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Delta progradation mechanisms and their stratigraphic record: the Namurian (Carboniferous) Kinderscout Delta, Pennine Basin, UK

Abbreviated title: Delta progradation mechanisms in the Pennine Basin

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Abstract: The Kinderscout Delta (Hebden Formation, Upper Carboniferous, Pennine Basin, UK) represents a well-exposed example of a river-dominated delta prograding into a relatively unconfined, deepening basin. High-resolution sedimentary logs, palaeoflow measurements, and UAV-derived virtual outcrop models were used to characterize facies, architectural elements and delta front depositional processes. This delta was fed by a major fluvial system, with distributaries 2–10 m deep, and mouth bars that prograded into water depths of at least ~50 m. Facies and architectural element analyses indicate a continuum of processes from shallow-water, bed-friction-dominated deposition in the landward clinothems to gravity-driven sedimentation on deeper-water foresets. Mass-failure of oversteepened foresets facilitated sediment bypass, contributing sediment directly to the prodelta and, in some cases, steering distributary avulsion. This behaviour resulted in the successive formation and abandonment of mouth bars at quasi-stationary fluvial outlets. This study highlights the interaction between two endmember conceptual models of deltaic deposition: “topset-dominated” autocycles driven by distributary avulsion in shallow, low-gradient settings, and “foreset-dominated” processes driven by gravitational instability on steep foresets in deeper-water shelf environments. Our results demonstrate that delta systems can hybridize, i.e., show downslope continuity between, endmember behaviours during progradation into deepening basins, with autogenic foreset processes and delta-top avulsion operating simultaneously.

Keywords: Pennine Basin, Carboniferous, Debris flows, Turbidity currents, Compound deltas

Introduction

Deltas – the deposits of rivers entering standing bodies of water such as seas and lakes – are important geomorphological elements in the landscape. They are fundamentally regressive, representing an important natural means of creating land (Bhattacharya and Walker, 1992; Reading and Collinson, 1996), and are major sinks of terrestrially derived sediment on marine shelves and beyond (Zavala *et al.*, 2021). Deltas show substantial morphodynamic variability as a function of, and sensitivity to: (1) the river conditions as it enters the standing body of water (flow velocity, concentration and calibre of bed and suspended sediment load); and (2) the conditions of the receiving basin (salinity, the presence of flocculated clay or “fluid mud” depth and gradient; Wright, 1977; Zavala *et al.*, 2021, 2024). Delta processes are often also influenced by waves and tides (Galloway, 1975; Nyberg and Howell, 2016).

Jet flow theory provides a unifying framework for deposition at river mouths that accounts for the interplay between riverine and basinal conditions (Bates, 1953; Wright, 1977; Zavala *et al.*, 2021, 2024). A flow exiting the river mouth can be treated as a turbulent jet which undergoes flow expansion and momentum diffusion into the standing water, thereby depositing its sediment load. When the density of the river discharge is less than that of the basin water, the jet is buoyed (a hypopycnal flow). Bedload is deposited at the river mouth in a feature termed a mouth bar (Wright, 1977; Edmonds and Slingerland, 2007) which aggrades and expands laterally and down-slope. The buoyant plume transports suspended load basinward, fallout from which contributes to delta front and more distal prodelta deposits. When river and basin waters have equal densities, total mixing or pure jet conditions (homopycnal flow) result in rapid flow deceleration and deposition of suspended load and bedload in a mouth bar, and delta front fines are limited. In practice, homopycnal conditions are most likely rare in marine settings, because maintaining the precise suspended-sediment concentration required to match the density of saline seawater is difficult (Wright, 1977; Zavala *et al.*, 2021, 2024). Finally, when river discharge density exceeds that of the receiving water, the flows plunge beneath the surface, bypassing the river mouth and transporting sediment in bottom-hugging hyperpycnal flows (Mulder and Syvitski, 1995; Wilson and Schieber, 2014). These deposits can form low-angle ramps that dip away from the river mouth (Zavala *et al.*, 2021, 2024).

In addition to buoyancy contrasts between the fluvial water input (the jet) and basin water, basin physiography plays an important role in controlling morphodynamics; that is sea/lake-floor gradient and bathymetry. In deltas building into relatively shallow basins (receiving water depth $\approx$ channel depth) with low bathymetric gradients, bed frictional interactions with the incoming fluvial jet are important (Wright, 1977; Edmonds *et al.*, 2011). In settings where tide and wave influence are negligible, delta morphodynamics are controlled primarily by mouth bar growth which shallows and

bifurcates the flow (Elliott, 1986; Wellner *et al.*, 2005; Olariu and Bhattacharya, 2006; Edmonds and Slingerland, 2007; Fagherazzi *et al.*, 2015; Daniller-Varghese *et al.*, 2020; Cole *et al.*, 2021), eventually resulting in mouth bar abandonment. Mouth bar abandonment may be followed by the formation of new bars in a more basinward position at the mouth of the relocated channels (Olariu and Bhattacharya, 2006; Edmonds and Slingerland, 2007). The concomitant reduction in flow efficiency resulting from progressive bar development and flow constriction induces backwater effects (*sensu* Cole *et al.*, 2021), thus sedimentation in the fluvial distributary, and triggers upstream avulsion (Hoyal and Sheets, 2009; Ganti *et al.*, 2016; Jerrett *et al.*, 2016; Cole *et al.*, 2021). This cycle of autogenic avulsion is considered fundamental to the formation of the dendritic, fractal plan-view of classical “river-dominated” deltas such as the modern Mississippi Delta (Coleman and Gagliano, 1964; Galloway, 1975; Edmonds and Slingerland, 2007).

Deltas building onto relatively steeply inclined basin slopes (though still perhaps $<5^\circ$), where bathymetries can increase to orders of magnitude greater than the depth of the feeding distributary channel, are common at shelf margins and in tectonically active settings. The incoming jet readily detaches from the bed at the “brinkpoint”, resulting in rapid aggradation of bedload in a barform that quickly attains or exceeds the natural angle-of-repose. The unstable bars undergo cycles of mass-failure – triggering density currents (grainflow, debris flows, and turbidity currents; Postma, 1984; Nemec, 1990; Nemec *et al.*, 1999; Winsemann *et al.*, 2009; Gobo *et al.*, 2015) that transport sediment down-slope – and progradation. Periodic mass-wasting of the mouth bar “brinkpoint” means that stable barforms tend not to be preserved at the river mouth. Autogenic behaviour, then, is expressed at the delta front, rather than on the delta top, hence Edmonds *et al.* (2011) refer to these as “foreset-dominated”, in contrast to “topset-dominated” deltas building into shallow-water, low-gradient settings. “Gilbert” or “fan” deltas (Gilbert, 1885; Budai *et al.*, 2021 and references therein) are characteristic examples of “foreset-dominated” deltas.

Because deltas can prograde across gently deepening shelves or ramps for several tens to hundreds of kilometres, the evolution from shallow-water, frictional “topset-dominated” deltas to deep-water, gravity-driven “foreset-dominated” deltas should in theory occur (e.g., Dixon *et al.*, 2013; Jerrett *et al.*, 2016; Cole *et al.*, 2021). Here, we aim to conduct an in-depth study of this transition by revisiting a well-exposed delta succession that was the subject of a classic early application of facies analysis – the Upper Carboniferous (Namurian) “Kinderscout Delta” (Pennine Basin, northern England, UK; Fig. 1A and B; Collinson, 1968; McCabe, 1978; Hampson, 1997). We do so with the benefit of improved data-collection techniques for architectural analysis (i.e., uncrewed aerial vehicles and orthorectified photography), and through the lens of several decades of advances in the understanding of deltaic and deep-water processes. By exploring the facies distribution and

architectures present in outcrops which record a shallow to deep-water transitional deltaic environment, we provide context for processes operating in similar systems in the modern day, as well as insights applicable to comparable rocks elsewhere.

Geological and stratigraphic context

During the early Carboniferous, the area that now comprises the Midlands and north of England was located on the southerly margin of the Laurussian continent at subequatorial latitudes in a hot and arid climatic setting (Jones, 2022). Northerly-directed subduction of the Rheic Ocean crust beneath the continental margin initiated a protracted episode of north-south extension and subsidence, forming an interconnected network of broadly west-east-oriented (half-) grabens known as the Pennine Basin (Collinson, 1988; Davies, 2008; Waters *et al.*, 2009; Woodcock and Strachan, 2012). The basin was bounded by two long-lived palaeogeographical highs to the north (Southern Uplands High) and south (Wales–London–Brabant Massif; Fig. 1A). The configuration of the interconnected grabens was strongly controlled by the geometries of pre-existing Caledonian structures in the lower Palaeozoic basement (Fraser and Gawthorpe, 1990, 2003). Within this context, initial deposition during Tournasian to Viséan times (~350–330 Ma) was dominated by carbonate platforms formed on horst blocks and half-graben flank highs, with deeper water siliciclastic and carbonate muds and calc-turbidites in the intervening bathymetric lows (Waters *et al.*, 2011).

By the end of the Viséan, active rifting gave way to a regime of slower post-rift thermal subsidence (Leeder, 1982; Leeder and McMahon, 1988; Fraser and Gawthorpe, 2003), and by Namurian times a major pulse of siliciclastic sediment delivered from the Laurussian craton to the north replaced carbonate deposition. Deeper troughs of the southern Pennine Basin were at this time accumulating thick mudstone successions (the Bowland Shale Formation) which, owing to their largely barren nature (with the exception of distinct marine bands), are considered to have formed under predominantly brackish or freshwater basin conditions, at least within the photic zone (Collinson, 1988; Holdsworth and Collinson, 1988). Remnant block-and-basin topography was progressively healed from the north through progradation of a major delta preceded by siliciclastic turbidites (the Millstone Grit Group; Fig. 1A and C; Collinson, 1988; Fraser and Gawthorpe, 2003), ultimately resulting in the establishment of a broad continental and paralic plain across the entire Pennine Basin by the Westphalian (the Coal Measures Supergroup; Fig. 1C). By this stage, the Pennine Basin province had drifted into equatorial latitudes in which warm and humid conditions prevailed (Davies, 2008). Peat accumulation was a feature of continental and delta-top environments in the Millstone Grit Group and, to a much greater degree, the Coal Measures Supergroup.

The focus of the present study is within the Hebden Formation (of the Millstone Grit Group), a lithostratigraphical unit that represents the first pulse of northerly-derived clastic sediment to the Edale Gulf (Fraser and Gawthorpe, 2003) or Basin (Yang *et al.*, 2015), which represents one of the fault-bounded Pennine Basin depocenters centred on what is now the north Derbyshire Peak District (Fig 1A and B). In the Edale Basin, the Hebden Formation comprises fine- to coarse-grained, pebbly, feldspathic sandstone interbedded with intervals of siltstone and claystone. The succession forms a broadly regressive, up to 600 m thick coarsening-upward package that has been traditionally (but informally) subdivided into the “Mam Tor Sandstone” or “Mam Tor Beds”, “Shale Grit”, “Grindslow Shales” and “Kinderscout Grit” (e.g., Allen, 1960; Reading, 1964; Walker, 1966b; Collinson, 1967; Collinson, 1969; McCabe 1978; Hampson, 1997; Waters *et al.*, 2007; Fig. 1C). A summary of these units and their classic Edale Basin outcrop localities is provided in Southern *et al.* (2014). The base of the formation is defined by the occurrence of the first significant sandstone bed above the Bowland Shale Formation (formerly, the “Edale Shales” in the Edale Sub-basin; e.g., Hudson and Cotton, 1945; Allen, 1960; Stevenson and Gaunt, 1971), at or above the *Hodsonites magistrorum* goniatite Marine Band (R_{1a}1). The top of the formation is defined by a shale comprising the *Bilinguites gracilis* goniatite Marine Band (R_{2a}1). These biozones bracket the R₁ stage (also termed the Kinderscoutian) of the Namurian in the British Carboniferous biostratigraphical framework (Walker, 1966b; Collinson, 1967; Waters and Davies, 2006; Waters *et al.*, 2007; Waters and Condon, 2012; Fig. 1C).

The Edale Basin was well-connected to the open ocean, as indicated by the fully marine fauna contained within the marine bands (e.g., Anita and Wood, 1977; Waters *et al.*, 2019) and “Mam Tor Beds” (Hudson and Cotton, 1945). The “Mam Tor Beds” (<120 m thick; Allen, 1960; Walker, 1966b) and “Shale Grit” (140–210 m thick; Allen, 1960; Walker, 1966b,) are interpreted as representing relatively unconfined basin-floor fan and deep-water slope turbidites respectively (Allen, 1960; Walker, 1966b; Collinson, 1968; McCabe 1978; Davies and McLean, 1996; Southern *et al.*, 2014) building into open marine conditions. It has been suggested that, at the time of this turbidite emplacement, the parent delta system was either positioned in the area immediately north of Rochdale, Greater Manchester, and Holmfirth, West Yorkshire (Allen, 1960), or around Bradford and Huddersfield, West Yorkshire (Walker, 1966b). Water depths at the time of the deposition of the upper part of the Bowland Shale Formation are thought to have been in the region of 100–200 m (Holdsworth and Collinson, 1988; Davies, 2008; Emmings *et al.*, 2020, 2024), which provides a broad context for basin bathymetry at this time. The turbiditic beds, in the Edale and Ashop valleys which flank the Kinder Scout plateau (Fig. 2), display broadly southward palaeocurrent indicators (Southern, 2015). The overlying “Grindslow Shales” (60–105 m thick; Collinson, 1969; Walker, 1966b) is characterized by channelized sand-prone elements encased within finer-grained material that Walker (1966a, 1966b) and McCabe

(1978) interpreted as “slope” fines associated with turbiditic slope channel- or gully-fills. By comparison, Collinson (1969) interpreted this unit as inter-distributary bay deposits with subordinate “fluvialite” channels. The uppermost “Kinderscout Grit” (~120 m thick; Walker, 1966b) is sand-prone, and is characterized by a lower unit displaying delta clinoform foresets up to 40–50 m thick and dipping by up to 30° (Collinson, 1968; McCabe, 1978; Hampson, 1997; this study), overlain by flat-lying cross-bedded channel-fill topsets that are generally agreed to represent fluvial distributaries (Walker, 1966b; Collinson, 1968; McCabe, 1978). Some workers locally distinguish an “Upper” and “Lower Kinderscout Grit” (e.g., Hampson, 1997), separated by the Butterly Marine Band. However, there are conflicting views on whether the Butterly Marine Band represents the base of the $R_{1c}4$ (and is thus equivalent to the *Reticuloceras coreticulatum* Marine Band; e.g., Davies and McLean, 1996) or the stratigraphically higher $R_{1c}5$ (e.g., Aitkenhead and Riley, 1996; Waters *et al.*, 1996; Hampson, 1997) interval.

Deposition of the “Lower Kinderscout Grit” foresets was thought by Collinson (1968) to have occurred in the form of Gilbert-type deltas building out into progressively deeper water (Hampson, 1997). The scale of the delta foresets (up to 50 m tall; Collinson, 1968; this study), and their facies context (building out over slope and slope-gully deposits; Walker, 1966b; McCabe, 1978), support the notion that these were “turbidite fronted” deltas (*sensu* Hampson, 1997) building into “deep” water. An alternative fluvial origin for this succession was proposed by McCabe (1975, 1977), whereby the large foresets were deposited by the downstream migration of bank-attached “alternate bars” whose heights represent the full depth of a 1–2 km wide river, a hypothesis which led Read (1991) to suggest that the “Lower Kinderscout Grit” represents the deposits of a “braid delta” (*sensu* McPherson *et al.*, 1987). This conceptual model of “alternate bars” has been questioned in the outcrop studies of Bijkerk (2014) in the Craven Basin.

Method and approach

The northern flank of the Kinder Scout plateau features a ~2 km long, ~25 m high escarpment – Ashop Edge (Fig. 2) – along which the “Lower Kinderscout Grit” (Hebden Formation) is extensively exposed (Collinson, 1968). Eight high-resolution (1:50) sedimentary logs were collected around the Kinder Scout plateau along well-exposed successions (predominantly stream gullies; Figs. 2 and 3). Grain size, sedimentary structures, and trace and body fossils were documented and used to define the range of lithofacies (Table 1; Fig. 4) and architectural elements (Figs. 5 and 6) within the Hebden Formation. Palaeoflow measurements were collected from the foresets of planar cross-beds and crests of asymmetrical current ripples. The dip directions of bar-scale accretion surfaces were measured to compare bar-accretion direction with palaeoflow measurements.

Uncrewed aerial vehicle (UAV) imagery of Ashop Edge was collected, then processed into a photomosaic and three-dimensional virtual outcrop model (Fig. 7A and B) using photoscan (Agisoft®) photogrammetric software and viewed within LIME® software (formerly by Virtual Outcrop Geology Group, now Fonix Geoscience; Buckley *et al.*, 2019). The virtual outcrop model permitted correlation between three logs collected along Ashop Edge (Fig. 3B, C and D), and aided observations of the architecture and interpretation of the panel.

Facies analysis

Six architectural elements, composed of nine lithofacies, have been recognized. The lithofacies are described in detail in Table 1 and representative photographs are shown in Figure 4. The architectural elements, two of which (C and F) possess a pair of sub-elements, are: (A) Delta Front Fines, (B) Delta Front Gully-fill, (C) Mouth Bar Sands, (D) Mouth Bar Fines, (E) Mouth Bar Mass-failure Deposit, and (F) Fluvial Distributary Channel-fill. The architectural elements are described and interpreted below, with representative images shown in Figure 5 and example facies stacking patterns shown in Figure 6. Where possible, correspondence of architectural elements with those identified by previous workers is indicated.

Architectural Element A (Delta Front Slope Fines)

Description: This architectural element is poorly exposed in discontinuous active landslide scarps and stream gullies cut into the grassy slopes beneath Ashop Edge. It consists of metre- to decimetre-thick successions of mudstone, siltstone, and more rarely claystone (M) in which subtle normal grading defines possible bed-scale features 0.1–1 m thick in places. The mudstones interstratify with lesser (<50%) discrete massive to normally-graded (SstT) or ripple cross-laminated (SstR) very fine- to fine-grained <0.5 m thick sandstone beds (Figs. 4A, 5A and 6). Detailed analysis of sandstone geometries is not possible, but beds are tabular at outcrop scale. *Planolites* and other trace fossils are recorded in equivalent strata by Collinson (1969), Davies and McLean (1996) and Hampson (1997).

Interpretation: These mudstones are the deposits of suspension-settling of clay and silt. The origin of these suspension-loads may be hypopycnal plumes derived from river mouths. However, subtle normal grading in the mudstones, as well as the presence of discrete graded sandstone beds, are indicative of deposition by low-concentration turbidity currents (or low-density turbidity currents *sensu* Talling *et al.*, 2012). These currents may have been sourced directly from river mouths (i.e., hyperpycnites; Mulder *et al.*, 2003), generated through the mass-failure of delta front mouth bars up-slope (e.g., Hughes Clarke, 2016), or by the collapse of buoyant plume sediments (Parsons *et al.*,

2001; Zavala *et al.*, 2016; Hage *et al.*, 2019). The generally tabular nature of the deposits suggests deposition in an unconfined setting. These observations point toward deposition on a prodelta shelf or low-gradient slope (Walker, 1966b; McCabe, 1975, 1978). The depositional context of this architectural element, alongside Architectural Element B (Delta Front Slope Gully-fill) and Architectural Elements C (Mouth Bar Sandstone) and D (Mouth Bar Fines; discussed below), suggests that it was developed on a slope or ramp setting.

Architectural Element B (Delta Front Slope Gully-fill)

Description: This element consists of erosionally-based 3–10 m thick crudely fining-upward successions. Although typically not well exposed, a basal master erosion surface defines a lithosome that has an irregular convex-down “scour-fill” geometry, which is typically 50–100 m across and has up to 10 m of erosional relief into Architectural Element A (Delta Front Slope Fines), C (Mouth Bar Sandstone) or D (Mouth Bar Fines; Figs. 5A and 7A). The scour fills are characterized by a more massive lower component of medium-to-very coarse-grained, granular, structureless or crudely stratified 0.5–2 m thick sandstone beds that show little grading (SstPM and SstMI; Figs. 4G and 4H). Subvertical dewatering fabrics (SstDW) are common (Fig. 4I). These in some cases pass upward into normally-graded (SstT; Fig. 4B) and ripple cross-laminated (SstR; Fig. 4C) fine-to-medium-grained sandstones that display an upward increase in mudstone interbeds (Fig. 3C). Rare planar cross-bedded sandstones (SstXbP; Fig. 4D) occur (Fig. 6). In some cases, lower order erosion surfaces, with up to 2–3 m of vertical relief, define bed-sets within these elements. These erosionally-based bed-sets stack generally vertically within the master erosion surface which defines the margin of the element. Casts of metre-scale tree branches and trunks filled with coarse sand are documented within this architectural element.

Interpretation: The major erosion surfaces define episodes of incision and the generation of a long-lived, up to 10 m deep, sediment conduit cut into Architectural Element A (Delta Front Slope Fines), C (Mouth Bar Sandstone) and D (Mouth Bar Fines). The existence of multiple internal erosionally-based bed-sets implies that the generation and fill of this feature were characterized by repeated episodes of excavation and fill by smaller-scale (2–3 m deep) channels. The general aggradational stacking of these erosive-based bed-sets is indicative of low rates of lateral accretion; the element is interpreted as having been generated by relatively low-sinuosity channels. The general fining-upward profile is likely a function of higher preservation of the youngest channel storeys before the ultimate abandonment of the conduit.

The character of the lower part of the fill – dominated by crudely-stratified or structureless sandstone – is characteristic of the rapid aggradation of traction carpets in the upper flow regime,

and/or of deposition by debris flows. Rapid deposition is indicated by the presence of dewatering fabrics (Fig. 4I); dewatering may also, in part, explain the absence of well-developed bed-scale grading. The dominance of normally-graded and ripple cross-laminated beds in the upper part of the fill is characteristic of deposition by high- and low-density turbidity currents. The presence of rare planar cross-beds implies that flows were at times sustained enough for dune formation as a result of sediment bypass through the gully system (Stevenson *et al.*, 2015), possibly hinting at a hyperpycnal (i.e. river-derived) origin for some of these flows.

The overall context of this element – eroding into Architectural Element A (Delta Front Slope Fines), C (Mouth Bar Sandstone) and D (Mouth Bar Fines) – suggests that it represents slope gully- or channel-fills. The flows were derived from up-slope failure of mouth bars and, at times, were possibly generated by hyperpycnal flows fed directly by river mouths. Collinson (1969) envisaged this association to represent “deep fluvial channels” but, for the reasons described above, we concur with McCabe’s (1978) interpretation of slope channels or gully-fills generated by multiple episodes of cut-and-fill.

Architectural Element C (Mouth Bar Sandstone)

This is the main cliff-forming element at Ashop Edge, where it forms successions up to 40 m thick (Figs. 3C and 7A). Beds are inclined broadly toward the west (Figs. 2 and 3). This architectural element is formed of the following two sub-elements.

Architectural Sub-element C1 (Gravity-dominated)

Description: Beds which are dominated by medium- to coarse-grained, massive or inverse-graded (SstMI; Fig. 4H), plane-bedded or massive, pseudo-stratified (SstPM; Fig. 4G), and plane-bedded, well-sorted (SstPW; Fig. 4F) sandstone, though a lesser proportion of dewatered sandstone (SstDW; Fig. 4I) also occurs (Fig. 6). Thicknesses in the region of 50 m are recorded at Kinder Downfall (Figs. 3E and 5B). The salient feature of this element is the inclination of strata, which dip by up to 30° (18–25° is common) and define a distinct clinoform geometry. Clinoforms dip towards the west or SW, broadly parallel to palaeoflow direction (Figs. 2 and 3). Clinoform-sets range between 5–50 m tall post-compaction and display a planar to weakly concave-up geometry, downlapping at relatively high angles onto the underlying substrate, and are truncated at high angles at their tops (Figs. 5B and 6A). That is, they form well-developed foresets and weakly-developed toesets. In places, a subtle sigmoidal geometry is also recognized as the preservation of the topset-foreset transition (or “rollover”; *sensu* Steel and Olsen, 2002; Helland-Hansen and Hampson, 2009). The sedimentary logs show that component lithofacies of this architectural element interstratify in a complex manner and their

proportions vary significantly from location to location. For example, well-sorted sandstone (SstPW; Fig. 4F) dominates at Fairbrook Naze (Fig. 3B), whereas massive to inverse-graded sandstones (SstMI; Fig. 4H) dominate at Nether Red Brook (Fig. 3C). Lithofacies may also grade or transition down-clinothem. Overall, this architectural element does not display a well-developed vertical grain size trend. Sand-filled tree branch and trunk casts occur, as well as intact tabular mud clasts.

Interpretation: The scale (5–50 m) and orientation of the clinoforms (dipping parallel to palaeoflow), and their downlap onto a basal surface, for which there is no evidence of erosion, implies that this element represents increments of bar progradation at the mouth of a river into a relatively unconfined setting. This interpretation is consistent with previous interpretations, with this element being a component of Collinson's (1969) and Hampson's (1997) "Deltaic Sedimentation Unit" and "Giant Foreset" associations, respectively. Units dipping in the flow-direction (i.e., basin-dipping clinoforms) were deposited on surfaces inclined by up to 30° and are as much as 40 m high at Ashop Edge (Fig. 7C). Both the angle and height of these clinoforms are likely to be underestimations of true bedform height and gradient due to post-depositional compaction. Deposition was dominated by gravity-driven flow processes; grainflows (SstPW), debris flows (SstMI), and high concentration (also known as high-density; *sensu* Talling *et al.*, 2012) turbidity currents (SstPM) are all represented. Rapid deposition is indicated by the occurrence of dewatering structures (SstDW).

Architectural Sub-element C2 (Friction-dominated)

Description: The geometric characteristics of this sub-element (e.g., height, dip direction and angle, relationship to the underlying and overlying strata) are the same as described above for Architectural Sub-element C1 (Gravity-dominated Mouth Bar Sandstone). This sub-element is comprised of randomly stacked beds of tabular cross-bedded (SstXbP; Fig. 4D) and ripple cross-laminated (SstR; Fig. 4C) sandstone (Fig. 6), and occurs predominantly in the upper half of clinothem. It is in lateral continuity with Architectural Sub-element C1 (Gravity-dominated Mouth Bar Sandstone) which typically lies in a down-clinothem, basinward position. Where measurable, SstXbP palaeocurrent indicators are generally concurrent with the direction of mouth bar progradation (Figs. 2 and 3).

Interpretation: The above interpretation of the basinward-dipping geometry of Architectural Sub-element C1 (Gravity-dominated Mouth Bar Sandstone) as incremental bar progradation also applies here. Sustained flow and traction-driven processes, represented by rippled and cross-bedded sandstones (SstXbP and SstR), occurred frequently on the upper foresets. The position of this sub-element on the upper part of the foreset indicates that bed traction was likely sustained by the

incoming fluvial jet at the nearby river mouth. This is supported by the similarity in palaeocurrent direction in the SstXbP facies and the general sense of delta progradation.

Architectural Element D (Mouth Bar Fines)

Description: This element is poorly exposed. It comprises recessive mud-dominated units (M; Fig. 4A). Ripple cross-laminated fine-grained sandstones (SstR; Fig. 4C) have been observed, and other lithofacies may be present (Fig. 6). Collinson (1968) describes bioturbation within units equivalent to this element. The element can form units that are tens of metres in thickness, which interstratify with Architectural Element C (Mouth Bar Sandstone). Consequently, clinoform geometries which are broadly conformable with, and at a similar scale to, those observed in Architectural Element C (Mouth Bar Sandstone) are interpreted rather than observed. Architectural Element C (Mouth Bar Sandstone) may also pass down-clinothem into Architectural Element D (Mouth Bar Fines). Owing to poor exposure, the occurrence of this element, and the nature of its basal contact with Architectural Element A (Delta Front Slope Fines) is largely inferred.

Interpretation: The geometrically conformable relationship of this element with Architectural Element C (Mouth Bar Sandstone) suggests that it represents periods of reduced sediment supply to mouth bar foresets, thus conditions of lower energy/less powerful fluvial currents, or avulsion and relocation of the distributary mouth. These elements were recognized by Collinson (1968) and are contained within their “Deltaic Sedimentation Unit” association. The element is thought to transition down-clinothem into Architectural Element A (Delta Front Slope Fines).

Architectural Element E (Mouth Bar Mass-failure Deposit)

Description: Asymmetric, “spoon-shaped” (e.g., Frey Martinez *et al.*, 2005) surfaces with marked west-facing concavity display incisional geometries as much as 10 m into Architectural Element C (Mouth Bar Sandstone; Fig. 5C). These surfaces contain medium-to-coarse, massive or inverse-graded (SstMI; Fig. 4H), plane-bedded or massive, pseudo-stratified (SstPM; Fig. 4G), plane-bedded, well-sorted (SstPW; Fig. 4F), and/or dewatered (SstDW; Fig. 4I) sandstone. That is, most of the lithofacies that also characterize Architectural Element C (Mouth Bar Sandstone), though dewatered sandstone (SstDW) facies, which are relatively uncommon in Architectural Element C (Mouth Bar Sandstone), tend to dominate (Figs. 3 and 6). Strata are less steeply inclined to the west than the subjacent mouth bar strata, or dip to the east (landward), or form a concave-down (mounded) geometry. The mounded or irregular geometry that marks the top of this element is downlapped or overlain by Architectural Element C (Mouth Bar Sandstone; Fig. 5C).

Interpretation: These elements are interpreted as slide and slump deposits. The west-facing (i.e., basin-facing) spoon-shaped basal surfaces are interpreted as listric normal faults, which are generally considered to be caused by gravitational processes (Jackson and Galloway, 1984; Imber *et al.*, 2003). Consistent with this interpretation, the subhorizontal or easterly-dipping strata above the fault can be explained in terms of the back-tilting of inclined Architectural Element C (Mouth Bar Sandstone) beds over the listric glide plane (i.e., half-graben). The development of mounded geometries and presence of substantially dewatered intervals suggest that plastic deformation to mass-fluidisation occurred as strata were translated. Therefore, the material was moved both as coherent slides and incoherent slumps (Middleton and Hampton, 1973; Postma, 1984; Pisarska-Jamroz̳y and Weckwerth, 2012). Transport distances are unlikely to have been more than the order of metres. This element has not been described by previous workers in the “Lower Kinderscout Grit” *per se*, but may be a re-interpretation of the same concave-upward erosion surfaces identified by Jones and McCabe (1980), which were previously thought to be created by either bar lee face scouring due to changes in flow geometry over the bar, or thalweg migration in a deep fluvial setting (McCabe 1975, 1977).

Slide and slump deposits are frequently described from oversteepened delta front deposits (e.g., Prior and Coleman, 1982; Dunne and Hempton, 1984; Nemec *et al.*, 1988; Pulham, 1989; Braathen *et al.*, 2018) including large-scale growth faults elsewhere in the Carboniferous of Derbyshire (Chisholm, 1977; Chisholm and Waters, 2012).

Architectural Element F (Fluvial Distributary Channel-fill)

This element forms a <15 m thick sheet-like or tabular unit on the upper part of Ashop Edge, and on Kinder Plateau itself. The succession largely overlies an erosion surface with up to 10 m of relief into the Architectural Element C (Mouth Bar Sandstone; Fig. 5B) and D (Mouth Bar Fines), but in places interfingers with these strata (Fig. 7C; see Architecture of Ashop Edge, below). It consists of the following two sub-elements which are differentiated based on their sedimentological character and cross-sectional geometry.

Architectural Sub-element F1 (Delta-top Distributary Network)

Description: This element consists overwhelmingly of planar and trough cross-bedded sandstones (SstXbP and SstXbT; Figs. 4D and 4E), with minor ripple cross-laminated sandstones (SstR; Fig. 4C). Cross-beds are typically 0.1–0.2 m thick, rarely as much as 0.5 m thick (Figs. 3C, 3E, 3G and 6). Angular to well-rounded quartz pebbles are common. The dominance of trough cross-bedded sandstone results in a complex architecture of crosscutting surfaces which is difficult to

define. However, crude, up to 1.5 m thick fining- and thinning-up bed-set packages, which overlie erosional surfaces that beds down- and onlap onto, can be discerned (Figs. 3 and 5B). Palaeoflow, as determined from cross strata and ripple-crests, is uniformly toward the west and SW (Figs. 2 and 3).

Interpretation: This element is interpreted as the deposits of delta-top fluvial distributary channels and is equivalent to the “Migrating Fluvial Channels” association of Collinson (1969). The difficulty in delineating an internal architecture for the deposits makes it challenging to reconstruct fluvial style (Hampson, 1997). The crude, up to 1.5 m thick fining- and thinning-upward bed-set packages, and their associated basal erosional surfaces, are likely individual storeys representing the fills of channels at least 1.5 m deep. However, cross-set heights of 0.1–0.2 m, and exceptionally 0.5 m, imply flow depths, therefore channel depths, of typically 2–4 m to as much as 10 m (Leclair and Bridge, 2001; Bradley and Venditti, 2017) not accounting for post-depositional compaction. The absence of preserved overbank, and of well-developed macroforms indicative of either lateral or downstream accretion, suggests the system may have been transitional between braided and anastomosing (Hampson, 1997). Intact mud clasts within Architectural Element C (Mouth Bar Sandstone; Fig. 3E and 3F), which indicate the possibility of subaerial exposure and subsequent fluvial flooding of the proximal upstream environment, indirectly support this hypothesis.

Architectural Sub-element F2 (Distributary Channel Mouth)

Description: Rare metre-scale tabular units of plane-bedded, well-sorted sandstone (SstPW; Fig. 6) which overlie concave-up erosion surfaces which incise Architectural Element C (Mouth Bar Sandstone; e.g., Figs. 3B, 4F and 5D). This incision cuts down on the order of ~5 m below the generally planar erosive base of Architectural Sub-element F1 (Delta-top Distributary Network; Figs. 5B and 7). This results in a lenticular geometry to these elements in cross-section. The elements are only found toward the very top of the succession and are not often downlapped by later Architectural Element C (Mouth Bar Sandstone) strata except at Kinder Downfall (Fig. 5B). They appear to be downlapped by Architectural Sub-element F1 (Delta-top Distributary Network) strata at Ashop Edge (Fig. 7C), placing their relative age as older than the delta-top distributary network.

Interpretation: Since the cross-sectional lenticular geometry of the channel is preserved, it can be inferred that lateral migration of these channels did not occur. This suggests either that they were short-lived conduits or were formed in areas where the bar-top gradient was locally high, or indeed both of the above. The coarse, plane bedded nature of the channel fill (Fig. 3B) suggests that deposition occurred in the upper flow regime under high discharge conditions capable of transporting coarse sediment and incising the mouth bar. There is no evidence for a multi-generational fill, suggesting that these channels were short-lived and quickly abandoned. We propose that these channels were incised

at the lip of the mouth bar, and they therefore represent the very terminal reaches of the distributary channel network during a flood event. The relative age of these sub-elements, younger than the Architectural Element C (Mouth Bar Sandstone) strata they incise, but older than Architectural Sub-element F1 (Delta-top Distributary Network), supports the theory that they formed between the mouth bar and the delta-top distributary network in plan-view, thus close to the delta front. We tentatively propose that they represent the landward continuation of Architectural Element B (Delta Front Slope Gully-fill) where these features resulted from hyperpycnal flows derived directly from the river mouth. Similar deposits have been interpreted from modern and ancient deltas by Olariu and Bhattacharya (2006). Termed “terminal distributary channels”, they form at the “last subaerial bifurcation and extend to the last channelised expression on the delta front” (Olariu and Bhattacharya, 2006, p.213).

Architecture of Ashop Edge

Palaeocurrent and clinoform dip data (Collinson, 1968; this study, Fig. 2) show that the Ashop Edge section (Fig. 7) is an oblique dip-oriented view through the architecture of the “Kinderscout Delta”. In this section, we examine how the architectural elements described above manifest in a two-dimensional exposure, and how they relate to one-another spatially. Logged sections provide verification of facies present at intervals across the exposure, allowing the cross-sectional geometry of architectural elements and their component facies to be robustly linked.

Description

Overall, the base of the succession is characterized by Architectural Element A (Delta Front Slope Fines) that encases Architectural Element B (Delta Front Slope Gully-fill) deposits: the succession that is traditionally assigned to the “Grindslow Shales”. Beds of Architectural Element B (Delta Front Slope Gully-fill) which are incised into and downlapped onto by Architectural Element A (Delta Front Slope Fines) appear to cluster vertically in the vicinity of the logged sections at Nether Red Brook and Upper Red Brook (Fig. 3C and 3D). These may be artefacts of exposure paucity introducing sampling bias, since these are the positions of two contemporary stream gullies which were used to log thicker accessible sections. Architectural Element B (Delta Front Slope Gully-fill) could therefore be substantially more abundant than is depicted in Figure 7C. Indeed, several such elements have been identified elsewhere by Walker (1966a) and Collinson (1969).

The overlying succession of Architectural Element C (Mouth Bar Sandstone), D (Mouth Bar Fines), E (Mouth Bar Mass-failure Deposit), B (Delta Front Slope Gully-fill) and F (Fluvial Distributary Channel-fill) elements comprise what is traditionally referred to as the “Lower Kinderscout Grit”. This succession displays a marked repeated architectural motif that is described below in terms of discrete

genetic units (MB1 to MB7; Fig. 7D), starting from the oldest deposits in the east with numbers increasing as deposits young toward the west. The architecture of sets of clinoforms is commonly described in terms of the trajectory of the inflection of the clinoform topset-foreset rollover. This is possible only in parts of this succession where well-developed and preserved sigmoidal geometries are present, as in places the topset-foreset rollover has been eroded by fluvial distributary channels. However, downlap trajectories of foresets are emphasized here because they are preserved throughout this succession, facilitating comparative analysis. Downlap trajectories are shown in Figure 7D.

The oldest unit, MB1, is not well exposed, but comprises Architectural Element C (Mouth Bar Sandstone) and D (Mouth Bar Fines) with relatively well-developed clinoform rollovers. Foresets are up to 10 m tall. Both rollover and foreset downlap geometries display a basinward ascending trajectory of 5 m over 100 m horizontal distance. MB1 therefore displays a geometry indicative of both progradation and aggradation.

MB1 is presumably onlapped by Architectural Element A (Delta Front Slope Fines) to the west, because the first clinoforms of MB2 downlap a surface that is at the same stratigraphic level as the rollover of the youngest clinoform in MB1. This represents an abrupt vertical and landward step in the clinoform downlap trajectory by 10 m and 50 m, respectively. In MB2, the trajectory of clinoform downlap descends vertically by 20 m over a distance of 800 m towards the basin. Topset-foreset rollovers are well-developed in places towards the east of MB2, but become truncated towards the western portion of the unit (or were never well-developed). They demonstrate an ascending trajectory of 30 m vertical change over a 300 m distance towards the basin. Therefore, in MB2, clinoform foresets increase in height from 5 m to more than 40 m towards the basin. Additionally, the maxima of inclination increase from 10° to 22°. This is the most westward expression of rollover geometries on Ashop Edge. As such, a broadly horizontal trajectory for all clinoform rollovers west of these is presumed based on the relatively horizontal aspect of the basal erosion surface of Architectural Element F (Fluvial Distributary Channel-fill; Fig. 7C). In the eastern part of MB2, Architectural Element C (Mouth Bar Sandstone) dominates the upper foreset, only rarely extending down to the foreset base where Architectural Element D (Mouth Bar Fines) tends to dominate. Erosional surfaces that define the bases of individual Architectural Sub-element F1 (Delta-top Distributary Network) sets can be traced westward into the clinoform foresets here, implying an intimate genetic relationship between the adjacent fluvial and mouth bar facies in MB2. In the westernmost part of MB2, Architectural Element C (Mouth Bar Sandstone) beds extend to the base of the foresets. However, a distinct package of Architectural Element D (Mouth Bar Fines) also extends to the top of the clinoform foresets. The upper and lower foresets display a number of instances of rotated and slumped Architectural Element E

(Mouth Bar Mass-failure Deposit). In places, relatively steeply- and westerly-dipping planar to listric surfaces truncate foresets that are healed or onlapped by younger clinoforms. These are interpreted as gravitational faults developed close to the depositional surface, and over which the hangingwall strata were partially or fully fluidized and evacuated down-slope.

The architectural and stratigraphic pattern established in MB2 is repeated, with some variation, in MB3 to MB7. A new unit is demarked by an abrupt backstepping of the clinoform downlap trajectories (by up to 30 m vertically and 200 m in a landward direction). In the cases of MB4 and MB7, this backstepping is coincident with the development of a relatively steep basin-facing ($\sim 30^\circ$) surface, which truncates the foresets of the underlying unit, that are interpreted as near-surface gravitational faults. In both examples, the faults are onlapped at their base by Architectural Element B (Delta Front Slope Gully-fill; Fig. 7C). Within MB3 to MB7, foreset downlaps show descending trajectories in the basinward direction (of up to 30 m over a 250–500 m distance). Clinoform foresets increase in height (from <10 m to >40 m) and inclination (from as little as around 15° to around 25°) basinward. Similar distributions of Architectural Element C (Mouth Bar Sandstone) and D (Mouth Bar Fines) occur, with the former preferentially occupying the upper foreset region and the latter the lower foresets. However, intervals of Architectural Element D (Mouth Bar Fines) also intermittently extend to the top of clinoform exposures. Mass-failure elements occur in places, as well as planar to listric basin-facing faults. Fluvial strata above MB3–MB7 overlie a major throughgoing erosion surface, so a strict genetic relationship between the mouth bar and overlying distributary network cannot be proven. MB7 is offset by a late basin-facing normal fault that extends into overlying fluvial strata. The displacement across this fault has been removed from the downlap trajectories shown in Figure 7D. Large inter-stratal faults (i.e., penetrating numerous architectural elements) of this nature have been observed elsewhere in the Namurian rocks of Derbyshire, where failure of the substrate onto which the delta prograded is cited as an initiation mechanism (Chisholm, 1977; Chisholm and Waters, 2012).

Interpretation

The main architectural motif at Ashop Edge: repeated sets of mouth bar clinoforms displaying basinward-descending downlap trajectories concomitant with increasing foreset height, has been recognized elsewhere in the “Lower Kinderscout Grit” (e.g., at Bamford Edge, Derbyshire; Collinson, 1969; Hampson, 1997). Each set of basinward-thickening clinoforms (i.e., MB1 to MB7) can be interpreted as the development and progradation of a discrete mouth bar. This is because they form a distinct, relatively similar set, bounded by a break or shift in clinoform trajectory that represents an abrupt repositioning of the distributary mouth (Steel and Olsen, 2002; Bhattacharya and Giosan, 2003; Olariu and Bhattacharya, 2006; Jerolmack and Swenson, 2007). Ashop Edge therefore records the

construction history of the Kinderscout Delta via the development and evolution of at least seven discrete mouth bars.

The systematic increase in foreset height in each mouth bar represents their advance into progressively deeper water. If foreset height provides a reliable indication of water depth, then bathymetry increased from typically 5–10 m to >40 m during the lifespan of each mouth bar moving from MB2 to MB7. Characteristic post-depositional compaction values for sandstone (e.g., Dasgupta and Mukherjee, 2019) suggest these bathymetries could be as much as doubled. Nevertheless, scaling relationships (e.g., a fourfold or greater increase in water depth during the lifespan of the mouth bar) are preserved. Water depths inferred from thicknesses in this study are viewed as minima. The trajectory of mouth bar clinoform topset-foreset rollovers in MB1 (Fig. 7C) broadly tracks the position of sea-level. Therefore, the ascending rollover geometry documented in MB1 and MB2 implies that water depth increased by at least 30 m during their deposition through relative sea-level rise (possibly glacio-eustatic; e.g., Maynard and Leeder, 1992). The broadly horizontal trajectory of rollovers implied by the architecture of MB3 to MB7 suggests that deposition of these mouth bars took place during a relative sea-level stillstand. By comparison, the trajectories of foreset downlaps (Fig. 7D) are a proxy for prodelta slope bathymetry. That is, the descending trajectories track the mouth bar base building into deepening water down a bathymetric slope. This partly (for MB2), or largely (for MB3 to MB7), explains the increasing water depth inferred from increasing foreset thicknesses. Further basinward, at Kinder Downfall (Fig. 2), foresets are as tall as 50 m (Fig. 3E), demonstrating further water depth increase beyond the western end of the Ashop Edge section. The abrupt vertical and backward stepping of the clinoform downlap trajectory, which marks the deposition of a new mouth bar, indicates that aggradation of seafloor sediment occurred between the deposition of each discrete mouth bar.

Discussion

Process regime of the Kinderscout Delta at Ashop Edge

The Kinderscout Delta formed at the terminus of a major, continent-scale fluvial system sourced in the Caledonian Highlands of what is now Greenland, NE Scotland, and western Norway (Sorby, 1849; Lancaster *et al.*, 2017; Jones, 2022). The system prograded southward across the greater Pennine Basin, healing an inherited “block-and-basin” bathymetry (Collinson, 1988; Fraser and Gawthorpe, 2003). At Ashop Edge, facies and architectural analysis suggest that the local distributaries in the delta were generally 2–4 m deep, although as much as 10 m in some places, and mouth bars prograded into water depths of 40 m or more. Indeed, at Kinder Downfall (Fig. 3E), water depths at the time of foreset progradation are inferred to be as great as 50 m not accounting for post-

depositional compaction. The absence of wave- or tide-generated sedimentary structures or bidirectional flow indicators suggests that the mouth bars were not noticeably affected by basin current re-working, either at the time of deposition or after it. This is consistent with the reconstruction of a marine connection, yet low tide and wave energy, in the Edale and other sub-basins in the broader Pennine Basin network (Collinson, 1969, 1988; McCabe, 1975; Martinsen, 1990; Hampson, 1997; Wells *et al.*, 2005), although analysis of borehole data has resulted in the suggestion of a tidal influence in the less geographically restricted Craven Basin (to the NW; Fig. 1A) during this time (Aitkenhead and Riley, 1996). Overall, the delta at Kinder Scout can be classified as being river-dominated (*sensu* Galloway, 1975) and is therefore a good candidate for the reconstruction of fundamental river mouth processes from its deposits.

McCabe (1975, 1977) suggested that the large foresets (or “giant cross-beds”) exposed at Kinder Scout (the Mouth Bar Sandstone – Architectural Element C of this study), and elsewhere in the Pennine Basin, represent the downstream migration of bank-attached “alternate bars” within a fluvial channel. For this to be the case, the fluvial channel would need to have attained depths in excess of 40 m and widths of between 0.5 and 1.5 km at Kinder Scout. In this conceptual model, the beds which overlie these foresets and exhibit “medium-scale” cross-bedding (Delta-top Distributary Network – Architectural Sub-element F1 of this study) represent the migration of dunes in shallower regions of the same deep river channel. Bijkerk (2014) rejected this hypothesis for similar Carboniferous strata in the Craven Basin based on sediment calibre and flux calculations, comparing these with modern analogues. We also contest the validity of this model at Kinder Scout for the following reasons: (1) the regionally widespread nature of beds with this character is at odds with deposition in an entrenched straight <2 km wide channel; (2) the foreset dip direction data presented in this study (Fig. 2) and those of Collinson (1968) create arcuate barform geometries in plan view (as opposed to an elongate in-channel bar morphology), characteristic of a mouth bar; (3) foresets up to around 50 m in height would imply river depths of at least this magnitude in some places, deeper still if post-depositional compaction is accounted for, which is far greater than the depths of modern rivers transporting comparable grain sizes, but with far greater volumes of bedload (e.g., the Jamuna River, Bangladesh; Ashworth *et al.*, 2000; Best *et al.*, 2003; Bijkerk, 2014); (4) there is no evidence for a significant erosion surface between the “Grindslow Shales” and “Lower Kinderscout Grit” at Kinder Scout, indeed Mouth Bar Sandstone (Architectural Element C) and Delta Front Slope Fines (Architectural Element A) interfinger (Fig. 7C); (5) Delta Front Slope Gully-fill (Architectural Element B) elements are incised into both Mouth Bar Sandstone (Architectural Element C) and Delta Front Slope Fines (Architectural Element A; Figs. 3, 5A and 7) which further suggests a genetic linkage between the two lithosomes; (6) the cross-sectional geometries observed at Ashop Edge (Fig. 7C), where the level at which Mouth Bar

Sandstone (Architectural Element C) beds downlap onto Delta Front Slope Fines (Architectural Element A) repeatedly steps upward and backward, are challenging to reconcile with a fluvial environment, as is the repeated rapid basinward thickening of the foreset package (thus deepening of water) over distances in the order of hundreds of metres (e.g., MB2 to MB7, Fig. 7D), and the raising of the base level by around 30 m between MB1 and MB2; finally, (7) if the “Lower Kinderscout Grit” foresets were to represent deposition in a deep fluvial setting, then the delta which likely formed at the terminus of this system is absent from the rock record.

Collinson (1968) considered the Kinderscout Delta to represent the deposits of a “Gilbert-type” delta. As originally described by Gilbert (1885), they are characterized by planar, steeply-dipping foresets (10–35°), built from fluvial bedload delivered to the “brinkpoint”, and resedimented down the foreset via grainflow. The topset-foreset transition is marked by an angular, erosional discordance between quasi-horizontal topset strata generated by tractional processes at the immediate river mouth, and the more steeply-dipping foresets. The bases of foresets downlap at high angles onto the basal substrate. Processes ascribed to “Gilbert-type” deltas have later been extended to include other gravity-driven processes, including slides (e.g., Gobo *et al.*, 2014), slumps (e.g., Sohn *et al.*, 1997; Gobo *et al.*, 2014), debris flows (e.g., Sohn *et al.*, 1997; Longhitano, 2008; Breda *et al.*, 2009; Gobo *et al.*, 2014; Budai *et al.*, 2021 and references therein), and turbidity currents (e.g., Sohn *et al.*, 1997; Breda *et al.*, 2007; Longhitano, 2008; Breda *et al.*, 2009; Gobo *et al.*, 2014, 2015; Okazaki *et al.*, 2020; Budai *et al.*, 2021 and references therein). Mud is required to provide the cohesion needed to translate material in blocks, or in plastically deforming masses, and is typically required to generate viscous suspensions in turbidity currents. Gilbert’s (1885) original conceptual model did not, however, envisage the deposition of mud (and therefore suspended load) at the river mouth. Nonetheless, deltas dominated by autogenic cycles of primary deposition accompanied by gravity-driven secondary re-sedimentation on the foreset, with or without a component of suspended load, can be collectively termed “foreset-dominated” (*sensu* Edmonds *et al.*, 2011).

The architecture and processes recorded by the Kinderscout Delta at Ashop Edge fulfil many of the criteria of “foreset-dominated” deltas. The steep (generally around 18–25°, though up to 30°) foresets are generally planar. Across much of the exposure, the foresets display a sharp, angular disconformable relationship with overlying fluvial distributary topsets, and downlap at high angles onto the underlying substrate. Foreset deposition was dominated by a continuum of gravity processes, including high concentration turbidity currents, debris flows, grainflows, slumps, and slides. The occurrence of a mud matrix in sandy intervals, and the notable development of mud-rich foreset intervals, which in places extend to the foreset top (Mouth Bar Fines – Architectural Element D), suggest that substantial settling of suspended sediment was at least periodically a feature of river

mouth processes. On this basis, it is unlikely that the Kinderscout Delta can be considered “Gilbert-type” (*sensu* Gilbert, 1885), as proposed by Collinson (1968).

Despite the predominance of relatively planar foresets through most of Ashop Edge, there are relatively well-developed topset-foreset rollovers developed in MB1 and in the more landward clinoforms of MB2 (Fig. 7C). This geometry, at this scale and in this subaerial setting, is more characteristic of classical mouth bars built into relatively shallow water (Anell, 2024) where bed-frictional deceleration of an expanding jet at the river mouth results in a high sedimentation rate that prompts the initiation, vertical ascension, and lateral expansion of a constructional bar, where accommodation allows (Van Heerden and Roberts, 1988; Van Wagoner *et al.*, 2003; Olariu and Bhattacharya, 2006; Hoyal and Sheets, 2009; Shaw *et al.*, 2018; Daniller-Varghese *et al.*, 2020). This scenario permits the ready preservation of mouth bar topsets, since the bar has not yet grown sufficiently to interfere with, and be eroded by, the fluvial jet. Unlike in “foreset-dominated” deltas, the shallow, low-gradient setting in which these processes occur does not permit the gravitational basinward evacuation of mouth bar deposits. Consequently, the aggrading and expanding obstacle increasingly interferes with the outflow jet (e.g., Cole *et al.*, 2021) until flow is bifurcated (Elliott, 1986; Wellner *et al.*, 2005; Olariu and Bhattacharya, 2006; Edmonds and Slingerland, 2007; Fagherazzi *et al.*, 2015; Daniller-Varghese *et al.*, 2020) or deviates (Martini and Sandrelli, 2015; Jerrett *et al.*, 2016) around it, and a new mouth bar begins to form at the terminus of the repositioned channel. Reduction in flow efficiency at the river mouth can also lead to upstream avulsion (Hoyal and Sheets, 2009; Ganti *et al.*, 2016). This sequence of autogenic processes results in a network of distributary channels and mouth bars developed over a low-gradient surface that dips basinward, and which Edmonds *et al.* (2011) termed “topset-dominated” deltas.

The interpretation of development of MB1, along with the more landward part of MB2, in relatively shallow water agrees with the foreset heights observed in these strata, which imply deposition in 5–10 m of water. This can be compared to the fluvial distributaries for which we reconstruct an in-channel water depth of at least 2–4 m, possibly as much as 10 m. Moreover, the more complex interfingering of fluvial distributary and mouth bar elements in MB2, whereby the erosional surfaces defining the bases of individual Delta-top Distributary Network (Architectural Sub-element F1) sets can be traced basinward into their respective clinothem, is characteristic of “topset-dominated” deltas (Olariu and Bhattacharya, 2006; Edmonds *et al.*, 2011).

Tabular cross-beds form a component of mouth bar foreset strata in places. For example, at the western (landward) expression of MB5, where foresets downlap on to MB4, cross-beds extend 20–25 m down the entirety of the foresets (i.e., at Nether Red Brook; Fig. 3C). This implies the maintenance of tractional currents at the bed down the delta foreset. Dunes resulting from sustained

turbulent bed traction are not usually generated by turbidity/hyperpycnal currents; it is more likely that dunes were driven by the inertia of the expanding river mouth jet which remained attached to the bed to depths of 25 m or more (Fig. 8A). “Foreset-dominated” deltas are characterized by the detachment of the jet from the bed at the “brinkpoint” (i.e., the rollover; Fig. 8B), and the transport of sediment down the foreset strictly through gravitational processes. In this regard, these strata represent deposition more typical of shallow-water, frictional “topset-dominated” deltas (Fig. 8A). It is notable, once again, that the development of cross-beds occurs in the most landward expression of MB5, where foreset height implies moderate water depth of at least 20–25 m.

Overall, the Kinderscout Delta at Ashop Edge displays features typical of both “topset-dominated” and “foreset-dominated” deltas (*sensu* Edmonds *et al.*, 2011), but the latter is predominant. Within individual mouth bars, architectures characteristic of bed-friction-driven processes are constrained to the early (landward) clinothem that were building into relatively shallow (<25 m) water depth, evolving basinward into gravity-dominated processes as the mouth bars entered deeper water (Fig. 8C).

Depositional model for the Kinderscout Delta at Ashop Edge: a hybrid topset/foreset-dominated delta

Using the Kinderscout Delta as a case example, an idealized depositional model is proposed for a delta and its component mouth bars developed in a monotonically-deepening, relatively unconfined basin.

After the avulsion of a distributary into shallow water (roughly equal to distributary depth) on the margin of a gently inclined basin floor, strong bed-friction at the immediate river mouth promotes rapid deposition of bedload (Wright, 1977). A classical lens- or pear-shaped (in plan-view) mouth bar begins to aggrade and expand laterally and down-slope (e.g., Van Heerden and Roberts, 1988; Van Wagoner *et al.*, 2003; Olariu and Bhattacharya, 2006; Hoyal and Sheets, 2009; Shaw *et al.*, 2018; Daniller-Varghese *et al.*, 2020). The initial aggradation phase can generate a mounded bar-form (e.g. Cole *et al.*, 2021), but the basinward expansion of the mouth bar results in low-gradient internal foresets (Wright, 1977; Edmonds and Slingerland, 2007). Sedimentary structures indicative of bed tractional processes, such as ripple cross-lamination, will be present (Fig. 8A; e.g., Friction-dominated Mouth Bar Sandstone – Architectural Sub-element C2).

Mouth bar expansion increasingly obstructs the river mouth, promoting bed-friction and jet deceleration (Cole *et al.*, 2021). The loci of deposition migrate away from the high-energy jet core as the bar expands, which when combined with progressive progradation into deeper water, ultimately results in the detachment of the jet from the bed at the foreset (Fig. 8B). Bedload deposition becomes

localized at this detachment point, which achieves or exceeds a stable angle of repose, triggering gravitational failure. Initially, these failures are likely to be in the form of grainflows and short run-out debris flows (e.g., Gravity-dominated Mouth Bar Sandstone – Architectural Sub-element C1). In principle, the jet–bed detachment line should be manifested as an inflection between gently-dipping friction-dominated foresets, and steeply-dipping gravity-dominated foresets; that is, the topset-foreset rollover, which is characteristic of mouth bars MB1 and parts of MB2 at Ashop Edge (Fig. 7C and D). Continued mouth bar expansion and progradation into deeper water will result in the landward retreat of the jet–bed detachment line (and geometric rollover) upslope along the clinoform, until a morphology characteristic of “Gilbert-type” fan-deltas is achieved. That is, an essentially horizontal platform where the jet interacts with the bed, a sharp “brinkpoint” inflection where the jet detaches from the substrate, and a steep close-to (but lower than) angle-of-repose front. Bedload deposition dominates at the “brinkpoint”, but long-run out distances permit re-sedimentation via a suite of fluid gravitational processes down the foreset (Fig. 8B).

In “Gilbert-type” deltas as originally described, mud is largely absent, the foresets lack cohesion, and are prone to high-frequency, low-magnitude cycles of aggradation and gravitational failure. These rapid autocycles tend to generate relatively uniform sediment distributions along the foreset and at the “brinkpoint”. Sediment is efficiently evacuated from the distributary mouth largely via gravity processes, thus backwater sedimentation, a trigger for upstream avulsion and distributary relocation (e.g. Cole *et al.*, 2021), is minimal. The distributary mouth instead tends to migrate or sweep laterally across the topset surface. Major avulsions resulting in wholesale distributary relocations are likely to be uncommon, and large portions of the delta may be constructed by a single distributary channel, producing simple, laterally extensive progradational clinoform sets (e.g. Van Dijk *et al.*, 2009). This behaviour and morphology could be described as “bifurcation dominant”, rather than “avulsion dominant” (Hoyal and Sheets, 2009). For this reason, these types of deposits are not usually termed “mouth bars”, instead the term “fan delta” is commonly employed (e.g. Rees *et al.*, 2017).

By contrast, many systems (such as the Kinderscout Delta) incorporate significant amounts of clay into river mouth deposits. In most cases, the origin of this fraction will be clay flocculation generated by salinity contrasts rather than suspension-settling (Walsh and Nittrouer, 2009). The associated increased sediment cohesion stabilizes the foresets, permitting the development of mass-transport processes such as slumps and slides. This also reduces the frequency but increases the magnitude of mass-failure events (e.g., Masson *et al.*, 2006; Elverhøi, *et al.*, 2011; Silver and Dugan, 2020). At locations such as Ashop Edge, this behaviour is expressed by faults, and slump and slide deposits (Figs. 3, 5C and 7). Scaling relationships exist between the magnitude of mass-failure scars — particularly those that retreat to the topset aided by fluvial scouring — and the distributary size and

discharge (c.f. Edmonds *et al.*, 2022). Major delta-lip failures therefore have the capacity to redirect tributary channel flow, steering discharge toward the failure site (Gales *et al.*, 2018; Fig. 9). This process describes a form of avulsion which is not usually considered characteristic of “Gilbert-type” deltas, *sensu stricto*. In the Kinderscout Delta, transient flow deflections of this type starved previously active foresets of sediment, resulting in the development of thick Mouth Bar Fines (Architectural Element D) packages that extend the full height of the foreset (Fig. 7C).

The architecture and stratigraphy of the delta front foresets in the Kinderscout Delta suggest the maintenance of avulsive (rather than constantly sweeping, “Gilbert-type”) processes, even as the mouth bar becomes dominated by gravity processes at the foreset. Quasi-stationary fluvial channels are therefore able to form discrete depositional elements at their mouths (mouth bars), which can be recognized in the rock record. It is notable that each fully preserved mouth bar at Ashop Edge (i.e., MB2 to MB6; Fig. 7D) prograded into approximately the same water depth (~40 m) and achieved similar maximum foreset gradient (>20°) before abandonment. This consistency implies that these values represent the maximum *viable* gradient and height attainable by these mouth bars (*sensu* Lambe and Whitman, 1969). It seems unlikely that another intrinsic or extrinsic factor would repeatedly result in the abandonment of a mouth bar once this height and gradient were achieved. Once a mouth bar’s depth and gradient threshold is attained, sediment is either stored temporarily on the metastable foreset slope, or bypasses it entirely (e.g., Gerber *et al.*, 2008), accumulating on the unconfined prodelta or delta front slope in the form of hyperpynites, turbidites and debrites. These processes aggrade the seafloor ahead of the mouth bars, elevating the bathymetry to permit the progradation of *viable* mouth bars. Alternatively, major mass-failures could deposit slide and slump masses at the toe of the foreset, generating powerful density currents capable of souring and filling gullies on the delta front slope (Fig. 9). As discussed, major mass-failures can also trigger avulsion of the tributary and abandonment of the mouth bar.

At the delta scale, the Kinderscout Delta was constructed via the progradation of multiple mouth bars that each record a transition from shallower-water frictional to deeper-water gravitational processes, followed by bypass and/or mass-failure, and abandonment. The consequence of delta construction in this manner is that each increment of mouth bar growth increases the bathymetric gradient at the coast. Therefore, early mouth bars are bound to be more friction-dominated (Fig. 8A), and delta-scale autocycles will at this time be characterized by upstream fluvial and delta-top tributary processes: bifurcation around existing mouth bars, initiation of new distributaries, and backwater-effects-driven upstream sedimentation-triggered “choking” and avulsion (i.e., a “topset-dominated” delta; Edmonds *et al.*, 2011; Cole *et al.*, 2021). As the whole delta progrades into progressively deeper water, the component of gravity-driven processes in each successive mouth bar

will increase. Here, autocycles are related to the aggradation and gravitational failure of mouth bar foresets (i.e., a “foreset-dominated” delta; Edmonds *et al.*, 2011; Fig. 8B). Edmonds *et al.* (2011) envisaged the “foreset-dominated” endmember of their conceptual model to largely preclude delta-top avulsive processes. The latter model’s condition of relatively steep, fixed distributary outlets emerging into steep and relatively confined basins is almost universally characteristic of true “Gilbert-type” deltas. By comparison, the Kinderscout Delta represents the deposit of a major (possibly continent-scale) river system (Waters and Davies, 2006; Jones, 2022) emerging into a relatively unconfined basin. In this case, we show that, despite the predominance of gravitational processes, avulsion leads to the deposition of a delta composed of multiple mouth bars. We propose an autogenic mechanism that links the low-frequency, high-magnitude mass-failure of relatively cohesive mouth bar foresets to the repositioning of distributaries and associated mouth bars (Fig. 9). The maintenance of sedimentation loci for sustained periods at the mouths of distributaries will result in backwater processes (e.g., Hoyal and Sheets, 2009; Ganti *et al.*, 2016; Jerrett *et al.*, 2016; Cole *et al.*, 2021). Consequently, upstream avulsion is likely also to be important in deltas of this nature.

Wider implications of this model for the interpretation of deltaic strata

A key implication of this model is that autogenic shoreline behaviour in active deltas prograding into (bathymetrically) deepening basins may be governed by coupled delta-top and delta-front processes, with distributary avulsion, foreset instability, and sediment bypass operating together. Additionally, subsurface facies architectures in these settings are likely to depart from those predicted by either topset- or foreset-dominated endmember models, producing hybrid stratigraphic successions in which fluvial, mouth-bar, mass-transport, and slope deposits are more intimately associated than conventional models would predict. Consequently, there may be similar deposits both at outcrop and in the subsurface elsewhere, deposited under comparable climatic conditions and in comparable tectonic regimes, which as a result of this revised model require revisiting and perhaps reinterpretation. Since the strata studied herein have been pivotal to current understanding of ancient delta deposits and the generation of depositional models for these settings globally, our reappraisal brings the interpretation of this classic and well-studied example in line with current understanding of modern gravity flow processes and delta systems.

Relationship of the “Grindslow Shales” to the “Lower Kinderscout Grit”

The stratigraphic significance of the surface onto which deltaic strata of the Kinderscout Delta downlap has been interpreted in different ways by previous workers. Walker (1966b) and McCabe (1978) considered the entire succession from the “Mam Tor Sandstones” to the “Lower Kinderscout

Grit” to represent a single regressive Waltherian continuum from basin floor fans, through a slope and slope channels, to a coastal delta. However, Collinson (1969) observed major channel-form elements in the upper part of the “Grindslow Shales” and interpreted these as fluvial distributaries set within finer interdistributary bay sediments. Collinson’s (1967, 1969) conceptual model required a major episode of relative sea-level rise to accommodate the deposition of the Kinderscout Delta, and it was therefore implicit that the “Grindslow Shales” to “Lower Kinderscout Grit” contact represented a drowned, formerly fluvial plain. McCabe (1978) re-interpreted the major channel-forms and associated coarse-grained strata in the entire “Grindslow Shales” as slope channels set within their finer-grained overbanks and did not see the need to invoke a major transgression. McCabe (1978) therefore saw the “Grindslow Shales” to “Lower Kinderscout Grit” contact as representing a west or SW-inclined submarine slope onto which the Kinderscout Delta prograded. Spectral gamma log signatures and palynological assemblages within these strata support a marine origin for the “Grindslow Shales” (Davies and McLean, 1996). Developing Collinson’s (1969) hypothesis for the relationship between the “Upper Grindslow Shales” channel-form sandbodies and the “Lower Kinderscout Grit”, Hampson (1997) noted their scale and multi-storey architecture and proposed that they were fluvially excavated during an episode of relative sea-level fall, then backfilled following an episode of relative sea-level rise that accommodated the Kinderscout Delta.

Our interpretations support McCabe (1978) and Davies and McLean’s (1996) conceptual models of “Grindslow Shales” deposition on a delta front slope with channel- or gully-fills which were genetically linked to the parent Kinderscout Delta (Fig. 9). Collinson (1969) and Hampson’s (1997) interpretations largely rest on the presence of cross-beds within the channel-fills, and on a shallow-water ichnofacies interpretation of *Planolites*, *Chondrites*, *Cochlichnus* and *Lockeia* in the fine-grained sediments that they interpret as interdistributary deposits. However, although present, cross-strata in the channel-fills are subordinate and, as McCabe (1978) recognized, the majority of facies are represented by high concentration turbidity current deposits and debrites (e.g. Fig. 3A, 3C and 3H). Trace fossils are also rare in the fines, and definitive evidence for subaerial exposure of what were suggested to be interdistributary regions, such as pedogenic features, is absent. Architectural and sedimentary log data presented here suggest that the upper part of the “Grindslow Shales” were deposited in water depths of 50 m or more. The repeated transport of oxygenated waters in gravity and turbidity flows has been shown to permit benthic colonisation at substantially greater water depths than this (e.g., Orr, 1995; Wetzel and Uchman, 2001; Rodríguez-Tovar *et al.*, 2010; Starek and Šimo, 2015; Botzolis *et al.*, 2022; Hovikoski *et al.*, 2025), so we do not see the occurrence of trace fossils as inconsistent with McCabe’s (1978), and our own, interpretations. The upward increase in bioturbation within the “Grindslow Shales” that Hampson (1997) records likely resulted from the increase in

delivery of oxygenated, nutrient-rich sediment reflecting increasing proximity to the advancing Kinderscout Delta. Finally, our architectural data does not show that the Kinderscout Delta downlaps a single planar stratigraphic surface (e.g. Collinson, 1967). Instead, Mouth Bar Sandstone (Architectural Element C) packages are, in places, onlapped by Delta Front Slope Fines (Architectural Element A) and Delta Front Slope Gully-fill (Architectural Element B) strata that characterize the “Grindslow Shales”, thus the base of the mouth bar sediments is undulating and the strata interfinger (Fig. 7C). High proportions of flocculated clay (or “fluid mud”) suspended within the fluvial input jet, as well as within the mouth bar deposits which are observed to have undergone mass-failure, likely increased the potential for hyperpycnal currents to travel great distances across the shelf, as is observed in active (e.g. the Amazon Delta, e.g., Sternberg *et al.*, 1996) and recent (e.g., the Orinoco Delta, e.g., Osman *et al.*, 2024) systems. Given that bathymetries are thought to have been in the region of 100–200 m during deposition of the upper part of the Bowland Shale Formation (Holdsworth and Collinson, 1988; Davies, 2008; Emmings *et al.*, 2020, 2024), and that we interpret a crudely rising relative sea-level during the Kinderscoutian Substage, we see the Hebden Formation as plausibly part of a conformable, regressive Waltherian continuum.

Hampson (1997) notes a minor ($<5^\circ$) increase in the inclination of bedding within the “Grindslow Shales” coincident with an increase in bioturbation and the proportion of sand-grade material. This transition marks the boundary between Hampson’s (1997) “Lower” and “Upper Grindslow Shales”. Although speculative, this surface may represent the downlap of low-angle clinoforms of a submarine delta (*sensu* Bates, 1953) or hyperpycnal subaqueous delta (*sensu* Zavala *et al.*, 2021, 2024) generated by gravity flows running-out ahead of the Kinderscout Delta mouth bars onto the delta front. As previously noted, the upward increase in bioturbation may represent an increase in oxygenated current input associated with these gravity flows. Therefore, the “Grindslow Shales” and “Lower Kinderscout Grit” represent components of a compound delta system that records progradation from subaqueous delta front “intrashelf” clinoforms (the “Grindslow Shales”) to proximal and subaerial mouth bars and distributaries (the “Lower Kinderscout Grit”; e.g., Hodgson *et al.*, 2017).

Conclusions

The Kinderscout Delta was fed by a major continent-scale fluvial system draining into the relatively unconfined, yet low-energy, marine Edale Sub-basin during the late Namurian. Facies and architectural analyses indicate that distributary channels were 2–4 m (locally 10 m) deep, and their associated mouth bars prograded into water depths of ~50 m or more. The delta experienced a largely river-dominated depositional regime. Mouth bar architecture records a continuum from shallow-water, bed-friction-dominated processes in early, landward clinoforms, to gravity-driven process dominance

as the delta front built out into deeper shelf water. Bypass and mass-failure of the oversteepened mouth bar terminated progradation; sediment in these circumstances may have been fed more-or-less directly to the prodelta or delta front region. Major mass-failure of oversteepened mouth bars may also have steered the avulsion of delta-top distributaries, effectively leading to mouth bar abandonment and the formation of a new bar at the mouth of the repositioned distributary. Recent summaries of delta processes distinguish between two endmember delta types based upon basin physiography and the locus of autogenic behaviour: “topset-dominated” deltas in shallow, low-gradient settings where autogenic behaviour is driven by distributary avulsive processes, and “foreset-dominated” deltas in steep, deeper-water settings, where autocyclicity is driven by gravitational processes at the foreset and avulsive processes are negligible. Here, we show that deltas building into progressively deeper water can hybridize both types of autocyclic behaviour. Moreover, we tentatively suggest that the onset of gravity-driven processes at the foreset and delta-top avulsion episodes are not necessarily mutually exclusive, and may have a cause-and-effect relationship.

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Author contributions

JH: conceptualization (equal), data collection (lead), data curation (lead), formal analysis (lead), investigation (lead), methodology (lead), visualization (lead), writing – original draft (lead), writing – review and editing (equal); **RMJ:** conceptualization (equal), data collection (supporting), formal analysis (supporting), investigation (supporting), methodology (supporting), visualization (supporting), writing – original draft (supporting), writing – review and editing (equal); **IAK:** conceptualization (equal), data collection (supporting), investigation (supporting), methodology (supporting), writing – review and editing (supporting); **ELS:** data collection (supporting), data curation (supporting), methodology (supporting).

Competing interests

The authors declare that they have no known competing financial interests or personal relationships that could have appeared to influence the work reported in this article.

Data availability

All data generated or analysed during this study are included in this published article.

ACCEPTED MANUSCRIPT

Figure captions

Figure 1. (A) Tectonic and palaeogeographical framework of the broader Pennine Basin Network in the mid Carboniferous, including the approximate location of the Kinder Scout study area (white star). Adapted from Fraser and Gawthorpe (2003) and Waters *et al.* (2009). The spatial extent of Figure 1B is shown by a black box. Abbreviations: SUH = Southern Uplands High, BB = Bowland Basin, CLH = Central Lancashire High, MWH = Market Weighton High, HB = Hull Basin, GT = Gainsborough Trough, EG = Edale Gulf, DH = Derbyshire High, WG = Widmerpool Gulf, EMS = East Midlands Shelf. **(B)** The location of the Kinder Scout study area (white star) within the southern Pennines, England. Notable population centres are shown in grey, major roads are indicated by black lines and the area covered by the Peak District National Park is shown in green. Abbreviations: H'field = Huddersfield; M'chester = Greater Manchester. **(C)** Chronostratigraphy and lithostratigraphy of part of the Carboniferous fill of the Pennine Basin in the area of the Kinder Scout plateau, based on Stevenson and Gaunt (1971) and Waters *et al.* (2009), recalibrated to Aretz *et al.* (2020). Inset, detail of the stratigraphy of the upper part of the Bowland Shale Formation and Hebden Formation, including informal lithostratigraphic subdivisions. Abbreviations: Fm. = Formation, Gp. = Group, Carb. = Carboniferous, Ks = Kinderscoutian, Wp = Westphalian.

Figure 2. A facies-based recoloration of a geological map of the Kinder Scout plateau by Stevenson and Gaunt (1971), where units are demarcated by interpreted depositional environment as opposed to geological unit name. The area that this map covers is broadly highlighted by the white star in Figures 1A and B. Mouth bar clinoform strike and dip, and palaeocurrent direction data collected in this study are shown. The locations of the eight measured sections (Fig. 3) and location of the Ashop Edge virtual outcrop model section (Fig. 7) are highlighted. Contains British Geological Survey materials © UKRI 2026.

Figure 3. Detail of eight measured sections through the “Lower Kinderscout Grit” gathered from stream gullies and accessible cliffs around the periphery of the Kinder Scout plateau. They are presented as two east-west (up- to down-system from left to right) transects which roughly track the northern (A-E) and southern (F-H) edges of the plateau. The three measured sections that lie along the Ashop Edge virtual outcrop model (Fig. 7) are highlighted. Sections are named according to nearby geographical features; often the modern-day stream gully section where the log was collected. Palaeocurrent direction and mouth bar clinoform strike and dip measurements collected in this study are shown alongside the bed where they were measured, as well as general relative dip (i.e., broadly horizontal versus inclined). The horizon of “truncation by distributaries” marks the base of what is interpreted as

the Architectural Element F (Fluvial Distributary Channel-fill) and, though this surface is unlikely to be planar, it has been used to correlate between logged sections where present. Lithofacies abbreviations as per Table 1. Architectural element abbreviations: A = Architectural Element A (Delta Front Slope Fines), B = Architectural Element B (Delta Front Slope Gully-fill), C = Architectural Element C (Mouth Bar Sandstone), D = Architectural Element D (Mouth Bar Fines), E = Architectural Element E (Mouth Bar Mass-failure Deposit), F = Architectural Element F (Fluvial Distributary Channel-fill).

Figure 4. Representative photographs of the lithofacies identified in this study area (Table 1). **(A)** Massive siltstones and rare claystones (M), **(B)** Massive to normally-graded very fine to medium-grained sandstone (SstT), **(C)** Ripple cross-laminated sandstone (SstR), **(D)** Planar cross-bedded sandstone (SstXBp), **(E)** Trough cross-bedded sandstone (SstXBt), **(F)** Plane-bedded, well-sorted sandstone (SstPW), with arrows highlighting concentrations of texturally mature outsized quartz clasts which create the “pinstripe” fabric, **(G)** Plane-bedded, poorly-sorted sandstone (SstPM), **(H)** Massive or inverse-graded sandstone (SstMI), **(I)** Dewatered sandstone (SstDW).

Figure 5. Representative photographs of architectural elements recognized in this study. White lines have been used to accentuate the geometry of major bedding surfaces. **(A)** Architectural Element B (Delta Front Slope Gully-fill) cutting into Architectural Element C (Mouth Bar Sandstone) and Architectural Element A (Delta Front Slope Fines) at Kinder Downfall. Inset: the erosional base of the gully. This image is of the southern wall of the Kinder Downfall gully and shows laterally equivalent strata to those logged on the northern wall of the gully between 0–17.7 m in Figure 3E. **(B)** Inclined Architectural Element C (Mouth Bar Sandstone) strata exposed at Kinder Downfall. There is an erosional surface which caps these deposits, above which Architectural Element F (Fluvial Distributary Channel-fill) deposits are present. Toward the top of the image, a second generation of inclined Architectural Element C (Mouth Bar Sandstone) strata can be seen downlapping onto Architectural Element F (Fluvial Distributary Channel-fill). These strata were likely emplaced by the vertical and backstepping (i.e. transgression) of the topset–foreset rollover, thus they represent a separate discrete younger mouth bar. This mouth bar, too, is truncated at its top by distributary channel deposits. This image is of the southern wall of the Kinder Downfall gully, though the strata pictured are broadly representative of those logged on the northern wall of the gully which are shown in Figure 3E from around 40 m upwards. **(C)** A rotational Architectural Element E (Mouth Bar Mass-failure Deposit) within the Ashop Edge section. Massively-bedded, disturbed-looking strata are back-rotated above a listric surface which dips basinward at a steeper angle than the foresets it truncates. At the top contact of this element, Architectural Element C (Mouth Bar Sandstone) beds downlap onto the mass-moved

material. Due to the inaccessibility of the cliff section, this particular element does not feature in any of the logged sections presented in Figure 3, though similar strata are logged elsewhere. **(D)** Overview of the geometry of a channelized Architectural Sub-element F2 (Distributary Channel Mouth) at Ashop Edge. An erosional contact with the underlying Architectural Element C (Mouth Bar Sandstone) is marked by the red line, above which around 3 m of horizontally-bedded SstPW facies are present. This particular image shows the strata from around 34–37 m in Figure 3B. Inset: the erosional base of this element.

Figure 6. Example lithofacies stacking patterns within each architectural element identified in this study. The example logs are taken from data presented in Figure 3, and are all displayed at the same scale. The depositional setting of each architectural element is highlighted in Figure 9.

Figure 7. (A) and (B) Two parts of the virtual outcrop model and interpretative linework used to make architectural sketches shown in parts (C) and (D) of this figure. **(C)** Architecture of Ashop Edge. The positions of the three sedimentary logs collected along the Edge at Fairbrook Naze (Fig. 3B), Nether Red Brook (Fig. 3C) and Upper Red Brook (Fig. 3D) are also shown. **(D)** Summary of the architecture (corrected for post-depositional dip) and trajectory of clinoform foreset downlap. Note that displacement by a late fault in MB7 has been restored in the clinoform foreset downlap trajectory.

Figure 8. Relative roles of friction and gravity processes in deltas. **(A)** “Topset-dominated” endmember of Edmonds *et al.* (2011) developed where the depth of the water in the receiving basin (d_b) is comparable to that of the distributary channel (d_c). The jet (or plume) remains attached to the whole length of a relatively low-gradient mouth bar foreset, driving sediment transport and bedform migration down the foreset. **(B)** “Foreset-dominated” endmember of Edmonds *et al.* (2011) developed where the depth of the water in the receiving basin (d_b) far exceeds that of the distributary channel (d_c). The jet detaches from the bed at the “brinkpoint” where sediment is thus deposited, and is transported down an angle-of-repose foreset via gravitational processes. The buoyant plume is lofted toward the top of the water column. **(C)** A proposed transitional style (e.g., the Kinderscout Delta at Ashop Edge) developed where the depth of the water in the receiving basin (d_{b1}) is initially comparable to that of the distributary channel (d_c) but deepens basinward (d_{b2}) to far exceed that of the distributary. The mouth bar thus evolves in its processes and geometry from “topset-dominated” to “foreset-dominated”. Crude grainsize trends at points along the delta are shown by white triangles.

Figure 9. Schematic diagrams demonstrating the proposed mechanisms for the emplacement of architectural elements and geological units recognized in this study. The depositional settings of architectural elements are labelled, and their deposits are coloured as per Figure 3. Note that the contact between the “Grindslow Shales” delta front and “Lower Kinderscout Grit” deltaic mouth bars is shown as planar for simplicity only, and in fact these two units interfinger (see Fig. 7C) representing the transition between the mouth bar foreset and the prodelta slope. **(A)** The depositional oversteepening of a “foreset-dominated” mouth bar “brinkpoint” sediment wedge results in the dewatering and mass-failure of Mouth Bar Sandstone (Architectural Element C) deposits, which are insufficiently cohesive to support the load atop them at such steep slope angles. The sediments are translated down-slope above a listric slip surface as intact slide blocks and incoherent slumped material, further amalgamated by fluid escape. These mass-movement events trigger erosive high concentration turbidity currents and/or debris flows that travel basinward down the foreset, carving high-relief, low-sinuosity gullies, and onto the delta front slope where they lose momentum and terminate in a fan lobe (i.e. a turbidite-fronted delta). The sediments infilling these erosive gullies, Delta Front Slope Gully-fill (Architectural Element B) elements, freeze en-masse or exhibit signs of backfilling behaviour, although there is evidence for subsequent reactivation of these gullies as sediment conduits in the form of internal nested erosion surfaces within the master basal erosion surface. These currents may, under especially high discharge conditions, be driven by hyperpycnal flows derived directly from the river mouth itself. Under these conditions, the mouth bar is largely bypassed as a depocenter. **(B)** Following mass-wasting, headwall retreat of the fault scarp in a landward direction, aided by fluvial scouring, provides the incoming jet with a new path of least resistance through the obstructing mouth bar. Here we tentatively suggest that, as a result, avulsion occurs at the river mouth, the old mouth bar and terminal distributaries are abandoned, and a new mouth bar forms at the relocated point of flow deconfinement (i.e. the river mouth). Mouth bar sediments downlap onto and heal the scoured topography created by the mass-movement event. **(C)** Once the scoured mouth bar slope topography is healed, new foresets descend the entire length of the foreset slope, covering the slope gully sediments. These gullies are preserved within the proximal mouth bar and more distal prodelta region of the delta front (the underlying “Grindslow Shales”) as Delta Front Slope Gully-fill (Architectural Element B) elements. **(D)** A cross-sectional conceptual illustration of the Kinderscout Delta system (i.e., the Hebden Formation) as a whole, placing the Kinderscout Delta mouth bars and distributary channels (the “Lower Kinderscout Grit”) in the context of the delta front “Grindslow Shales”, and proximal and distal slope and basin-floor fan turbidites of the “Shale Grit” and “Mam Tor Beds” respectively. Note that, for clarity, thicknesses, distances and water

depths in this illustration are not to scale. Abbreviations: “Lr KG” = “Lower Kinderscout Grit”; “GS” = “Grindslow Shales”; “SG” = “Shale Grit”; “MTB” = “Mam Tor Beds”.

Table captions

Table1. Lithofacies of the studied deposits.

References

- Aitkenhead, N. and Riley, N.J. (1996). Kinderscoutian and Marsdenian successions in the Bradup and Hag Farm boreholes, near Ilkley, west Yorkshire. *Proceedings of the Yorkshire Geological Society*, 51(2), pp.115–125. doi:<https://doi.org/10.1144/pygs.51.2.115>.
- Allen, J.R.L. (1960). The Mam Tor sandstones, a ‘turbidite’ facies of the Namurian deltas of Derbyshire, England. *Journal of Sedimentary Research*, 30(2), pp.193–208. doi:<https://doi.org/10.1306/74d709fc-2b21-11d7-8648000102c1865d>.
- Allen, J.R.L. (1982). Orientation of Particles During Sedimentation: Shape-fabrics. In: J. R. L. Allen, *Sedimentary structures, their character and physical basis Volume 1*. Elsevier, pp.179–235.
- Anell, I. (2024). The quintessential s-shape in sedimentology: A review on the formation and controls of clinoform shape. *Earth-Science Reviews*, 254, p.104821.
- Anita, D.D.J. and Wood, B.A. (1977). Some aspects of the palaeoecology of the E1b zone of the Edale Shales (Namurian) in the Edale Valley, Derbyshire. *Mercian Geologist*, 6(3), pp.179–196.
- Aretz, M.R., Herbig, H.-G., Wang, X., Gradstein, F.M., Agterberg, F.G. and Ogg, J.G. (2020). The Carboniferous period. In: F.M. Gradstein, J.G. Ogg, M.D. Schmitz and G.M. Ogg, eds., *The Geological Time Scale 2020*. Amsterdam: Elsevier, pp.811–874.
- Ashworth, P.J., Best, J.L., Roden, J.E., Bristow, C.S. and Klaassen, G.J. (2000). Morphological evolution and dynamics of a large, sand braid-bar, Jamuna River, Bangladesh. *Sedimentology*, 47(3), pp.533–555. doi:<https://doi.org/10.1046/j.1365-3091.2000.00305.x>.

Bates, C.C. (1953). Rational theory of delta formation. *AAPG Bulletin*, 37(9), pp.2119–2162. doi:<https://doi.org/10.1306/5CEADD76-16BB-11D7-8645000102C1865D>.

Best, J.L., Ashworth, P.J., Bristow, C.S. and Roden, J. (2003). Three-dimensional sedimentary architecture of a large, mid-channel sand braid bar, Jamuna River, Bangladesh. *Journal of Sedimentary Research*, 73(4), pp.516–530. doi:<https://doi.org/10.1306/010603730516>.

Bhattacharya, J.P. and Giosan, L. (2003). Wave-influenced deltas: geomorphological implications for facies reconstruction. *Sedimentology*, 50(1), pp.187–210. doi:<https://doi.org/10.1046/j.1365-3091.2003.00545.x>.

Bhattacharya, J.P. and Walker, R.G. (1992). Deltas. In: R.G. Walker and N.P. James, eds., *Facies Models: Response to Sea-Level Change*. St. John's: Geological Association of Canada, pp.157–177.

Bijkerk, J.F. (2014). *External controls on sedimentary sequences: a field and analogue modelling-based study*. PhD Thesis, University of Leeds, pp.182–183.

Botziolis, C., Maravelis, A.G., Uchman, A. and Zelilidis, A. (2022). Trace fossils from Upper Eocene to Lower Oligocene deep-sea deposits of the foreland Pindos Basin, western Greece. In: *Bulletin of the Geological Society of Greece*. 16th International Congress of the Geological Society of Greece, pp.626–627.

Braathen, A., Midtkandal, I., Mulrooney, M.J., Appleyard, T.R., Haile, B.G. and van Yperen, A.E. (2018). Growth-faults from delta collapse - structural and sedimentological investigation of the Last Chance delta, Ferron Sandstone, Utah. *Basin Research*, 30(4), pp.688–707. doi:<https://doi.org/10.1111/bre.12271>.

Bradley, R.W. and Venditti, J.G. (2017). Reevaluating dune scaling relations. *Earth-Science Reviews*, 165, pp.356–376. doi:<https://doi.org/10.1016/j.earscirev.2016.11.004>.

Breda, A., Mellere, D. and Massari, F. (2007). Facies and processes in a Gilbert-delta-filled incised valley (Pliocene of Ventimiglia, NW Italy). *Sedimentary Geology*, 200(1-2), pp.31–55. doi:<https://doi.org/10.1016/j.sedgeo.2007.02.008>.

Breda, A., Mellere, D., Massari, F. and Asioli, A. (2009). Vertically stacked Gilbert-type deltas of Ventimiglia (NW Italy): The Pliocene record of an overfilled Messinian incised valley. *Sedimentary Geology*, 219(1-4), pp.58–76.

doi:<https://doi.org/10.1016/j.sedgeo.2009.04.010>.

Buckley, S.J., Ringdal, K., Naumann, N., Dolva, B., Kurz, T.H., Howell, J.A. and Dewez, T.J.B. (2019). LIME: Software for 3-D visualization, interpretation, and communication of virtual geoscience models. *Geosphere*, 15(1), pp.222–235. doi:<https://doi.org/10.1130/ges02002.1>.

Budai, S., Colombera, L. and Mountney, N.P. (2021). Quantitative characterization of the sedimentary architecture of Gilbert-type deltas. *Sedimentary Geology*, 426, p.106022. doi:<https://doi.org/10.1016/j.sedgeo.2021.106022>.

Chisholm, J.I. (1977). Growth faulting and sandstone deposition in the Namurian of the Stanton Syncline, Derbyshire. *Proceedings of the Yorkshire Geological Society*, 41(3), pp.305–323. doi:<https://doi.org/10.1144/pygs.41.3.305>.

Chisholm, J.I. and Waters, C.N. (2012). Syn-sedimentary deformation of the Ashover Grit (Pennsylvanian, Namurian, Marsdenian Substage) deltaic succession around Wirksworth, Derbyshire, UK. *Proceedings of the Yorkshire Geological Society*, 59(1), pp.25–36. doi:<https://doi.org/10.1144/pygs.59.1.303>.

Cole, G., Jerrett, R. and Watkinson, M.P. (2021). A stratigraphic example of the architecture and evolution of shallow water mouth bars. *Sedimentology*, 68(3), pp.1227–1254. doi:<https://doi.org/10.1111/sed.12825>.

Coleman, L.M. and Gagliano, S.M. (1964). Cyclic sedimentation in the Mississippi River deltaic plain. *Gulf Coast Association of Geological Societies Transactions*, 14, pp.67–80.

Collinson, J.D. (1967). *Sedimentation of the southern part of the Grindslow shales and the Kinderscout grit*. PhD Thesis, University of Oxford, pp. 1–12.

Collinson, J.D. (1968). Deltaic sedimentation units in the Upper Carboniferous of northern England. *Sedimentology*, 10(4), pp.233–254. doi:<https://doi.org/10.1111/j.1365-3091.1968.tb00833.x>.

Collinson, J.D. (1969). The sedimentology of the Grindslow Shales and the Kinderscout Grit: A deltaic complex in the Namurian of northern England. *SEPM Journal of Sedimentary Research*, Vol. 39. doi:<https://doi.org/10.1306/74d71c17-2b21-11d7-8648000102c1865d>.

Collinson, J.D. (1988). Controls on Namurian sedimentation in the Central Province basins of northern England. In: B.M. Besly and G. Kelling, eds., *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe*. Glasgow: Blackie, pp.85–101.

Daniller-Varghese, M.S., Kim, W. and Mohrig, D.C. (2020). The effect of flood intermittency on bifurcations in fluviodeltaic systems: Experiment and theory. *Sedimentology*, 67(6), pp.3055–3066. doi:<https://doi.org/10.1111/sed.12732>.

Dasgupta, T. and Mukherjee, S. (2019). Compaction of sediments and different compaction models. In: T. Dasgupta and S. Mukherjee, eds., *Sediment compaction and applications in petroleum geoscience*. Switzerland: Springer, pp.1–8.

Davies, S.J. (2008). The record of Carboniferous sea-level change in low-latitude sedimentary successions from Britain and Ireland during the onset of the late Paleozoic ice age. In: C.R. Fielding, T.D. Frank and J.L. Isbell, eds., *Special Paper 441: Resolving the Late Paleozoic Ice Age in Time and Space*. Geological Society of America, pp.187–204.

Davies, S.J. and McLean, D.I. (1996). Spectral gamma-ray and palynological characterization of Kinderscoutian marine bands in the Namurian of the Pennine Basin. *Proceedings of the Yorkshire Geological Society*, 51(2), pp.103–114. doi:<https://doi.org/10.1144/pygs.51.2.103>.

Dixon, J.F., Steel, R.J. and Olariu, C. (2013). A model for cutting and healing of deltaic mouth bars at the shelf edge: Mechanism for basin-margin accretion. *Journal of Sedimentary Research*, 83(3), pp.284–299. doi:<https://doi.org/10.2110/jsr.2013.25>.

Dunne, L.A. and Hempton, M.R. (1984). Deltaic sedimentation in the Lake Hazar pull-apart basin, south-eastern Turkey. *Sedimentology*, 31(3), pp.401–412. doi:<https://doi.org/10.1111/j.1365-3091.1984.tb00868.x>.

Edmonds, D., Chadwick, A.J., Lamb, M.P., Lorenzo-Trueba, J., Murray, B., Nardin, W., Salter, G. and Shaw, J.B. (2022). 10.09 – Morphodynamic modeling of river-dominated deltas: A

review and future perspectives. In: J. F. Shroder, *Treatise on Geomorphology (Second Edition)*. Academic Press, pp.110–140. doi:<https://doi.org/10.1002/essoar.10507512.1>.

Edmonds, D.A., Shaw, J.B. and Mohrig, D. (2011). Topset-dominated deltas: A new model for river delta stratigraphy. *Geology*, 39(12), pp.1175–1178. doi:<https://doi.org/10.1130/g32358.1>.

Edmonds, D.A. and Slingerland, R.L. (2007). Mechanics of river mouth bar formation: Implications for the morphodynamics of delta distributary networks. *Journal of Geophysical Research*, 112(2), pp.1–14. doi:<https://doi.org/10.1029/2006jf000574>.

Elliott, T. (1986). Deltas. In: H.G. Reading, ed., *Sedimentary Environments and Facies*. Oxford: Blackwell Scientific Publications, pp.113–154.

Elverhøi, A., Breien, H., De Blasio, F.V., Harbitz, C.B. and Pagliardi, M. (2010). Submarine landslides and the importance of the initial sediment composition for run-out length and final deposit. *Ocean Dynamics*, 60(4), pp.1027–1046. doi:<https://doi.org/10.1007/s10236-010-0317-z>.

Emmings, J.F., Davies, S.J., Vane, C.H., Moss-Hayes, V. and Stephenson, M.H. (2020). From marine bands to hybrid flows: Sedimentology of a Mississippian black shale. *Sedimentology*, 67(1), pp.261–304. doi:<https://doi.org/10.1111/sed.12642>.

Emmings, J.F., Hennissen, J.A.I., Vane, C.H., Damaschke, M., Marvin, L., Moss-Hayes, V., Lamb, A., Lacey, J., Leng, M.J. and Riley, N.J. (2024). The Bowland Shale Formation in the Blacon Basin: syngenetic processes, stacking patterns and heat productivity. *Geological Society London Special Publications*, 534(1), pp.61–90. doi:<https://doi.org/10.1144/sp534-2022-262>.

Fagherazzi, S., Edmonds, D.A., Nardin, W., Leonardi, N., Canestrelli, A., Falcini, F., Jerolmack, D.J., Mariotti, G., Rowland, J.C. and Slingerland, R.L. (2015). Dynamics of river mouth deposits. *Reviews of Geophysics*, 53, pp.642–672.

Falk, P.D. and Dorsey, R.J. (1998). Rapid development of gravelly high-density turbidity currents in marine Gilbert-type fan deltas, Loreto Basin, Baja California Sur, Mexico. *Sedimentology*, 46(4), pp.331–349. doi:<https://doi.org/10.1046/j.1365-3091.1998.0153e.x>.

Fraser, A.J. and Gawthorpe, R.L. (1990). Tectono-stratigraphic development and hydrocarbon habitat of the Carboniferous in northern England. *Geological Society, London, Special Publications*, 55(1), pp.49–86. doi:<https://doi.org/10.1144/gsl.sp.1990.055.01.03>.

Fraser, A.J. and Gawthorpe, R.L. (2003). *Atlas of Carboniferous Basin Evolution in Northern England*. Geological Society of London.

Frey Martinez, J., Cartwright, J. and Hall, B. (2005). 3D seismic interpretation of slump complexes: examples from the continental margin of Israel. *Basin Research*, 17(1), pp.83–108. doi:<https://doi.org/10.1111/j.1365-2117.2005.00255.x>.

Gales, J.A., Talling, P.J., Cartigny, M.J.B., Hughes Clarke, J., Lintern, G., Stacey, C. and Clare, M.A. (2018). What controls submarine channel development and the morphology of deltas entering deep-water fjords? *Earth Surface Processes and Landforms*, 44(2), pp.535–551. doi:<https://doi.org/10.1002/esp.4515>.

Galloway, W.E. (1975). Process framework for describing the morphologic and stratigraphic evolution of deltaic depositional systems. In: M.L. Broussard, ed., *Deltas: Models for Exploration*. Houston Geological Society, pp.87–98.

Ganti, V., Chadwick, A.J., Hassenruck-Gudipati, H.J. and Lamb, M.P. (2016). Avulsion cycles and their stratigraphic signature on an experimental backwater-controlled delta. *Journal of Geophysical Research: Earth Surface*, 121(9), pp.1651–1675. doi:<https://doi.org/10.1002/2016jf003915>.

Gerber, T.P., Pratson, L.F., Wolinsky, M.A., Steel, R., Mohr, J., Swenson, J.B. and Paola, C. (2008). Clinoform progradation by turbidity currents: Modeling and experiments. *Journal of Sedimentary Research*, 78(3), pp.220–238. doi:<https://doi.org/10.2110/jsr.2008.023>.

Gilbert, G.K. (1885). *The Topographic Features of Lake Shores*. Department of the Interior – U. S. Geological Survey, pp.104–108.

Gobo, K., Ghinassi, M. and Nemec, W. (2014). Reciprocal changes in foreset to bottomset facies in a Gilbert-type delta: Response to short-term changes in base level. *Journal of Sedimentary Research*, 84(11), pp.1079–1095. doi:<https://doi.org/10.2110/jsr.2014.83>.

Gobo, K., Ghinassi, M. and Nemec, W. (2015). Gilbert-type deltas recording short-term base-level changes: Delta-brink morphodynamics and related foreset facies. *Sedimentology*, 62(7), pp.1923–1949. doi:<https://doi.org/10.1111/sed.12212>.

Hage, S., Cartigny, M.J.B., Sumner, E.J., Clare, M.A., Hughes Clarke, J.E., Talling, P.J., Lintern, D.G., Simmons, S.M., Silva Jacinto, R., Vellinga, A.J., Allin, J.R., Azpiroz-Zabala, M., Gales, J.A., Hizzett, J.L., Hunt, J.E., Mozzato, A., Parsons, D.R., Pope, E.L., Stacey, C.D. and Symons, W.O. (2019). Direct monitoring reveals initiation of turbidity currents from extremely dilute river plumes. *Geophysical Research Letters*, 46(20), pp.11310–11320. doi:<https://doi.org/10.1029/2019gl084526>.

Hampson, G.J. (1997). A sequence stratigraphic model for deposition of the Lower Kinderscout Delta, an Upper Carboniferous turbidite-fronted delta. *Proceedings of the Yorkshire Geological Society*, 51(4), pp.273–296. doi:<https://doi.org/10.1144/pygs.51.4.273>.

Helland-Hansen, W. and Hampson, G.J. (2009). Trajectory analysis: concepts and applications. *Basin Research*, 21(5), pp.454–483. doi:<https://doi.org/10.1111/j.1365-2117.2009.00425.x>.

Hodgson, D.M., Browning, J.V., Miller, K.G., Hesselbo, S.P., Poyatos-Moré, M., Mountain, G.S. and Proust, J.-N. (2017). Sedimentology, stratigraphic context, and implications of Miocene intrashelf bottomset deposits, offshore New Jersey. *Geosphere*, 14(1), pp.95–114. doi:<https://doi.org/10.1130/ges01530.1>.

Holdsworth, B.K. and Collinson, J.D. (1988). Millstone Grit cyclicity revisited. In: B.M. Besly and G. Kelling, eds., *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe*. Glasgow: Blackie, pp.132–152.

Hovikoski, J., Virtasalo, J.J., Wetzel, A., Muthre, M., Strasser, M., Proust, J.-N. and Ikehara, K. (2025). Bioturbation in the hadal zone. *Nature Communications*, [online] 16(1). doi:<https://doi.org/10.1038/s41467-025-56627-x>.

Hoyal, D.C.J.D. and Sheets, B.A. (2009). Morphodynamic evolution of experimental cohesive deltas. *Journal of Geophysical Research*, 114(2), pp.1–18.
doi:<https://doi.org/10.1029/2007jf000882>.

Hudson, R.G.S. and Cotton, G. (1945). The Carboniferous rocks of the Edale Anticline, Derbyshire. *Quarterly Journal of the Geological Society of London*, 101(1-4), pp.1–36.
doi:<https://doi.org/10.1144/gsl.jgs.1945.101.01-04.02>.

Hughes Clarke, J.E. (2016). First wide-angle view of channelized turbidity currents links migrating cyclic steps to flow characteristics. *Nature Communications*, 7(1).
doi:<https://doi.org/10.1038/ncomms11896>.

Imber, J., Childs, C., Nell, P.A.R., Walsh, J.J., Hodgetts, D. and Flint, S. (2003). Hanging wall fault kinematics and footwall collapse in listric growth fault systems. *Journal of Structural Geology*, 25(2), pp.197–208. doi:[https://doi.org/10.1016/s0191-8141\(02\)00034-2](https://doi.org/10.1016/s0191-8141(02)00034-2).

Jackson, M.P.A. and Galloway, W.E. (1984). Thin-skinned gravity sliding as a mechanism for growth faulting: Unit 10: Principles. In: *CN 25: Structural and Depositional Styles of Gulf Coast Tertiary Continental Margins: Application to Hydrocarbon Exploration*. American Association of Petroleum Geologists, pp.37–45.

Jerolmack, D.J. and Swenson, J.B. (2007). Scaling relationships and evolution of distributary networks on wave-influenced deltas. *Geophysical Research Letters*, 34(23), pp.1–5.
doi:<https://doi.org/10.1029/2007gl031823>.

Jerrett, R.M., Bennie, L.I., Flint, S.S. and Greb, S.F. (2016). Extrinsic and intrinsic controls on mouth bar and mouth bar complex architecture: Examples from the Pennsylvanian (Upper Carboniferous) of the central Appalachian Basin, Kentucky, USA. *Geological Society of America Bulletin*, 128(11-12), pp.1696–1716. doi:<https://doi.org/10.1130/b31429.1>.

Jones, C.M. (2022). Climatic influences on Upper Carboniferous (Serpukhovian to Mid-Bashkirian) sedimentary sequences in the UK Pennine and other European basins. *International Journal of Geosciences*, 13(08), pp.715–778.
doi:<https://doi.org/10.4236/ijg.2022.138038>.

Jones, C.M. and McCabe, P.J. (1980). Erosion surfaces within giant fluvial cross-beds of the Carboniferous in northern England. *SEPM Journal of Sedimentary Research*, Vol. 50(2), pp.613–620. doi:<https://doi.org/10.1306/212f7a63-2b24-11d7-8648000102c1865d>.

Lambe, T.W. and Whitman, R.V. (1969). *Soil mechanics*. New York: Wiley, p. 149.

Lancaster, P.J., Daly, J.S., Storey, C.D. and Morton, A.C. (2017). Interrogating the provenance of large river systems: multi-proxy in situ analyses in the Millstone Grit, Yorkshire. *Journal of the Geological Society*, 174(1), pp.75–87. doi:<https://doi.org/10.1144/jgs2016-069>.

Leclair, S.F. and Bridge, J.S. (2001). Quantitative interpretation of sedimentary structures formed by river dunes. *Journal of Sedimentary Research*, 71(5), pp.713–716. doi:<https://doi.org/10.1306/2dc40962-0e47-11d7-8643000102c1865d>.

Leeder, M.R. (1982). Upper Palaeozoic basins of the British Isles—Caledonide inheritance versus Hercynian plate margin processes. *Journal of the Geological Society*, 139(4), pp.479–491. doi:<https://doi.org/10.1144/gsjgs.139.4.0479>.

Leeder, M.R. and McMahon, A.H. (1988). Upper Carboniferous (Silesian) basin subsidence in northern Britain. In: B.M. Besly and G. Kelling, eds., *Sedimentation in a Synorogenic Basin Complex: the Upper Carboniferous of Northwest Europe*. Glasgow: Blackie, pp.43–52.

Longhitano, S.G. (2008). Sedimentary facies and sequence stratigraphy of coarse-grained Gilbert-type deltas within the Pliocene thrust-top Potenza Basin (Southern Apennines, Italy). *Sedimentary Geology*, 210(3-4), pp.87–110. doi:<https://doi.org/10.1016/j.sedgeo.2008.07.004>.

Lowe, D.R. (1982). Sediment gravity flows: II depositional models with special reference to the deposits of high-density turbidity currents. *SEPM Journal of Sedimentary Research*, Vol. 52. doi:<https://doi.org/10.1306/212f7f31-2b24-11d7-8648000102c1865d>.

Martini, I. and Sandrelli, F. (2015). Facies analysis of a Pliocene river-dominated deltaic succession (Siena Basin, Italy): Implications for the formation and infilling of terminal distributary channels. *Sedimentology*, 62(1), pp.234–265. doi:<https://doi.org/10.1111/sed.12147>.

- Martinsen, O.J. (1990). Fluvial, inertia-dominated deltaic deposition in the Namurian (Carboniferous) of northern England. *Sedimentology*, 37(6), pp.1099–1113.
doi:<https://doi.org/10.1111/j.1365-3091.1990.tb01848.x>.
- Masson, D.G., Harbitz, C.B., Wynn, R.B., Pedersen, G. and Løvholt, F. (2006). Submarine landslides: processes, triggers and hazard prediction. *Philosophical Transactions of the Royal Society A: Mathematical, Physical and Engineering Sciences*, 364(1845), pp.2009–2039.
doi:<https://doi.org/10.1098/rsta.2006.1810>.
- Maynard, J.R. and Leeder, M.R. (1992). On the periodicity and magnitude of Late Carboniferous glacio-eustatic sea-level changes. *Journal of the Geological Society*, [online] 149(3), pp.303–311. doi:<https://doi.org/10.1144/gsjgs.149.3.0303>.
- McCabe, P.J. (1975). *The Sedimentology and Stratigraphy of the Kinderscout Grit Group (Namurian, R1) between Wharfedale and Longdendale*. PhD Thesis, Keele University, pp. 82–92.
- McCabe, P.J. (1977). Deep distributary channels and giant bedforms in the Upper Carboniferous of the central Pennines, northern England. *Sedimentology*, 24(2), pp.271–290.
doi:<https://doi.org/10.1111/j.1365-3091.1977.tb00257.x>.
- McCabe, P.J. (1978). The Kinderscoutian Delta (Carboniferous) of Northern England: a slope influenced by density currents. In: D.J. Stanley and G. Kelling, eds., *Sedimentation in Submarine Canyons, Fans and Trenches*. Stroudsburg: Dowden, Hutchinson & Ross, pp.116–126.
- McPherson, J.G., Shanmugam, G. and Moiola, R.J. (1987). Fan-deltas and braid deltas: varieties of coarse-grained deltas. *Geological Society of America Bulletin*, 99(3), pp.331–340.
doi:10.1130/0016-7606(1987)99<331:fabdvo>2.0.co;2.
- Miall, A.D. (1978). Lithofacies types and vertical profile models in braided river deposits: a summary. In: A.D. Miall, ed., *Fluvial Sedimentology*. Canadian Society of Petroleum Geology, pp.597–604.

Miall, A.D. (1985). Architectural-element analysis: A new method of facies analysis applied to fluvial deposits. *Earth-Science Reviews*, 22, pp.261–308.
doi:<https://doi.org/10.2110/scn.85.19.0033>.

Miall, A.D. (1996). *The Geology of Fluvial Deposits: Sedimentary Facies, Basin Analysis, and Petroleum Geology*. New York: Springer-Verlag, p.582.

Middleton, G.V. and Hampton, M.A. (1973). Part 1. Sedimentary gravity flows: Mechanics of flow and deposition. In: G.V. Middleton and A.H. Bouma, eds., *Turbidites and Deep Water Sedimentation*. SEPM Pacific Section Short Course Notes.

Mulder, T. and Syvitski, J.P.M. (1995). Turbidity currents generated at river mouths during exceptional discharges to the world oceans. *Journal of Geology*, 103(3), pp.285–299.
doi:<https://doi.org/10.1086/629747>.

Mulder, T., Syvitski, J.P.M., Migeon, S., Faugères, J.-C. and Savoye, B. (2003). Marine hyperpycnal flows: initiation, behavior and related deposits. A review. *Marine and Petroleum Geology*, [online] 20(6-8), pp.861–882.
doi:<https://doi.org/10.1016/j.marpetgeo.2003.01.003>.

Nemec, W. (1990). Aspects of sediment movement on steep delta slopes. In: A. Colella and D.B. Prior, eds., *Coarse-Grained Deltas*. Special Publications of the International Association of Sedimentologists, pp.29–73.

Nemec, W. (1995). The dynamics of deltaic suspension plumes. In: M.N. Oti and G. Postma, eds., *Geology of Deltas*. Rotterdam: A. A. Balkema, pp.31–93.

Nemec, W., Lønne, I. and Blikra, L.H. (1999). The Kregnes moraine in Gauldalen, west-central Norway: anatomy of a Younger Dryas proglacial delta in a palaeofjord basin. *Boreas*, 28(4), pp.454–476. doi:<https://doi.org/10.1080/030094899421975>.

Nemec, W., Steel, R.J., Gjelberg, J., Collinson, J.D., Prestholm, E. and Oxnevad, I.E. (1988). Anatomy of collapsed and re-established delta front in Lower Cretaceous of eastern Spitsbergen: Gravitational sliding and sedimentation processes. *AAPG Bulletin*, 72(4), pp.454–476. doi:<https://doi.org/10.1306/703c8eb6-1707-11d7-8645000102c1865d>.

Nyberg, B. and Howell, J.A. (2016). Global distribution of modern shallow marine shorelines. Implications for exploration and reservoir analogue studies. *Marine and Petroleum Geology*, 71(71), pp.83–104. doi:<https://doi.org/10.1016/j.marpetgeo.2015.11.025>.

Okazaki, H., Isaji, S. and Kurozumi, T. (2020). Sedimentary facies related to supercritical-flow bedforms in foreset slopes of a Gilbert-type delta (middle Pleistocene, central Japan). *Sedimentary Geology*, 399, p.105613. doi:<https://doi.org/10.1016/j.sedgeo.2020.105613>.

Olariu, C. and Bhattacharya, J.P. (2006). Terminal distributary channels and delta front architecture of river-dominated delta systems. *Journal of Sedimentary Research*, 76(2), pp.212–233. doi:<https://doi.org/10.2110/jsr.2006.026>.

Oliveira, C.M.M., Hodgson, D.M. and Flint, S.S. (2011). Distribution of soft-sediment deformation structures in clinoform successions of the Permian Ecca Group, Karoo Basin, South Africa. *Sedimentary Geology*, 235(3-4), pp.314–330. doi:<https://doi.org/10.1016/j.sedgeo.2010.09.011>.

Orr, P.J. (1995). A deep-marine ichnofaunal assemblage from Llandovery strata of the Welsh Basin, west Wales, UK. *Geological Magazine*, 132(3), pp.267–285. doi:<https://doi.org/10.1017/s0016756800013601>.

Osman, A., Steel, R.J., Ramsook, R. and Olariu, C. (2024). Impact of wave, tides and fluid mud on fluvial discharge across a compound clinoform (Pliocene Orinoco Delta). *Sedimentology*, 71(4), pp.1113–1148. doi:<https://doi.org/10.1111/sed.13167>.

Parsons, J.D., Bush, J.W.M. and Syvitski, J.P.M. (2001). Hyperpycnal plume formation from riverine outflows with small sediment concentrations. *Sedimentology*, 48(2), pp.465–478. doi:<https://doi.org/10.1046/j.1365-3091.2001.00384.x>.

Pisarska-Jamroży, M. and Weckwerth, P. (2012). Soft-sediment deformation structures in a Pleistocene glaciolacustrine delta and their implications for the recognition of subenvironments in delta deposits. *Sedimentology*, 60(3), pp.637–665. doi:<https://doi.org/10.1111/j.1365-3091.2012.01354.x>.

Postma, G. (1984). Slumps and their deposits in fan delta front and slope. *Geology*, 12(1), p.27. doi:[https://doi.org/10.1130/0091-7613\(1984\)12%3C27:satdif%3E2.0.co;2](https://doi.org/10.1130/0091-7613(1984)12%3C27:satdif%3E2.0.co;2).

Postma, G. (1986). Classification for sediment gravity-flow deposits based on flow conditions during sedimentation. *Geology*, 14(4), p.291. doi:[https://doi.org/10.1130/0091-7613\(1986\)14%3C291:cfsgdb%3E2.0.co;2](https://doi.org/10.1130/0091-7613(1986)14%3C291:cfsgdb%3E2.0.co;2).

Postma, G., Nemec, W. and Kleinspehn, K.L. (1988). Large floating clasts in turbidites: a mechanism for their emplacement. *Sedimentary Geology*, 58(1), pp.47–61. doi:[https://doi.org/10.1016/0037-0738\(88\)90005-x](https://doi.org/10.1016/0037-0738(88)90005-x).

Prior, D.B. and Coleman, J.M. (1982). Active slides and flows in underconsolidated marine sediments on the slopes of the Mississippi Delta. In: S. Saxov and J.K. Nieuwenhuis, eds., *Marine Slides and Other Mass Movements*. New York: Springer, pp.21–49.

Pulham, A.J. (1989). Controls on internal structure and architecture of sandstone bodies within Upper Carboniferous fluvial-dominated deltas, County Clare, western Ireland. *Geological Society London Special Publications*, 41(1), pp.179–203. doi:<https://doi.org/10.1144/gsl.sp.1989.041.01.14>.

Read, W.A. (1991). The Millstone Grit (Namurian) of the southern Pennines viewed in the light of eustatically controlled sequence stratigraphy. *Geological Journal*, 26(2), pp.157–165. doi:<https://doi.org/10.1002/gj.3350260207>.

Reading, H.G. (1964). A review of the factors affecting the sedimentation of the Millstone Grit (Namurian) in the central Pennines. *Developments in Sedimentology*, 1, pp.340–346. doi:[https://doi.org/10.1016/s0070-4571\(08\)70506-5](https://doi.org/10.1016/s0070-4571(08)70506-5).

Reading, H.G. and Collinson, J.D. (1996). Clastic coasts. In: H.G. Reading, ed., *Sedimentary Environments: Processes, Facies and Stratigraphy*. Oxford: Blackwell, pp.154–231.

Rees, C., Palmer, J. and Palmer, A. (2017). Gilbert-style Pleistocene fan delta reveals tectonic development of North Island axial ranges, New Zealand. *New Zealand Journal of Geology and Geophysics*, 61(1), pp.64–78. doi:<https://doi.org/10.1080/00288306.2017.1406377>.

Rodríguez-Tovar, F.J., Uchman, A., Payros, A., Orue-Etxebarria, X., Apellaniz, E. and Molina, E. (2010). Sea-level dynamics and palaeoecological factors affecting trace fossil distribution in Eocene turbiditic deposits (Gorrondatxe section, N Spain). *Palaeogeography, Palaeoclimatology, Palaeoecology*, 285(1-2), pp.50–65.

doi:<https://doi.org/10.1016/j.palaeo.2009.10.022>.

Shanmugam, G. (1997). The Bouma Sequence and the turbidite mind set. *Earth-Science Reviews*, 42(4), pp.201–229. doi:[https://doi.org/10.1016/s0012-8252\(97\)81858-2](https://doi.org/10.1016/s0012-8252(97)81858-2).

Shaw, J., Miller, K.S. and McElroy, B. (2018). Island formation resulting from radially symmetric flow expansion. *Journal of Geophysical Research: Earth Surface*, 123(2), pp.363–383. doi:<https://doi.org/10.1002/2017jf004464>.

Silver, M.M.W. and Dugan, B. (2020). The influence of clay content on submarine slope failure: insights from laboratory experiments and numerical models. *Geological Society, London, Special Publications*, 500(1), pp.301–309. doi:<https://doi.org/10.1144/sp500-2019-186>.

Sohn, Y.K. (1997). On traction-carpet sedimentation. *Journal of Sedimentary Research*, Vol. 67. doi:<https://doi.org/10.1306/d42685ae-2b26-11d7-8648000102c1865d>.

Sohn, Y.K., Kim, S.B., Bahk, J.J., Choe, M.Y. and Chough, S.K. (1997). Characteristics and depositional processes of large-scale gravelly Gilbert-type foresets in the Miocene Doumsan Fan Delta, Pohang Basin, SE Korea. *SEPM Journal of Sedimentary Research*, Vol. 67, pp. 130–141. doi:<https://doi.org/10.1306/d4268513-2b26-11d7-8648000102c1865d>.

Sorby, H.C. (1849). The structure and origin of the Millstone-grit of South Yorkshire. *Proceedings of the Yorkshire Geological Society*, 3(0), pp.669–675. doi:<https://doi.org/10.1144/pygs.3.669>.

Southern, S. J. D. (2015). *Influence of basin physiography on the evolution and sedimentation from flows transitional between turbidity current and debris flow*. PhD Thesis, University of Leeds, p.82.

Southern, S.J., Mountney, N.P. and Pringle, J.K. (2014). The Carboniferous Southern Pennine Basin, UK. *Geology Today*, 30, pp. 71–78. <https://doi.org/10.1111/gto.12044>.

Starek, D. and Šimo, V. (2015). Trace fossils from Eocene turbiditic deposits: A case study from the Slovak-Moravian Carpathians. *Acta Geologica Slovaca*, 7(2), pp.129–138.

Stauffer, P.H. (1967). Grain-flow deposits and their implications, Santa Ynez Mountains, California. *SEPM Journal of Sedimentary Research*, 37, pp.487–508.

doi:<https://doi.org/10.1306/74d716f9-2b21-11d7-8648000102c1865d>.

Steel, R.J. and Olsen, T. (2002). Clinoforms, clinoform trajectory and deepwater sands. In: J.M. Armentrout and N.C. Rosen, eds., *Sequence Stratigraphic Models for Exploration and Production: Evolving Methodology, Emerging Models and Application Histories*. GCS-SEPM Special Publication (2002), pp.367–381.

Sternberg, R.W., Cacchione, D.A., Paulso, B., Kineke, G.C. and Drakez, D.E. (1996). Observations of sediment transport on the Amazon subaqueous delta. *Continental Shelf Research*, 16(5-6), pp.697–715. doi:[https://doi.org/10.1016/0278-4343\(95\)00045-3](https://doi.org/10.1016/0278-4343(95)00045-3).

Stevenson, C.J., Jackson, C.A.-L., Hodgson, D.M., Hubbard, S.M. and Eggenhuisen, J.T. (2015). Deep-water sediment bypass. *Journal of Sedimentary Research*, 85(9), pp.1058–1081. doi:<https://doi.org/10.2110/jsr.2015.63>.

Stevenson, I.P. and Gaunt, G.D. (1971). *Geology of the country around Chapel en le Frith (Explanation of one-inch geological sheet 99, New Series)*. *Memoirs of the Geological Survey of Great Britain England and Wales*. London: HMSO.

Talling, P.J., Masson, D.G., Sumner, E.J. and Malgesini, G. (2012). Subaqueous sediment density flows: Depositional processes and deposit types. *Sedimentology*, 59(7), pp.1937–2003. doi:<https://doi.org/10.1111/j.1365-3091.2012.01353.x>.

Todd, S.P. (1989). Stream-driven, high-density gravelly traction carpets: possible deposits in the Trabeg Conglomerate Formation, SW Ireland and some theoretical considerations of their origin. *Sedimentology*, 36(4), pp.513–530. doi:<https://doi.org/10.1111/j.1365-3091.1989.tb02083.x>.

Van Dijk, M., Postma, G. and Kleinhans, M.G. (2009). Autocyclic behaviour of fan deltas: an analogue experimental study. *Sedimentology*, 56(5), pp.1569–1589.
doi:<https://doi.org/10.1111/j.1365-3091.2008.01047.x>.

Van Heerden, I.L. and Roberts, H.H. (1988). Facies development of Atchafalaya Delta, Louisiana: A modern bayhead delta. *AAPG Bulletin*, 72(4), pp.439–453.
doi:<https://doi.org/10.1306/703c8eb1-1707-11d7-8645000102c1865d>.

Van Wagoner, J.C., Hoyal, D.C.J.D., Adair, N.L., Sun, T., Beaubouef, R.T., Deffenbaugh, M., Dunn, P.A., Huh, C. and Li, D. (2003). Energy dissipation and the fundamental shape of siliciclastic sedimentary bodies. In: *American Association of Petroleum Geologists 2003 Annual Convention Exhibition, Salt Lake City, Utah*. American Association of Petroleum Geologists, pp.1–7.

Walker, R.G. (1966a). Deep channels in turbidite-bearing formations. *AAPG Bulletin*, 50, pp.1899–1917. doi:<https://doi.org/10.1306/5d25b6a5-16c1-11d7-8645000102c1865d>.

Walker, R.G. (1966b). Shale Grit and Grindslow Shales: Transition from turbidite to shallow water sediments in the Upper Carboniferous of northern England. *SEPM Journal of Sedimentary Research*, 36, pp.90–114. doi:<https://doi.org/10.1306/74d71415-2b21-11d7-8648000102c1865d>.

Walker, R.G. (1976). Facies models 2. Turbidites and associated coarse clastic deposits. *Geoscience Canada*, 3(1), pp.25–36.

Walsh, J.P. and Nittrouer, C.A. (2009). Understanding fine-grained river-sediment dispersal on continental margins. *Marine Geology*, 263(1-4), pp.34–45.
doi:<https://doi.org/10.1016/j.margeo.2009.03.016>.

Waters, C.N. (2009). Carboniferous geology of northern England. *Journal Open University Geological Society*, 30(2), pp.5–16.

Waters, C.N., Browne, M.A.E., Dean, M.T. and Powell, J.H. (2007). *Lithostratigraphical framework for Carboniferous successions of Great Britain (Onshore)*. Keyworth: British Geological Survey, pp. 25–27 and 58.

Waters, C.N. and Condon, D.J. (2012). Nature and timing of Late Mississippian to Mid-Pennsylvanian glacio-eustatic sea-level changes of the Pennine Basin, UK. *Journal of the Geological Society*, 169(1), pp.37–51. doi:<https://doi.org/10.1144/0016-76492011-047>.

Waters, C.N. and Davies, S.J. (2006). Carboniferous: extensional basins, advancing deltas and coal swamps. In: P.J. Brenchley and P.F. Rawson, eds., *The geology of England and Wales*. London: Geological Society of London, pp.173–223.

Waters, C.N., Jones, N.S., Collinson, J.D. and Cleal, C.J. (2011). Craven Basin and southern Pennines. In: C.N. Waters, I.D. Somerville, N.S. Jones, C.J. Cleal, J.D. Collinson, R.A. Waters, B.M. Besly, M.T. Dean, M.H. Stephenson, J.R. Davies, E.C. Freshney, D.I. Jackson, W.I. Mitchell, J.H. Powell, W.J. Barclay, M.A.E. Browne, B.E. Leveridge, S.L. Long and D. McLean, eds., *A Revised Correlation of Carboniferous Rocks in the British Isles*. London, UK: Geological Society, pp.74–81.

Waters, C.N., Maynard, J.R. and Wignall, P.B. (1996). New developments in the Late Carboniferous geology of the central Pennines, northern England: a review. *Proceedings of the Yorkshire Geological Society*, 51(2), pp.81–86. doi:<https://doi.org/10.1144/pygs.51.2.81>.

Waters, C.N., Vane, C.H., Kemp, S.J., Haslam, R.B., Hough, E. and Moss-Hayes, V.L. (2019). Lithological and chemostratigraphic discrimination of facies within the Bowland Shale Formation within the Craven and Edale basins, UK. *Petroleum Geoscience*, 26(2), pp.325–345. doi:<https://doi.org/10.1144/petgeo2018-039>.

Waters, C.N., Waters, R.A., Barclay, W.J. and Davies, J.R. (2009). *A lithostratigraphical framework for the Carboniferous successions of southern Great Britain (Onshore)*. Keyworth: British Geological Survey Research Report RR/09/01.

Wellner, R., Beaubouef, R., Wagoner, J.V., Roberts, H. and Sun, T. (2005). Jet-plume depositional bodies—The primary building blocks of Wax Lake Delta. *Transactions of the Gulf Coast Association of Geological Societies*, 55(55), pp.867–909.

Wells, M.R., Allison, P.A., Piggott, M.D., Pain, C.C., Hampson, G.J. and De Oliveira, C.R.E. (2005). Large sea, small tides: the Late Carboniferous seaway of NW Europe. *Journal of the Geological Society*, 162(3), pp.417–420. doi:<https://doi.org/10.1144/0016-764904-128>.

Wetzel, A. and Uchman, A. (2001). Sequential colonization of muddy turbidites in the Eocene Beloveža Formation, Carpathians, Poland. *Palaeogeography, Palaeoclimatology, Palaeoecology*, 168(1-2), pp.171–186. doi:[https://doi.org/10.1016/s0031-0182\(00\)00254-6](https://doi.org/10.1016/s0031-0182(00)00254-6).

Wilson, R.D. and Schieber, J. (2014). Muddy prodeltaic hyperpycnites in the Lower Genesee Group of central New York, USA: Implications for mud transport in epicontinental seas. *Journal of Sedimentary Research*, 84(10), pp.866–874. doi:<https://doi.org/10.2110/jsr.2014.70>.

Winsemann, J., Hornung, J.J., Meinsen, J., Asprion, U., Polom, U., Brandes, C., Bußmann, M. and Weber, C. (2009). Anatomy of a subaqueous ice-contact fan and delta complex, Middle Pleistocene, North-west Germany. *Sedimentology*, 56(4), pp.1041–1076. doi:<https://doi.org/10.1111/j.1365-3091.2008.01018.x>.

Woodcock, N.H. and Strachan, R.A. (2012). Regional geological history: why and how? In: N.H. Woodcock and R.A. Strachan, eds., *Geological History of Britain and Ireland (Second Edition)*. Oxford: Blackwell, pp.1–18.

Wright, L.D. (1977). Sediment transport and deposition at river mouths: A synthesis. *Geological Society of America Bulletin*, 88(6), p.857. doi:[https://doi.org/10.1130/0016-7606\(1977\)88%3C857:stadar%3E2.0.co;2](https://doi.org/10.1130/0016-7606(1977)88%3C857:stadar%3E2.0.co;2).

Yang, S., Horsfield, B., Mahlstedt, N., Stephenson, M. and Könitzer, S. (2015). On the primary and secondary petroleum generating characteristics of the Bowland Shale, northern England. *Journal of the Geological Society*, 173(2), pp.292–305. doi:<https://doi.org/10.1144/jgs2015-056>.

Zavala, C., Arcuri, M., Di Meglio, M., Zorzano, A., Otharón, G., Irastorza, A. and Torresi, A. (2021). Deltas: a new classification expanding Bates's concepts. *Journal of Palaeogeography*, 10(1). doi:<https://doi.org/10.1186/s42501-021-00098-w>.

Zavala, C., Arcuri, M., Di Meglio, M., Zorzano, A., Hugo, V. and Roberto, L. (2016). Prodelta hyperpycnites: Facies, processes and reservoir significance. Examples from the lower cretaceous of Russia. In: *International Conference and Exhibition, Barcelona, Spain, 3–6 April 2016*. doi:<https://doi.org/10.1190/ice2016-6356637.1>.

Zavala, C., Arcuri, M., Zorzano, A., Trobbiani, V., Torresi, A. and Irastorza, A. (2024). Deltas: New paradigms. *The depositional record*, 10(5), pp.600–636.
doi:<https://doi.org/10.1002/dep2.266>.

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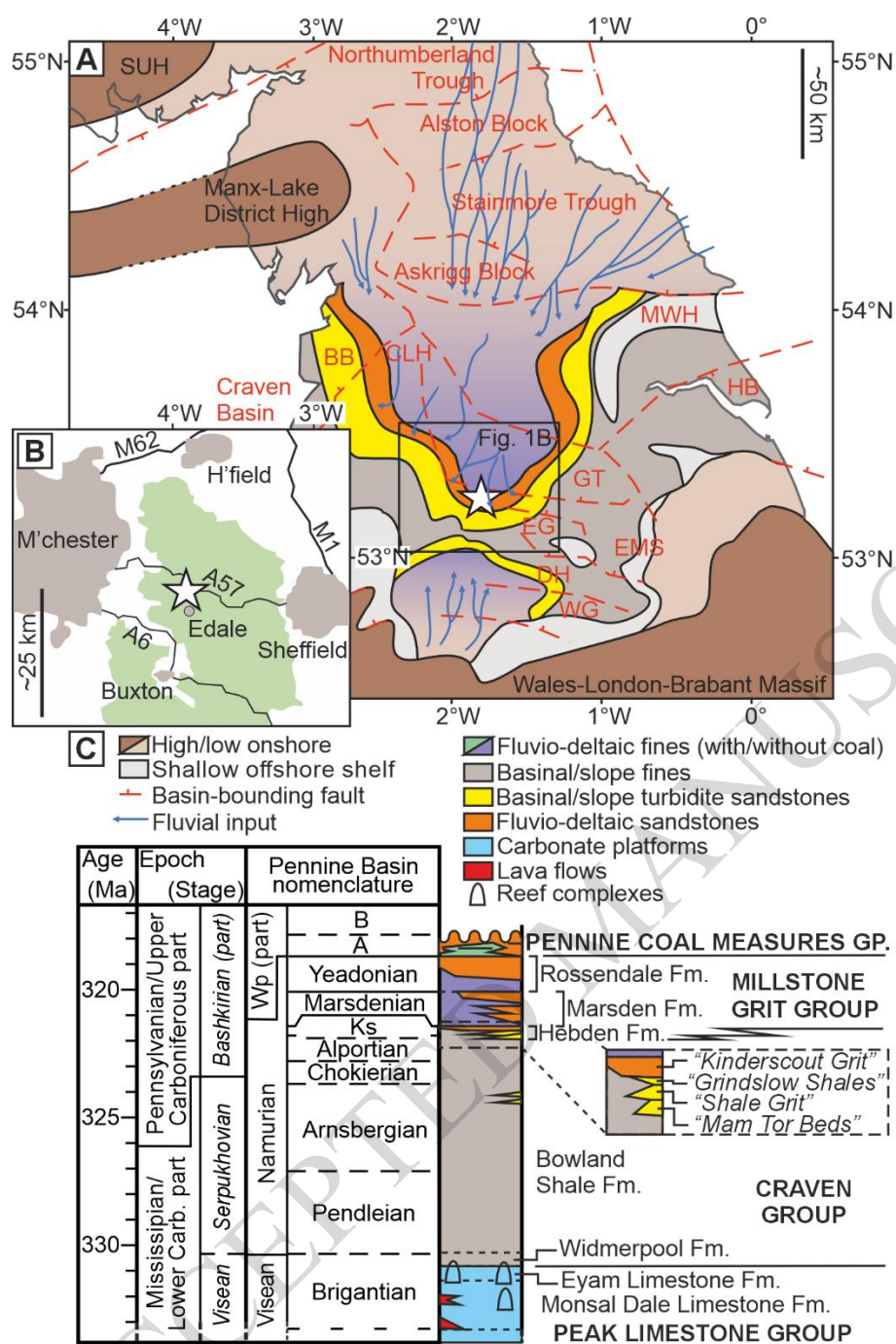


Figure 1

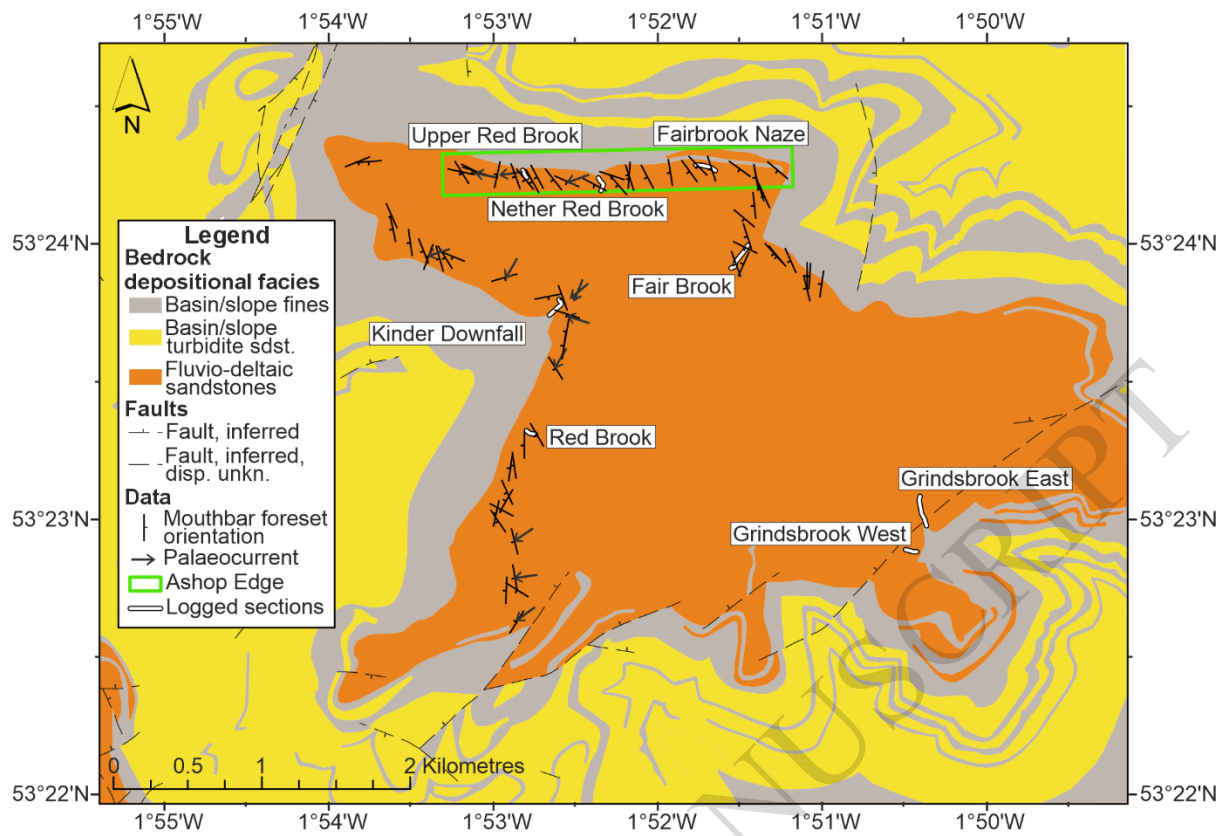


Figure 2

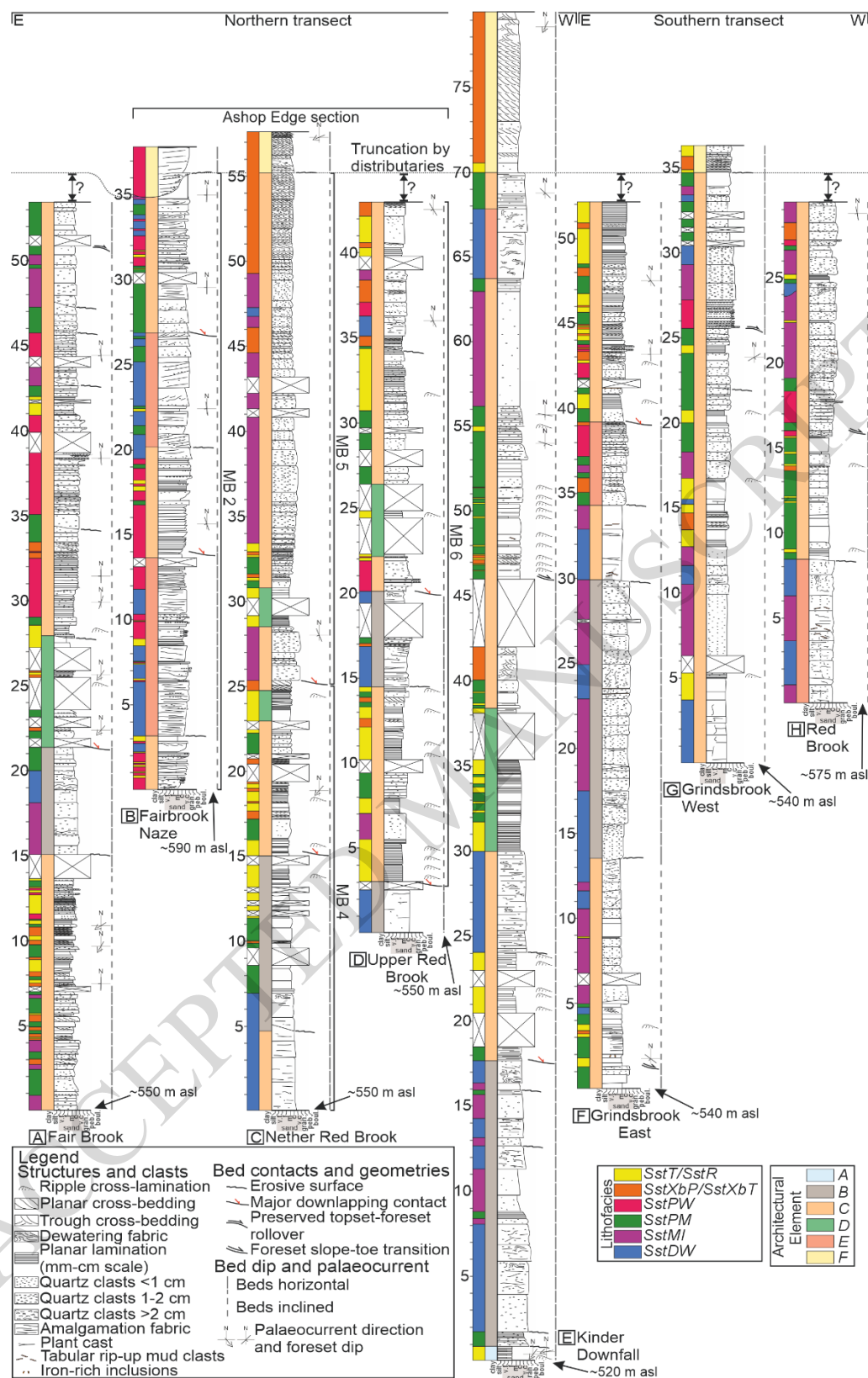


Figure 3

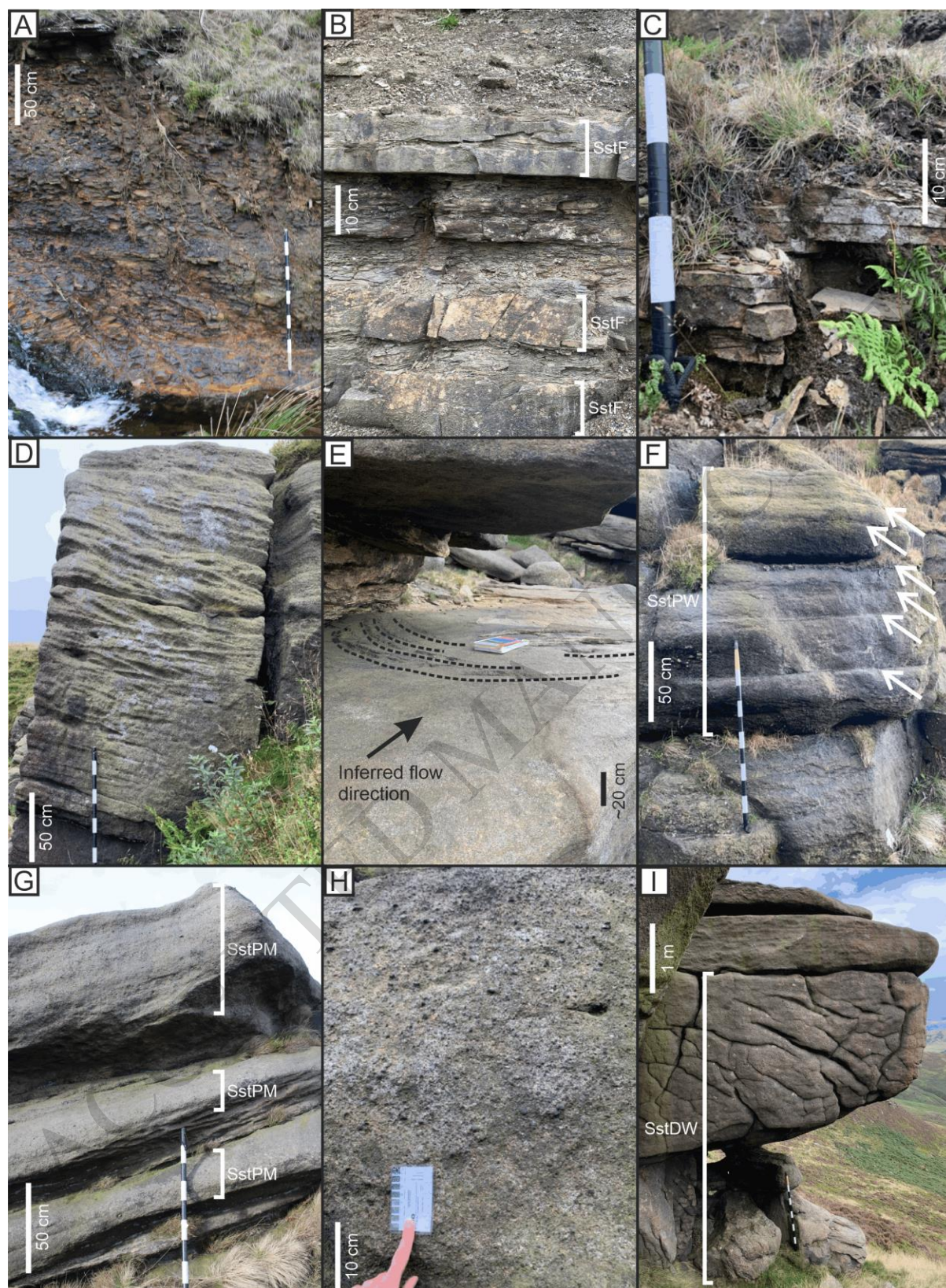


Figure 4

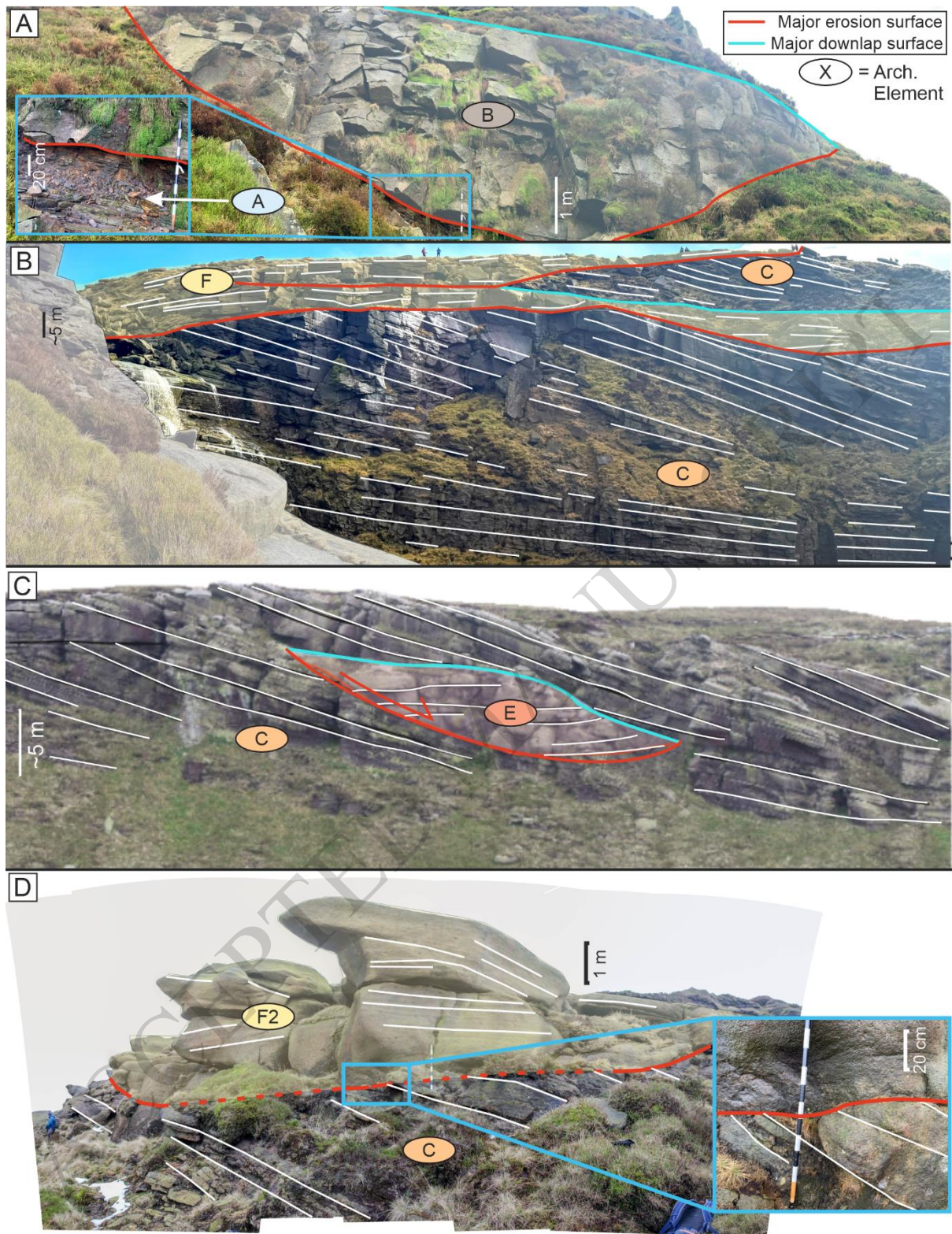


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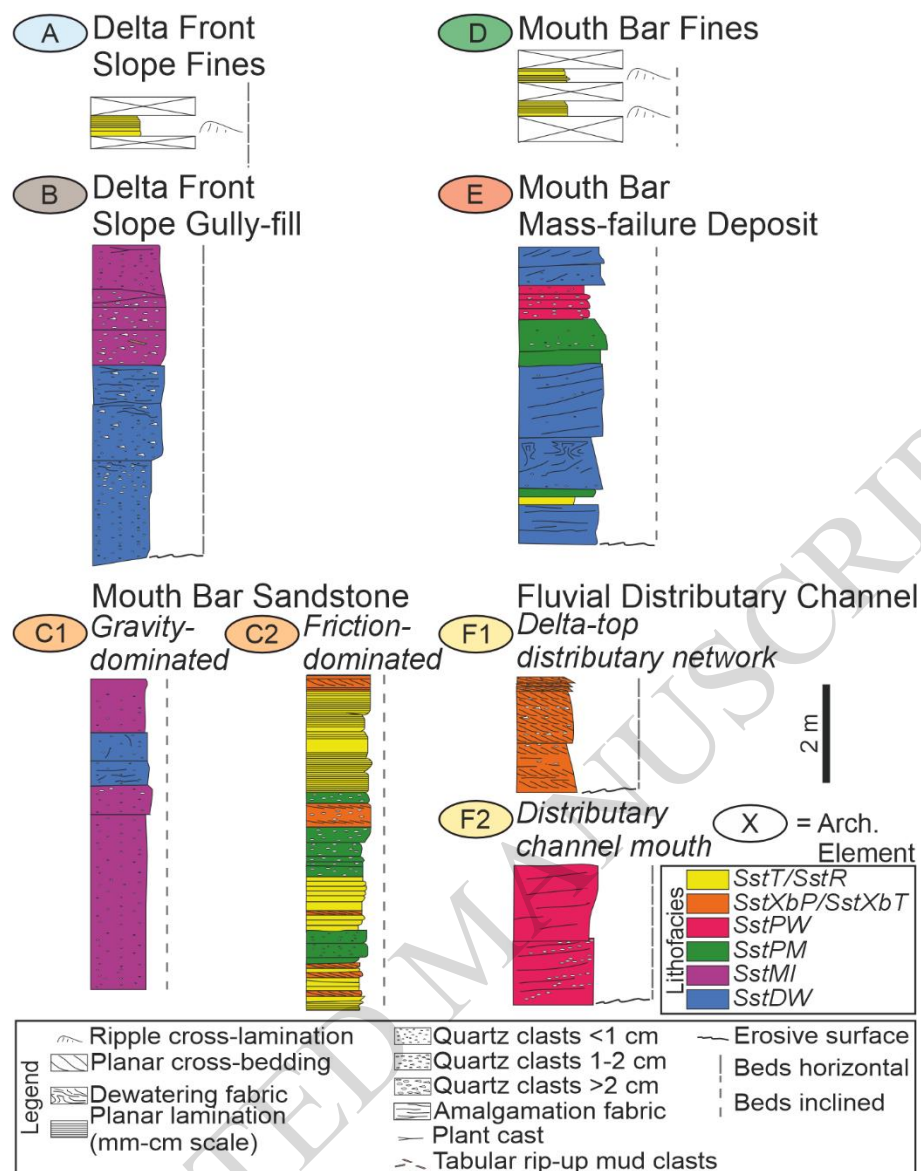


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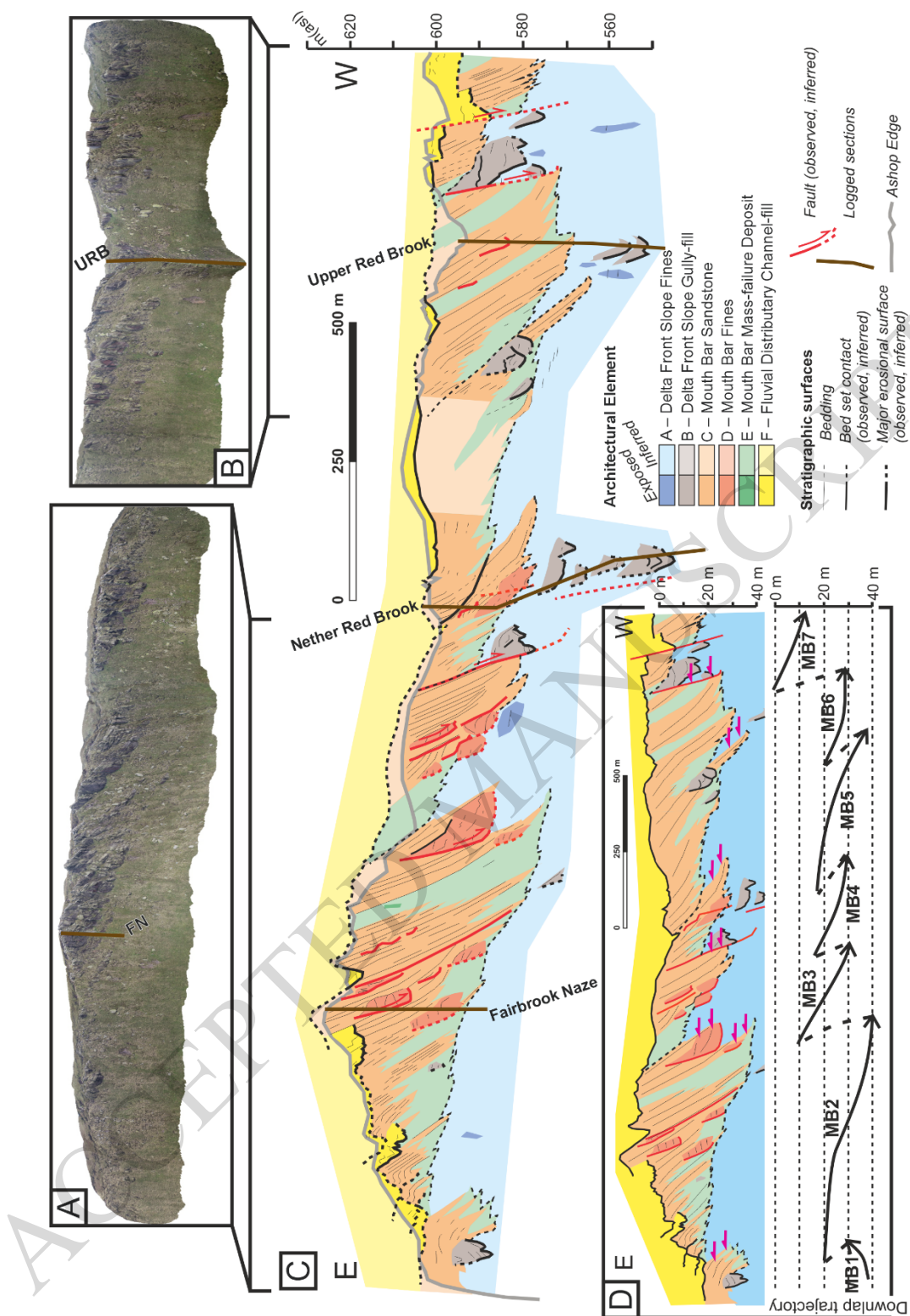


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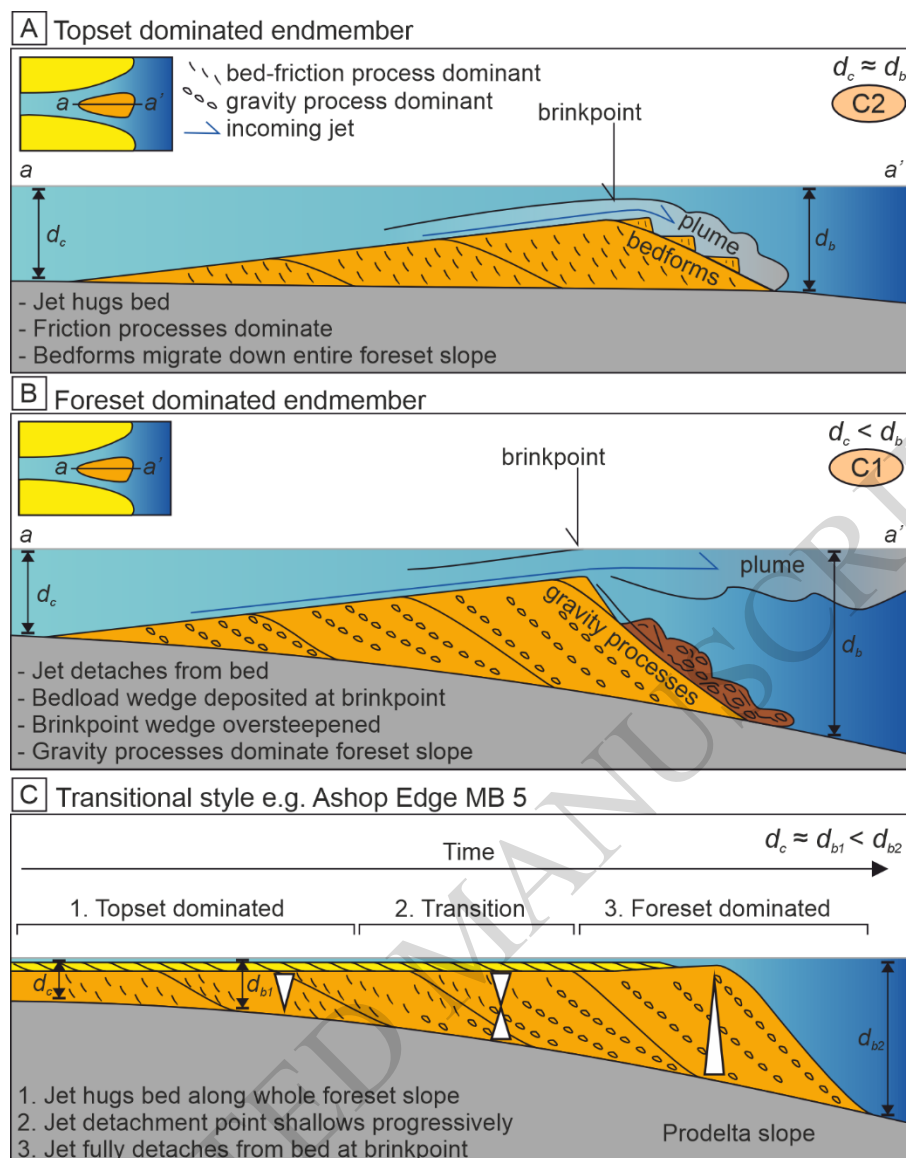


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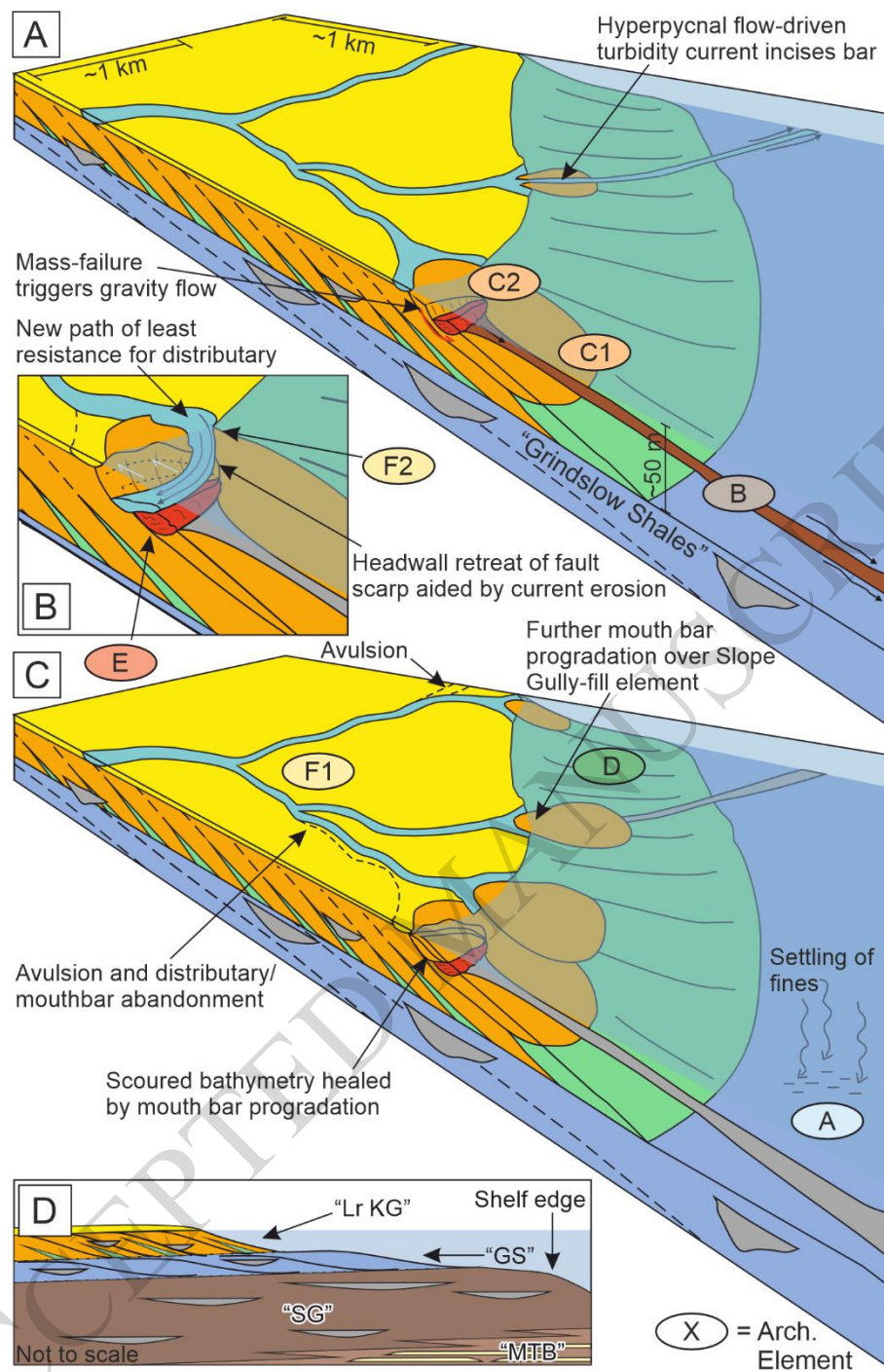


Figure 9