bedload (m$^3$/s)</th>
<th>$\bar{u}$ Avg. Velocity (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>E1 Min.</td>
<td>150</td>
<td>10</td>
<td>$6.1 \times 10^{-5}$</td>
<td>0.14, 0.20, 0.27, 0.30</td>
<td>$1.6 \times 10^3$</td>
<td>0.08</td>
<td>1.0</td>
</tr>
<tr>
<td>E1 Max.</td>
<td>230</td>
<td>15</td>
<td>$4.1 \times 10^{-5}$</td>
<td>0.14, 0.20, 0.27, 0.30</td>
<td>$3.7 \times 10^3$</td>
<td>0.12</td>
<td>1.0</td>
</tr>
</tbody>
</table>

FIG. 12.—The global climate-zone classification of the Cenomanian period. The Dunvegan deltaic complex was located in the northern mid-latitude warm humid zone (NMW) (modified after Hay and Floegel 2012).
continued throughout the falling-stage systems tract, resulting in sediment starvation due to incision on the interfluves and delivery of all available sediment to the delta front. Thus, each valley system might record half of the depositional duration of the allmembers of river entrenchment. Allomember E was further divided into four parasequences, and the individual trunk river parasequences likely represent deposition in less than 25,000 years (Bhattacharya et al. 2016).

**Evaluation of Cumulative Sediment Load**

The paleodischarge of the trunk river in Allomember E, as well as the total sediment load passing through the trunk-river fulcrum point, can now be calculated using the estimated paleohydrologic parameters and total duration. The results and estimated parameters are summarized in Table 1. The results indicate that the trunk river of Allomember E was 10–15 m deep and 150–230 m wide, carried mainly fine-grained (<200 microns) sand and flowed over a low-gradient paleoslope of 4.1–6.1 \( \times 10^{-5} \). Annual total sediment load is estimated to be in the range of 5.2 \( \times 10^{10} \) to 3.2 \( \times 10^{11} \) m\(^3\)/yr, including both bedload and suspended load. Within 25,000 years, the river is estimated to have transported 1.4 \( \times 10^{11} \) m\(^3\) to 3.1 \( \times 10^{11} \) m\(^3\) of sediment into the basin (Table 1).

<table>
<thead>
<tr>
<th>Channel</th>
<th>Bankfull Suspended-Load Discharge, ( Q_{tb} ) (Van Rijn 1984) (m(^3)/s)</th>
<th>Mean Annual Bedload ( Q_{sb} ) (m(^3))</th>
<th>Mean Annual Suspended Sediment ( Q_{tsb} ) (Van Rijn (m(^3))</th>
<th>Total Bedload ( Q_{tb} + Q_{sb} ) (km(^3))</th>
<th>Total Suspended Load ( Q_{tsb} ) (km(^3))</th>
<th>Total Sediment Load ( Q ) (km(^3))</th>
</tr>
</thead>
<tbody>
<tr>
<td>E1</td>
<td>2.1</td>
<td>1.9 ( \times 10^{8} )</td>
<td>5.2 ( \times 10^{9} )</td>
<td>4.8</td>
<td>130</td>
<td>135</td>
</tr>
<tr>
<td></td>
<td>4.8</td>
<td>3.0 ( \times 10^{8} )</td>
<td>1.2 ( \times 10^{10} )</td>
<td>7.4</td>
<td>300</td>
<td>307</td>
</tr>
</tbody>
</table>

**Sediment Accumulation in the Sink and Mass Balance**

An isopach map of Allomember E (Bhattacharya 1993) was used to assess total sediment volume deposited downstream of the trunk river (Fig. 14). The volume of mapped deposits in the basin, down-dip of the fulcrum point, is approximately 4.3 \( \times 10^{11} \) m\(^3\), assuming 20% average porosity (Plint et al. 2009), thus the total volume of sediment deposited in the basin is about 3.4 \( \times 10^{11} \) m\(^3\), including both sandstones and mudstones. A 10% compaction factor (Holbrook and Wanas 2014) is also used to reconcile the original uncompactsed sediment volume prior to compaction that subsequently approximates the sediment volume of 3.8 \( \times 10^{11} \) m\(^3\) (380 km\(^3\)). This sediment volume is estimated from the whole of Allomember E, whereas the lowstand delta fed by the trunk river was deposited only during the time represented by Parasequence E1. Four parasequences were mapped in Allomember E in total (Bhattacharya 1993); the sediment volume of Parasequence 1 is very likely greater than the other three parasequences, based on the paleogeographic maps of Bhattacharya (1993) and an approximate thickness estimate from Figure 2. Therefore, here we assume that the sediment volume of Parasequence E1 accounts for one third of the total estimated sediment volume of Allomember E, which is 1.3 \( \times 10^{11} \) m\(^3\) (130 km\(^3\)). A sandstone isopach map of Parasequence E1 delineates sandstone distribution and shows a lobate geometry (Bhattacharya 1993) (Fig. 15). The estimated sandstone volume is approximately 3.2 \( \times 10^{10} \) m\(^3\) (32 km\(^3\)) beyond the fulcrum point. Assuming 20% average porosity yields a sand sediment volume of ca. 2.6 \( \times 10^{10} \) m\(^3\) (26 km\(^3\)).

The estimated sediment volume delivered and passed through the fulcrum based on the paleohydrologic evaluation of the trunk river is ca. 1.4 \( \times 10^{11} \) m\(^3\)–3.1 \( \times 10^{11} \) m\(^3\) (135–307 km\(^3\)), while the documented sediment volume transported through the trunk river and accumulated in the sink area is about 1.3 \( \times 10^{11} \) m\(^3\) (130 km\(^3\)). The sediment budget and accumulation volumes are on the same order of magnitude, whereas the upper-range estimate is 2.5 times greater than the sediment mapped in the study area. This indicates minimal sediment balance between source and sink, although the higher values indicate a possibly underbalanced system (Fig. 16).

**DISCUSSION**

**The Mass Balance**

The upper-range estimate of sediment delivered from the source is 2.5 times the measured sediment volume in the sink area, which, if accurate, would suggest significant sediment escape (Bhattacharya et al. 2016). Because the Cretaceous Western Interior Seaway was a ramp margin during the Cenomanian, and no shelf–slope break existed, there are no submarine fan deposits. Hyperpycnal flows have been documented in the proximal part of the shelf (Bhattacharya and MacEachern 2009), but owing to the low gradient, it would be difficult for sediment to accelerate or to be transported far from the shoreline to deep water (Varban and Plint 2008). The alternative sediment transport mechanism is wave reworking, but Allomember E is one of the most river-dominated systems in the Dunvegan (Bhattacharya 1993). The greater sediment input versus the accumulation supports the hypothesis that in Dunvegan time, mud was widely dispersed southward along the Alberta Foreland Basin by geostrophic currents.

![Figure 13](image)  
**Fig. 13.**—Plot of percentage of total suspended-load discharged within 2% time of a year by rivers with various drainage areas and in different climate regimes (modified after Meybeck et al. 2003).
associated with storm processes and counterclockwise oceanic gyres (see Plint et al. 2009). Along-shelf geostrophic flows towards the southwest would result from Coriolis deflection of the water mass in the northern hemisphere (Slingerland et al. 1996).

Plint et al. (2009) state that the younger allomembers (F–D) of the Dunvegan Formation are low-gradient systems, and formed laterally extensive muddy clinothems, suggesting relatively shallow water (< 40 m). The shallow water would allow storm-driven geostrophic flows reaching the sea floor and resuspending mud for further transport. Flocculated and pelleted muds are likely to be repeatedly resuspended by concomitant storm waves and advected obliquely across and along the shelf. The long-term large-scale counterclockwise oceanic circulation in the Western Interior Seaway initiated a south-directed oceanic gyre flow that also exerted a force for mud transport along the shelf. Combined and geostrophic flows have been emphasized as the principal mechanisms likely responsible for the offshore dispersal of both sand and mud across storm-influenced shelves (Fig. 17). Mud dispersal mechanisms are also reviewed and discussed by Blum and Hattier-Womack (2008), showing that mud transported by hyperpycnal plumes can be merged with alongshore currents and resuspended for transport basinward. Mud therefore can be transported over hundreds of kilometers along the shelf. The study of the Kaskapau Formation shales, which lie directly above the Dunvegan Alloformation, suggests that mud was resuspended by storm waves and transported as much as 250 km offshore by wind-driven currents (Varban and Plint 2008).

The modern Amazon S2S system shows that < 50% of the sediment delivered by the Amazon River is deposited on the adjacent inner-shelf subaqueous delta, whereas the remaining sediment is transported northwards by the Guyana Current towards the Orinoco Delta (Warne et al. 2002; Somme et al. 2009b). Romans et al. (2015) also revealed that more than 50% of the sediment delivered through the Eel River to the sea was transported beyond the shelf. Also the Po, Mississippi, and Atchafalaya delta show significant along-shelf export, even though the latter two are river-dominated (Allison et al. 2000; Draut et al. 2005; Cattaneo et al. 2007). Similar mud transport mechanisms may be invoked to explain the difference between the volumes of estimated sediment supplied and accumulated in the proximal part of the Dunvegan shelf. In addition, the isopach maps represent the volumes of sediment accumulated proximal to the river mouth and do not incorporate along-shelf or across-shelf mud belts.

In addition to sediment escape on the shelf, sediment sequestration and partitioning inside the transfer systems (e.g., deposited in the floodplain)
may be a secondary mechanism attributed to the volume difference between the source and sink (Allen 2008a; Romans et al. 2015). In sediment routing systems, although fluvial systems serve as the transfer zones, where sediment is considered not purely eroded, nor accumulated, there is still erosion and storage, thus a proportion of sediment can be sequestered or permanently stored within a river and its floodplain (Walling and Collins 2008; Holbrook and Bhattacharya 2012; Romans et al. 2015); unfortunately, the sequestration ratio remains equivocal in ancient systems.

In the Dunvegan Alloformation, the tectonically active drainage basin is estimated to have been on the order of 10^5 km² in area, which is considered to be a moderate drainage system and suggests a relatively long sediment routing system. As a consequence, sediment sequestration could be considerable in the transfer systems (Milliman and Syvitski 1992; Covault et al. 2010; Romans et al. 2015). Moreover, landward tectonic flexural subsidence has been demonstrated to be important in Allomember E, as indicated by landward thickening of the nonmarine strata updip (Plint and Wadsworth 2003). Landward sequestration may be important, but it remains equivocal as to when this occurred (i.e., which parasequence of Allomember E). Timing of flexural subsidence is critical to unravel this issue and may have been more important in the deposition of the older parasequences in Allomember E, which do not show signs of degradational stacking that would indicate falling sea level. Rather, interfluve paleosols, described by Plint and Wadsworth (2003) suggest degradation during the falling stage of the valley that occurs in the latter stage of deposition of Allomember E, and this limits the possibility of mud sequestration in floodplains during the falling stage associated with valley erosion.

**Estimates of Paleodischarge and Sediment Volumes**

Paleodischarge of the ancient trunk river was estimated to be in the range of 1.6–3.7 × 10^3 m³/s, which is in agreement with the paleodischarge estimated by Bhattacharya and MacEachern (2009). The discharge of the Rhine River is approximately 3.5 × 10^3 m³/s, which represents a likely modern analogue for the Dunvegan trunk river, as was also suggested by Davidson and North (2009).

Based on the estimates of sediment transport in the trunk river of Allomember E, the bedload is calculated to be about 4% of the total sediment load, compared with a more common value of 10% reported in the literature on modern rivers (Turowski et al. 2010). The low percentage calculated here for the Dunvegan can be attributed to low shear stress due to the low-gradient in contrast to smaller-scale and steeper-gradient mountainous rivers, the bedload ratio of which can be up to 40% (Turowski et al. 2010). On the other hand, the estimated bedload sediment volume is about 5–7 km³, whereas the sand volume estimated from the isolith map is about 26 km³. This indicates that a significant amount of very-fine-grained...
sandy sediments (i.e., < 62–125 microns) were transported as suspended load. Core data (Bhattacharya 1992) shows that much of the sand in the lowstand delta is in the very fine upper range (88–125 microns), supporting the idea of significant transport in suspension.

The average thickness of Parasequence E1 downstream is about 6–8 m (Fig. 15), which represents an approximate sedimentation rate of 0.3 m/kyr, given the depositional duration of 25,000 years. The Dunvegan corresponds to Unit 8 on the "Sedimentation Rate Scale" (SRS) of Miall (2014), characterized as channel belts or delta deposits (his fig. 1). The sedimentation rate reflects that each parasequence of Allomember E was deposited under the control of short-term Milankovitch cycles (Miall 2014) and supports the hypothesis that allogenic Milankovitch cyclicity produced the high-frequency sequences in the Dunvegan Formation.

According to Davidson and North (2009), assuming a northern mid-latitude warm humid climate, the drainage basin area of Allomember E can be estimated with the regional curve parameters:

\[ d_m = 1.11(DA)^{0.51} \]  
\[ Q_w = 100.6(DA)^{0.76} \]

where \( d_m \) is bankfull flow depth, \( Q_w \) is water discharge, and \( DA \) is drainage-basin area. The estimated drainage-basin area is on the order of 1.4–1.8 \( \times 10^5 \) km\(^2\), which is consistent with the drainage-basin area estimated by both Plint and Wadsworth (2003) and Bhattacharya et al. (2016) based on paleotectonic and paleogeographic reconstructions.
The modern Cfb climate zone is located between 40° N and 55° N. Similarly, the latitude of the Maritime Temperate climate zone, which represents the climate thought to be the most characteristic of the Dunvegan Formation, is between 45° N and 55° N (Davidson and North 2009). Nonetheless, McCarthy et al. (1999) suggested that the Dunvegan Formation was deposited at a paleolatitude of 65° N, which is higher than its modern climate analog. This is explained by the idea that the Cenomanian represented a “greenhouse” time in earth history with a correspondingly warmer climate (Glancy et al. 1993).

The Rhine River, which is considered as a modern analog to the Dunvegan system, based on the regional climate classification, has a basin area of $1.6 \times 10^5$ km² and a discharge of $3.5 \times 10^3$ m³/s. The drainage-basin area is quite similar to the estimated drainage-basin area of the Dunvegan, but the estimated discharge of the Dunvegan trunk river, using the regional curve equation listed above, is significantly larger than the Rhine River, by a factor of 5 (Davidson and North 2009). River discharge is substantially related to drainage area, relief, and climate. The discharge of the Dunvegan trunk river seems to be rather overestimated by the regional curve method. Davidson and North (2009) admitted that the regional-curve method is less reliable in predicting discharge than estimating drainage-basin area.

The estimated sediment load of the Dunvegan trunk river ranges from 12.5 to 28.6 MT/yr, given a bulk density of 2.4 g/cm³ (Beaumont 1981), and sediment-yield estimates range from 89.3 to 158.9 t/km²/yr (Table 2). The correlations between estimated drainage-basin area and sediment load and sediment yield of the Dunvegan trunk fluvial system are plotted on a global modern-river dataset (Fig. 18A, B). The Dunvegan E1 trunk fluvial system can be categorized as a moderate-size mountain-river drainage system, according to modern-river analogs (Milliman and Syvitski 1992). Appropriate modern-analog selection, based on an accurate quantitative approach, allows better prediction of the size and geometry of an ancient system. Linking modern-analog systems to outcrops and subsurface data helps to reconstruct paleogeography and scaling of depositional environments, as well as estimating the scales and dimensions of paleodrainage networks. With the knowledge of the geometry and paleohydrological elements of the paleodrainage system, paleoclimate can also be reconstructed.

**Comparison to BQART Method**

Syvitski and Milliman (2007) introduced the “BQART” method of sediment-discharge estimation based upon discharge ($Q$), drainage area ($A$), relief ($R$), temperature ($T$), and $B$, which is a factor determined by multiple geological elements that are discussed below. We also applied this method to estimate sediment discharge independently of the fulcrum approach. Paleoclimate analogized by the Köppen-Geiger climate classification and associated with the regional curves of Davidson and North (2009) indicates that the average temperature of the Dunvegan area was between −3°C and 18°C. Fossil flora described by Bell (1963) show a dominance of tropical plants, such as *Ficus* and *Magnolia*, characterized by broad, smooth-edged leaves. This suggests that, despite being at a temperate latitude, the Cretaceous Dunvegan system was actually in a tropical–subtropical paleoclimate zone, the average temperature of which would have been around 18°C, according to the Köppen-Geiger climate classification. Dennis et al. (2013) estimated that mean annual temperature of the Late Cretaceous Western Interior Seaway of 20°C using a carbonate-isotope thermometer. We used 20°C as the mean temperature in our estimation. Therefore, a more applicable equation is

$$Q_s = 0.02BQ^{0.31}A^{0.5}RT$$

and

$$B = IL(1 - T_E)E_h$$

where $I$ is the glacier erosion factor ($I \geq 1$), $L$ is an average basin-wide lithology factor, $T_E$ is the trapping efficiency of lakes and man-made reservoirs, such that $(1 - T_E) \leq 1$, and $E_h$ is a human-influenced soil erosion factor.

The Late Cretaceous is well accepted as a “greenhouse” world without significant ice cover; therefore, $I$ equals 1. $L$ is set at 2, as suggested by

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**Table 2.—Estimated drainage-basin areas, sediment loads, and sediment yields for the trunk river of Allomember E.**

<table>
<thead>
<tr>
<th>Channel</th>
<th>Drainage Basin Area ($\times 10^5$ km²)</th>
<th>Sediment Discharge ($\times 10^3$ m³/s)</th>
<th>Sediment Load (Mt/yr)</th>
<th>Sediment Yield (t/km²/yr)</th>
<th>Remark</th>
</tr>
</thead>
<tbody>
<tr>
<td>E1</td>
<td>0.14–0.18</td>
<td>5.2–11.9</td>
<td>12.5–28.6</td>
<td>89.3–158.9</td>
<td>Moderate-size mountain river</td>
</tr>
</tbody>
</table>

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**Fig. 18.—** A) Sediment load vs. basin area. B) Sediment yield vs. basin area of the global river database. Note the strongly normal trend between sediment load and basin area, whereas the generally inverse relationship between yield and basin area (modified Milliman and Syvitski 1992).
Syvitski and Milliman (2007) for a clastic soft sedimentary basin. $T_c$ and $E_a$ are set at 1 as there could not have been any anthropogenic effects in the Cretaceous. The current height of the Laramide-age Rocky Mountains is about 4,400 m after erosion, and the peak is less than 3000 m in Canada, but Laramide orogenic activity occurred after the Dunvegan (Dickinson and Snyder 1978). Hence the continental relief during Dunvegan time would likely be less than today. Beaumont (1981) modeled the formative mechanisms of the Alberta Foreland Basin of Western Canada and implied that the Rocky Mountains were about 2000 m above sea level during the Late Cretaceous, and we thus assume a relief of 2,000 m for the “BQART” analysis. As the discharge estimated from trunk-river dimensions is the bankfull discharge, and likely only occurs 2% of a year, the annual sediment load is used for comparison. Therefore, the estimated sediment discharge using the BQART formula ranges from 5.2 to 7.6 $\times 10^6$ m$^3$/s, the lower end of which is the same as the lower range of the annual sediment discharge estimated by the fulcrum method. The upper limit of the BQART estimate is about 60% of the upper-range estimate using the fulcrum method (Fig. 19; Table 2). The BQART calculation is sensitive to estimates of relief, temperature, glacial cover, and anthropogenic effects. Anthropogenic effects and glaciers are almost certainly not relevant factors for the Dunvegan, and we believe that the hinterland relief was not likely much higher than our estimate. However, the average temperature may have been higher than 20 °C. Slingerland et al. (1996) discussed that the temperature of the Western Interior Seaway during the Late Cretaceous could have been higher than 20 °C. Using a higher temperature value would increase the BQART estimate, which results in an upper-range value more similar to the upper-end estimate using the fulcrum approach detailed above.

Errors and Uncertainties

The fulcrum approach involves a number of uncertainties, including field measurements, numerical assessments used to estimate paleohydrologic parameters, paleomorphodynamics derived from stratigraphic records, applicability of empirical equations, chronologic estimates, and selection of modern-analogue data. The integration of these variables constrains the accuracy of sediment-volume estimation to at worst one order of magnitude (see discussion in Hajek and Wolinski 2012; Holbrook and Wanas 2014; Bhattacharya et al. 2016) (Fig. 20).

An analysis of error and uncertainty shows that the most sensitive error is the annual-discharge estimate related to annual flood duration, and can show an error of an order of magnitude (Fig. 20). This agrees with the most uncertain element in the study of Holbrook and Wanas (2014). The annual discharge estimated is a function of estimates of bankfull event duration, recurrence, and sediment transport capacity, which remain difficult to constrain in deep-time systems. For example, bankfull floods are easily masked or overprinted in S2S systems (Wang et al. 2011; Cramer et al. 2015; Romans et al. 2015). The bankfull flood is defined as the moment when the channel is fully filled with water. Durations and return periods of bankfull flood vary from river to river, and they are still controversial in the hydrologic community. Bankfull floods are highly dependent on climate regime and tectonic topography, associated with seasonality and annual cyclicity (Arora and Boer 2001). Durations of bankfull events in modern rivers are also highly variable, ranging from a couple of days to 80 days per year (Andrews 1984; Holbrook and Wanas 2014). The proportion of sediment load carried and transported during bankfull flood events is also highly variable, as it is controlled by bankfull-event duration, sediment supply, drainage-basin scale, fluvial diffusivity, annual flood frequency, river normal flow condition, and high flow frequency and magnitude, and may range from 25 to 90% (Nolan et al. 1987; Andrews and Nankervis 1995; Moog and Whiting 1998; Sichomgabula 1999; Davide et al. 2003; Meybeck et al. 2003; Powell et al. 2006). Flood events of the Dunvegan trunk river may transport a low proportion of annual sediment due to a long sediment route and low gradient, compared to more vigorous smaller river systems (Meybeck et al. 2003).

It may not be feasible to estimate long-term sediment volumes based on simply summing short-term estimates. Miall (2014) emphasizes the fragmentary nature of sedimentary systems and the difficulty of assigning time to rocks. This argument leads to a question as to whether the 25,000 years assumed for Parequence E1 represents the effective sedimentation time, or whether there are significant time gaps at the top flooding surface or associated with the valley sequence boundary. If the time scale is shorter, the annual sediment discharge may not represent an average value, especially when flood magnitudes with different return periods are considered (Sadler 1999; Holbrook and Wanas 2014; Miall 2014; Sadler...
and Jerolmack 2015). Nonetheless, when we estimated annual sediment discharge, we considered both short-term peaks of concentrated sediment transport and the percentage of the concentration of total annual discharge. In other words, we have assumed a constant averaged annual sediment discharge of the system by applying the concept of peak-flood-duration sediment concentration and related proportion of total annual discharge. Therefore, the annual discharge can be used to estimate total sediment volumes.

The fulcrum approach does not take sediment stored landward of the fulcrum into consideration, but only accounts for sediment passing the fulcrum, which therefore inevitably introduces uncertainties. Multiple fulcrum assessments along longitudinal profiles of trunk valleys could account for variable storage of sediment in a valley system (Holbrook and Wanas 2014). Multiple drainage systems may coexist in the Dunvegan deltaic complex of Allomember E (Fig. 3), which can be analyzed by dividing different drainage systems, estimating and combining multiple feeding rivers, and comparing them to the documented sediment accumulation volumes in the sink. If the estimated sediment volume delivered by a single fulcrum trunk channel was less than the sediment accumulated, then multiple trunk channels can be inferred (Holbrook and Wanas 2014).

Alternatively, Sadler (1999) suggested that over time scales of Milankovitch cycles, the average effective sediment accumulation rates (the rate of sediment preserved in the basin over total source sediment) are in the range of 10–20%. Given the instantaneous sediment discharge of 2.1–4.8 m³/s, including both bedload and suspended load, and the total chronometric duration of 25,000 years, sediment accumulations in 10% of total depositional duration are estimated to be in the range of 1.7–3.7 × 10¹⁲ m³ (166–370 km³), which is consistent with the results of 1.4–3.1 × 10¹² m³ (135–307 km³). If we use 20% as effective sedimentation rate, the results are still on the same order of magnitude.

**Hyperpycnal Flows**

The estimated bankfull flood discharge of the Dunvegan trunk river is in the range of 1.6–3.7 × 10¹³ m³/s. Milliman and Syvitsky (1995) suggest that rivers with average discharge lower than 6000 m³/s routinely generate hyperpycnal flows during floods (Bhattacharya and MacEachern 2009). The Dunvegan drainage system is adjacent to a tectonically active mountain belt in a warm humid climate and drained into a shallow ocean (less than a few hundred meters deep), albeit it has a relatively low slope. In addition, examination of cores through delta-front and prodeltaic facies, especially in Allomember E, shows abundant hyperpycnites, turbidities, and wave-enhanced sediment-gravity-flow deposits (Bhattacharya and MacEachern 2009). This suggests that the Dunvegan rivers routinely generated hyperpycnal flows. However, it is unlikely that direct river-driven hyperpycnal flows were able to flow very far on such low slopes of the Dunvegan shelf. Longer-scale mud transport and dispersal on the shelf in the form of fluid mud was likely enhanced by storm waves. Abundant and ubiquitous hummocky cross-stratification and wave ripples observed in shoreface and deltaic successions indicate a storm-dominated seaway (Pattison and Walker 1992). This further confirms the hypothesis of postdepositional mud dispersal accounting for the volume difference between sediment influx and sediment accumulation. It is plausible that hyperpycnal flows contributed to the mud dispersal along the shelf, combined with other oceanic currents.

**CONCLUSIONS**

The fulcrum approach assumes that the total mass balance in a closed S2S system should match the sediment volume delivered from the source area through a fulcrum point (i.e., trunk river) to that deposited in the sink. Ancient trunk rivers are prominent fulcrum candidates, paleohydraulic parameters of which can be used to estimate paleodischarge and instantaneous sediment volumes passing through them. Consequently, the total sediment volumes can be estimated with a deep-time depositional duration. The workflow comprises trunk-river dimension and paleoslope reconstruction, grain-size evaluation, instantaneous and annual discharge estimates, analysis of long-term depositional duration, and measurements of sink-area sediment accumulation.

The fulcrum approach was tested in the Upper Cretaceous Dunvegan Alloformation by comparing the total mass balance between source sediment production and sink accumulation through paleohydrologic assessment of the trunk river. The results indicate that the trunk river of Allomember E was 10–15 m deep and 150–230 m wide, flowed over a low-gradient paleoslope of about 4.1–6.1 × 10⁻⁵, and carried fine- to medium-grained sand. The river is estimated to have transported 1.4 × 10¹¹ m³–3.1 × 10¹¹ m³ (135–307 km³) of sediment into the basin within 25,000 years, which is consistent with the 1.3 × 10¹¹ m³ (130 km³) of sediment documented in the sink area. The upper-range estimate of sediment delivered to the sink is 2.5 times the measured sediment-sink volume and suggests sediment escape on the shelf or sediment sequestration inland. This supports the hypothesis of mud dispersal in Dunvegan time and accommodation increase landward due to tectonic subsidence. The concomitant storm-initiated hyperpycnal flows could also help to explain the mud dispersal. These estimates also allow more robust comparison of ancient and modern analogs and their scaling relationships.

**SUPPLEMENTAL MATERIAL**


**ACKNOWLEDGMENTS**

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Allison, M.A., Kineke, G.C., Gordon, E.S., and Goni, M.A., 2000, Development and Alloformation by comparing the total mass balance between source sediment production and sink accumulation through paleohydrologic assessment of the trunk river. The results indicate that the trunk river of Allomember E was 10–15 m deep and 150–230 m wide, flowed over a low-gradient paleoslope of about 4.1–6.1 × 10⁻⁵, and carried fine- to medium-grained sand. The river is estimated to have transported 1.4 × 10¹¹ m³–3.1 × 10¹¹ m³ (135–307 km³) of sediment into the basin within 25,000 years, which is consistent with the 1.3 × 10¹¹ m³ (130 km³) of sediment documented in the sink area. The upper-range estimate of sediment delivered to the sink is 2.5 times the measured sediment-sink volume and suggests sediment escape on the shelf or sediment sequestration inland. This supports the hypothesis of mud dispersal in Dunvegan time and accommodation increase landward due to tectonic subsidence. The concomitant storm-initiated hyperpycnal flows could also help to explain the mud dispersal. These estimates also allow more robust comparison of ancient and modern analogs and their scaling relationships.

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