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Supplemental Material

Supplemental Text S1. Estimation of channel geometries, paleoflow conditions, uncertainty analysis, and HadCM3L General Circulation Model.

Table S1. Field localities visited in this study.

Table S2. Measured paleoflow depth indicators in terrestrial fluvial sandstone bodies of the Last Chance Ferron Sandstone.

Table S3. Field data collected in this study. The table is located in the attached Excel spreadsheet in the sheet named “Data”.

Supplement to: *Constraining flow and sediment transport intermittency in the geological past*

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S1. Estimation of channel geometries

i. Bedform heights

We used field measurements of dune-scale cross-set thicknesses, b_{xs} , to estimate mean original bedform (i.e., dune) heights, b_d . In doing so, we assumed the bedform preservation ratio, defined as the ratio of b_{xs} to b_d , i.e., b_{xs}/b_d , is a value between 0.3 and 0.7. Typically, b_{xs}/b_d is assumed to be a constant of ~ 0.3 in steady-state conditions, but can be up to 0.7 or higher in disequilibrium conditions (Paola & Borgman, 1991; Leclair & Bridge, 2001; Leclair, 2002; Jerolmack & Mohrig, 2005; Reesink et al., 2015; Ganti et al., 2020; Leary & Ganti, 2020). Lyster et al. (2022) presented evidence for enhanced bedform preservation under disequilibrium conditions in channel deposits of the Last Chance Ferron Sandstone, and we therefore assumed that b_{xs}/b_d can vary between 0.3 and 0.7 in our analyses.

ii. Flow depths

To estimate flow depth, H , we used field measurements of select channel architectural elements as proxies for H . At field localities, which are reported in Table S1, we measured bar-scale clinoform heights and maximum thicknesses of single channel storeys (Tables S2 and S3). We also gathered data from literature (Tables S2 and S3). Where bar-scale clinoforms are fully preserved, they provide a minimum estimate for the maximum flow depth (e.g., Hajek & Heller, 2012). However, we note that measured bar-scale clinoforms were not necessarily fully-preserved, and may have been partially-preserved (c.f. Chamberlin and Hajek (2019)), and that our measurements do not account for compaction, which may add c. 10% to the estimated height (e.g., Allen, 1965). Importantly, we highlight that estimates of H derived from the

heights of channel architectural elements have previously been corroborated using independent bedform-scale approaches to estimate H (c.f. Lyster et al., 2022).

iii. Channel width

In the Last Chance Ferron Sandstone, the most paleo-landward localities preserve terrestrial fluvial deposits of major meandering trunk channels that fed the Last Chance fluvial-deltaic complex. For these deposits, Bhattacharya and Tye (2004) reported channel widths of 250 m. Further, Garrison Jr and van den Bergh (2004) noted that channels had average widths of 250 m, and that measured channel-belt widths did not exceed 2 km. These authors collectively found no evidence to suggest that channel widths in Last Chance Ferron trunk channels exceeded a few hundred metres. To allow for uncertainty, we prescribed a range of channel widths spanning 200–300 m.

iv. Paleoslope

To reconstruct paleoslope we used the empirical approach of Trampus et al. (2014) which, for sand-grade deposits, has been demonstrated to recover paleoslope values that are similar to paleoslope values recovered using a Shields stress inversion (Ganti et al., 2019; Lyster et al., 2021). Following Trampus et al. (2014), we estimated paleoslope, S , as

$$\log S = \alpha_0 + \alpha_1 \log D + \alpha_2 \log H, \quad \text{Eq. S1}$$

where $\alpha_0 = -2.08 \pm 0.036$, $\alpha_1 = 0.254 \pm 0.016$, and $\alpha_2 = -1.09 \pm 0.044$ are constants, and D is grain size (as described in the main text).

S2. Estimation of paleoflow conditions

i. Flow characteristics

To calculate the instantaneous channel-forming water discharge, $Q_{w(cf)}$, and bed material load, $Q_{bm(cf)}$, we first calculated flow velocity, U , using the Chézy formulae for hydraulic flow resistance:

$$C_z \equiv \frac{U}{u_*} = C_f^{-\frac{1}{2}}, \quad \text{Eq. S2}$$

where C_z is a Chézy friction coefficient, C_f is a dimensionless bed resistance coefficient, and u_* is the bed shear velocity ($u_* = gHS^{0.5}$, where g is acceleration due to gravity). To solve for U , we calculated C_f using the Manning–Strickler formulation:

$$C_f^{-1/2} = \alpha_r \left(\frac{H}{k_s} \right)^{\frac{1}{6}}, \quad \text{Eq. S3}$$

where α_r is a dimensionless constant between 8 and 9 and k_s is the skin friction height (or, the grain roughness height). For sand-bed streams $\alpha_r = 8.32$ is often used (Wright & Parker, 2004). Meanwhile, k_s is

approximated as $k_s \cong n_k D_{90}$, where n_k is a dimensionless number between 1.5 and 3 and D_{90} is the 90th percentile of grain size. We assume $n_k=3$ following Wright and Parker (2004) and van Rijn (1984), and we substituted D for D_{90} , given the difficulty of measuring the D_{90} of sand-grade deposits in the field — we anticipate the effect of this assumption is negligible given that we implement generous uncertainty margins (see Section S3).

We then calculated instantaneous channel-forming water discharge, $Q_{w(cf)}$, as:

$$Q_{w(cf)} = UHW, \quad \text{Eq. S4}$$

where U is the flow velocity, which we calculated using Chézy formulae (Equations S2 and S3), and where H and W were determined from primary and secondary field data, described previously.

ii. Form drag correction

Equations S2 and S3 (and, therefore, S4) assume that all drag force exerted on the river bed is skin friction, i.e., they are skin friction predictors. In the absence of bedforms all drag force exerted on riverbeds is skin friction, however the presence of bedforms exerts additional form drag which acts normal to river beds. Use of a skin friction predictor is problematic as it acts to overestimate shear stress on the river bed (e.g., Andrews, 1984; Kean & Smith, 2006) and is in direct conflict with ubiquitous cross-bedding in terrestrial fluvial sandstone bodies of the Last Chance Ferron Sandstone.

In sediment transport models, form drag is accounted for analytically by “removing” the portion of flow depth affected by form drag. The flow depth of a river, H , can be considered a composite flow depth, H_c , which is the flow depth due to both skin friction and form drag. It is possible to calculate a skin friction flow depth, H_{sk} , which is the flow depth due to skin friction alone, i.e., the portion of the flow depth that is unaffected by form drag. The predictor of Wright and Parker (2004) is an empirical predictor of the Shields stress due to skin friction, τ_{sk}^* , where

$$\tau_{sk}^* = 0.05 + 0.7(\tau^* Fr^{0.7})^{0.8}, \quad \text{Eq. S5}$$

and where Fr is the Froude number ($Fr=U/gH^{0.5}$). We solved for τ_{sk}^* iteratively. We iterated values of H_{sk} between 0 and H and for each value of H_{sk} we calculated: (1) the skin friction bed shear velocity, u_{sk}^* , as $u_{sk}^*=gH_{sk}S^{0.5}$; (2) the skin friction Shields stress, τ_{sk}^* , as $\tau_{sk}^*=H_{sk}S/RD$; (3) a constant T , as $T=(\tau_{sk}^*-0.05/0.7)^{5/4}$; (4) the skin friction flow velocity, U_{sk} , using H_{sk} and Equations S1 and S2; and (5) the composite flow depth, H_c , as $H_c=(T(RD/S)(g^{0.5}/U_{sk})^{0.7})^{20/13}$. We iterated through values of H_{sk} until we found the value of H_{sk} such that H_c is equal to H . This methodology is outlined in detail in Parker (2004).

As k_s is a skin friction roughness height, we then calculated the composite roughness height, k_c , following Parker (2004) as

$$k_c = \frac{11H}{e^{\kappa C_z}}, \quad \text{Eq. S6}$$

where κ is the von Karman constant, taken as 0.4, and where C_z is the skin friction C_z (Equation S1).

iii. Suspended fraction of the bed material load

Prior to calculating the suspended fraction of the bed material load, we first calculated the sediment settling velocity, w_s , following Ferguson and Church (2004), as

$$w_s = \frac{RgD^2}{C_1\nu + (0.75C_2RgD^3)^{0.5}}, \quad \text{Eq. S7}$$

where C_1 and C_2 are constants associated with grain sphericity and roundness ($C_1 = 18$ and $C_2 = 1$ for natural grains; c.f. Ferguson and Church (2004)), and subsequently calculated the Rouse number, Z , as

$$Z = \frac{w_s}{\beta\kappa u_*}, \quad \text{Eq. S8}$$

where β is a constant that correlates eddy viscosity to eddy diffusivity, typically taken as 1.

Several relations have been proposed to calculate the entrainment, E , of uniform material (see review by García and Parker (1991)), which is effectively the concentration of suspended sediment at the reference height or level, a . While various entrainment relations exist (e.g., van Rijn, 1984; García & Parker, 1991; Wright & Parker, 2004), we used the relation of Wright and Parker (2004) which best suits larger, low-sloping sand-bed rivers. Using the value of u_{*sk}^* that resulted in $H_c=H$, Wright and Parker (2004) calculate entrainment as

$$E = \frac{AZ_u^5}{1 + \frac{A}{0.3}Z_u^5}, \quad Z_u = \frac{u_{*sk}}{w_s} Re_p^{0.6} S^{0.7}, \quad \text{Eq. S9}$$

where $A=5.7 \times 10^{-7}$. We then computed the Rouse-Vanoni profile for suspended sediment as

$$I = \int_{\zeta_b}^1 \left[\frac{(1-\zeta)/\zeta}{(1-\zeta_b)/\zeta_b} \right]^Z \ln \left(30 \frac{H}{k_c} \zeta \right) d\zeta, \quad \text{Eq. S10}$$

where b is a/H , $a = 0.05H$ (so $b=0.05$), and ζ is the dimensionless vertical coordinate in the channel cross-section (i.e., $\zeta=0$ on the bed surface and $\zeta=1$ at the flow depth, H). Finally, we calculated the instantaneous suspended fraction of the bed material load, $Q_{bm(s)}$, for channel-forming conditions in units of m^2/s as

$$Q_{bm(s)} = \frac{u_* EH}{\kappa} I. \quad \text{Eq. S11}$$

This framework recovers an instantaneous channel-forming discharge that is specific to the suspended bed material load, i.e., the portion of the bed material load that is intermittently suspended in the water column. It is not appropriate to refer to this value as a suspended sediment load. At present, the suspended bed material load has been reconstructed per unit width, which we multiplied by channel width, W , to recover the total suspended bed material load.

iv. Bedload fraction of the bed material load

To calculate the instantaneous bedload fraction of the bed material load, $Q_{bm(b)}$, for channel-forming conditions, we used the relation of Mahon and McElroy (2018). This model is a bedform-scale model in which the unit bedload flux is calculated geometrically, per unit width. To implement this model, we first calculated the characteristic bedform migration velocity, V_c , prior to calculating the unit bedload flux (i.e., $Q_{bm(b)}$). These variables are given as:

$$\log V_c = \beta_0 + \beta_1 \log S, \quad \text{Eq. S12}$$

$$Q_{bm(b)} = (1 - \varphi) \frac{h_d V_c}{2}, \quad \text{Eq. S13}$$

where $\beta_0 = 0.6113 \pm 0.144$ and $\beta_1 = 1.305 \pm 0.0515$ are constants, and where φ is a dimensionless bed porosity of 0.5 (c.f. Mahon & McElroy, 2018). At present, the bedload fraction of the bed material load has been reconstructed per unit width, which we multiplied by channel width, W , to recover the total bedload fraction of the bed material load.

v. Total bed material load

With estimates of both $Q_{bm(b)}$ and $Q_{bm(s)}$ for channel-forming conditions, we calculated the instantaneous channel-forming bed material load, $Q_{bm(cf)}$, as:

$$Q_{bm(cf)} = Q_{bm(b)} + Q_{bm(s)}. \quad \text{Eq. S14}$$

S3. Uncertainty analysis

To account for variability and uncertainty in model inputs and parameters we implemented a Monte Carlo uncertainty propagation scheme. For each model input and parameter throughout Equations 1–8 in the main text, and in Equations S1–S14, we implemented a range of values. Specifically, we generated 10^6 random samples between bounds defined by this range. We generated samples from a uniform distribution in order to be as conservative as possible with uncertainty, and to avoid introduction of additional assumptions where the shape and the scale of the full distribution of the data is unknown (e.g., Equations S1 and S12). In propagating these randomly generated samples through Equations 1–8 and S1–S14, we recovered 10^6 plausible values for each reconstructed parameter and, therefore, we recovered 10^6 plausible values for the flow intermittency factor, I_w , and the sediment transport intermittency factor, I_s .

For model inputs that reflect field data, we determined this range using the mean (μ) and standard deviation (σ) of the data. We extracted μ and σ of sand-fraction grain sizes, gravel-fraction grain sizes, cross-set thicknesses, and channel architectural element thicknesses (i.e., flow depth proxies). For each of these datasets, we set bounds defined by μ and σ . To calculate sand-transporting flow conditions, which we

interpreted as the dominant channel-forming condition (given the predominance of sand-grade deposits throughout channel-fill facies), these bounds were defined by $\mu - \sigma$ and $\mu + \sigma$. We interpreted gravel-transporting flow conditions as the least dominant channel-forming condition (due to the rare occurrence of gravel-grade deposits in channel-fill facies). However, we considered that gravel-transporting conditions potentially reflect the largest formative flow events preserved in the fluvial stratigraphy. To potentially simulate the largest formative flow events, we calculated gravel-transporting flow conditions using bounds defined by μ and $\mu + \sigma$.

Meanwhile, for model inputs that reflect topographic data, we determined this range based on independent observations or constraints in published literature, as outlined in the main text. Further, for paleoclimate data, we determined this range using results of HadCM3L simulations, which is also outlined in the main text. Finally, for model parameters in Equations S1 and S12, we defined the bounds for constants α_0 , α_1 , α_2 , β_0 , and β_1 as $\mu - \sigma$ and $\mu + \sigma$, using the values given for μ and σ .

S4. HadCM3L General Circulation Model

Here we employed HadCM3L, a coupled Atmosphere–Ocean General Circulation Model (AOGCM) similar to the widely used UKMO model HadCM3, but with a lower resolution ocean. HadCM3L (specifically HadCM3BL-M2.1aD; Valdes et al. (2017)) has a resolution of 3.75° longitude \times 2.5° latitude in the atmosphere and ocean (equivalent to a cell size of 278×417 km at the equator and 278×295 km at 45° latitude), with 19 hybrid levels in the atmosphere and 20 vertical levels in the ocean with equations solved on the Arakawa B-grid with sub-grid scale processes (such as convection, cloud, orographic variance) parameterized. A dynamic vegetation model, TRIFFID (Top-down Representation of Interactive Foliage and Flora Including Dynamics; Cox et al. (1998)), predicts the life cycle and the distribution of vegetation using a plant functional type (PFT) approach for 5 different PFTs: broadleaf trees, needleleaf tree, C3 grass, C4 grass and shrubs. Grid-boxes are fractional and can contain a mixed coverage.

The model used in this study is very similar to the HadCM3BLM2.1aD model that is described and evaluated under modern climate configuration in Valdes et al. (2017), except that it includes a modification to the ozone profile which ensures that the model does not develop a runaway warming at $\times 4$ preindustrial atmospheric CO_2 , as discussed in Lunt et al. (2016). Relative to more recent and/or higher resolution GCMs, HadCM3L is fast and allows millennial and multi-millennial-scale integrations (Farnsworth et al., 2019), which is essential for deep-time modeling work where the initial condition may be far from the final equilibrium state. Recent work shows that deep-time GCMs require multi-thousand year integrations to fully represent applied boundary forcings (Farnsworth et al., 2019), and HadCM3L has been used successfully in numerous pre-Quaternary paleoclimate studies (Lunt et al., 2007; Tindall et al., 2010; Craggs et al., 2012; Lunt et al., 2016).

Boundary conditions for the HadCM3L model used here are those described in Farnsworth et al. (2019), with boundary conditions (topography, bathymetry, ice sheet, solar luminosity) set for the Turonian stage of the Late Cretaceous. Turonian HadCM3L results are also identical to those presented in Farnsworth et al. (2019). Atmospheric CO₂ was set at 1120 ppmv ($\times 4$ preindustrial atmospheric CO₂), which is within the range of the Foster et al. (2017) reconstruction. The simulation was run for 11,422 model years and has reached full equilibrium with ocean integral temperatures showing insignificant trends and a top-of-the-atmosphere net energy balance of 0.1 W/m².

Results from Turonian HadCM3L simulations were resampled to a spatial resolution of 0.1° latitude \times 0.1° longitude, which equates to a cell size of $\sim 11 \times 11$ km, using a bilinear resampling technique. This facilitates visualization of spatial variation and enables selection of an appropriate range of values for paleoclimate variables required in this study.

S5. Field data

Table S1 | Field localities visited in this study

Locations	Elevation, m ($\pm 3-4$)
N38 40 18.9, W111 24 52.5	2255
N38 40 20, W111 24 45.3	2241
N38 40 21.7, W111 24 17.1	2218
N38 40 17.5, W111 24 12	2209
N38 40 12, W111 24 2.5	2190
N38 40 7.7, W111 23 50.3	2179
N38 40 9.1, W111 23 44.8	2187
N38 40 8.9, W111 23 53.6	2215
N38 34 50.9, W111 28 6.2	2668
N38 34 49, W111 28 6.5	2636
N38 34 48.9, W111 28 4.5	2631
N38 34 47.6, W111 28 5.4	2592
N38 34 35.1, W111 27 48.4	2537
N38 44 0.4, W111 18 47.2	1965
N38 43 37.4, W111 18 46.5	1926
N38 43 25.2, W111 18 45.9	1895

Table S2 | Measured paleoflow depth indicators in terrestrial fluvial sandstone bodies of the Last Chance Ferron Sandstone. The data that are listed as sourced in this study are the same data as those presented in Lyster et al. (2022)

Paleoflow depth proxy	Thickness (m)	Source
Point bar deposit	9.1	Cotter (1971)
Point bar deposit	<8	Gardner et al. (2004)
Channel-fill deposits (maximum thickness)	~9	Gardner et al. (2004)
Channel-fill deposits (maximum thickness)	~9	Garrison Jr and van den Bergh (2004)
Point bar deposits	8, 7.5, 9, 3.2, 4.8, 3.6, 6.5, 7.5, 3.6, 4.1, 2.7, 6.4, 5.5, 2.8, 1.1, 1.9, 7.5, 2.7, 7.1, 1.2, 4.4, 3.7, 3.1, 3.4, 3, 2.5, 5.9, 2.5, 4.7, 10, 4.2, 1.6, 3, 6.5, 10	This study/Lyster et al. (2022)
Single channel stories (maximum thickness)	8.6, 11.1, 12.2, 9, 7.6, 7.1, 3.9, 5.6, 2.6, 7.3, 12, 9.3	This study/Lyster et al. (2022)

Table S3 | Field data collected in this study. The table is located in the attached Excel spreadsheet in the sheet named “Data”.

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