Invited review

Estimation of source area, river paleo-discharge, paleoslope, and sediment budgets of linked deep-time depositional systems and implications for hydrocarbon potential

Janok P. Bhattacharya a*, Peter Copeland b, Timothy F. Lawton c, John Holbrook d

a School of Geography and Earth Sciences, McMaster University, 1280 Main St., West, Hamilton, Ontario L8S 4L8, Canada
b Department of Earth and Atmospheric Sciences, University of Houston, Houston, TX 77204, United States
c Centro de Geociencias, Universidad Nacional Autónoma de México (UNAM), Blvd. Juriquilla No. 3001, Querétaro 76230, Mexico
d School of Geology, Energy, and the Environment TCU, Box 29330, Fort Worth, Texas 76129, United States

A R T I C L E   I N F O

Article history:
Received 25 September 2014
Received in revised form 28 September 2015
Accepted 27 October 2015
Available online 31 October 2015

Keywords:
Source to sink
Sequence stratigraphy
Rivers
Deltas
Submarine fans
Paleodischarge
Provenance
Detrital zircons
Paleoslope

A B S T R A C T

The source-to-sink (S2S) concept is focused on quantification of the various components of siliciclastic sedimentary systems from initial source sediment production areas, through the dispersal system, to deposition within a number of potential ultimate sedimentary sinks, and has more recently been applied to deep-time stratigraphic systems. Sequence-stratigraphic analysis is a key first step that allows depositional systems to be correlated and mapped, within a time-stratigraphic framework, such that fluvial transport systems can be linked to down-dip shorelines, shelf and deep-water deposits and interpreted in the context of allogenic processes. More recently, attempts have been made to quantify catchment areas for ancient depositional systems, using scaling relationships of modern systems. This also helps predict the size of linked depositional systems, such as rivers, deltas and submarine fans, along the S2S tract. The maximum size of any given depositional system, such as a river, delta, or submarine fan, is significantly controlled by the area, relief, and climate regime of the source area, which in turn may link to the plate tectonic and paleogeographic setting. Classic provenance studies, and more recent use of detrital geochronology, including zircons, provide critical information about source areas, and place limits on catchment area. Provenance studies, especially when linked to thermochronometry also provide key information about rates of exhumation of source areas and the link to the tectonic setting. In this paper the techniques for estimation of water and sediment paleodischarge and paleo-drainage area are outlined, and sediment budgets are calculated for a number of Mesozoic systems, primarily from western North America. The relevance for hydrocarbon exploration and production is discussed for each example. In Mesozoic Western Interior basins of North America, extensive outcrop and subsurface data allow the largest trunk rivers to be identified, typically within incised valleys. Thickness, grain size, and sedimentary structures can be used to infer slope and flow velocities, and using width estimations, water and sediment paleodischarge can be calculated. River paleoslope can also be independently measured from stratigraphic-geometric considerations and used to assess paleo-river flow. Paleodischarge in turn is used to estimate the size of the catchment source area. Paleodischarge of rivers can also be estimated independently by integrating estimates of catchment source area, for example by using detrital zircons integrated with paleoclimate. The catchment areas of North America evolved significantly during the late Mesozoic. During the Jurassic–Early Cretaceous, fluvial systems consisted of continental-scale low slope ($S = 10^{-4}$), axially drained rivers, forming the 40-m-deep channels in the Mannville Group in Canada, which now host the supertgiant heavy-oil-sands reserves. During the times of maximum transgression of the Cretaceous Seaway, such as the Turonian and Campanian, the western North American foreland basin was characterized by smaller-scale (typically 10-m deep), steeper gradient ($S = 10^{-3}$) sand and gravel bedload rivers, dominated by transverse drainages in the rising Cordilleran. This created a number of smaller river-delta S2S systems along the coast. As the Laramide Orogeny progressed, the Western Interior Seaway receded, and by the Paleocene the modern continental-scale drainage of North America was largely established with a major continental division separating south-flowing Mississippi drainage from north-flowing systems. The integration of paleodischarge estimates with provenance analysis enables the improved use of the sedimentary record to make estimates about the entire S2S system, as opposed to primarily the depositional component.

* Corresponding author.
E-mail address: bhattaj@mcmaster.ca (J.P. Bhattacharya).

http://dx.doi.org/10.1016/j.earscirev.2015.10.013
0012-8252/© 2015 Elsevier B.V. All rights reserved.
Clinothem stacking relationships and isopach mapping of stratigraphic volumes have also been integrated with chronofractographic data to analyze long-term S2S sediment budgets. A more quantitative approach to estimating the scale of erosional, transport and depositional components of sedimentary systems, especially in the context of linked source and depositional areas, also puts constraints on the size and scale of potential hydrocarbon reservoirs and thus has significant economic value.

© 2015 Elsevier B.V. All rights reserved.

Contents

1. Introduction: S2S in deep-time systems ........................................................... 78
2. Sequence-stratigraphy overview .................................................................. 79
3. Critical considerations in applying S2S concepts to deep-time systems .......... 80
4. Components of S2S dispersal and depositional systems ................................ 82
   4.1. Alluvial systems, catchments and transfer ............................................. 82
   4.2. Shorelines, shelves, and along- and across-shelf transfer of sediment ...... 84
   4.3. Deep-water systems, the ultimate sink .................................................. 84
5. Source area versus discharge ........................................................................ 84
6. Paleodischarge, paleoslope, sediment budgets, and source-area estimates in deep-time systems ................................................................. 86
   6.1. Characterizing the trunk river ................................................................. 86
   6.2. Estimating channel dimensions and sediment characteristics ............... 86
   6.3. Paleoslope estimates ............................................................................. 88
   6.4. Velocity and paleodischarge estimates ................................................... 89
   6.5. Paleodischarge estimates using regional hydraulic geometry curves ........ 90
   6.6. Estimates of sediment budgets from paleodischarge ................................ 91
   6.7. Drainage area and reservoir quality from provenance analysis .............. 92
   6.8. Use of detrital zircons for formation age determination and thermochronology (uplift and exhumation rates) ........................................... 93
6.9. Tectonics, subsidence and sediment supply .............................................. 94
7. Examples of Mesozoic S2S systems in Canada and the USA ....................... 94
   7.1. Triassic–Jurassic ................................................................................... 95
   7.2. Big rivers in the early Cretaceous ......................................................... 95
   7.3. Late–Early Cretaceous, end of the big rivers .......................................... 96
   7.4. Late Cretaceous (Cenomanian), rivers and deltas of the Dunvegan Formation and equivalents ................................................................. 97
       7.4.1. The Dunvegan delta ........................................................................ 97
       7.4.2. Dunvegan paleodischarge estimates ............................................. 97
       7.4.3. Dunvegan sediment budgets ......................................................... 100
       7.4.4. Dunvegan petroleum perspective ................................................ 100
   7.5. Turonian: a time of many deltas ............................................................ 101
   7.6. The Ferron sandstone, a case study ........................................................ 101
   7.7. Santonian to Paleocene, the mega-bajada ............................................. 103
8. Conclusion and future work ......................................................................... 105
Acknowledgements ......................................................................................... 105
References ...................................................................................................... 105

1. Introduction: S2S in deep-time systems

The source-to-sink (S2S) concept focuses on quantification of the various components of sedimentary systems from source areas, through dispersal systems, and thence to deposition in a number of potential sedimentary sinks (Fig. 1). Boundaries of particular interest include the gravel–sand transition in the alluvial realm, the shoreline, the shelf-slope break, and the continental slope to deep-water transition. The NSF-funded S2S program focused on a 10-year analysis of a pair of high-gradient modern systems (Keuhl et al., 2015), the Fly-Strickland River and offshore systems of Papua New Guinea, and the Waipaoa River S2S system in New Zealand (e.g., Driscoll and Nittouer, 2000; Aalto et al., 2008; Francis et al., 2008; Jorry et al., 2008; Slingerland et al., 2008; Carter et al., 2010; Marsaglia et al., 2010; Kuehl et al., 2015). The S2S results invited a closer look at ancient sedimentary systems and, in particular, how quantitative analysis of a given part of the stratigraphic archive can be used to make inferences or predictions about upstream control, or downstream limits of strata. Such predictions are of keen interest to the hydrocarbon industry, especially in the prediction of coarser-grained reservoir sediment volumes and quality, prediction of deep-water reservoirs, and estimations of mud-prone hydrocarbon source-rocks and sealing facies (now also unconventional reservoirs), which are key stratigraphic components of all petroleum systems (Magoon and Dow, 1994). A key technique in characterizing and modeling subsurface hydrocarbon reservoirs is the use of both modern and outcrop analogs to constrain the size, shape, and complexity of sedimentary bodies (e.g., Reynolds, 1999; Bridge and Tye, 2000; Bhattacharya and Tye, 2004; Tye, 2004; Pranter et al., 2009; Deveugle et al., 2014). Picking a correctly scaled analog can be critical in reservoir prediction, and can be greatly informed by S2S concepts that help define the nature of linked up-dip to down-dip depositional systems as a function of the catchment parameters (Bhattacharya and Tye, 2004).

The focus of this paper is to review how S2S concepts are applied to quantitative analysis of deep-time strata, with an emphasis on paleodischarge and sediment budgets, to determine the scaling relationships among the various components of the S2S system. The paper also provides examples of how S2S concepts can be applied to petroleum systems, especially in predicting sediment volumes, sandstone–shale ratios, and heterogeneity of conventional and unconventional reservoirs. Insights about sediment source areas, and especially climate and tectonics, provide predictions about reservoir quality, which is largely a function of sediment composition. The scale of a depositional system is ultimately limited by factors such as catchment area, climate, and relief, which in turn control river discharge (Svistrik and Milliman,
2007) and the size and character of downstream linked depositional systems, such as deltas and submarine fans, and their components, such as lobes, channels, and bars.

A fundamental question is how S2S concepts can improve prediction of sediment partitioning among the alluvial, shoreline, shelf and deep-water environments, as embodied in the fulcrum approach of Holbrook and Wanas (2014). This approach can improve prospectivity of reservoir volumes in different parts of a linked source-to-sink sedimentary system. This paper emphasizes examples from the Mesozoic Interior Seaway of Western North America because they are very well studied; they contain prolific petroleum systems, and most examples were deposited in relatively closed systems, allowing sedimentary budgets to be estimated.

2. Sequence-stratigraphy overview

Sequence stratigraphy provides a practical geometric approach to stratigraphic analysis which, when integrated with chronostratigraphic or biostratigraphic data, allows magnitudes of base-level change to be quantified (e.g., Posamentier et al., 1988; Nystuen, 1998; Catuneanu, 2002; Bhattacharya, 2011) (Fig. 2). Sequence stratigraphy also provides a theoretical framework that explains how depositional sedimentary systems are linked and how they evolve as a consequence of changes in accommodation. Initial sequence stratigraphic studies posited that depositional systems (such as rivers, deltas, and submarine fans; which also form key hydrocarbon reservoirs) show predictable 3D stratigraphic organization (Fig. 2), which were primarily related to cycles of changes in accommodation. Many submarine fans, for example, were hypothesized to largely form in the distal areas of sedimentary basins during lowstands of sea level (Posamentier et al., 1988; Van Wagener et al., 1990; Nystuen, 1998). During sea-level hightstands, more-distal environments were generally inferred to experience slower deposition, resulting in organic-rich condensed sections, which may make excellent hydrocarbon source rocks, sealing facies, or unconventional reservoirs (Fig. 2).

In the original sequence stratigraphic models (e.g., Posamentier et al., 1988) sediment supply was assumed to be constant; lateral variability was seldom considered, and most sequence stratigraphic surfaces were considered to be chronostratigraphic. Since its original formulation, there have been many modifications to these models (e.g., Posamentier and Allen, 1993; Catuneanu, 2002; Strong and Paola, 2008; Catuneanu et al., 2009; Covault and Graham, 2010; Bhattacharya, 2011; Holbrook and Bhattacharya, 2012). There are now numerous studies, especially of modern systems, which demonstrate that climate, sediment supply, and shelf width have a significant influence on sequence formation and timing (e.g., Anderson et al., 2004; Walsh and Nittrouer, 2009; Covault and Graham, 2010). Sediment may be delivered directly to some deep-water systems during sea-level hightstands, especially in settings where shelves are narrow or where sediment supply is very high (e.g., Walsh and Nittrouer, 2003; Covault et al., 2007; Carvajal et al., 2009; Covault and Graham, 2010).

Inasmuch as S2S recognizes boundaries between depositional systems, the shoreline is fundamental in sequence-stratigraphic analysis. The concept of shoreline trajectory (Helland-Hansen and Martinsen, 1996) involves tracking the migration of shoreline facies throughout stratigraphic successions (Fig. 2). The stacking and trajectory of shoreline deposits and their linked fluvial and offshore shelf and deep-water deposits, can be analyzed to estimate magnitudes of accommodation change (e.g., sea-level rise and fall), and potentially sediment supply (Neal and Abreu, 2009) and more recently sediment budgets (e.g., Carvajal and Steel, 2012; Guilloucheau et al., 2012; Hampson et al., 2014). The shelf-slope break (Figs. 1 and 2) is also an important boundary that can be tracked in many stratigraphic data; and is commonly observed in regional seismic lines. Cyclic alternations of strata have been observed in which sedimentary wedges that lie basinward of the shelf-slope break alternate with more extensive strata that extend landward of the shelf-slope break. The onlapping basinal wedges were interpreted as lowstand deposits, whereas the landward deposits have been interpreted as hightstands (Vail et al., 1977; Fig. 2). Such regional changes in onlap patterns were used to reconstruct the history and magnitude of sea-level change throughout geologic time (e.g., Vail
et al., 1977; Haq et al., 1987), although not without controversy (e.g., Miall, 1997; Dickinson, 2003).

A fundamental aspect of the S2S perspective is that observations from one part of the system, such as an incised river and its valley, potentially permit predictions to be made about other parts of the system, such as the downstream shorelines or submarine fans fed by these valleys. This, of course assumes that sediment can be transferred across these boundaries and that effects of downstream sea-level changes or upstream climate changes can also be transmitted across these boundaries (Blum et al., 2013; Romans et al., 2015). Identification and correlation of key stratal-bounding surfaces permit time-equivalent depositional systems to be correlated and mapped across a systems tract, which potentially allows consideration of sediment budgets across S2S boundaries (e.g., Hampson et al., 2014). As we increasingly understand the role of sediment supply as a control on stratigraphy, it becomes more important to quantify water and sediment fluxes in deep-time systems.

Other boundaries that are important landward of the shoreline include the bayline and backwater limits (Fig. 3; Blum et al., 2013). The bayline commonly marks the landward limit of marine facies and tidal deposits. The backwater length is the ratio of flow depth divided by dimensionless river slope (Blum et al., 2013) and may control the gravel–sand boundary in some fluvial systems (Fig. 3). Backwater lengths range from hundreds of kilometers in low gradient continental-scale systems, such as the Mississippi (flow depth is 40 m and slope is $5 \times 10^{-5}$), down to a few kilometers in steep gradient systems with shallow rivers (Blum et al., 2013).

There has been much recent work that focuses on the relative importance of upstream and downstream controls on sedimentary systems (e.g., Blum and Törnqvist, 2000; Holbrook et al., 2006; Holbrook and Bhattacharya, 2012; Blum et al., 2013). Sea level may be far less important in controlling sediment accumulation and the behavior of rivers landward of the backwater length where climate and tectonics strongly impact discharge of sediment (Adams and Bhattacharya, 2005; Lawton et al., 2015) (Fig. 4). In fact, these recent studies suggest that sea-level rise or fall may have no effect landward of the backwater limit, indicating that sea-level changes may not be the dominant control across the entire S2S transect. However, these concerns must be tempered with the recognition that sediment produced by erosion of uplands is nevertheless routinely transferred into basinal sinks. However, not all grades of sediment are transferred downstream of the backwater limit, and upstream perturbations (i.e., signals) may be dampened or shredded during downstream transit (see Romans et al., 2015). There are numerous open questions regarding how sediment transfer occurs. For example what is the degree of intermediate staging of sediment along the transfer route, how do environmental signals propagate upstream and downstream across key environmental boundaries, such as the shoreline and the backwater length, what are the time lags in sediment transfer across these boundaries, and what are the partition coefficients between adjacent sinks (e.g., floodplain versus shelf). A number of these questions are addressed in the companion papers in this volume.

3. Critical considerations in applying S2S concepts to deep-time systems

A critical aspect of deep-time systems is that over geological time (i.e., $10^9$ years) much sediment may eventually be transferred across an entire S2S system, even if temporarily stored in intermediate sedimentary sinks at shorter time scales (see also Romans et al.,
Conversely, a significant proportion of sediment can be temporarily or permanently stored in these intermediary sinks (e.g., the alluvial plain), and may subsequently or never be transported across an entire S2S transect. For example, there are numerous well-documented examples of extensive non-terminal fluvial and other non-marine deposits, stored in fossilized intermediate sinks that span the rock record (e.g., Davies and Gibling, 2010; Gibling and Davies, 2012; Long, 2004, 2011; Ielpi and Ghinassi, 2015). Intermediate continental sediment storage can be important in low-gradient, continental-scale systems, and it is also important in the smaller-scale Cretaceous systems that we review in this paper. Syndepositional structural features, such as growth faults and folds, can also accommodate significant volumes of sediment, especially on continental margins, and may preclude or reduce transfer of sediment to deep water, as long as such faults are active (Galloway, 1986; Kuehl et al., 2015). These considerations affect how the stratigraphic record is used to make inferences about sediment source areas, or downstream sediment volumes, which may aid in prediction of down-dip sandstone reservoirs.

Another critical issue in deep-time systems is the chronometric resolution of the deposition of strata (see also Romans et al., 2015) and their subsequent diagenesis or deformation. In the modern, sediment transfer may occur at a variety of time scales, from days up to a few hundred years, but deep-time chronometers rarely have precision that is better than a few hundred thousand years. In some cases, unique events, such as volcanic eruptions, large-magnitude turbidity flows, meteor impacts and the like, which deposit individual beds, can be tracked and correlated in deep-time records (Ricci-Lucchi, 1995; Bhattacharya, 2011). Today's chronometric techniques can now routinely achieve an uncertainty of 0.5% of the total age, but this nevertheless can still represent a large time span, e.g., half a million years for the Cretaceous rocks we are going to consider. A precision of 0.1%, which is becoming the norm for modern geochronology, leaves an uncertainty of 94,000 years for strata whose estimated age is equivalent to the
Cenomanian-Turonian boundary (93.9 Ma). The precision obtained from isotopic methods will depend on the method used, and the method used depends on the type of rock or sediment present (Dickin, 2005). Without appropriate lithology (e.g., a zircon-bearing rhyolite; bentonite layer), the temporal resolution in a sedimentary sequence may be limited to understanding informed by the included fossils. As a consequence, linking short-term sedimentary processes, such as terrace formation in fluvial systems, to longer-term processes, such as the formation of sequence boundaries over the Milankovitch time-band, has long been problematic for deep-time studies (Sadler, 1981; Holbrook and Bhattacharya, 2012; Blum et al., 2013; Romans et al., 2015) But, it has been successfully achieved in Quaternary and some other examples (e.g., Semme et al., 2011, 2013; Semme and Jackson, 2013). Most recently, Singer et al. (2015) have achieved precision of 0.02% using 40Ar/39Ar Multicollectors in Cretaceous systems, which promises the ability to resolve individual Milankovitch Astrochronological cycles.

Lastly, time-stratigraphic analysis of deep-time systems shows that in any given vertical succession, very little geological time is actually represented by strata. Rather, many surfaces of non-deposition and erosion commonly separate stratigraphic units (Barrell, 1917; Ager, 1973; Sadler, 1981, Miall, 2015; Spencer et al., 2014 and references therein) (Fig. 2). However, this does not mean that the record is not preserved elsewhere in the basin (Sadler, 1981). As a consequence, documentation of the full stratigraphic record ideally requires 4D analysis, wherein Wheeler diagrams add the 4th dimension of time (Qayyum et al., 2015). Such 4D approaches of ancient S2S systems, have recently been attempted in several cases (e.g., Guillocheau et al., 2012; Semme et al., 2013; Hampson et al., 2014).

Quantitative S2S approaches also may use theoretical limits of sediment production (e.g., BQART model of Syvitski and Milliman, 2007; Hinderer, 2012) to constrain the volumes of mud, sand, and gravel that can be preserved in basins, as has recently been attempted by Holbrook and Wanas (2014) for Cretaceous strata in Egypt. This approach will be described for some of the Cretaceous examples discussed below. Prediction of sediment volumes is of fundamental importance to hydrocarbon resource estimates, and particularly conventional sand-gravel reservoir volumes, but may impact estimates of hydrocarbon-source rock volumes (which addresses volume of hydrocarbon resources) but this requires estimation of organic sequestration and concentration. Leithold et al. (2015) specifically address carbon cycling and organic sequestration in a variety of S2S systems, and this is critical to predicting petroleum systems, especially source-rock quality and unconventional shale reservoirs.

4. Components of S2S dispersal and depositional systems

The building blocks of all S2S systems include the catchment (source area), a transfer system (usually rivers), and a series of linked down-dip depositional systems, including shoreline, shelf and deep-water systems (Fig. 1). The first-order control on the ultimate size of any given S2S system is the size of the catchment area available to produce sediment (e.g., Matthai, 1990). To a lesser degree, the relief and climate of the catchment area play a role (Figs. 5 and 6). The area and relief of the catchment can be measured directly in modern systems (e.g., Syvitski and Milliman, 2007), but are typically more difficult to estimate in deep-time systems. Reconstructing the source area may require reconstruction of drainage basins based on drainage divides and plate-tectonic reconstructions (Fig. 7), especially in areas where tectonics have deformed or removed the source area, such as commonly occurs along rifted margins. Despite this limitation, there are a number of approaches to estimating the area, relief, and sediment yield from an ancient system (e.g., Bhattacharya and Tye, 2004; Miller et al., 2013; Holbrook and Wanas, 2014; Carvajal and Steel, 2012). In addition, source areas may be identified by provenance analysis, such as the dating of detrital zircons (e.g., Dickinson and Gehrels, 2008; Miller et al., 2013; Blum and Pecha, 2014) (Fig. 7), as discussed in Section 6.7.

Ideally, estimating S2S parameters should take a statistical or probabilistic approach in which a range of values may be estimated to place error bars on possible maximum and minimum sediment volumes and consequently, the scaling of linked depositional systems along a S2S transect. Particularly useful are scaling relationships, such as recently explored by Semme et al. (2009), who demonstrated correlations among a number of measurable parameters based on a compilation of Quaternary S2S systems. They showed positive correlations between river-length and submarine fan length, and an increase in submarine fan area with increasing catchment areas (e.g., Fig. 6), and a negative correlation between catchment area and shelf gradient. The scaling relationships are typically empirical in nature and involve searching for relationships (e.g., power laws) that demonstrate how parameters such as catchment area and river discharge are related.

4.1. Alluvial systems, catchments and transfer

In this paper we treat the catchment (i.e., drainage basin) as an area of net sediment production (i.e., export), especially where there is net

![Fig. 5. River discharge versus drainage area. From Mulder and Syvitski, 1995 after Matthai, 1990.](image1)

![Fig. 6. “Conveyor-belt” vs. “vacuum-cleaner” models for sediment supply. Conveyor-belt model indicates long-lived sediment supply from a large hinterland drainage basin in a piedmont valley, vs. local excavation of a coastal-plain paleovalley. The conveyor-belt model represents an extra-basinal river system, draining a larger catchment area, which should build a larger fan, compared to the smaller-scale, intra-basinal vacuum-cleaner model. Modified from Blum and Törnqvist (2000) and Blum et al. (2013).](image2)
uplift. However, we recognize that the catchment area may include its own depositional elements, such as alluvial fans and talus deposits (Fig. 1). For the purpose of this paper, we focus on rivers as the primary transfer mechanism for sediment. See Jaeger et al. (2015) for insights on cryogenically controlled S2S systems. Rivers often show a tributary pattern in the catchment area; channels join downstream to form a single trunk stream in the middle reaches (Fig. 1). On the flatter coastal plain, rivers commonly form a distributive pattern, creating either an aggradational distributive fluvial system (e.g., Weissmann et al., 2010) or a deltaic deposit. The length of the medial to distal parts of the S2S system (i.e., the area between a mountain front and the shoreline), and the area and relief of the potential catchment area, control the scale of the largest trunk rivers (Figs. 5 and 6) and this appears to affect the size of deltas and submarine fans (Sømme et al., 2009).

Blum et al. (2013) point out that continental-scale, integrated drainage basins, such as the Mississippi, are more likely to develop in low-gradient settings during icehouse times (e.g., Fig. 8). Sea-level falls across a low-gradient shelf during an icehouse, expose large areas resulting in the development of a wide coastal and alluvial plain and thus gives more opportunity for development of integrated drainage basins at times of low sea level. During Greenhouse times, higher sea levels, and consequent flooding of continents, created smaller, less...

---

**Fig. 7.** A. Paleogeographic reconstruction of early Triassic circum-Arctic drainage areas based on B. Detrital zircon analysis of Triassic outcrops around the circum-Arctic Ocean. Vertical axis in B refers to the percentage of age-specific detrital zircons versus the total population measured. Variations in age-spectra are associated with age of source areas. From Miller et al. (2013).

---

**Fig. 8.** A. Campanian versus B. Paleocene paleo-drainage patterns in North America. During the Cretaceous “Greenhouse” when global sea-levels were high, (A), the continent is flooded resulting in numerous separate small delta systems and associated local drainage areas. During up-lift of North America during the later Laramide orogeny (B), the smaller drainages become integrated and coalesce to form much larger S2S systems that feed deep-water fans, such as in the Gulf of Mexico. This is also analogous to Icehouse times when sea retreat and larger areas of land are exposed. Green outlines of paleo-land-masses largely based on reconstructions of R. Blakey, and Colorado Plateau Geosystems (2015); Paleodrainage in B is based on Blum and Pecha (2014). Gulf of Mexico Paleogeography from McDonnell et al. (2008).
integrated drainage basins resulting in smaller and more isolated S2S systems, such as those that characterized much of the Cretaceous of North America (Fig. 8). Of course, tectonics also play a major role in catchment development and scale, such as during early break of Pangea when continents were at higher elevations or during major orogenic plateau development.

For any given sedimentary basin, the contributing rivers may drain different areas. These fluvial systems can range from the largest-scale extra-basinal rivers that are linked to bedrock catchment areas to smaller-scale basin-fringe rivers, or the smallest-scale intra-basinal streams lying entirely within a depositional basin (Galloway, 1982; Blum and Törnqvist, 2000), pointed out that extra-basinal river systems act as conveyer belts, transferring sediment from upland catchment areas across and into the depositional basin, whereas intra-basinal valleys simply provide local material to more distal downstream sinks forming smaller-scale depositional systems (Fig. 6). Extra-basinal systems have also been described as piedmont incised-valley systems (Boyd et al., 2006) in which a “fall line” typically marks the transition from steeper to lower gradients. Downstream of the “fall line” both deposition and transfer of sediment can occur, whereas upstream erosion dominates (Figs. 1 and 6). Piedmont-valley systems are distinguished from coastal-plain valleys, in that the latter occupy the lower depositional reaches of a S2S system (Fig. 2). This paper will primarily focus on extra-basinal, piedmont-valley systems, as the associated trunk streams contain information about the entire upstream source area, and can be used to determine the area, relief, and provenance, but we emphasize here that it is important to attempt to distinguish these valley types in ancient systems.

The total sediment load, or volume, associated with a S2S system is primarily related to the size of the drainage area, although relief and climate play a secondary role (Matthai, 1990; Milliman and Syvitski, 1992), whereas the sediment yield (i.e., production of sediment per unit area) is more sensitive to climate and relief. This is critical in determining which type of S2S systems are more effective in delivering sediment to downstream sinks. Sediment yield plays a big role in the nature of oceanic delivery processes, such as hypopycnal versus hyperpycnal flows (Milliman and Syvitski, 1992; Mulder et al., 2003). Estimates of sediment yield and sediment load, integrated with the scale of the S2S system, can be used to predict the nature of downstream facies, with implications for unconventional oil and gas resources.

4.2. Shorelines, shelves, and along- and across-shelf transfer of sediment

A significant sink for much riverine sediment is the shoreline and adjacent shelf (Fig. 1). There are a variety of sub-environments within these depositional systems, including subaerial deltas, shorefaces, estuaries and their coeval offshore muddy shelves, typically manifested as mid-shelf mud belts (Hill et al., 2007). Sedimentation may occur relatively close to the river mouth (Walsh and Nittrouer, 2009), such as in a river-dominated delta system, and there is increasing recognition that small rivers that drain mountainous areas can deliver sediment directly onto the shelf via hyperpycnal flows when they experience flood conditions (Mulder and Syvitski, 1995; Parsons et al., 2001; Mulder et al., 2003). Sediment gravity flows, (such as turbidity currents, fluid muds, and hyperpycnal flows), are fundamental mechanisms that transfer sediment to shelf and deep water, and S2S analysis can help predict which types and scales of river systems generate these types of flows. In addition, a variety of shelf processes (chiefl y waves, tides, and oceanic currents) can transport significant sediment far from the point of fl uvial entry to seaward areas (Hill et al., 2007; Walsh and Nittrouer, 2009). Sandy bedload sediment may be moved and deposited alongshore as wave-dominated shoreface deposits (Plint, 2010; Bhattacharya and Giosan, 2003). Mud can also be transported along-shore to form a chenier plain (Warne et al., 2002; Rine and Ginsburg, 1985; Bhattacharya, 2010; Bentley et al., in this volume) or may be deposited farther offshore to form a mid-shelf mud belt (Fig. 9) (Walsh and Nittrouer, 2009 and references therein).

There have been a recent spate of studies documenting inner-to mid-shelf mud belts in which exceedingly large volumes of fine-grained sediment (chiefly clay and silt) may be transported over distances that may exceed a thousand kilometers from the point of riverine or deltaic depositional input (Hill et al., 2007; Liu et al., 2009; Walsh et al., 2015). In the 1980’s, it was thought that mid-shelf geostrophic flows, typically linked to storms, moved significant quantities of sand and gravel (e.g., Bergman and Snedden, 1999), but these models were later revised with the recognition that so-called “shelf” sands are more likely to be deposited during sea-level falls and left behind as top-truncated shoreface and delta deposits resulting in basin-isolated sandstones (Plint, 1988; Bhattacharya and Willis, 2001; Martinsen, 2003a, 2003b). It is now clear that incremental geostrophic transport is an important process for sediment transport on marine shelves (e.g., Suter, 2006; Bentley, 2003; Hill et al., 2007), but it is primarily responsible for the transport of mud rather than sand or gravel. This along-shelf transport must be considered in S2S budget analyses.

Other breakthroughs in research on mud transport include the pioneering work of Schieber et al. (2007), who suggested that bedload transport of fl oocule-mud ripples may be far more important in shelf systems than previously thought. These ideas are now being applied to the re-interpretation of ancient mudstones, previously thought primarily to reflect passive settling by suspension from hypopycnal flows during times of quiescence in the associated sea or lake sink (Bhattacharya and MacEachern, 2009; Plint et al., 2009). Wave-supported sediment gravity flows have also been related to ancient strata (Macquaker et al., 2010).

This array of processes have enormous implications for S2S systems, because of the potential extent of along-strike and across-shelf sediment transport and the need to account for this “export” in sediment budget studies (Draut et al., 2005; Hill et al., 2007; Kuehl et al., 2015). This is also important from the perspective of sediment budgets. In most S2S systems 70–95% of all the sediment carried by rivers comprises fine-grained to muddy suspended load, and accounting for all of this mud can be challenging in both modern and ancient systems (Walsh and Nittrouer, 2009; Kuehl et al., 2015). With the rise in importance of unconventional reservoirs, and especially shale plays, there has been renewed economic interest in the sedimentology and stratigraphy of mudstones, and this will be discussed with respect to several of the Cretaceous examples in Section 7.

4.3. Deep-water systems, the ultimate sink

Deep-water systems, such as submarine fans, are often considered to be the ultimate sink (Figs. 1, 2, 6 and 8). Sediment delivery to deep water is favored during lowstand conditions, when rivers debouch at the shelf edge (Mallarino et al., 2006; Posamentier and Walker, 2006). However, large river-deltas may reach the shelf edge or connect to canyons during highstand times, especially if the shelf is narrow or where submarine canyons tap into the littoral drift system, such as occurs today along the California coast (e.g., Anderson et al., 2004; Walsh and Nittrouer, 2003, 2009; Covault et al., 2007; Covault and Graham, 2010; Romans et al., 2015). Volumes of sediment in submarine fans can be linked and compared to long-term river discharge and have been used to estimate sediment budgets and sediment partitioning in Quaternary and older systems (e.g., Goodbred and Kuehl, 1999, 2000; Carvajal and Steel, 2012; Petter et al., 2013; Somme et al., 2011, 2013; Somme and Jackson, 2013).

5. Source area versus discharge

A significant aspect of the S2S concept is linking sediment source areas to the depositional record. In many modern S2S systems, much of the water (and sediment) is routed through a small number of incised
trunk rivers. Discharge measurements from these rivers can in turn be linked to the area, relief, and climate conditions of the drainage basin. For example, Matthai (1990) defined empirical power-law relationships that showed how drainage area could be used to predict discharge of downstream trunk rivers (Fig. 5), although these discharge estimates vary over an order of magnitude for any given area.

Syvitski and Milliman (2007) recently introduced the more sophisticated empirical BQART equation, based on analysis of hundreds of modern systems:

\[ Q_s = \omega B Q_w^{0.31} A^{0.55} R T \]

In this formulation, sediment discharge \( Q_s \) \((10^9 \text{t/yr})\) is related to water discharge \( Q_w \) \((\text{km}^3/\text{yr})\), basin area \( A \) \((\text{km}^2)\), relief \( R \) \((\text{km})\), and average temperature \( T \) \((\text{in } \text{°C})\). The B term accounts for lithology, glacial erosion, and anthropogenic effects (see discussion in Syvitski and Milliman, 2007), but glacial erosion and anthropogenic effects did not occur in the Cretaceous systems that we discuss below and can thus be ignored. The equation was developed for environments with average catchment temperatures greater than 2 °C, where \( \omega = 0.0006 \). Somme et al. (2009) used this approach to estimate S2S budgets in the Quaternary Golo River in Corsica, France, and its linked submarine fans, and showed significant improvements in the uncertainty of estimates by incorporating relief and temperature versus just catchment area. Carvajal and Steel (2012) used this equation inversely to estimate relief in the late Cretaceous Washakie basin in Wyoming, given that they had some control on the long-term sediment discharge from analysis of clinoform stacking patterns.

Paleogeographic and plate tectonic reconstructions of an ancient system may be used to characterize the sediment source area, especially when linked with provenance data, such as detrital zircons and other heavy minerals (e.g., Dickinson and Gehrels, 2008; Miller et al., 2013; Blum and Pecha, 2014). Relief may be estimated from thermochronometry, landward projection of slope estimates, or from geodynamic modeling (Allen and Allen, 2013), and temperature may be estimated from climate proxies (e.g., paleosols).

In many sedimentary basins, paleorivers drained the same areas as overlying modern rivers, allowing easier estimations of paleodischarge. The Mississippi drainage basin, for example, has been supplying sediment to the Gulf of Mexico for much of the Cenozoic (Winker, 1982; Galloway et al., 2011; Blum and Pecha, 2014, Bentley et al., in this volume; Fig. 8). In other systems, paleorivers and their associated drainage basins may be separated by rifting or other tectonic processes. For example, the source area that supplied the Permo-Triassic sandstones of the Sadlerochit Formation in the Brooks Range, Alaska, was rifted away from the depositional basin during the Jurassic opening of the Canada basin (Grantz et al., 1990), making it difficult to link the source to its sink (Fig. 7). The Sadlerochit Formation now lies on the Arctic–Alaska microplate, and interpretations about the original source area must be inferred from stratigraphic analysis, facies distributions, palaeocurrent analysis, plate tectonic reconstruction, and provenance analysis, including detrital zircons (e.g., Miller et al., 2013, Fig. 7). Linking source and sink in this case have important resource implications because farther to the northwest, the Sadlerochit Formation contains the Ivishak Sandstone, which hosts the supergiant, 24 billion-barrel Prudhoe Bay Field (Tye et al., 1999; Bhattacharya and Tye, 2004).

Rivers can also change course, as a consequence of local tectonics. For example, Wernicke (2011) suggested that during the Late Cretaceous, the paleo-Colorado River originally drained California, cutting the initial phase of the Grand Canyon and flowing east. Later Laramide uplift changed the direction of the Colorado River to flow southwest and finishing the carving of the modern Grand Canyon. A good example of river-course changes in older strata can be found in the Campanian fluvial deposits in the western interior of the USA, where renewed uplift caused significant paleogeographic reorganization and provenance change of rivers (e.g., Lawton et al., 2003; Adams and Bhattacharya, 2005).

**Fig. 9.** Orinoco and Amazon S2S systems. 50% of the mud in the Orinoco delta is derived from the Amazon river mouth (dark gray), representing 1100 km of along-shelf transport, primarily by wind-driven waves (Warne et al., 2002). Mud is also sequestered along the Suriname Chenier plain and inner shelf (brown color) (Rine and Ginsberg1985).
6. Paleodischarge, paleoslope, sediment budgets, and source-area estimates in deep-time systems

There have been a number of attempts to constrain S2S budgets in deep-time systems using a variety of approaches. One approach is to estimate both water and sediment paleodischarge using outcrop and subsurface data (Bhattacharya and Tye, 2004; Davidson and North, 2009; Li et al., 2010; Musial et al., 2012; Holbrook and Wanas, 2014). Water discharge (Q_w) of modern rivers varies over about 6 orders of magnitude (1 m³/s up to 10⁵ m³/s), and comparison with modern analogs is a key component in testing predictions. The BQART method of Syvitski and Milliman (2007) has also been applied to Quaternary systems (e.g., Sømme et al., 2009) although this requires information about the relief and area of the catchment basin. Sediment budgets and partitioning can also be calculated from chronostratigraphically constrained stratigraphic studies, in which the degree of aggradation versus progradation of clinothems is linked to sedimentation rates and can be compared with mapped volumes of sediment in distal shelf and/or submarine fan sinks (e.g., Hampson et al., 2013; Reijenstein et al., 2011; Musial et al., 2012; Miall, 2014) and, depending on seismic resolution, width and depth of channels may be extracted (Fig. 11).

6.2. Estimating channel dimensions and sediment characteristics

Key channel dimensions, including bankfull depth, width, and cross-sectional area may be measured or estimated. Grain size, sorting, and sedimentary structures (Fig. 12) can be used to quantify bed shear stress and to interpret bedforms and flow regime. Channel depth may be estimated on the basis of the thickness of fully preserved fining-upward facies successions, which represent individual channel stories, as seen in outcrop (Figs. 2, 10 and 12), cores or well logs. Maximum bar height typically represents about 80% of flow depth. Fully preserved bars may be marked by finer-grained drapes or a distinct rollover in accretion bedding, both of which may indicate the bar top (e.g., Hajek and Heller, 2012; Blum et al., 2013; Wu et al., 2015; Ullah et al., 2015, Holbrook and Wanas, 2014). Flow depths can also be calculated using the average thickness of dune-scale cross-sets (LeClair and Bridge, 2001; Adams and Bhattacharya, 2005; Li et al., 2010). Channel width can be calculated based on empirical equations (see references in Bridge, 2003), or may be estimated in well-exposed outcrops (Figs. 10 and 11C), such as the flow-perpendicular cliff exposures that can commonly be found in the outcrops of the western interior of North America. In cases where point bars are well exposed, the width of lateral accretion surfaces typically represents 50% to 80% of channel width (Bridge, 2003; Donselaar and Overeem, 2008). Channel width can also be estimated using plan-view images from 3D seismic data sets (e.g., Reijenstein et al., 2011; Musial et al., 2012; Miall, 2014), and rarely, plan view exposures in outcrops (Ielpi and Chinassi, 2014; Wu et al., 2015) (Fig. 11C). Holbrook and Wanas (2014) suggest that where

![Fig. 10](image_url) Erosional scour marks a base-of-valley sequence boundary (SB) overlain by a floodplain and its meanderbelt. Teal-colored dotted line distinguishes the smaller meandering channel (about 5 m deep and 20 m wide) from the larger valley-scour, Ferron Notom delta system, Utah.
cross sectional or plan-view data are absent, empirical width estimates may be accurate to a factor of 2–4 (Fig. 13), but when present width estimate errors are considerably lower (<2), but need to consider the orientation of the exposure with respect to paleocurrents.

Bedload sediment is easily measured in the field with a standard grain size card, usually providing a 1/2-phi precision. Sorting can also be estimated visually. Bedforms are interpreted based on observation of sedimentary structures, such as cross bedding, which is formed by subaqueous dunes, or cross lamination, which is formed by ripples. Thickness of dune-scale cross sets should also be compiled as they can be used to calculate flow depths (Leclair and Bridge, 2001). Paleocurrent data is used to determine the orientation of the outcrop face with respect to flow direction, which is critical for interpreting bar and channel types (i.e., braided versus meandering, Bridge, 2003), and simple trigonometry can be used to adjust apparent widths to true widths, especially for morphological reconstructions of channel aspects from point-bars.

Measurements are typically compiled in standard measured sections, showing thickness, grain size, and sedimentary structures (Fig. 12). The depth of channels is relatively easy to measure in the field, with a precision of about 1 m using standard tools (e.g., Jacob’s staff or measuring tape). Channel and bar dimensions can be obtained from facies architectural analysis based on bedding diagrams of photomosaics (Fig. 14). Lidar and photorealistic data can be used to avoid parallax, allowing more accurate measurements of channel and bar geometries (e.g., Pranter et al., 2009; Olariu et al., 2011; Hajek and Heller, 2012). Errors associated with field measurements, especially grain size and thickness are low, typically on the order of <10% (Fig. 13). Errors in width estimations are higher (typically 50–100%), and depend on the quality of exposures (Fig. 13). Errors in empirical equations that relate width and depth (e.g., Williams, 1988) also have an error of about a factor of 2 (Fig. 13). Width estimates are much less certain in braided streams than in single-thread meandering rivers, and the distinction between braided and sinuous channels is not always easy to make, especially in subsurface.
6.3. Paleoslope estimates

Slope is a fundamental control on S2S systems. Alluvial fans typically exhibit slopes in excess of 1.5° (Blair and McPherson, 1994) whereas natural river gradients are much lower and range over about 3 orders of magnitude \((10^{-3} - 10^{-5})\) (Fig. 3). Slope controls many aspects of the morphology of rivers and also influences the position of specific S2S boundaries, such as the shoreline, backwater, and bayline (Figs. 3 and 4). For example, slope affects river plan form (e.g., braiding is favored by higher slopes and meandering by lower slopes), depth of river incision, and facies boundaries, such as the gravel–sand and limit of tidal influence (Bridge, 2003; Blum et al., 2013).

Measuring low slopes in modern systems is easy; one simply calculates the elevation drop over some distance; however, measuring slopes in ancient systems is more challenging. Slope may be calculated using physics-based methods, such as the Shield’s thresholds for bedload movement, especially where the grain size, flow depth, and width of a paleo-river are known (Bhattacharya and Tye, 2004, Bhattacharya and MacEachern, 2009, Davidson and Hartley, 2010; Musial et al., 2012; Holbrook and Wanas, 2014), as described in Section 6.4.

Blum and Törnqvist (2000) suggest a simple geometric method for estimating paleoslope based on onlap limits of river-fed clastic wedges formed as a consequence of sea-level changes (Fig. 4). An alternate stratigraphic approach measures the elevation drop of coeval channel
or valley floors (i.e., parts of the fluvial long profile), with respect to a flat stratigraphic datum in flow-parallel cross-sections (Figs. 4 and 15). Estimates of slope, obtained from such stratigraphic data, can be used to estimate flow velocity and paleodischarge and provide an independent estimate that can then be compared to those made using the methods based on grain size, bedforms, and channel depths, as described below.

The errors in these geometric-stratigraphic slope estimates depend on the precision and accuracy of measurements of the thickness and length of onlapping clastic wedges. The precision of length and thickness measurements in stratigraphic studies using standard measuring tools (e.g., Jacob’s-staffs) depends on the scale of measurement, but are typically about 10% (e.g., thickness of 10 m ± 1 m, and lengths of 1 km ± 100 m). The accuracy of the measurement also is related to the orientation (i.e., obliqueness) of the cross section, especially if taken from 2D cross sections. Although the thickness of the wedge is not sensitive to orientation of the cross section, the horizontal extent of the wedge is strongly dependent. Longer apparent widths yield erroneously low slope estimates. Error may be resolved if the wedge can be mapped in 3D, and thus measurements can be made perpendicular to the shoreline. Also, if the degree of obliqueness can be determined, a geometric correction can be made using simple trigonometry. Compaction can also affect thickness estimates and may result in underestimation of slope. Compaction of about 50% can cause underestimation of slope by a factor of about 2 or more, but probably not an order of magnitude. The use of these stratigraphic techniques yields slope estimates within an order of magnitude, and perhaps better in high-quality datasets. Geometric-stratigraphic methods can also be compared with physics-based estimates, as described below.

6.4. Velocity and paleodischarge estimates

Flow velocity may be estimated by plotting grain size, bedform, and flow depth of a fluvial deposit on the 3D bedform phase diagram of Rubin and McCulloch (1980) (Fig. 16). This yields an estimate that typically ranges over a factor of 2. Alternatively Law-of-the-Wall relationships, or estimates based on Shields criteria for grain movement can be made based on grain size and sorting measurements (Musial et al., 2012; Holbrook and Wanas, 2014). For such calculations, bedload grain-size distribution (e.g., $D_{50}$) is required. The average velocity ($U$) is multiplied by the estimated channel cross-sectional area to obtain a discharge estimate ($Q_w$). If slope is known, the Manning or Brownlie (1983) equations can also be used to estimate velocity and discharge.
respectively, for a river of known depth, slope, and roughness coefficient:

\[ \overline{U} = \frac{1}{n} R_h^{0.66} S^{0.5} \]

Where \( \overline{U} \) is average velocity, \( n \) is the Manning coefficient (in \( \text{s/m}^{0.33} \)), \( R_h \) is the hydraulic radius (m), and \( S \) is the dimensionless slope.

The Brownlie (1983) equations can also be used:

\[ Q^{0.6539} = R/0.3724S^{-0.2542} \sigma_5^{1.035} D_{50} \]

\[ Q = \frac{Q_w}{W_g^{0.5} D_{50}^{1.5}} \]

wherein the dimensionless discharge \( (Q) \), is calculated based on the hydraulic radius \( (R) \), channel gradient \( (S) \), grain sorting \( (\sigma_5) \) mean grain size \( (D_{50}) \), paleodischarge \( (Q_w) \), channel width \( (W) \) and gravitational constant \( (g) \). Where river depth and velocity are already known, these equations can be inverted to calculate slope. Table 1 shows estimations for a number of examples.

Musial et al. (2012) used the Brownlie (1983) equation to estimate paleodischarge of the river deposits of the Cretaceous McMurray Formation in Canada (Fig. 17). They had data on grain size, point-bar thickness (30 m) and channel width (750 m), and estimated slopes of \( 10^{-4} \). Using these numbers they calculated a \( Q_w \) of \( 15 \times 10^3 \) m\(^3\)/s, about the same as the modern Mississippi River. We have independently estimated the velocity using the 3D bedform diagram approach described above. The 40 m deep McMurray channel comprise dune-scale cross bedded, medium-grained sands. Plotting these on the bedform phase diagram yields a bankfull velocity estimate of between 0.75 m/s and 1.5 m/s. The corresponding \( Q_w \) ranges from \( 17 \times 10^3 \) m\(^3\)/s to \( 34 \times 10^3 \) m\(^3\)/s, concurring with the \( Q_w \) calculations of Musial et al. (2012).

Velocity estimates using the Manning or Brownlie equations are sensitive to uncertainties in slope estimates and grain-size distribution. Grain size of the bedload fraction can be estimated quite precisely, but slope estimates can typically be made only to within an order of magnitude (see below). Although there is little uncertainty in measuring grain size and determining bedform type, both bedforms and grain size can vary considerably in any given channel deposit, reflecting both vertical and horizontal variations in flow. Maximum velocities, however, will be reflected in bedforms moving in channel thalwegs and transporting the coarsest grain sizes as bedload (rather than lag deposits). Ideally, velocities should be estimated using both the Manning and Brownlie equations as well as the bedform phase-diagram to test for congruity of values. Velocity estimates made using the bedform-phase diagram (which does not incorporate slope as a dependent variable) can also be used to estimate slopes.

6.5. Paleodischarge estimates using regional hydraulic geometry curves

Davidson and North (2009) use regional hydraulic geometry curves to estimate the discharge and drainage area of a number of ancient river systems based on the drainage area. These relationships are developed from analysis of rivers from different climatic and tectonic areas. Application to an ancient system requires some knowledge of the climate and morphological regime of the ancient drainage area (e.g., arid or humid, temperate or tropical, high or low latitude, mountainous or lowland).

Table 1

<table>
<thead>
<tr>
<th>River</th>
<th>Bankfull thalweg depth (m)</th>
<th>Mean bankfull depth (m)</th>
<th>Width (m)</th>
<th>Slope</th>
<th>( U ) m/s</th>
<th>Water</th>
<th>Sediment</th>
<th>DA</th>
<th>Fulcrum calculation of sed. volume in 20,000 years km(^3)</th>
<th>Sandstone isopach volume km(^2)</th>
<th>Total Sediment isopach volume km(^2)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mississippi</td>
<td>40.0</td>
<td>-</td>
<td>1500</td>
<td>0.00005</td>
<td>1.0</td>
<td>15</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>300,000</td>
</tr>
<tr>
<td>Lancer</td>
<td>7</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>66</td>
</tr>
<tr>
<td>Blackhawk</td>
<td>9.8</td>
<td>7.2</td>
<td>-</td>
<td>0.0007</td>
<td>0.4-2.2</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Ferron V1</td>
<td>8.4</td>
<td>4</td>
<td>140</td>
<td>0.0008</td>
<td>0.75-2.3</td>
<td>1.3-3</td>
<td>0.70</td>
<td>11.2</td>
<td>50</td>
<td>17.60</td>
<td>20</td>
</tr>
<tr>
<td>Ferron V2</td>
<td>5.5</td>
<td>3</td>
<td>70</td>
<td>0.0007</td>
<td>1.9</td>
<td>0.4</td>
<td>0.33</td>
<td>2.8</td>
<td>50</td>
<td>0.83</td>
<td>5</td>
</tr>
<tr>
<td>Dunvegan</td>
<td>16-20</td>
<td>11</td>
<td>170-200</td>
<td>0.0003-0.0005</td>
<td>0.75-1.6</td>
<td>1.6-6</td>
<td>-</td>
<td>125</td>
<td>-</td>
<td>-</td>
<td>20-75</td>
</tr>
<tr>
<td>Paddy</td>
<td>7.0</td>
<td>-</td>
<td>100</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>McMurray</td>
<td>40</td>
<td>80</td>
<td>-</td>
<td>0.0003</td>
<td>0.75-1.5</td>
<td>15-34</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>300</td>
</tr>
<tr>
<td>Paddy</td>
<td>7.0</td>
<td>-</td>
<td>100</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>Chinde/Dockum</td>
<td>10-15</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
</tbody>
</table>
Such climate information may be derived from direct observation of the sedimentary facies (e.g., analysis of floodplain paleosols) or plate-tectonic and paleogeographic reconstructions, which place a system within a corresponding paleo-climate belt. The data for this approach require estimates of bankfull channel depth, which can be estimated from point-bar dimension, or channel-fill thickness, as well inferred climatic and tectonic regime. Regional hydraulic geometry curves (Fig. 18) use power-law relationships of the form:

\[ D = aA^b \]

or

\[ Q_w = aA^b \]

(Davidson and North, 2009), where \( D \) = channel depth, \( Q_w \) = discharge, and \( A \) = drainage area. The constants, \( a \) and \( b \), are empirically determined using survey and stream gauge data from analog catchment areas (Davidson and North, 2009 and references therein); one example is shown in Fig. 18. \( Q_w \) and \( D \) may be estimated from the field or subsurface using the approaches described above. Errors and uncertainties associated with use of these curves typically ranges over a factor of 2–4, and up to an order-of-magnitude.

6.6. Estimates of sediment budgets from paleodischarge

Expanding on the techniques described above, Holbrook and Wanas (2014) calculated longer-term sedimentation rates to estimate sediment budgets of Cretaceous paleo-rivers, exposed in Egyptian outcrops. This required estimation of annual sediment discharge, which in turn required estimating flood frequency. The bulk of sediment transfer in modern rivers typically occurs during large

---

**Fig. 17.** Point Bars of the McMurray, Formation, Alberta, Canada. A. seismic time slice taken from about 400 m below the surface. B. Interpretation of seismic time slice showing point bars and abandoned channels. C. Three-dimensional reconstruction showing paleochannel and paleohydraulic estimates. Inclined point-bar lateral accretion surfaces are locally draped by shale forming inclined heterolithic strata (IHS) that inhibit fluid flow and contribute to the reservoir heterogeneity. From Musial et al. (2012).

**Fig. 18.** Plot of drainage area versus bankfull discharge for climate regimes in Florida. From Davidson and North (2009).
floods, with recurrence rates of 1.5 times per year (Svitytski and Milliman, 2007), and modern rivers generally experience flood conditions for <10 days per year, which is less than 3% of the time on an annual basis. Holbrook and Wanas (2014) used empirically established scaling relationships (e.g., Wright and Parker, 2004; Van Rijn, 1984) to evaluate instantaneous bedload and suspended load discharge. These instantaneous values were then integrated over longer geological time periods, based on chronostratigraphic analysis and sediment flux estimates. Assuming that there is good age control, such sediment discharge estimates can then be compared with sediment yields in the upstream drainage basin and the amount of sediment in the downstream sinks, such as might be obtained from isopach maps of deltas or submarine fans (e.g., Carvajal and Steel, 2012; Guillocheau et al., 2012). Guillocheau et al. (2012) present methods to reconstruct or extrapolate cross-sectional geometries of stratal units in areas where distal sinks are incompletely sampled or imaged. Many 2D seismic lines, for example, may not sample the entire extent of a basin. The reconstructed stratal cross-sections can then be used to extrapolate isopach maps, allowing total sediment volumes to be estimated.

In their example from the Cretaceous Bariyah Formation of Egypt, Holbrook and Wanas (2014) show that estimated sediment delivery exceeded the calculated volumes of sediment transferred through the trunk rivers, leading to speculation that some sediment was exported out of the basin. Their analysis also reveals that the scale of Cretaceous rivers in Egypt was considerably smaller than the modern Nile system, and helps explain the preponderance of associated reeval carbonate formations that would likely be absent had the area been drained by a much larger river.

Holbrook and Wanas (2014) provide a detailed analysis of the uncertainties associated with each step of their analysis, including many of the parameters discussed above, such as channel width and depth (Fig. 13). The largest uncertainties in their study are associated with their empirical estimates of channel width from 1D-measured sections, as they lacked good cross-sections to better constrain channel width estimates. Another area of uncertainty is the conversion from short-term estimates of water and sediment discharge to longer-term rates (Fig. 13), which is required to link paleo-discharge estimates to the stratigraphic record. It is generally known that most fluvial sediment discharge occurs during large-magnitude, and rare flood events, but estimating the frequency of events is difficult in a deep-time system. Discharge obviously varies as function of climate regime (e.g., monsoon versus arid, Davidson and North, 2009) but also varies with decadal climate events, such as ENSO or El Nino cycles, as well as longer-term glacial Milankovitch events. Analysis of floodplain paleosols and the use of analog regional hydraulic geometry curves, or integration with detailed paleo-climate models may aid in making more accurate assumptions about intermediate- and long-term climate controls on sediment delivery.

6.7. Drainage area and reservoir quality from provenance analysis

Estimations of drainage basin area, lithology, locations of former drainage divides, and tectonic setting can be made from provenance analysis, a field of research that has typically focused on sandstones rather than mudstones (Dickinson and Suczek, 1979; Mack, 1984; Dickinson, 1985). Provenance analysis by means of sandstone petrology also has important implications for estimations of reservoir lithology and quality in hydrocarbon prospects. Volcanic arc terranes, for example, typically result in clay-rich volcanic litharenites, with an abundance of labile framework grains, such as volcanic rock fragments, and unstable minerals, such as plagioclase, mica, pyroxenes, and amphiboles (e.g., Marsaglia and Ingersoll, 1992; Critelli and Ingersoll, 1995). These can undergo extensive diagenesis during burial and compaction, resulting in widespread occlusion of initial porosity forming low-quality-reservoir sandstones. In contrast, erosion of shield terranes, such as stable continental interiors, typically results in multi-cycle, quartz-rich sandstones that generally have higher reservoir quality (e.g., Dickinson, 1985; Johnsson, 1993). Sediment composition of reservoirs is ultimately related to its provenance, which relates to source rock composition (e.g., Pettijohn et al., 1987). Composition also depends on weathering and uplift rate in the source area, and weathering of sediment in intermediate sinks (e.g., Franzinelli and Potter, 1983; Johnsson, 1990a, 1990b, 1993). For example, humid tropical climates are typically characterized by high rates of chemical weathering, such that carbonate rocks experience extensive dissolution. Labile minerals in siliciclastic and igneous rocks more readily weather to produce clay, instead of sand or gravel.

Sandstone composition is traditionally determined based on the proportion of quartz, feldspar, and rock fragments (QFL) plotted on a ternary diagram (Dickinson and Suczek, 1979). Other approaches use isotopic geochemical methods to determine the geochronology of specific grains (e.g., feldspars, micas, apatites, and zircons) including $^{180}O$, $^{87}Sr/^{86}Sr$, and $^{207}Pb/^{204}Pb$.

Analysis of age-spectra of detrital zircons has recently become one of the most-used provenance analysis techniques (e.g., Rainbird et al., 1992; Gehrels et al., 1995; Dickinson and Gehrels, 2003). Because the closure temperature of Pb in zircon is very high (> 700 °C, Lee et al., 1997), and zircon is very chemically stable in the sedimentary environment, U–Pb zircon ages are difficult to change during metamorphism, burial, or weathering. Thus, chronometric analysis of the age-range of detrital zircons in any given sample can fingerprint the rocks contributing to the sediment in a basin at a given time. The fingerprint reflects the age of formation of the rocks or the age of formation of the provenance, in the case of zircons in sandstones (e.g., Gehrels et al., 1995). Detrital zircon age-spectra can be compared to plate tectonic and general geological reconstructions to locate source areas, and thus offers great promise for improved quantification of drainage basin area and location.

![Fig. 19. Paleodrainage of river systems developed in the Triassic. Compilation based on Gibson and Barclay (1989) for the Alberta systems; Dubiel (1994) and Dickinson and Gehrels (2008) for the Chine-Dockum, and Miller et al. (2013) for Arctic Canada. Green outlines of paleo-land-masses largely based on reconstructions of R. Blakey and Colorado Plateau Geosystems (2015).](image-url)
Hf isotopes. Because Hf and Zr are similar geochemically, most consensus in combination with a recent addition to the detrital geochemistry deed were continental-scale rivers that drained most of North America. Musial et al. (2012) estimate to exhibit a Mississippi-scale discharge, in the McMurray Formation, which were 30 to 40 m deep and which axial river system transported sediment parallel to the Sierra Madre northeastern Mexico (Lawton et al., 2009), and indicate that a long the Difunta Group in the foreland basin of the Sierra Madre Oriental in western Laurentia are present in Cretaceous-Paleogene sandstones of way, detrital zircon ages characteristic of basement provinces in south-sible source regions of the Chinle-Dockum system. Used in a similar (Fig. 19). These distinctively-aged zircons are rare in rocks of other pos-

orth and the resultant access of the drainage basins to detrital zircon grains.

Recent maps of drainage-basin evolution for North America (Figs. 7, 8, 19 and 20) specify the extent of paleo-drainage basins on the basis of distinctive detrital zircon populations in sandstones (e.g., Dickinson and Gehrels, 2003, 2009; Dickinson et al., 2010). The headwaters of these river systems can potentially be defined by integration of grain ages and basement-rock distribution in time-equivalent orogenic belts or uplifted rift flanks, as inferred from extant outcrop belts (Moecher and Samson, 2006). The resulting drainage basin estimates, however, depend upon correct interpretation of headwater source-rock ages and the resultant access of the drainage basins to detrital zircon grains.

Recent maps of drainage-basin evolution for North America (Figs. 7, 8, 19 and 20) specify the extent of paleo-drainage basins on the basis of distinctive detrital zircon populations in sandstones (e.g., Dickinson and Gehrels, 2003, 2009; Dickinson et al., 2010). The headwaters of these river systems can potentially be defined by integration of grain ages and basement-rock distribution in time-equivalent orogenic belts or uplifted rift flanks, as inferred from extant outcrop belts (Moecher and Samson, 2006). The resulting drainage basin estimates, however, depend upon correct interpretation of headwater source-rock ages and the resultant access of the drainage basins to detrital zircon grains.

6.8. Use of detrital zircons for formation age determination and thermochronology (uplift and exhumation rates)

Sometimes, the ages of detrital zircons can also help determine the depositional age of a sandstone. For example, Spencer et al. (2014) analyzed zircons from the Van Horn Formation in west Texas. Because of its stratigraphic position and lack of fossils, most workers had long considered the Van Horn to be Precambrian. In the upper portion of the Van Horn, Spencer et al. (2014) found several Cambrian zircons, one grain with an age of 521 ± 6 Ma. Because the U–Pb system in these zircons could not have been reset by subsequent events, we now understand that at least the upper part of the Van Horn must be Cambrian. This technique has been used to better understand the time of deposition of other unfossiliferous rocks; in such cases, care must be taken to analyze a sufficient number of zircons to give reasonable confidence that the extremes of the distribution have been identified (Vermeech, 2004).
Because chronometric analysis using the U-Pb method in zircons is insensitive to the subsequent thermal history of the grains, this approach is useful in identifying the formation age of the material in the source area of a sandstone. This is, however, complicated by the possibility of multi-cycle grains (e.g. Campbell et al., 2005). Other chronologic methods, which are sensitive to resetting at temperatures as high as 400 °C to as low as ~70 °C, have been used to further understand the tectonics of source areas. In these cases, the apparent age of the mineral does not indicate the age of formation but rather the last time the mineral passed through a certain temperature range, which can in turn be linked to its burial depth and time of subsequent uplift. The most commonly used isotopic tools for this sort of provenance evaluation are, in order from lowest temperature sensitivity (i.e., high closure temperature) to highest temperature sensitivity, muscovite (^{40}Ar/^{39}Ar), K-feldspar (^{40}Ar/^{39}Ar), zircon (U + Th/He and fission track), apatite (fission track) and apatite (U + Th/He).

Copeland and Harrison (1990) determined the ^{40}Ar/^{39}Ar ages of detrital K-feldspar and muscovite from strata of ODP sites 717–719 at the distal end of the Bengal fan. Despite analyzing what today would be considered a very small number of grains from each stratum (deposited from ~16 Ma to ~2 Ma), they found some individual muscovite and K-feldspar grains whose cooling ages were indistinguishable from the depositional age of the fan sediment. This indicated that some portion of the source area of the Bengal fan was being exhumed and transported to the deep sea at very rapid rates. More specifically, this gives evidence that the closure of these systems to the loss of Ar occurred several km below the surface of the Himalaya, followed by uplift, exhumation, transport, and deposition, 3000 km to the south over a very short time window. Although these results show that some parts of the Himalaya were experiencing an extreme rate of uplift and erosion during the Middle to Late Miocene, it is unlikely that this is representative of the whole mountain range. If such rates of uplift and erosion were regional, this would result in exhumation and extensive exposure of more deeply buried high metamorphic grade granulite rocks in the Himalaya, but such granulite outcrops are largely absent.

Similar studies have been undertaken on detrital material deposited closer to the Himalaya. For example, Bernet et al. (2006) analyzed zircons from ~14 Ma old to modern day foreland-basin deposits of the Siwalik Group in western Nepal by both the U-Pb and fission-track dating methods. They found that the difference between the depositional age and the youngest fission-track zircon age (closure temperature ~220–240 °C) was around 4 million years, suggesting parts of western Nepal were being exhumed at rates of about 2 mm/y and transported to downstream sinks during this time.

In the Himalaya, grains with cooling ages equal to or very close to the age of deposition necessarily indicate order-of-magnitude faster rates of erosion (up to 1 cm/y) taking place over at least brief intervals of time. Cooling ages of grains equal to or very close to the age of deposition could alternatively represent volcanic zircons, erupted at the time of sedimentation. However, during the Indo-Asian continent-continent collision, there has been little volcanism in the Himalaya, and consequently there is no documented volcanic detritus in the sediments of the Siwalik Group and Bengal Fan.

In continent-ocean convergent tectonic settings, where knowledge of erosion rates is also of great value, interpretation of such “zero-age” grains is complicated by the possibility that the cooling age was not attained during the relatively slow transition past some critical isotherm many kilometers below the surface, but rather during the extremely rapid transition from magmatic temperatures to surface temperatures during an explosive volcanic eruption. Painter et al. (2014) addressed this problem in a study of Jurassic-Cretaceous foreland basin deposits of the North American Cordillera in Utah, Colorado, Wyoming, and South Dakota. (U-Th)/He ages from zircon and fission-track ages from apatite were determined from some of these sandstones. In many cases, the same material was analyzed for the U-Pb age as well. Grains in which the crystallization-age, as determined from U-Pb analysis, matches the cooling age, as obtained from He or fission-track analysis, and is also the same age as the sediments, are very likely volcanic, and therefore should not be used in evaluating the erosion rates of the source area of these deposits. Conversely, grains in which He or fission-track ages are similar to the depositional age of the sediment, but show much older U-Pb ages (i.e., crystallization age), indicate older grains that have not been buried for long, and these can be used to argue for significant rates of erosion in the source area and deposition as sediments a short-time later. Painter et al. (2014) use populations of detrital zircons and apatites not contaminated with a volcanic component to suggest erosion rates in excess of 1 mm/y in the Sevier fold-and-thrust belt during the Cretaceous.

Another concern in using thermochronology to inform us about the tectonics of the source area of a sandstone, is the potential for post-depositional resetting. This is most likely with low-closure-temperature systems, such as the fission-track and (U-Th)/He methods. During burial, the grains of interest may achieve temperatures sufficient to produce open-system behavior and lessening of the apparent ages, making the ages a reflection of the thermal history of the basin, rather than the tectonics of the source area. Painter et al. (2014) found this not to be a concern for the Cretaceous foreland basin deposits in the central Rockies, but van der Beek et al. (2006) determined that the Neogene foreland basin deposits of the Siwalik Group of western Nepal had been buried to a depth such that the pre-Siwalik history of many of the apatite grains has been essentially erased. As such, the fission-track ages of these apatites now reflect the exhumation of these foreland basin deposits, rather than the exhumation of the material that now fills these basins.

6.9. Tectonics, subsidence and sediment supply

Uplift and erosion are intimately linked processes (Beaumont et al., 1992), and of course, subsidence within sedimentary basins (which ultimately controls preservation) includes both compactional subsidence, isostatic depression caused by sediment loading, and larger-scale lithospheric subsidence, such as may be related to larger-scale geothermal or plate-tectonic processes. One of the simplifying assumptions in much of the early sequence stratigraphic literature (e.g., Posamentier et al., 1988; Jervy, 1988) was of constant sediment supply, but of course in nature, this is never true. Rates of erosion, and consequently sediment production, may vary over an order of magnitude, especially passing from glacial to interglacial cycles (e.g., Hinderer, 2001), but changes in sediment supply were rarely invoked to explain stratigraphic changes in sedimentary systems in the early sequence stratigraphic literature. The S2S approach has revitalized a focus on sediment supply and its short- and long-term controls, as indicated by the papers in this volume.

7. Examples of Mesozoic S2S systems in Canada and the USA

The Mesozoic history of North America records the pulsed evolution of the collision and subduction of the Pacific and associated oceanic plates. This resulted in a complex history of terrane accretion in the Western Cordillera, development of the Rocky Mountains and associated foreland basins, and coincided with globally high sea-levels. There are corresponding profound changes in the scale, organization, and paleogeographic development of associated drainage basins, their rivers, and linked down-dip depositional systems. These changes also reflect variations in play types and hydrocarbon systems. This section of the paper reviews and synthesizes the paleogeographic history of central North America, with a focus on examples from the USA and Canada. We document changes in the size of rivers, and show how changes in channel depth and paleo-discharge can be used to interpret changes in drainage basin area and position through time, and the implications for the size of down-dip depositional systems and hydrocarbon potential.
7.1. Triassic–Jurassic

Paleo-rivers in the Triassic Chinle and Dockum formations, in the southwestern U.S.A. that formed during the early break-up of Pangea, lowly flowed to the northwest draining Laurentian and Appalachian source areas (Dubiel, 1992; Dickinson and Gehrels, 2008; Miller et al., 2013) (Fig. 19). Sediment contributions from western North American and Pacific island arc systems were minor. A major drainage divide lay to the southeast (Fig. 19), roughly in the position of the older Laurentian transcontinental arch corresponding to the Paleozoic Appalachian–Ouachita orogenic belt, which was in the early stages of rifting. Catchment areas are correspondingly interpreted to be on the order of 6 × 10^4 km^2 for rivers draining the Mogollon highlands to the south, and 2 × 10^6 km^2, for rivers draining eastern Laurentian source areas.

Facies analysis of corresponding incised valley-fills in associated Triassic strata, such as the Dockum Formation in Texas and its equivalents in the Chinle Formation of Utah (e.g. Dubiel, 1992, 1994), suggests that the largest rivers were on the order of 10–15 m in depth. These rivers fed their sediments into the deforming continental margin of Idaho and Nevada (Lupe and Silberling, 1985; Dickinson and Gehrels, 2008). Extensive marine Triassic facies are also documented in western Canada (Gibson and Barclay, 1989; Edwards et al., 1994) and thick fluvo-deltaic systems produced from draining of the northern Laurentian continent are found in Alaska (Fig. 19) (Miller et al., 2013). However, detailed palaeohydraulic analyses are lacking.

The Jurassic Morrison Formation is the next major Mesozoic fluvial unit exposed in the region and reflects a significantly greater contribution of sediment supplied from the emerging Western Cordilleran fold and thrust belt (Fig. 20) (Hansley, 1986). Extensive exposures, and a number of corresponding studies focused on facies architectural analyses, suggest that rivers of the lower Salt Wash Member reached a maximum of about 18 m deep (Robinson and McCabe, 1997). Owen et al. (2015) quantified changes in regional thickness of channel stories associated with the east-flowing, transverse river systems, which drained the rising Cordillera. Channel stories (and by inference channel depths), change from a maximum of about 7.7 m in more proximal locales to 2.3 m deep in downstream areas, suggesting a distributive fluvial system with an apex near Las Vegas, Nevada. The largest transverse streams were on the order of 8 m deep, indicating relatively small drainage systems compared to the 18-m-deep axial river systems that drained the Mogollon highlands to the south.

In a study of the Jurassic rivers in the Alberta basin, which were in the early stages of foreland basin development, Raines et al. (2013) integrated new detrital zircon age dates from the Jurassic Monteeite Formation, equivalent in age to the fluvial Brushy Basin Member of the Upper Morrison in Utah. The Monteeite includes two fundamentally different fluvial systems. A larger system drained both the Mogollon highlands to the south and an Appalachian source to the southeast, and these catchments collectively fed a north-flowing river system that drained parallel to the axis of the foreland basin and fed shorelines of the growing Cordilleran forebulge (Fig. 20). Unfortunately, the record of this major river was largely eroded by the younger Cretaceous Cadomin Formation (Kukulski et al., 2013), but it is interpreted to have been continental in scale. A secondary source of sediment was delivered from the emerging Cordillera to the west, which formed a smaller series of transverse drainages. Sediment derived from the western source areas became more dominant as the foreland basin filled. The western catchment areas are estimated to have been about 100,000 km^2, delivering sediment to coarse-grained downstream distributive fluvial systems (e.g., Weissmann et al., 2010; Hartley et al., 2010). Channels were up to 12 m deep, and show variations in the proportion of channel belt versus floodplain facies (i.e., net-to-gross) through time.

From the hydrocarbon perspective, correlation and modeling of channel belt and floodplain stratigraphic patterns, and determination of the size of channels and channel belts, were based on estimated thickness, width and measured net-to-gross ratios, and were then used to identify prospective sandstone-prone reservoir bodies. These sandstones form part of the deeply-buried tight gas deposits in the Alberta Basin, in which a higher degree of channel-belt amalgamation is required to provide economic wells. Inasmuch as the thickness of channel fills reflects the drainage area, estimation of channel size allows estimates of channel belt width to be made, which is essential in predicting reservoir connectivity (Larue and Hovadik, 2006).

7.2. Big rivers in the early Cretaceous

The Late Jurassic drainage patterns continued into the Early Cretaceous, and are characterized by a number of amalgamated compound valley systems reflecting foreland basin tectonics and the complex heterogeneity of the North American craton, inherited from its assembly (Fig. 20). The Lower Cretaceous is marked by a major unconformity associated with a number of incised fluvial units, including the Cloverly, Lakota and Dakota formations in the USA, and the Mannville Group in Canada. In the Alberta basin, a series of NW-SE trending compound valley systems separated by major interfluves have been identified (Fig. 20). These valleys are tens of kilometers wide and up to 50 m deep. The easternmost Assiniboia valley is linked to the southern continental-scale catchment network, whereas the westernmost Spirit River Valley and the middle Edmonton Valley include a greater contribution from the Western Cordillera (Smith, 1994; Benyon et al., 2014; Blum and Pecha, 2014). Like the Jurassic systems, the larger early Cretaceous rivers flowed parallel to the foreland basin, forming an axial system that drained the eastern Appalachian Mountains of Laurentia, and the southern Mogollan highlands, with lesser contributions from the Western Cordillera (Fig. 20). Coarse-grained river systems of the Cloverly Formation consisted of moderate-sized rivers, up to 11 m deep, with slopes ranging from 6 × 10^{-3} up to 6 × 10^{-4}, and discharges up to 1000 m^3/s (Zaleha, 2013). The largest rivers in the Cloverly Formation drained axially, whereas the smaller rivers drained transversely, and the overall eastward increase in discharge was partly attributed to an increase in precipitation away from the rain shadow of the Early Cretaceous Sevier mountain belt (Zaleha, 2013).

Analysis of fluvial deposits in the supergiant heavy-oil reserves of the McMurray Formation in Canada indicate axially drained, continental-scale meandering rivers on the order of 40 m deep (Fig. 17), with paleo-discharge estimated to be on the order of 15 × 10^3 m^3 (Musial et al., 2012), comparable to the modern Mississippi in scale. The transcontinental drainage divide lay in the southern USA and sediment was derived from the southern extension of the Appalachians (e.g., Ouachita Mountains in Oklahoma and the Marathon uplift in west Texas, Fig. 20(,)). Detrital zircon analysis of McMurray sandstones indicate Appalachian and Grenville basement sources in eastern North America and corroborate the continental scale of the McMurray drainage basin (Benyon et al., 2014).

A key aspect that affects the productivity of Mannville heavy oil petroleum systems (Fig. 17) is the heterogeneity within channels (Fustic et al., 2012). In subsurface production, oil is recovered with paired subsurface horizontal wells using the steam assisted gravity drainage (SAGD) technology. Steam injected in the upper well liquefies the oil and allows it to drain into the lower well and then pumped to the surface. Extensive muddy inclined heterolithic layers (IHS), associated with point bars (Fig. 17C) are barriers that impede steam injection and inhibit flow of the oil, and prediction of these barriers is thus critical (Thomas et al., 1987). In areas where oil sands are mined, the IHS mudstone layers slow the mining operation, as they do not contain oil and must be excavated and disposed of before processing of sands can resume. Understanding that Mannville point bars are formed by a continental-scale river is reflected in the large-scale of channel belts. As a consequence, individual inclined layers are laterally more extensive than the typical well spacing, thus allowing them to be correlated and mapped explicitly, versus modeled stochastically (e.g., Fustic et al., 2012).
sodes (Smith, 1994; Hayes et al., 1994). At times shorelines lay in
Mannville Group contains a number of important transgressive epi-
complexes of the Bluesky Formation extended along strike for over
an order of 3 × 104 km2 (Leckie, 1986; Hayes et al., 1994). The areal scale
parasequences, tens of meters in thickness, and cover areas on the
Columbia.

...Mannville Rivers were deposited within a few hundred kil-
ometers of the shoreline and were thus likely within the backwater
length of the system. Blum (2015)in contrast, used 3D seismic images
to document the migration patterns of Mannville channels and sug-
gested that the high sinuosity and the high degree of lateral migration
of meanders require the channels to have been landward of the backwa-
ter reach and thus beyond the limit of tidal influence, although this is
incompatible with the observation of marine microfossils in associated
mudstone drapes.

The Mannville valleys include major axial systems, as well as smaller
steeper-gradient transverse tributary systems, draining the Western
Cordillera, which fed the extensive shorelines of the sand- and gravel-
prone Fahlr Formation and associated prodeltaic and shelf mudstones
of the Spirit River Formation in Northern Alberta and British Columbia.
These reach up to 300 m thick. The Edmonton valley system also fed
shorelines of the Bluesky Formation in northern Alberta and British
Columbia.

The Fahlr shoreface and deltaic sandstones form individual
parasequences, tens of meters in thickness, and cover areas on the
order of 3 × 105 km2 (Leckie, 1986; Hayes et al., 1994). The areal scale
of the associated sandy delta systems (Fig. 20) is broadly similar to mod-
ern Mississippi or Niger, but thickness is quite a bit less, and is consistent
with the smaller catchments of the Western Cordillera. Isopach sand-
stone volume of an individual mapped deltaic system of the Fahlr
(i.e., a parasequence) is estimated to be on the order of 3 × 105 km3. Total sedimentary volume, including mudstone and sandstone, is likely
an order of magnitude greater (e.g., 106 km3). Paleogeographic mapping
by Smith (1994) demonstrated that Upper Mannville shoreline
complexes of the Bluesky Formation extended along strike for over
1000 km and along dip for 600 km, suggesting widespread shoreline
deposits consistent with sediment supply associated with the larger
river systems of the Assiniboia valley.

...Mannville Group contains a number of important transgressive epi-
sodes (Smith, 1994; Hayes et al., 1994). At times shorelines lay in
central Alberta, and at other times shorelines prograded several hun-
dred kilometers seaward into British Columbia and beyond. Unfortu-
nately, the distal limits of deep-water systems fed by the largest
Mannville rivers are largely eroded as a result of later foreland
basin tectonics and Quaternary glacial erosion, consequently no re-
cord exists of linked shelf clinofoms or deeper-water submarine fans.
Much of the distal sediment delivered to this Seaway is either
now contained within the Rocky Mountain fold and thrust belt, or
eroded away. However, comparison with the volumes of sediment
on modern continental-scale submarine fans, such as the Mississippi
(Walker, 1992) may be used to suggest that Mannville rivers deliv-
ered Mississippi-scale volumes (3 × 105 km3) to the Boreal Seaway,
and would indicate that the bulk of sediment delivered to the ocean
by the Mannville Rivers was transferred out of these systems and
is thus no longer preserved in the rocks of western North America.

There were certainly muds delivered to the shelf by Mannville rivers,
but these did not serve as the major source of heavy oil for the
Mannville, although they may source hydrocarbons in the Bluesky and
Fahlr shoreline sandstones farther west. The source rocks that charged
the Mannville heavy oil deposits are thought to be Jurassic to Paleozoic
in age (Creaney et al., 1994). These source rocks matured in the deeper
parts of the foreland basin, producing oil that migrated up-dip,
across the erosional valley floor, and into the Mannville. In terms of
petroleum systems, the oil-sands source rock is disassociated
(i.e., derived from older formations), although locally, the Mannville
Group does contain self-sourced gas, some oil, and extensive coal,
especially in the more deeply-buried western parts of the Alberta basin
(Hayes et al., 1994).

In terms of the heavy oil resource, evaporation and degradation of
the oil at the surface and in the shallow subsurface dispersed the lighter
hydrocarbon compounds leaving heavy oil in the reservoirs (Creaney
et al., 1994). Because the deposits are shallow, they are easily accessible,
but the high viscosity results in difficulty in flowing the oil requiring ex-
tensive reservoir engineering, such as the use of horizontal wells and
steam-assisted gravity drainage (SAGD). The scale of the river systems
and the associated heterogeneity is a major factor in determining the
economic viability of this vast resource.

7.3. Late–Early Cretaceous, end of the big rivers

The top of the Mannville, at around 100 Ma, is marked by a regional
transgression, associated with high global sea level and flexural depres-
sion of the North American Plate, enhancing the developing foreland
basin. The transgression is marked by the Joli Fou Shale in Canada and
its equivalents in the USA (Reinson et al., 1994). This transgressive
shale forms a major hydraulic barrier to oil and gas migration such
that no oil and gas from older source rocks migrates across this shale
boundary (Creaney et al., 1994). Despite this transgression, rivers
systems in the overlying formations also flowed northwest, delivering
deposit into the Boreal Seaway. Incised valleys in the Viking
Formation and Paddy Member of the Peace River Formation in Canada
were associated with a Late Albian lowstand, and are equivalent in age
to the Muddy Sandstone in Colorado and Wyoming. U-Pb dating of
zircons from the Paddy Member (Buechmann, 2013) shows similar
provenance to the Mannville, with Laurentian and Cordilleran source
areas. The Paddy valley was large, tens of kilometers wide, and
hundreds of kilometers long. Lateral accretion bedding and associated
tidally influenced point bars are on the order of <10 m thick. Abandoned
channel mud plugs in the Paddy are about 7 m thick (Leckie and Singh,
1991), indicating rivers were significantly smaller than the Mannville
trunk streams (<10 m deep) and suggesting a less well-integrated,
probably more segmentated drainage network. Outcrops of the Viking
are relatively rare, and a comprehensive analysis of river scales has
not been attempted.

The western Canada foreland basin was formed above a heteroge-
nous lithosphere characterized by a number of subtle arches, such as
the Peace River and Sweetgrass structures, and at times these caused
segmentation of the basin and associated drainages, especially at
lowstand when the basin was characterized by fluvial deposition.
Motion on these structural lineaments may reflect intra-plate stresses
or activation of a peripheral bulge (e.g., Stelck, 1975; O’Connell, 1994;
Wright et al., 1994; Plint and Wadsworth, 2006).

The sandstones of the Viking and Muddy formations constitute im-
portant oil and gas reservoirs in Canada and the USA, including fields
within incised valleys and shallow marine facies (Dolson et al., 1991;
Reinson et al., 1994). The depth of incision of valley fills is typically asso-
ciated with shoreface deposits, exposed during falls of sea-level
(e.g., Wescott, 1993). The exposed wedge of highstand sediment typi-
cally has a steep-fronted shoreface, forming a point-ridge that allows ini-
ciation of about 20 m. Variation in the width of valley-fills, and their
associated reservoirs, is controlled by the duration of sea-level
lowstands (Wescott, 1993). Longer-lived lowstands result in wider
and more complex compound valley fills, as appears to be the case of
the older Mannville and Paddy valleys, which were tens of kilometers

2013). Muddy inclined drapes may form as counter-point bars in up-
stream reaches (Smith et al., 2009) and by tidal processes in more
downstream reaches (Thomas et al., 1987). In the more-distal reaches
of the ancient river systems, within the backwater limit, tides were
more important as reflected in the abundance of bioturbation and
other marine indicators. Given the continental scale of the river systems,
gradients were certainly low, and probably on the order of 10–4 (Musial
et al., 2012). Backwater lengths would correspondingly have been on
the order of 400 km, allowing tidal effects to propagate far upstream.
Regional paleogeographic maps of Smith (1994) show that the heavy
oil deposits in the McMurray were deposited within a few hundred kil-
ometers of the shoreline and were thus likely within the backwater
length of the system. Blum (2015) in contrast, used 3D seismic images
to document the migration patterns of Mannville channels and sug-
gested that the high sinuosity and the high degree of lateral migration
of meanders require the channels to have been landward of the backwa-
ter reach and thus beyond the limit of tidal influence, although this is
incompatible with the observation of marine microfossils in associated
mudstone drapes.
in width. In contrast, smaller valleys in the Viking were typically only a few kilometers in width and may indicate shorter lowstands (Reinson et al., 1994).

### 7.4. Late Cretaceous (Cenomanian), rivers and deltas of the Dunvegan Formation and equivalents

In general, the Cenomanian marks a time of segmentation of drainages in the Seaway and a profound transition from northwest- to southeast-flowing rivers (Fig. 21). Tectonic salients and promontories in the Sevier orogenic front, as well as the tectonic lineaments discussed above, exerted a control on localization of discrete delta systems and their fluvial feeder systems, such as the Dunvegan Formation in Canada and the Frontier Formation in Wyoming (e.g., Lawton et al., 1994). The transcontinental arch was still a positive feature at this time, and the Boreal (northern) Seaway is interpreted to have been closed to the south (Williams and Stelck, 1975). The confined southern arm of the Seaway may have led to anoxic substrate conditions forming hydrocarbon-rich source rocks, including the Fish-Scales unit in Canada and the Mowry Shale in the USA.

The Woodbine Formation in Texas represents a modest-sized delta system on the southern side of the drainage divide that drained the Appalachian and prograded to the southwest into the paleo-Gulf of Mexico Seaway (Fig. 21). Rivers in the Woodbine were small, less than a few meters deep, reflecting a localized drainage area (Oliver, 1970; Ambrose et al., 2009; Adams and Carr, 2010), but rich source rocks ensure that the Woodbine is the host reservoir for the giant East Texas oil fields.

#### 7.4.1. The Dunvegan delta

The Cenomanian Dunvegan Formation, in Canada marks the first appearance of major Cretaceous rivers that flowed southeast (Figs. 21 and 22), representing a 180° paleogeographic reversal from the preceding Viking and Paddy shorelines. Conglomeratic alluvial facies of the Dunvegan are exposed in the Northwest Territories of Canada (Stott, 1982), and record accretion of continental terranes to northwestern North America (Johnston, 2008).

Extensive outcrop and subsurface well data through the Dunvegan, allow mapping of individual valleys and their associated down-dip sandy delta lobes and prodeltas (e.g., Bhattacharya, 1993, 1994) enabling a comprehensive S2S analysis (Figs. 22–25). Extensive subsurface mapping shows that the Dunvegan valley systems form an extensive tributary network that coalesces into trunk valleys (Plint, 2000, 2002; Plint and Wadsworth, 2003, 2006) (Fig. 22). Mapped individual valley segments, including the trunk valley, are typically less than a few kilometers in width, and the depth of incision was generally less than about 35 m. Age-equivalent deltaic sandstones and prodelta mudstones can be mapped downlapping onto a distal peripheral bulge (Plint et al., 2009; Fig. 25), in a mostly closed depositional sink that allows estimation of bulk sediment volumes.

Allostratigraphic analysis shows that the entire Dunvegan Formation was deposited in about 2 million years. The ten mapped allomembers represent deposition in no more than 200,000 years, and probably as few as 100,000 years. Individual parasequences likely represent deposition in much less than 50,000 years, assuming depositional rates were on the order of 1–10 cm/ky. These chronometric constraints allow long-term sediment budgets to be estimated.

Plate tectonic considerations (Plint and Wadsworth, 2006) and recent detrital zircon dating (Buechmann, 2013) show the bedrock source area was about 150,000 km² (Fig. 22). Outcrops and cores samples of valley fills were used to document key paleohydraulic parameters (Figs. 24 and 26). Outcrops of tributary channels show well-developed lateral accretion, indicating single-thread meandering rivers, ~16 m deep and ~150 m wide (Fig. 24). Trunk streams must have been correspondingly larger, and perhaps closer to 20 m deep, although there are no outcrop exposures of the trunk-valley rivers, so width of trunk rivers is estimated from empirical equations.

Trunk valley fills are filled primarily with decimeter-thick sets of dune-scale cross-bedded fine- to medium-grained (125–350 μm) sandstone (Fig. 26). Using the bedform phase diagram, as described in Section 6.4, flow velocities on the order of <1.5 m/s have been estimated (Bhattacharya and MacEachern, 2009; Fig. 15).

Plint et al. (2009) used correlations and modern analogs to estimate regional paleo-slopes for the Dunvegan rivers of 3–4 × 10⁻⁴, suggesting a moderate- to low-gradient system. Recalculations of slopes, based on grain-size analysis and water depth estimates (using the Manning and Brownlie equations as described in Section 6.4) indicate a significantly lower slope of 3–5 × 10⁻⁵. Using the higher slope and assuming channel depths of up to 20 m, backwater lengths of about 50–65 km are estimated. Using the lower slope estimate yields a backwater length of 400–600 km.

Regional considerations, including plate-tectonic position from palaeomagnetics as well as analysis of floodplain paleosols, indicate deposition at about 70° latitude, within a temperate-humid climate regime (Irving et al., 1993; McCarthy and Plint, 1999; McCarthy, 2002). The elongate, axial drainage area was about 1.25 × 10⁶ km² (Fig. 22A).

#### 7.4.2. Dunvegan paleodischarge estimates

Based on flow velocities determined from the sedimentological observations (grain size, channel thickness, channel width, and sedimentary structures), and using the bedform phase diagram approach (Fig. 15), bankfull paleodischarge of Dunvegan rivers was estimated to be 4–6 × 10⁴ m³/s (see Table 1). Paleodischarge estimates of 3.5 × 10⁴ m³/s–1 × 10⁵ m³/s, were based on regional hydraulic geometry curves (Davidson and North, 2009) assuming a drainage basin of...
about 1.25 × 10^5 km^2, and are consistent with sedimentological observations. A discharge estimate of 6 × 10^3 m^3/s was calculated using the Manning Equation (with n = 0.035). The corresponding velocity estimate of 2.4 m/s, was based on a 20 m deep, 200 m wide meandering channel, with a calculated hydraulic radius of about 11 m, flowing over a slope of 3 × 10^{-4}. This velocity is well within the range predicted using other techniques, such as the bed form phase diagram. Using the lower slope (3–5 × 10^{-5}) yields a velocity of 0.8 m/s and a Q_w of 1600–2300 m^3/s, which is a factor of 3 lower. In general, the discharge estimated using sedimentologic criteria, such as channel depth, grain size, and slope, as outlined in Section 6 range from 1.6–6 × 10^3 m^3/s, or over a factor of 3.75. The largest deviation appears to be associated with use of the regional hydraulic geometry curves (factor of 6.25 difference compared to lowest estimate using sedimentological criteria), and we suggest that this technique has the largest uncertainty. Bhattacharya and MacEachern (2009) used paleodischarge estimates of Dunvegan trunk rivers, and compared these to empirical relationships based on modern rivers to predict their propensity for generating hyperpycnal flows (Mulder and Syvitski, 1995). Mulder and Syvitski (1995) showed that small rivers, with an average discharge

Fig. 22. Dunvegan paleogeography, A. tectonic regime and drainage basin (125,000 km^2), B. Tributive valley systems feed trunk rivers and major delta lobes, based on over 3000 well logs, cores and outcrops in fold and thrust belt (after Plint and Wadsworth, 2003, 2006). Cross section A–A’ shown in Fig. 23.

Fig. 23. Depositional dip cross section through the Dunvegan Formation, Alberta, Canada. Sequence boundary in Allomember E dives under attached lowstand delta (E1). From Bhattacharya and Walker (1991).
of <6000 m$^3$/s, routinely produce hyperpycnal flows during subsequent higher-discharge floods because the suspended sediment concentrations typically exceed the capacity of the flow. In contrast, they show that rivers with average discharge, >1 × 10$^4$ m$^3$/s, never produce hyperpycnal flows, even during exceptional floods, because the suspended sediment concentration is unable to reach the required threshold. The estimates of paleo-discharge for the Dunvegan, likely reflect flood conditions, rather than average flow conditions (see discussion in Bhattacharya and MacEachern, 2009), but in either case do not exceed the 6000 m$^3$/s threshold of Mulder and Syvitski (1995). Whether or not a given paleo-river actually generated hyperpycnal flows can be evaluated by examining associated delta front and prodelta facies, which may record sediment delivery mechanisms, especially in settings where sedimentation rates are high and levels of bioturbation low (e.g., Bhattacharya and MacEachern, 2009; Li et al., 2015; Keuhl et al., 2015; Romans et al., 2015). In the case of the Dunvegan, examination of prodelta facies in core shows low levels of bioturbation, rare wave ripples and some hummocky cross stratification (HCS), and abundant inverse and normal graded beds with current ripples and Bouma sequences. The inverse graded bed and Bouma sequences are diagnostic of hyperpycnal flows and turbidity currents respectively (Mulder et al., 2003). The low bioturbation indicates high sedimentation rates.

Fig. 24. Large-scale lateral accretion sets in a point bar within a Dunvegan valley fill. Photo of cliff section (above) and interpreted bedding diagram (below) lowest panel shows reconstructed channel (modified after Plint and Wadsworth, 2003).

Fig. 25. A. Isolith map of Parasequence E1 in Allomember E, Dunvegan Formation. B. Isopach map of Allomember E. From Bhattacharya (1993).
and suggests a river-dominated prodelta. The wave ripples and HCS indicates periodic storms that aided in along-shelf and across-shelf shelf transport (Plint et al., 2009).

### 7.4.3. Dunvegan sediment budgets

Isolith mapping of sandy delta deposits within various allomembers (Fig. 25) shows that individual lobes (i.e., parasequences) cover an area ranging from 2 to 5 × 10³ km², with sandstone thickness up to 15 m (Bhattacharya, 1993). Individual delta lobes thus contain sand volumes on the order of 10 km³. Isopach maps of Dunvegan allomembers, which include sandstone and shales, indicate sediment bodies that are about 200 km in length, from their updip to down-dip pinchout limit, and extend along depositional strike for at least 300 km, representing an area of at least 6 × 10⁴ km² (Fig. 25). Allomember E is about 50 m thick, and the total mapped sedimentary volume (including porosity) is estimated to be about 3000 km³. The landward to seaward extent is well constrained by the onlap and downlap limits of Allomembers, thus defining a closed system, although the southwestern limit has not been constrained and thus it is possible that along-shelf export of sediment may have occurred.

Time stratigraphic analysis suggests that Dunvegan valleys may have persisted for 10,000–50,000 years. Fulcrum analysis (i.e., Holbrook and Wanas, 2014) of the Dunvegan valley fills integrating instantaneous discharge estimates over these longer time periods, suggests that total sediment delivery was between 200 and 800 km³ of sediment. This would correspond to a single parasequence within an allomember. The rock rock volume of Parasequence 1 in Allomember E, which represents the lowstand delta deposits fed by the valley, is about 750 km³ (i.e., about 1/4 of the total sediment volume in the Allomember), and matches remarkably well with the volume of sediment estimated to be delivered out of the trunk valley.

### 7.4.4. Dunvegan petroleum perspective

From the petroleum perspective, the Dunvegan Formation contains numerous small conventional and mature oil and gas fields, primarily associated with incised valleys and lowstand delta lobes (Bhattacharya, 1994). Because the Dunvegan progrades southeast but the basin is structurally tilted to the southwest, traps typically require a local lateral stratigraphic pinchout to the northeast. Such trapping configurations occur along the northern margins of incised meandering valleys or northern margins of delta lobes, resulting in localized traps and small fields. More recent exploration in the Dunvegan Formation is focused on unconventional resources in the down-dip heterolithic prodelta facies. The nature and productivity of these thin-bedded reservoirs is controlled by the geomechanical properties, which in turn were created by the depositional processes. For example, knowing the proportion of preserved hyperpycnal deposits, and turbidites, versus reworked storm beds or more slowly deposited and typically more bioturbated suspension deposit may control the sorting, thickness, grain size and textural properties of beds, which are in turn important for predicting the productivity of horizontal wells and associated fracture jobs in these thin-bedded reservoirs. The approach presented above may help to predict the propensity of a particular deltaic delivery processes (e.g., hyperpycnal versus hypopycnal flows) and, when integrated with knowledge of shelf processes (e.g., storms, oceanic currents) and
B. Turonian

Fig. 27. North American paleogeography during the Turonian. A series of separate delta systems drained portions of the Western Cordillera. Drainage Areas are 50,000 km², K-Kaiparowits, NM – New Mexico. Based on Gardner, 1995 and Bhattacharya and Tye, 2004. Green outlines of paleo-land-masses largely based on reconstructions of R. Blakely and Colorado Plateau Geosystems (2015).

sedimentation rates may help determine the nature of prodelta facies and along-strike variability (Li et al., 2015).

7.5. Turonian: a time of many deltas

The Turonian age was marked by an overall expansion of the Cretaceous Seaway, which finally connected the newly formed Gulf of Mexico and Atlantic Ocean with the Boreal Seaway to the north. A series of separate S2S deltaic-shoreline systems persisted along the western margin of the Seaway (Fig. 27).

One of these includes the Cardium Formation in Alberta, which comprises a series of parasequences deposited in broadly elongate wave-dominated shorefaces fed by numerous east-flowing rivers (e.g. Krause et al., 1994 and references therein; Plint et al., 1988). Because of the paleogeographic reorientation, the wave-dominated shoreline sands of the Cardium Formation pinch out to the northeast, in the opposite direction to the tectonic tilt of the foreland basin. This produces a near-perfect structural/stratigraphic trap, resulting in giant oil and gas accumulations. This is in distinct contrast to the Dunvegan, where the paleogeographic dip parallels tectonic strike resulting in much smaller accumulations, despite probably being fed by a larger river.

The Cardium parasequences were likely fed by numerous small rivers, which flowed transverse to the foreland axis. Localized conglomerates cap several of the parasequences, and have been interpreted to mark entry points of gravelly-bedload rivers (Walker and Eyles, 1988; Amrott, 1991, 1992; Hart and Plint, 2003). The abundance of conglomerate suggests that rivers were steep-gradient (e.g., Slope ~10⁻³), and the thickness of fluvial conglomerates suggest channels were ~6 m deep (Hart and Plint, 2003). Wave reworking formed shorefaces that were amalgamated along strike, resulting in a relatively homogenous reservoir body, locally capped by the conglomerates (Bergman and Walker, 1987). The conglomerates also mark regional transgressive marine erosion surfaces, which supports the interpretation that rivers were shallow, as deeper channels would have been more likely to survive the transgression (Bhattacharya, 2011).

In addition to the sandstones, the marine conglomerates in the Cardium form prolific hydrocarbon reservoirs (Krause et al., 1987, 1994). Exploration efforts in the Cardium are now focused on thin-bedded facies on the seaward margin of the shoreface sandstone body. Along-strike variability in these facies likely depend on the position and lateral spacing of fluvial entry points, which should be marked by lower bioturbation intensities and an increase in beds formed by river plumes.

The Frontier Formation in Wyoming (Gani and Bhattacharya, 2007; Lee et al., 2007; Vakarelov and Bhattacharya, 2009) similarly formed a series of extensive wave-dominated and locally conglomeratic shorefaces with local deltaic input (Fig. 27). Like the Cardium, steep-gradient, small rivers are indicated by the abundant transgressively reworked conglomerates that cap sandstone bodies. In addition to the sandstones, the conglomerates create locally prolific reservoirs.

The Ferron Sandstone Member in Utah (detailed below) and the Gallup Sandstone Formation in New Mexico are more deltaic in nature. River deposits are preserved and both systems show an axial component, despite their catchments being contained entirely within the Cordillera (Fig. 27). Sediment sinks derived from eastern North America were confined to the eastern margin of the Seaway, as indicated by the appearance of Appalachian-derived zircons in the Dakota sandstone in Iowa and Nebraska at the eastern Seaway margin as early as late Albian time (Finzel, 2014).

The Cardium, Frontier, and Gallup all contain important oil, gas, and, in some cases, coal deposits, and have all received extensive study (e.g., Plint et al., 1986; Krause et al., 1994; papers in Van Wagoner and Bertram, 1995). Source rocks for these formations are derived from underlying and overlying condensed section shales. These formed at maximum highstands, and probably were enhanced in organic matter by restricted circulation in the Seaway and development of anoxic substrates (e.g., Schlanger and Jenkyns, 1976; Jenkyns, 1980; Creaney et al., 1994).

7.6. The Ferron sandstone, a case study

One of the most extensively studied Turonian units from a S2S perspective is the Ferron Sandstone Member. The Ferron Sandstone, a member of the Mancos Shale Formation, comprises a series of three broadly deltaic clastic wedges fed by rivers draining the emergent paleo-Rocky Mountains during the mid- to late-Turonian (Garrison and van den Bergh, 2004). The Notom, Last Chance, and Vernal delta complexes have been recognized (Cotter, 1975), and the Last Chance and Notom have been extensively studied (Fig. 27; see papers in Chidsey et al., 2004; as well as papers by Fielding, 2010, 2011, 2015; Li et al., 2010; Zhu et al., 2012; Li and Bhattacharya, 2013; Ullah et al., 2015). Paleosol and palynological analyses indicate that the climate was subtropical and probably ever wet (Akyuz et al., 2015). Bhattacharya and Tye (2004) estimated the area of Ferron drainage basins as approximately 5.0 × 10⁴ km² (Fig. 27).

Detailed sequence-stratigraphic studies suggest incised valleys occur with the Notom and Last Chance systems (see Li et al., 2010 and Garrison and van den Bergh, 2004; Fig. 15), and detailed facies-architecture studies have documented the style and scale of channels and channel belts (e.g., Corbeau et al., 2004; Li et al., 2010; Bhattacharyya et al., 2015; Ullah et al., 2015; Wu et al., 2015), allowing trunk rivers to be pinpointed (Figs. 10 and 14).

The depth of the channels in the Last Chance delta complex were about 11 m at their deepest (Garrison and van den Bergh, 2006), and these lie within valleys exhibiting up to 32 m of erosional relief. Channel widths, measured from outcrop data, range from 100 to 174 m. Channel-belt sandstones are dominated by medium-grained, cross-bedded sandstones, which are locally conglomeratic. Extraformational
conglomerate phases are typically pea-grade gravel, and indicate that Ferron rivers drained the Sevier hinterland. Valleys within the Notom wedge also show relief of up to about 30 m, but they are up to several tens of kilometers wide and compound in nature suggesting a long-lived and complex history associated with prolonged falls and lowstands of sea-level (Li et al., 2010; Li and Bhattacharya, 2013). Slope estimates in the Notom deltaic complex associated with these valleys are on the order of $10^{-3}$, and were made based on onlap-distance and height of aggrading coastal prisms, as well as the elevation drop of valley floors along a depositional dip-oriented regional cross section (Fig. 15B). These slopes suggest a relatively steep fluvial gradient.

Plan-view exposures of channels and channel belts (Fig. 11C) indicate low-sinuosity ($b < 1.44$), single-thread, meandering sand and gravel bed rivers, containing bars built mostly with medium-grained cross-bedded sandstones (Fig. 12). Paleo-thalwegs of channels within valleys in the Notom system appear to be about 5–7 m deep (Li et al., 2010; Ullah et al., 2015.), with average channels 2–3 m deep. Plotting these on the bedform phase diagram indicates flow velocities of between 0.75–1.3 m/s (Li et al., 2010) (Fig. 16). Calculations using the Manning Equation yield velocities of up to 2.3 m/s. Assuming widths of 100–174 m, Li et al. (2010) estimated bankfull paleodischarge of the trunk rivers to be $1.3–3.0 \times 10^3$ m$^3$/s. Davidson and North (2009) used hydraulic geometry curves to estimate discharges ranging from 66 to 2833 m$^3$/s. Bhattacharya and MacEachern (2009) employed these estimates to predict that even the largest Ferron rivers were routinely capable of generating hyperpycnal flows during high magnitude flood events, suggesting high rates of delivery of sediment to the shelf. This hypothesis is also consistent with the observation of inverse-graded hyperpycnites as a common component of prodeltaic facies within the underlying Tununk Shale, fed by Ferron rivers (Bhattacharya and MacEachern, 2009; Li et al., 2015).

We used the fulcrum technique of Holbrook and Wanas (2014) to estimate both short- and long-term sediment discharge rates and consequent sediment volumes delivered by Ferron paleorivers (Table 1).
Slopes of $0.7 \sim 0.8 \times 10^{-3}$, calculated based on grain-size analysis, are similar to slope estimates using stratigraphic considerations above. The larger rivers in the youngest sequence of the Ferron Notom wedge, show a paleodischarge of $1.3 \times 10^4$ m$^3$/s bedload discharge of 0.7 m$^3$/s and a suspended-load discharge of $10.5 \times 10^3$ m$^3$/s. We assume that Ferron rivers flowed at bankfull rates for about 10–14 days per year, comparable to modern sub-tropical rivers. This yields an annual mean bedload estimate of $9 \times 10^3$ m$^3$/y. Owing to the coarse grade of the paleo-river bedload, we used the equations of Wright and Parker (2004) rather than Van Rijn (1984) to calculate suspended load. Assuming deposition occurred across one 20,000-year Milankovitch cycle, results in a total bedload sediment volume of about 20 km$^3$ and a total sediment volume of about 300 km$^3$. Correlation and mapping of sandstones delta lobes within Ferron parasequences show that they were approximately 500–1000 km$^2$ in area and about 10–20 m thick (Figs. 15 and 28), representing a sandstone volume of 5–20 km$^3$, which matches remarkably well with the bedload estimate above. Unfortunately, the limits of prodelta and shelf mudstone facies within individual Ferron parasequences have not been mapped, but if these are assumed to be roughly an order of magnitude more than the bedload, total sediment volumes would be 50–200 km$^3$, consistent with the estimates made above using the fulcrum method (Holbrook and Wanam, 2014).

The Ferron has long been regarded as an analog for paralic petroleum reservoirs worldwide (Bhattacharya and Tye, 2004; Deveugle et al., 2014), but there has been debate as to the scale of Ferron rivers and suitable modern analogs. The Ferron is also important in containing extensive coals deposits and some gas (see papers in Chidsey et al., 2015). The coal deposits are largely associated with times of positive accommodation, and likely formed in wetland environments behind linked shoreline deposits that that show an aggradational to retrogradational component to their shoreline trajectory.

The Ferron and its age-equivalents along the Cretaceous Interior Sea- way are overlain by deposits of a major transgression (Niobrara transgression of Kauffman, 1977). The transgressive strata include the Blue Gate Member of the Mancos Shale Formation in Utah, the Niobrara Shale in Wyoming and Montana, and the First White Speckled Shale in Western Canada, and correlative strata. Within these shales, a number of stacked thin shale units are documented along the western margin of the Seaway (e.g., the Emery Sandstone in Utah and the Medicine Hat Sandstone in Canada), but no major fluvial deposits of this age have been described. However, the cyclic alternation of progradational clastic, fluvial-deltaic wedges and transgressive organic-rich shales form numerous reservoir, source and seal triplets that that make up the extensive Cretaceous petroleum systems along the margins of the Cretaceous Interior Seaway.

Presently, greater attention is being paid to these shalier formations as they are the target for unconventional oil and gas reserves. Key to production is the ability to fracture the shaly formations. The richest source rocks commonly are associated with marine highstand condensed sections, but these contain abundant bentonites and high organic content, which results in elastic rocks that do not easily fracture. Downlapping river-fed prodeltaic or along-strike mud-belt facies, which prograde over these condensed sections, while leaner in organics, may contain higher proportions of silt and quartz, improving their geomechanical properties for induced fracturing. Superposition of downlapping river-derived prodeltaic deposits (which also have better primary porosity) with pelagic and hemipelagic source intervals may yield more prospective unconventional reservoir-source pairs.

7.7. Santonian to Paleocene, the mega-bajada

The next major phase of recorded fluvial deposition (and last for discussion) began in the Santonian and continued throughout the remainder of the Cretaceous Period with two major cycles, an earlier Santonian to Campanian cycle (~84 Ma) and a younger Maastrichtian to Paleocene phase (~70–60 Ma) interrupted by transgression of the Bearpaw-Claggett shale (Kauffman, 1977).

Thick Campanian fluvial successions include the uppermost Straight Cliffs and Wahweap Formations in southern Utah (Eaton, 1991; Lawton et al., 2014a; Lawton et al., 2014b), the Mesaverde Group, including the Blackhawk-Castlegate Formations in central Utah (Hampson et al., 2013; Hajek and Heller, 2012), the Ericson Formation in Wyoming (Martsinsen et al., 1999), the Williams Fork Formation in Colorado (Pranter et al., 2009), and the Belly River/Judith River in Canada. Maastrichtian fluvial successions include the Lance Formation in Wyoming (Carvaljal and Steel, 2009) and the Horseshoe Canyon Formation in Canada (Dawson et al., 1994). The Campanian units formed a relatively contiguous band forming a giant bajada-like alluvial fringe, a few hundred kilometers wide, which likely extended for several thousand kilometers along most of the western margin of the Seaway (Fig. 8A). Paleo-river systems in southern Utah comprise transverse distributive fluvial fans, eroded from the thrust-belt, which connected with north flowing axial rivers that drained the rejuvenated Mogollon highlands in Arizona (Lawton et al., 2003). Detrital zircon studies indicate that these north-flowing axial systems drained basement rocks that lay south of the Cordilleran foreland basin (Lawton et al., 2014a, 2014b; Szwarc et al., 2015). The sand-rich distributive fluvial fans of the Straight Cliffs and Wahweap Formations comprised upstream fluvial systems that were sourced from weakly-lithified Mesozoic eolian units in the frontal parts of the Sevier orogenic belt, with erosion enhanced by a monsoonal climate (Lawton et al., 2014a, b). The extensive uplift was likely caused by the overall transition into thick-skinned tectonics of the Laramide orogeny, forming an extensive western plateau which culminated in the Eocene-age “Nevadaplano” of DeCelles (2004). The transverse component became increasing important as deformation progressed (Lawton et al., 2014b). Eventually, Laramide deformation caused the Cretaceous Seaway to retreat, and the foreland basin was segmented into a series of largely non-marine Cenozoic Laramide sub-basins.

Campanian rivers in Central Utah, Wyoming and Canada flowed broadly east to southeast, reflecting dominantly transverse drainage with little axial transport. Hampson et al. (2012, 2013) and Flood and Hampson (2015) compiled data on the facies architecture and paleodischarge of river deposits in the Blackhawk Formation in Utah along an approximately 100 km stretch of near-continuous exposures and also attempted a mass-balance S2S analysis (Hampson et al., 2014). Single-thread meandering rivers exhibit a maximum depth of 9.8 m. Multi-thread, braided rivers show depths ranging from 2.4 to 14.2 m. Paleoslope estimates of $0.7 \times 10^{-3}$, suggesting steep gradients. Estimates of peak water discharge range from 0.42–2.2 $10^3$ m$^3$/s for meandering streams and 9.0–11 $10^3$ m$^3$/s for the braided channels. The largest values are similar to the modern Mississippi, but almost certainly drained a smaller area. Numerous rivers, spaced several tens up to about 70 km apart, fed sediment to the shoreline and were reworked into an extensive, and relatively smooth, wave-dominated coastline, similar to the Cardium Formation discussed in Section 7.5. The Blackhawk alluvial plain thus consisted of a series of separate rivers and floodplains, probably drained from separate segments of the Western Cordillera, but coalescing to form a single amalgamated coastal plain. Collectively, these rivers would have supplied massive amounts of sediment to the Cretaceous Seaway. The Belly River Wedge in the Alberta basin covers about 500,000 km$^2$ and is about 300 m thick representing $1.5 \times 10^2$ km$^2$ of preserved sediment, but this was deposited over about 5 Ma. Mass-balance calculations of Hampson et al. (2014) suggest significant along-strike loss of sediment.

Studies of the Campanian Ericson Formation in Wyoming show similar scale channels, averaging several meters deep, with the deepest channels up to about 15 m (Martsinsen et al., 1999). Pranter et al. (2009) compiled sand-body dimensional data for the Williams Fork Formation and showed the thickest single-story sand bodies to be up to 9 m deep, suggesting similarity to the Blackhawk Formation. To
date there have been no attempts to estimate paleodischarge or paleodrainage for the Ericson or William's Fork systems.

By the Maastrichtian, a new drainage divide associated with the Laramide orogeny emerged that closely paralleled the Canada-USA border (Fig. 8B). Rivers of the Lance Formation flowed largely south, draining off this divide (Carvajal and Steel, 2009). A distinct fluvial, shelf-slope and linked deep-water system developed for the first time in the Washakie and Great Divide basins of southern Wyoming during the Maastrichtian and the basin reached up to 500 m deep (Fig. 29). Carvajal and Steel (2009) undertook detailed sequence stratigraphic analysis of the Lance, Fox Hills and Lewis Shale, which represent the fluvial, shoreline, and deep-water components respectively. Fourteen separate clinothem units were correlated and mapped (Fig. 29). The total clinoform set was deposited in 1.8 Ma, suggesting each clinothem unit represents about 10,000 years. These clinothem units typically contain several parasequences deposited in a few thousand years, suggesting rates typical of modern systems (Walsh and Nittrouer, 2009). The clinothems were also grouped into two distinct stages (Carvajal and Steel, 2009, 2012). Stage 1 is dominated by aggradational stacking and correlated to a time of higher subsidence. Stage 2 is dominantly progradational and correlates to a time of increased tectonic uplift which is inferred to have generated higher sediment supply.

Fluvial trunk channels in the Lance Formation were about 7 m deep. Associated sandy deep-water fans in the Lewis Shale were quite small, covering an area of 500–750 km². Fan deposits reach thicknesses of up to 100 m, with up to 60 m of sandstone (Fig. 29). Carvajal and Steel (2012) estimated sandstone volumes of up to 66 km³. Total clinothem volumes of 630 km³ were calculated by Carvajal and Steel (2012), but represent minimal estimates, as the proximal portions of clinothems were not included. Drainage areas were estimated to range from 5 to 20 × 10⁶ km² during stage 1, and expanded to 20–35 × 10⁶ km², during stage 2. These drainages are comparable to the Ferron systems discussed in Section 7.6, and river depths are correspondingly similar (e.g., 5–7 m deep). They also calculated sediment yields ranging from 200 to 2000 t/km² y and sediment loads ranging from 4 to 16 Mt./y (Mt = megatons). Carvajal and Steel (2012) also used the BQART model (Syvitski and Milliman, 2007) to estimate that the relief of the source area changed from lowland (100–500 m) to upland (500–1000 m) in stage 1, to mountainous (1000–3000 m) in stage 2. The calculated sediment loads indicate modest rivers, similar in scale to those in Southeast Asian islands (e.g., New Guinea, Sumatra, Borneo), and consistent with small drainage basins. The higher sediment yields are also comparable with analogous small mountainous rivers and could also have generated frequent hyperpycnal flows (Milliman and Syvitski, 1992). Soyinka and Slatt (2008) identified abundant hyperpycnites and turbidites in the Lewis Shale in Wyoming, which were fed by these rivers. Petter et al. (2013) present a simpler alternate method for estimating sediment paleoflux using stratigraphic criteria. Clinoform dimensions and stacking patterns were used to measure the degree of aggradation and progradation, which was equated to changes in sediment discharge rates. Their estimates of sediment discharge for the Lance-Fox Hills-Lewis system were 8.8Mt/y, well within the 4-16Mt/y range estimated by Carvajal and Steel (2012).

The fluvial Lance sandstones form significant tight gas reservoirs in the Jonah and Pinedale Field in Wyoming (e.g., Longman et al., 2015)
and references therein), and the degree of sandstone amalgamation as well as scale of channels are important in controlling recovery. The submarine fan sandstones in the Lewis host gas reservoirs in the Green River Basin, Wyoming (Muller and Winkler, 2004) and the southern extension of the Lewis shale forms important gas reservoirs in Colorado and New Mexico (Dubiel, 2007). The quantitative S2S approach suggests that the moderate-paleodischarge Lance paleorivers directly fed the interpreted hyperpycnites and turbidites of the deepwater Lewis Shale (Soyinka and Slatt, 2008) and illustrates the value of this approach in understanding the nature of linked fluvial to deepwater depositional systems, with implications for hydrocarbon discovery and recovery.

8. Conclusion and future work

This paper shows how quantitative source-to-sink analysis can be used to make estimates regarding the paleodischarge of ancient river systems and in turn predictions about the volume and characteristics of deposits in downstream sinks, as well as the size, relief, slope, and climate regime of upstream areas. This approach requires analysis of the scale (e.g., depth), and/or sedimentology (e.g., grain size, sedimentary structures) of rivers to determine their velocity and discharge. For example, documentation of the relatively small nature of trunk rivers in the Ferron, Dunvegan and Lance, indicate they could have routinely produced hyperpycnal flows during large magnitude flood events (Mulder and Svytitski, 1995). Examination of associated downdip prodelta and deep-water facies show locally abundant hyperpycnites and turbidites, likely fed by these rivers (Bhattacharya and MacEachern, 2009; Soyinka and Slatt, 2008). These heterolithic facies are now being exploited as thin-beded reservoirs that fringe the tapped-out conventional sandstone play in the Dunvegan Formation, and form deep-water gas reservoirs in the Lewis Shale.

In steep gradient S2S systems (Slope \( \approx 10^{-3} \)), such as the Frontier and Cardium formations, backwater lengths were short (<10 km), which allows gravel to reach the coast, and these shallow marine conglomerates form important oil and gas reservoirs. In lower gradient systems (S = 10^{-4} to 10^{-5} ), such as the McMurray and Dunvegan backwater lengths are higher (>100 km), and gravel phases are confined to the inland alluvial system, and coastal systems are sandy, precluding the development of coarser-grained reservoirs. The Lower Cretaceous McMurray Formation, was a continental-scale, low-gradient river system enabling development of an extensive tidal backwater zone. Abundant tidal features and lack of gravel-grade sediment support this prediction. Because channel deposits are large, internal heterogeneity, and especially tidally-influenced point bar mud drapes, can be mapped and modeled explicitly, which is important in managing both steam floods in subsurface and mining operations at the surface.

Dimensions and paleo-discharge estimates of Mesozoic paleorivers was integrated with provenance analysis and regional paleotectonic maps to document the evolution of drainage areas in North America throughout the Mesozoic. These show three major phases. The first phase as associated with continental-scale rivers draining southern and easterly (Appalachian source areas) in the Triassic through Albian. The second phase is marked by a transition to smaller drainage basins in the Cenomanian-Campanian, primarily draining the Western Cordillera, and reflecting the development of the Sevier Orogeny, which formed an extensive foreland basin along the thrust front that was flooded by the Cretaceous Interior Seaway. The third phase shows a return to a continental-scale drainage in the Paleocene as the Seaway retreated and the foreland basin rebounded.

During the Cenomanian to Maastrichtian, river and delta systems, such as the Dunvegan, Ferron and Lance, were relatively small, allowing closure of S2S sediment budgets. Comparison of predicted sediment transfer to mapped sediment volumes, such as in the Dunvegan, and Lance-Fox Hills-Lewis systems, show remarkable concordance, despite the uncertainties and errors that may be associated with the S2S approach.

Although it was not the focus of this paper, future regional stratigraphic S2S analysis must be integrated with geodynamic analysis to better link tectonic processes with the sedimentary record. This is critical to determine the relative role of tectonic versus climatic processes in forming the stratigraphic record. To that end, integration of detailed analysis of floodplains and their paleosols may help to better determine paleoclimate variability. A remaining challenge is high-resolution chronometry. The sequence stratigraphic approach is critical in that it enables subdivisions of formations into smallest-scale, mappable genetic units, identified by integrating environmental lithofacies and bounding surfaces, (i.e., parasequences, clinoflumes) from which estimates of sediment volumes associated with a specific paleo-river can be made. These parasequences, however, are typically finer-scale than the precision of isotope chronometers, and assigning absolute durations to these units involves significant uncertainty. Recent higher-precision methods of Singer et al. (2015) provide immense promise in determination of high-resolution astrochronological (Milankovitch) cycles. These improved chronostratigraphic tools could offer much insights, but despite this, in a number of examples discussed in this paper, there is remarkable conformity between predictions of sediment transfer and mapped sedimentary volumes, indicating that the S2S approach is robust and can be applied to deep-time systems. Morever, such efforts may provide new knowledge that can be economically beneficial to oil and gas exploration and production.

Acknowledgements

This paper was written with the generous support of the McMaster University/University of Houston Quantitative Sedimentology Laboratories, including current funding from BP, and Inpex and a Natural Sciences and Engineering Research Council of Canada Discovery Grant to Bhattacharya. We also acknowledge former support from Anadarko, BHP Billiton, Chevron, Ecopetrol, ExxonMobil, Pioneer Natural Resources, and Shell. We thank Steve Hubbard and Christian Carvajal for their thorough reviews, and Special Edition editor J.P. Walsh for his insightful comments and meticulous editing.

References


formation: applications for development of Prudhoe Bay Field, Alaska. AAPG Bull. 83, 1588–1623.


