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Caledonian foreland basin sedimentation: A new depositional model for the Upper Silurian-Lower Devonian Lower Old Red Sandstone of the Midland Valley Basin, Scotland

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Abstract

Reconstruction of the geological history of orogenic events can be challenging where basins have limited and/or fragmentary preservation. Here, we apply understanding gained from modern analogues to the sedimentological analysis of the succession of Upper Silurian to Lower Devonian Lower Old Red Sandstone (LORS), northern Midland Valley, Scotland, in order to reconstruct the foreland to the Caledonian orogeny. A new depositional model is presented which differs significantly from current understanding. Using facies analysis, grain size distribution and palaeocurrent data a large distributive fluvial system is reconstructed. Three lithofacies and nine sublithofacies are identified, forming fluvial channel and floodplain facies associations. The system was derived from an emerging mountain range in the Caledonian foreland undergoing constant tectonic rejuvenation to produce 9 km of coarse-grained sediment, exhibiting an overall decrease in thickness towards the west and a large-scale downstream reduction in grain size. Conglomerate sublithofacies dominate proximal areas in the east where amalgamated fluvial channel facies association is abundant, with a downstream increase in the dominance of floodplain facies. Additionally, observed grain size cyclicity is attributed to a pulsatory tectonic influence. The LORS records the time-period between the late phases of the Caledonian Orogeny and the onset of post-orogenic collapse in the mid-Devonian and the presented model allows improved understanding of the north-Atlantic Caledonian foreland.

1 | INTRODUCTION

Analysis of basins with limited and/or fragmentary preservation is often problematic, resulting in difficulties reconstructing the geological history of orogenic events. While events of the Caledonian Orogeny have been well documented (e.g. Coward, 1990; Coward, Dewey, Hempton, & Holroyd, 2003; Leslie, Smith, & Soper, 2008; McKerrow, MacNiocaill, & Dewey, 2000) the Caledonian foreland is poorly preserved in the rock record. As such, it is important to utilize generic models and modern analogues to identify and reconstruct the orogenic history of the area. By applying this understanding alongside observations of Caledonian sedimentary lithologies it is possible to begin to reconstruct the foreland to the Caledonian orogeny. The 9 km thick Lower Old Red Sandstone (LORS) fluvial succession in the northern Midland Valley

Basin (MVB), Scotland, presents an opportunity to apply modern analogues to an ancient problem. Tectonics and associated deformation significantly impact landscapes and act as primary drivers of surface processes (Duvall et al., 2020). River systems are particularly susceptible to tectonic influence, responding to mountain uplift (e.g. Bishop, 2007; Whipple & Tucker, 1999), lateral crustal movements (e.g. Miller & Slingerland, 2006) and material weakening along fault lines (e.g. Roy, Koons, Upton, & Tucker, 2015). The margins of both ancient and modern continental sedimentary basins are thus dominated by fluvial and alluvial fan systems (Ventra & Clarke, 2018). These depositional landforms occur due to large volumes of clastic debris sourced from areas of topographic relief being shed into lower, open terrains (e.g. Blair & McPherson, 1994a; Horton & DeCelles, 2001; Weissmann et al., 2010). These systems occur in a variety of climatic, geomorphological and tectonic settings across a range of timescales (Ventra & Clarke, 2018). Studies have revealed that fluvial sedimentation patterns in modern continental sedimentary basins are dominated by distributive fluvial systems (DFS; Weissmann et al., 2010; Weissmann et al., 2011). As such, a significant part of the geological record is consequently expected to comprise significant DFS deposits (Hartley, Weissmann, Nichols, & Warwick, 2010; Owen, Nichols, Hartley, Weissmann, & Scuderi, 2015) and examples have been described in several locations, including the Devonian systems of Greenland and Ireland (Kelly & Olsen, 1993), the Jurassic Salt Wash Member of the Morrison Formation (Owen et al., 2015) and the Cenozoic deposits of the Andes (Horton & DeCelles, 2001).

Highlights

- Three lithofacies with nine sublithofacies are identified and assigned to two facies associations, fluvial channel and floodplain.
- The MVB LORS was deposited by a large distributive fluvial system in the foreland to the Caledonian Orogeny.
- The system was derived from an emerging mountain range that was being constantly rejuvenated.
- Reconstruction of the system allows improved understanding of the Caledonian foreland prior to post-orogenic collapse.

This paper describes the sedimentology of the LORS of the northern MVB, Scotland, the region lying between the Highland Boundary Fault (HBF) to the north and the Southern Upland Fault (SUF) to the south (Figure 1a). Previous

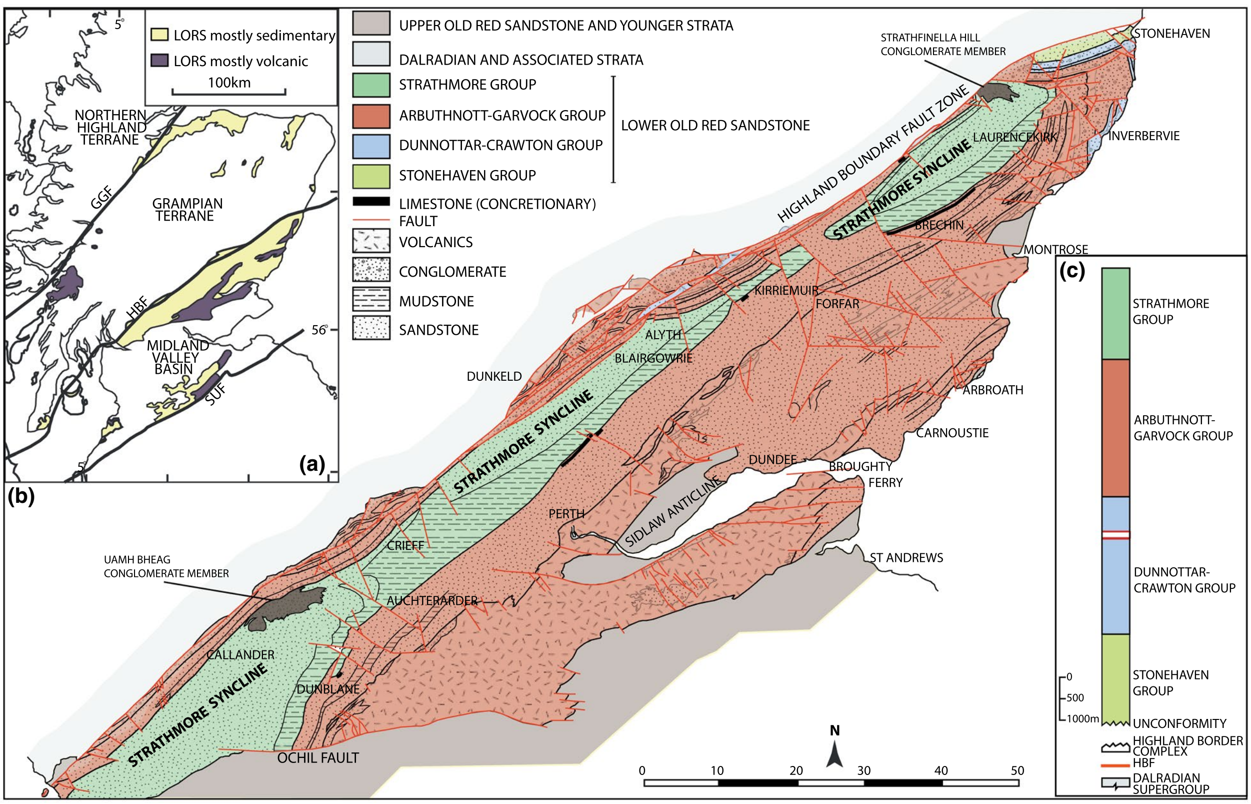

FIGURE 1 Location of the study area (a) Sites of LORS across Scotland (After Hartley & Leleu, 2015). (b) Geological map of the northern MVB (modified after Armstrong & Paterson, 1970). (c) MVB LORS stratigraphical chart (modified after Browne et al., 2002)

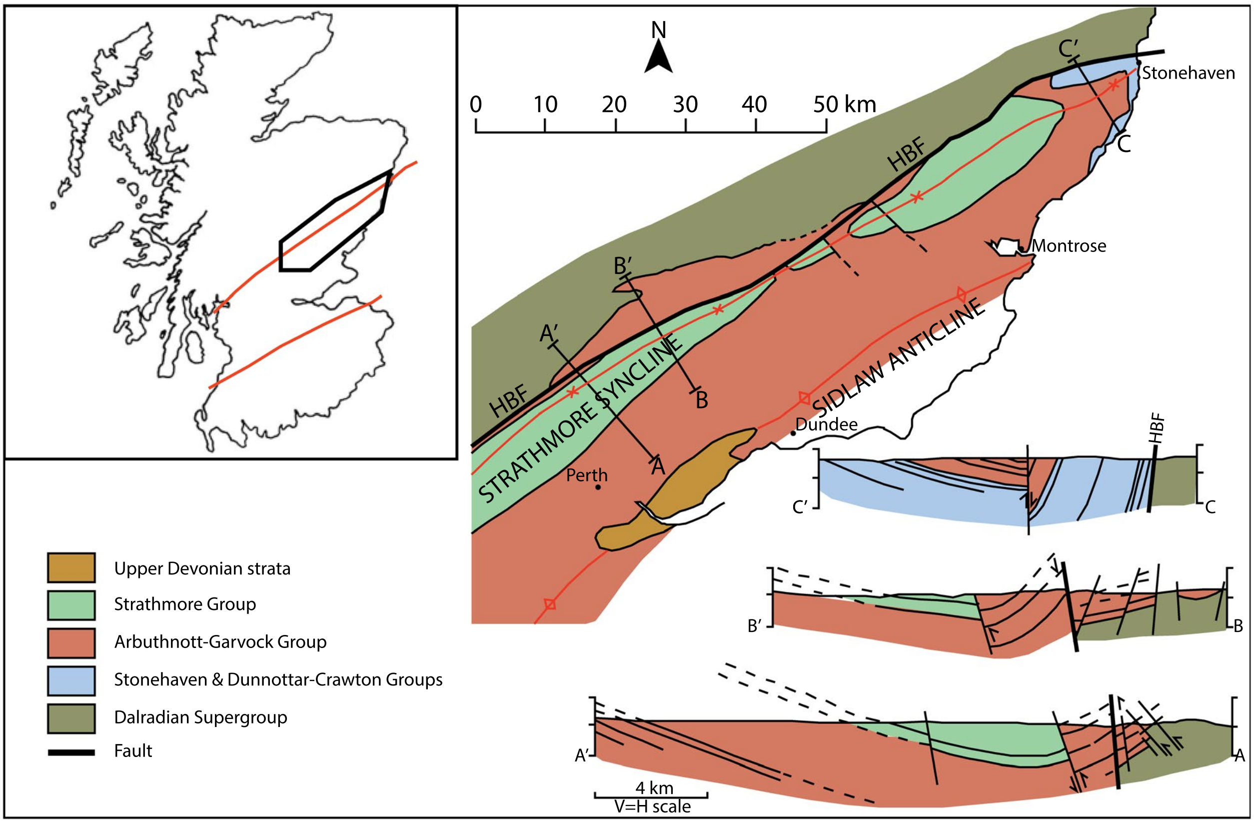

FIGURE 2 Simplified map and cross section of the northern MVB showing distribution of LORS stratigraphic groups across major folds. After Hartley and Leleu (2015)

studies have focussed on specific areas within the basin (e.g. Haughton, 1988; Haughton & Farrow, 1989; Haughton & Halliday, 1991; Haughton, Rogers, & Halliday, 1990), the stratigraphy of the Lower Old Red Sandstone (e.g. Browne, Smith, & Aitken, 2002) or discussion of structural controls on the basin itself (e.g. Bluck, 1986, 2000, 2010, 2013, 2015; Haughton & Bluck, 1988). However, these have not adequately consolidated an understanding of the larger tectonic controls on the basin nor the sedimentary response to these influences. In this study, a combination of new field data and the correlation scheme of Armstrong and Paterson (1970) are used to describe and interpret a thick, pebbly, fluvial succession deposited across the northern MVB. From this analysis, depositional models are presented and discussed together with implications for basin development within the wider tectonic framework.

2 | GEOLOGICAL SETTING

Early descriptions of the LORS of the northern MVB date back over a century (Imrie, 1812), however, other than the in-depth local studies (e.g. Haughton, 1988; Haughton & Farrow, 1989; Haughton & Halliday, 1991; Haughton et al., 1990) the sedimentology of the wider area remains poorly constrained due to fragmentary preservation and limited exposure. While the provenance signature of the MVB LORS suggests sources in the Dalradian Supergroup and Caledonian volcanic rocks (McKellar, Hartley, Morton, & Frei, 2020), the physical sedimentological processes involved require further assessment in order to establish the depositional history of the basin.

The MVB LORS is a key tectonostratigraphic unit in the Palaeozoic history of Scotland. Evolution of the Midland Valley followed closure of the Iapetus Ocean during the mid to late Silurian (Oliver, Wilde, & Wan, 2008), after which subduction gave way to fluvial sedimentation and local volcanism in the NE Midland Valley (Hole et al., 2013). This interval of stratigraphy presents a rare opportunity to assess sedimentation in a time period between the final stages of the Caledonian Orogeny in the mid-late Silurian through to late-post orogenic Caledonian collapse in the latest Lower to early Middle Devonian (McKellar et al., 2020). This study provides a stratigraphic and sedimentological analysis of the LORS succession located in the northern MVB, adjacent to the HBF (Figure 1).

Old Red Sandstone successions are recognized throughout Britain in several basins and are broadly classified with respect to their location relative to the Caledonian Mountains (Kendall, 2017). Basins to the north, including the Orcadian and Midland Valley basins are considered internal or intermontane basins, with little or no access to the ocean. External basins were located to the south, with river networks that drained to coastlines to the east and south and include the Anglo-Welsh Basin and the Dingle-Shannon Basin of southern Ireland (Allen & Crowley, 1983). As such, while sedimentation to the south was subject to marine influence, the northerly basins including the Midland Valley and the largely lacustrine Orcadian, are devoid of any marine sediments or marine influence (Friend, Williams, Ford, & Williams, 2000).

During the period of LORS deposition, the MVB lay between latitudes 10° to 30° south of the equator. The climate was seasonally wet, but overall arid to semi-arid (Browne et al., 2002). Although the true thickness of the succession remains unclear, the preserved thickness measures ~9 km in the east, varying laterally across the basin to around 4km in the west (Armstrong & Paterson, 1970), and contains a diverse sequence of continental clastic sedimentary and intercalated volcanic rocks. These strata lie with a basal unconformity on the Arenig–Caradoc Highland Border Complex, which forms the basement of at least the northern Midland Valley (Tanner, 2008; Figure 1b,c). Southerly derived LORS deposits of equivalent age in the southern Midland Valley are distinct from those in the northern MVB, with the two areas being geographically separate during late Silurian to Early Devonian times (Phillips, Barron, Smith, & Arkley, 2004). The current HBF is a reverse fault of mid-Devonian to Carboniferous age (Bluck, 1984), separating the Lower Old Red Sandstone and Highland Border Complex from the Dalradian Supergroup, a sequence of Late Pre-Cambrian metamorphic lithologies deformed during the early Ordovician Grampian Orogeny (Soper, Ryan, & Dewey, 1999). The MVB was subject to large-scale folding during Middle Devonian Acadian deformation (e.g. Friend et al., 2000), resulting in development of the Strathmore Syncline and Sidlaw Anticline. These structures dominate the MVB, running parallel to the HBF (Figure 2).

3 | STRATIGRAPHIC OVERVIEW

In the northern MVB, the LORS is divided into the Stonehaven, Dunnottar-Crawton, Arbuthnott-Garvock and Strathmore Groups (Figure 3). First described by Campbell (1913), the Stonehaven Group is confined to the NE of the basin, cropping out on the steeply dipping northern limb of the Strathmore Syncline. Due to the limited exposure and post-depositional deformation, the true thickness is difficult to determine but minimum thickness has been estimated at around 1500–1800 m (Armstrong & Paterson, 1970; Carroll, 1995b). The group is subdivided into the Cowie and Carron Sandstone formations, both of which are dominated by cross-stratified and horizontally laminated sandstones, with some intercalations of silt- and mudstone and locally conglomeratic units. The oldest of the four groups, its true depositional age is contentious. Previously, the age has been described as mid to late Silurian based on

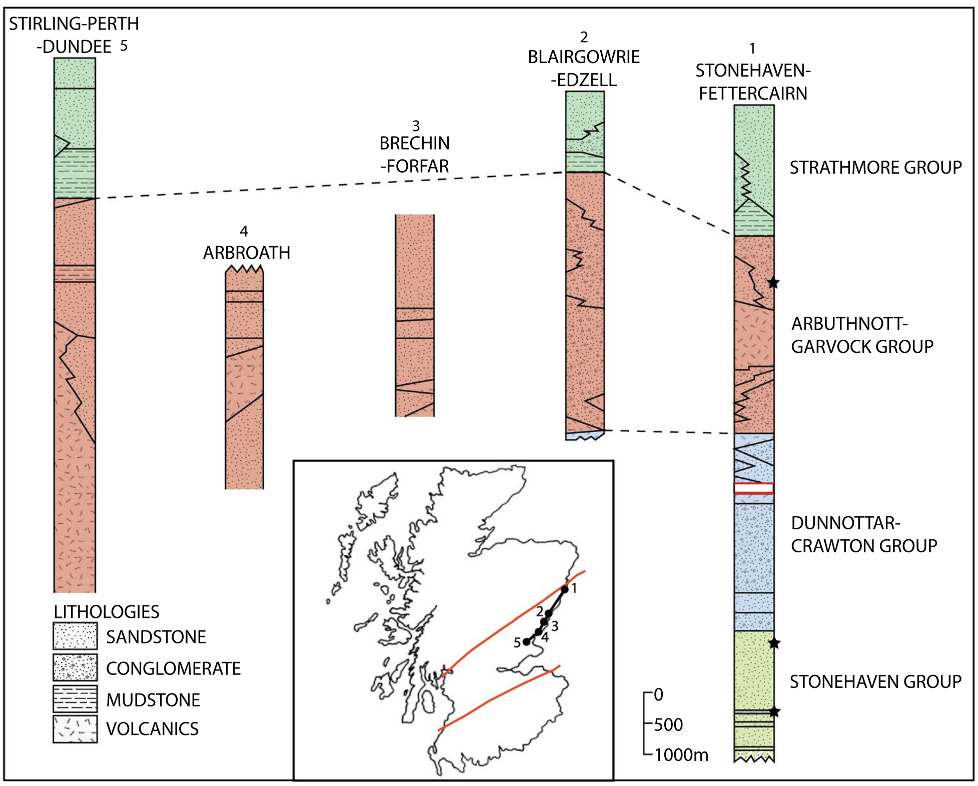

FIGURE 3 Correlation of LORS stratigraphy across the northern MVB (modified from Browne et al., 2002). Stars show locations of example logs for Figures 5, 6 and 7

TABLE 1 Description and interpretation of lithofacies

Lithofacies	Description	Interpretation
1. Conglomerate	**1.1 Breccia.** Poorly stratified, grain-supported angular conglomerate in medium to coarse sandy matrix. Coarse mix of pebbles and cobbles of sandstone, chert and metabasalt up to 0.2 m in size	Lack of oriented clasts, crude bedding and poor sorting indicative of subaerial debris flow. Derived from adjacent basement
	1.2 Moderately sorted conglomerate. Massive to poorly stratified, clast-rich conglomerate with moderate sorting. Rounded clasts reach up to 1 m in size with lack of well-defined clast fabrics. Bedsets up to 9-m thick	Indicate low-relief barforms developed at the base of a channel. Represent bedload deposits of a large, gravelly braided river system
	1.3 Imbricated conglomerate. Poorly sorted, clast-rich conglomerate with A-B axis imbrication. Clasts range from pebble to boulder size, up to 2 m, and are subrounded with blade- and disc-shaped clasts common locally	Represent bedload deposits of a gravelly braided river system composed of channel fills and barforms representing the upstream portion or bar head of gravel bars
	1.4 Cross-stratified conglomerate. Large, laterally extensive cosets between 0.5 and 5-m thick, with individual sets from 0.25- to 1-m thick. Overall increase in grain size from cobble to boulder through cosets. Abundant trough with subordinate planar cross-stratification. Clasts rounded to sub-rounded with palaeocurrents predominantly towards the SW	Prepresent bedload of channel fills and barforms. Likely formed in bar tail region by accretion of the bar platform face forming deposits of a gravelly braided river system
2. Sandstone	**2.1 Fine to medium cross-stratified sandstone.** Sandstone comprising subangular to subrounded grains of quartz, feldspar and lithic fragments with grain size range from fine to medium, with occasional local granule- to pebble-sized material. Abundant trough with subordinate planar cross-stratification with beds 0.5- to 5-m thick	The sandstones represent transverse barforms of a downstream accreting fluvial system
	2.2 Fine to medium planar-laminated sandstone. Composed of subangular to subrounded grains of quartz, feldspar and lithic fragments. Grain sizes range from fine to medium with rare granule to pebble grade material. Tends to occur at the top of cross-stratified units or conglomerate sequences. Beds range from 0.3- to 2-m thick	Transportation as planar sheets during upper flow regime while the channel floor becomes a tractional carpet, with almost continual particle movement. These sandstones represent sheet deposits of a large river system and occur towards bar tops
	2.3 Medium to very coarse lithic, pebbly sandstone. Largely composed of subangular to subrounded grains of quartz, feldspar and lithic fragments. Grain size ranges from medium to very coarse with abundant lithic fragments of pebble to cobble size. Beds range from 0.4- to 2-m thick and are massive or trough cross-stratified. Crude fining upwards from coarse/very coarse to medium sand often observed within foresets. Where medium grain size is observed this lithofacies is often observed alongside sublithofacies 2.1, with individual sets containing a higher proportion of pebbles	Sandstone deposited as channel-lag gravels and cross-stratified channel bar sands. Trough cross-stratified sets are often arranged in multi-story cosets, occasionally in conjunction with sublithofacies 2.1, but may also represent parts of bar deposits through dune migration in unit bars
	2.4 Massive sandstone. Composed of subangular to subrounded grains of quartz, feldspar and lithic fragments. Medium to very coarse-grained, locally granule to pebble-sized, sandstone beds from 0.4- to 3-m thick. Typically associated with disruption features towards bed tops and overlain by planar-laminated sandstone	Massive sandstones represent periods of rapid deposition and dewatering and suggest flood deposits associated with flow surge or bank collapse in a large river system
3. Mudstone	**3.1 Interbedded mudstone and sandstone.** Beds range from 0.1- to 7-m thick. Comprise massive to finely laminated (mm scale) mudstone (silt- and claystone grade material) with common interlaminations of fine- to very fine-grained sandstone. Occasional asymmetrical ripples observed. Sandstone beds (0.4- to 3.5-m thick) are parallel laminated or exhibit climbing ripples, with current ripples or desiccation cracks towards bed tops. Locally, a few surfaces containing small sand-filled arthropod burrows are observed	Sandstone interbeds form as sheet-like deposits during flooding events, whereas mudstones represent deposition in a floodplain complex developed lateral to the main channel system. Desiccation cracks indicate periods of subaerial exposure

palynomorph evidence from inland outcrops (Marshall, 1991; Wellman, 1993). However, additional ages have been suggested by Suarez, Brookfield, Catlos, and Stöckli (2017), with new ^{238}U–^{206}Pb zircon data from better-exposed coastal exposures yielding a youngest single grain age of 413.7 ± 4.4 Ma, placing the group within the Lower Devonian. This is in contrast to further ^{238}U–^{206}Pb data from the same coastal exposures yielding a youngest single grain age of 439 ± 5 Ma (McKellar et al., 2020). Furthermore, age indicators from ichnofauna from coastal exposures suggest deposition around the Silurian – Devonian boundary and indicate that the inland outcrops yielding palynomorph evidence may reflect a different stratigraphic unit (Shillito & Davies, 2017). While there is variation in these dates there is some overlap, suggesting that deposition is likely to have commenced at some point between the latest Silurian and the Lower Devonian.

The overlying Dunnottar-Crawton Group is again largely restricted to the NE of the basin. The true extent of the group is not known, but it is presumed to be progressively overlapped to the SW by the younger strata of the overlying Arbuthnott-Garvock Group (Armstrong & Paterson, 1970; Browne et al., 2002). Minimum thickness may be variable across the group and exposure limitations make thickness estimates difficult. However, previous work (e.g. Armstrong & Paterson, 1970; Browne et al., 2002; Carroll, 1995b) suggest ranges from around 3300 to 3700 m. The Group is subdivided into the Dunnottar Castle Conglomerate, Tremuda Bay Volcanic, Gourdon Sandstone, Whitehouse Conglomerate and Crawton Volcanic formations, however, the combined group name is used for inland areas due to the difficulty in distinguishing individual formations in these locations (Browne et al., 2002). The group is dominated by conglomerates with a diverse clast assemblage of volcanic and metamorphic lithologies (Browne et al., 2002; McKellar et al., 2020; Phillips, 2007), with intercalated sandstones and volcanic lithologies (e.g. Hole et al., 2013; McKellar et al., 2020; Phillips, 2007). The true age of the group is uncertain, however, Rb/Sr isotopic ages derived from volcanic lithologies in the Crawton Volcanic Formation suggest an age of 415.5 ± 5.8 Ma (Thirlwall, 1988), placing the group in the Lower Devonian.

The Arbuthnott–Garvock Group is by far the most extensive of the four groups, being deposited across the entire basin with volcanic lithologies prevalent. The groups are dominated by sandstones with a number of localized volcanic formations and mudstones, although conglomerates are common towards the east (Armstrong & Paterson, 1970; Browne et al., 2002; McKellar et al., 2020). Formation names vary across the basin, as does the thickness, from around 4000 m in the east to ~2400 m in the west (Armstrong & Paterson, 1970; Browne et al., 2002). Again, the depositional age is uncertain, though Rb/Sr isotopic ages from volcanic lithologies suggest an age of 410 ± 5.6 Ma (Thirlwall, 1988).

The topmost Strathmore Group is generally poorly exposed in the northern MVB but is again laterally extensive (Armstrong & Paterson, 1970). Age indicators are largely absent, however, it is believed to be latest Lower Devonian in age (Richardson, Ford, & Parker, 1984). It is dominated by sandstones and sandy mudstones, with formation names varying laterally across the basin (e.g. Browne et al., 2002). True thickness remains unknown due to exposure and post-depositional deformation, however, a minimum thickness of ~1800 m is indicated (Armstrong & Paterson, 1970). Following deposition of the Strathmore Group there is a break in sedimentation during the mid-Devonian prior to deposition of the unconformably overlying Upper Old Red Sandstone (Bluck, 1967; Burgess, 1961).

4 | Sedimentology of the northern Midland Valley Basin

This study focuses on fluvial conglomerates and sandstones of the LORS that crop out in the northern MVB (Figure 1). Exposure is largely restricted to coastal outcrops, with occasional roadcuts and river or stream exposures. After extensive examination of accessible outcrops, representative facies associations were identified and logged in detail. Eleven measured stratigraphic sections, totalling 350 m, have been documented and used together with the stratigraphic framework established by Armstrong and Paterson (1970) to correlate facies variation laterally across the basin.

4.1 | Sedimentary lithofacies

Nine sublithofacies and three lithofacies forming two facies associations are recognized and documented in Table 1 and illustrated in Figure 4.

4.1.1 | Lithofacies 1: conglomerates

Sublithofacies 1.1 breccia

Poorly horizontally stratified, matrix-supported conglomerate in a medium to coarse sandstone matrix forms the basal part of the entire LORS section (Figure 4a). The breccia ranges from 0- to 50-m thick, although laterally attenuated by faulting. Beds are ~1- to 2-m thick but are poorly defined. The breccia is a chaotic mix of angular cobbles and pebbles of metabasalt, black mudstone and red chert with low sphericity that are similar to underlying strata, alonside angular blocks of sandstone that still retain the remnants of cross-stratification. Clast sizes range from pebbles to cobbles up to

0.2 m in diameter with no discernible development of clast fabric. The breccia is massive, with an overall lack of sedimentary structures. Bed contacts are rarely visible, but where present are erosive with relief reaching 0.5–1 m (Figure 4; Figure 5).

Interpretation

The lack of orientated clasts in the breccia, poor sorting and crude bedding are suggestive of transport and deposition by subaerial debris flow (Blair & McPherson, 1994, 2009; Pierson, 1980). Clasts were mainly locally derived from the

FIGURE 4 Lithofacies photographs. (a) Breccia, sublithofacies 1.1, Cowie Formation, Stonehaven; hammer for scale. (b) Moderately sorted conglomerate, sublithofacies 1.2, Dunnottar-Crawton Group, Crawton; rucksack for scale. (c) Imbricated conglomerate, sublithofacies 1.3, with clast A/B axes dipping upstream (to left), Dunnottar-Crawton Group, Crawton; camera pouch for scale. (d) Cross-stratified conglomerate, sublithofacies 1.4, with sub-horizontal bedding, Arbuthnott-Garvock Group, Arbroath; camera lens for scale. (e) Fine to medium cross-stratified sandstone, sublithofacies 2.1, Cowie Formation, Stonehaven; notebook for scale. (f) Fine to medium planar laminated sandstone, sublithofacies 2.2, Cowie Formation, Stonehaven; compass clinometer for scale. (g) Medium to very coarse lithic pebbly sandstone, sublithofacies 2.3, Carron Formation, Stonehaven; compass clinometer for scale. (h) Massive sandstone, sublithofacies 2.4, Carron Formation, Stonehaven; compass clinometer for scale. (i) Interbedded sandstone and mudstone, sublithofacies 3.1, Cowie Formation, Stonehaven; compass clinometer for scale

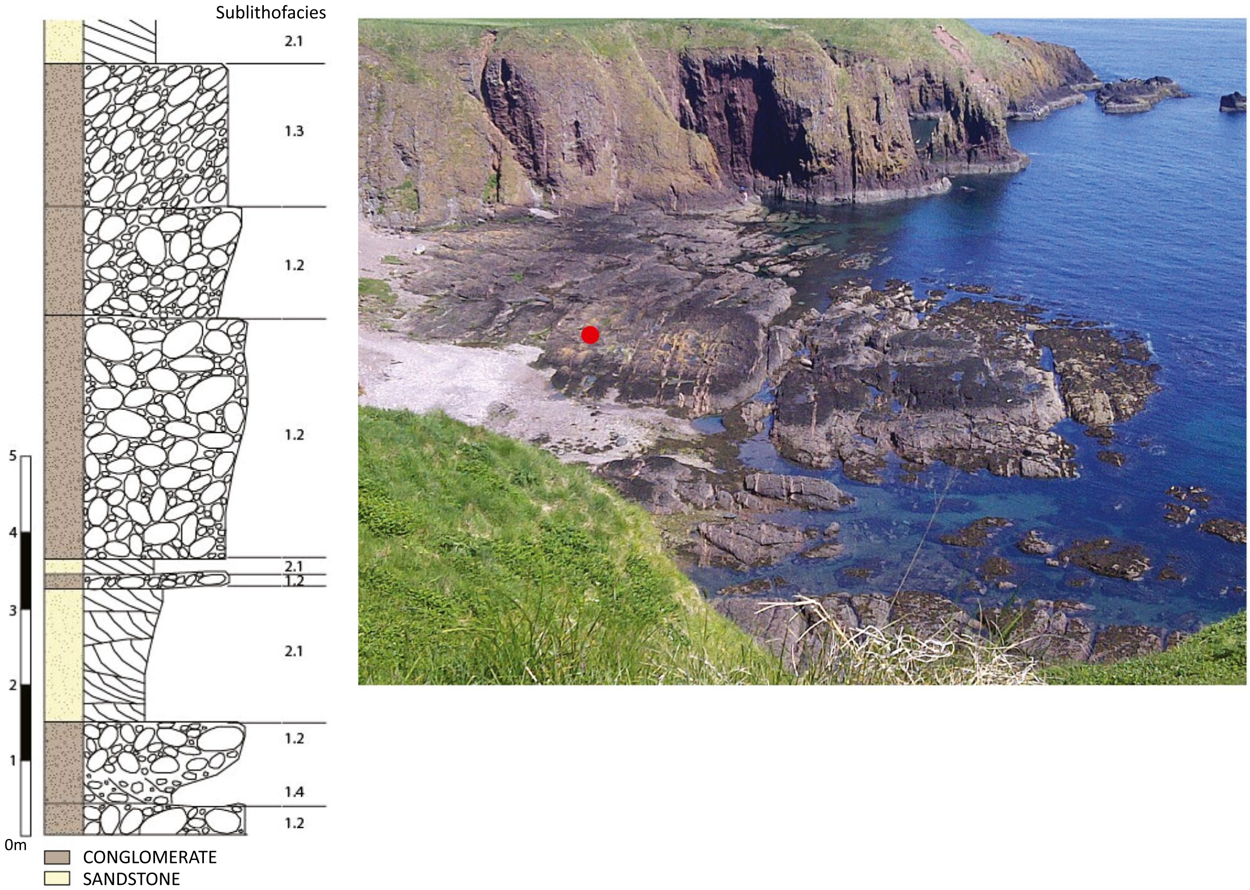

FIGURE 5 Conglomerate example log from the Dunnottar Conglomerate (NO 8801 8430) showing associated sublithofacies. Photograph shows Dunnottar Conglomerate exposed at Castle Haven, Stonehaven (NO 879 841) looking NE. Beds are vertically orientated to locally overturned. Sand lenses appear as thin, raised structures running vertically through the cliffs (~50-m high) in the background and across the wave-cut platform in the foreground. The red dot shows the location of the example log

underlying/adjacent basement lithologies, with the breccia being deposited unconformably on an irregular erosional surface.

Sublithofacies 1.2 moderately sorted conglomerate

Conglomerates in this sublithofacies are better sorted than those of sublithofacies 1.1, with more rounded clasts with moderate sphericity (Figure 4b). They form sheet-like beds with tabular geometries, ranging in thickness from 0.5 to 6 m. Where composite beds occur, grading up from matrix-supported conglomerate at the base to clast-supported, the beds reach up to 9 m. Lateral exposure is limited but beds may reach 50–100 m in width. Beds are massive or poorly horizontally stratified, polymict, clast-rich (60%–90%) conglomerate with a medium to coarse sandy matrix where present, with locally sparry calcitic cement. The conglomerates are generally moderately sorted with a lack of well-defined clast fabrics. Clasts are rounded to subrounded, sometimes reaching over a metre in size. A number of lithologies are represented in the conglomerates, including quartzite, jasper, psammite, granite and schist. Additionally, the conglomerates contain an abundance of volcanic clasts, in excess of 50%. Bases within this sublithofacies tend to be erosional and uneven with relief reaching 1–2 m.

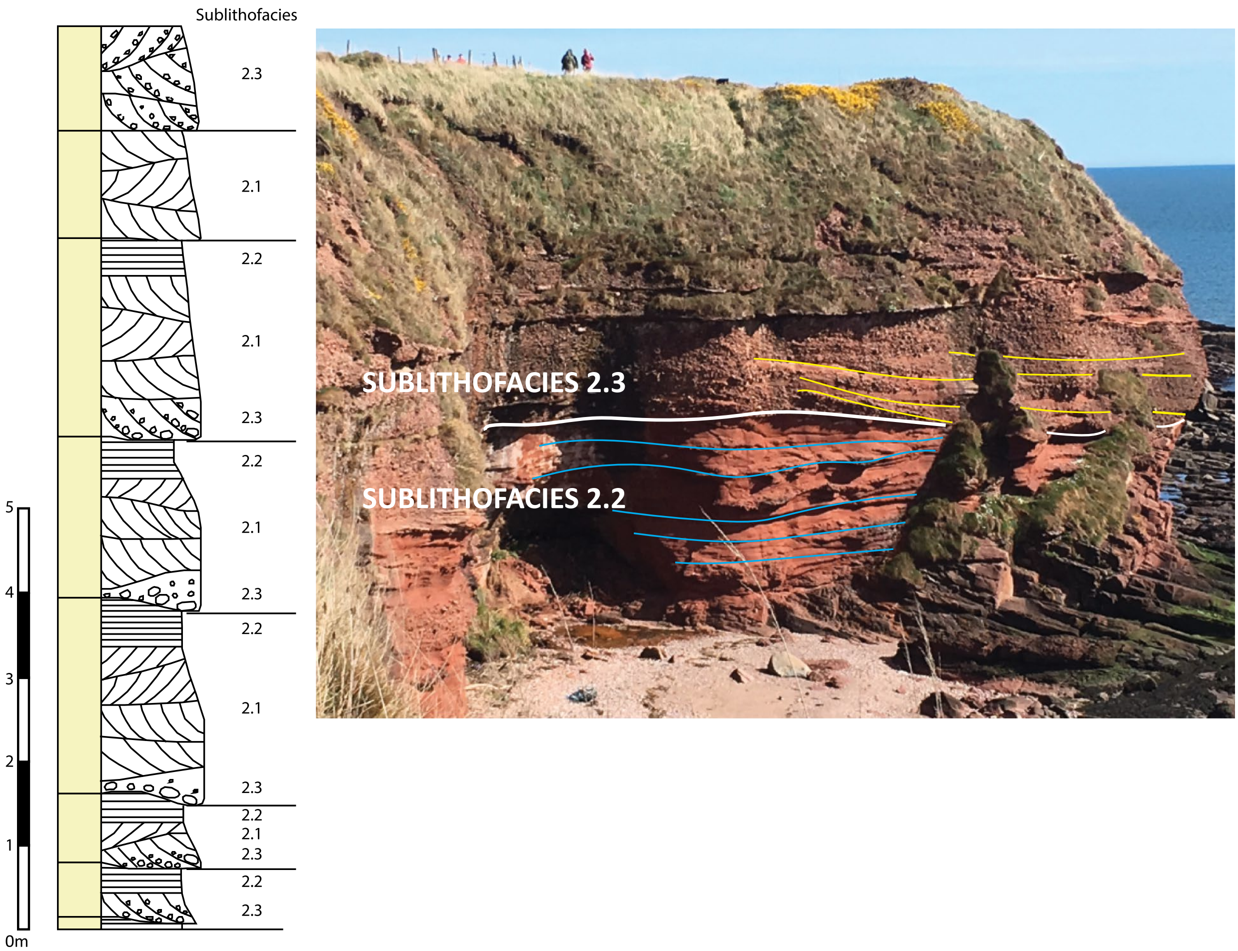

FIGURE 6 Sandstone example log from the Carron Formation of the Stonehaven Group, exposed at Stonehaven Harbour (NO 8786 8526) showing associated sublithofacies. Photograph shows an example outcrop of the Scone Sandstone Formation exposed at Whiting Ness, Arbroath (NO 660 410) looking east, exhibiting planar and cross-stratification (Sublithofacies 2.2 and 2.3). Beds dip slightly towards the east. The cliff face is ~40-m high

Interpretation

The sheet-like nature of the beds suggests the development of low relief barforms, and where erosional bases are present this indicates development at the base of a channel (Rust, 1972a). Conglomerates in this sublithofacies mainly represent bedload deposits of a large, gravelly braided river system composed of channel fills and low-relief barforms (Bluck, 1976).

Sublithofacies 1.3 Imbricated Conglomerate

Beds in these clast-supported conglomerates reach up to 6 m in thickness and are massive or rarely horizontally stratified (Figure 4c). The matrix, where present, is medium to coarse sand. The conglomerates are generally poorly sorted, with A-B axis imbrication. Locally blade- and disc-shaped clasts are common, along with rounded to subrounded clasts. Several clast lithologies are observed including psammite, quartzite, schist, granite and abundant volcanics, and clast sizes range from pebble to boulder size, averaging cobble sized, with some reaching up to two metres. Erosional bases reaching 1–2 m are common.

Interpretation

Clast imbrication suggests bedload transport, with the A-axis orientated perpendicular to flow, the AB axis dipping upstream, thus a(t)b(i) (notation of Walker, 1975) clast fabrics are well defined (Rust, 1972a). The conglomerates largely represent bedload deposits of a gravelly braided river system composed of channel fills and barforms representing the upstream portion (or bar head) of gravel bars (Haughton, 1989). The sheet-like nature of the beds suggests the development of low relief barforms at the channel base where overlying an erosion surface.

Sublithofacies 1.4 cross stratified conglomerate

These conglomerates occur as large, laterally extensive cosets, typically between 0.5- and 5-m thick, and 50- to 100-m wide, although this is a minimum due to limited lateral exposure (Figure 4d). Individual sets range from 0.25 to 1 m in thickness and often display an upward increase in overall grain size from cobble to boulder through the cosets. Trough

FIGURE 7 Mudstone example log from the Castle of Cowie Member of the Cowie Formation, Stonehaven Group, exposed at Cowie (NO 8,841 8711). Photograph shows the outcrop looking NE; notebook for scale. Beds are steeply dipping to vertical and exposure in the cliff and foreshore exhibits interbedded sandstone and mudstone (sublithofacies 3.1). The lithologies young to the right of the photograph

cross-stratification is abundant, with planar cross-stratification accounting for up to 10% of cross-stratified beds. This sublithofacies typically occurs beneath structureless clast-supported conglomerate units. The conglomerates are grain-supported, with clast percentages ranging from 40% to 50%. Maximum clast sizes range from abundant pebbles and cobbles to occasional boulder grade material with a medium to coarse sand matrix. Clasts are rounded to subrounded and include quartzite and volcanic lithologies with subordinate psammite. Individual foresets often display a crude fining upwards profile. Erosional bases are common.

Interpretation

Trough cross-stratification represents either sinuous-crested dunes associated with bar deposits (e.g. Bluck, 1979; Lunt, Bridge, & Tye, 2004), lunate dunes of in-channel bedforms (e.g. Smith, 1990) or progressive accretion of successive unit bars (Bridge & Lunt, 2004). Where planar cross-stratification occurs, these conglomerates likely formed by downstream accreting straight-crested bedforms in fluvial bars that were initially pebble-rich but display progressive decrease in pebble grade material downstream (Smith, 1990; Steel & Thompson, 1983). The cross-stratified conglomerates largely represent bedload deposits of a gravelly braided river system composed of channel fills and barforms (Bluck, 1979, 1980; Church, 2010). The trough cross-stratified conglomerates likely formed in the bar tail region by accretion of the bar platform face (Bluck, 1976; Haughton, 1989).

4.1.2 | Lithofacies 2: sandtones

Sublithofacies 2.1 fine to medium cross-stratified sandstone

Sandstone comprising subangular to subrounded grains of quartz, feldspar and lithic fragments with grain sizes ranging from fine to medium with occasional local granule to pebble grade material constitute this sublithofacies (Figure 4e). Pebbles, where present, are subangular to rounded and comprise a range of lithologies such as mudstone, quartzite, metasandstone, volcanics and rare granitoids. Beds in this lithofacies are 0.5- to 5-m thick and constitute 35%–40% of the entire MVB LORS. Beds are laterally extensive, extending tens of metres, although lateral exposure is often limited. Trough cross-stratification is abundant, with subordinate sub-horizontal to low-angle planar cross-stratification, and local ripple cross-stratification with individual sets ranging from 0.2- to 1-m thick. Foreset dip angles vary between 15 and 30 degrees. Locally, trough cross-strata are lined with angular mudclasts that are imbricated in places, with abundant concretions (Figures 4 and 6).

Interpretation

Trough cross-stratification is formed through dune migration at the base of channels (Bridge & Lunt, 2004), whereas planar cross-stratification records the migration of transverse bars (Ore, 1964). Trough cross-stratified cosets are organized into multi-storey channel fills and may represent migration of barform deposits in a downstream accreting fluvial system (Bridge & Lunt, 2004).

Sublithofacies 2.2 fine to medium planar laminated sandstone

Sandstone composed of subangular to subrounded grains of quartz, feldspar and lithic fragments form this sublithofacies (Figure 4f). Grain sizes range from fine to medium with rare granule to pebble grade material. Beds in this sublithofacies are typically up to a metre in thickness. Beds extend laterally for tens of metres, however, exposure is laterally restricted. Planar lamination is generally centimetre-scale, varying with grain size. This sublithofacies occurs mainly at the top of trough cross-bedded units or above conglomerate units.

Interpretation

In fluvial systems, horizontal lamination can occur in shallow water or during flood stage (Miall, 1977). The sandstones are interpreted as flood-emplaced sands deposited when sand-laden flood water spilled over gravel or sand bars into abandoned channels, or onto upper bar surfaces (Rust, 1972b). Rapid deposition may impair the formation of other bedforms that may be expected under the flow regime (Harms, 1975). These sandstones represent upper flow regime sheet deposits within a large river system and formed close to or on bar tops.

Sublithofacies 2.3 medium to very coarse lithic pebbly sandstone

In this sublithofacies, the sandstones are largely composed of subangular to subrounded grains of quartz, feldspar and lithic fragments (Figure 4g). Grain sizes range from medium to very coarse, with abundant pebble- to cobble-sized lithic fragments. Beds in this lithofacies range from 0.4 to 2 m and are massive or trough cross-stratified. Crude fining upwards from coarse or very coarse-grained to medium sand is often observed within foresets. Individual sets range from 0.2 to 1 m in thickness. Beds are laterally extensive and may extend for some tens of metres, although exposure is often limited. Where medium grain size is present this lithofacies is often observed alongside sublithofacies 2.1, with individual sets containing a higher proportion of pebbles.

Interpretation

These sandstones were deposited as channel-lag gravels and cross-bedded channel bar and braid bar sands. Where structureless units occur in sheet-like form this suggests the development of low relief barforms, and where erosional bases are present this indicates development at the base of a channel (Rust, 1972a). Trough cross-stratified sets are often arranged in multi-storey cosets, occasionally in conjunction with sublithofacies 2.1, but may also represent parts of bar deposits through dune migration in unit bars (Bridge & Lunt, 2004).

Sublithofacies 2.4 massive sandstone

Massive sandstone composed of subangular to subrounded, moderately sorted grains of quartz, feldspar and lithic fragments form this sublithofacies (Figure 4h). Medium to very coarse-grained, locally granular to pebble grade sandstone beds, ranging from 0.4 to 3m thick, typically occur alongside disruption features towards bed tops, overlain by planar laminated sandstone.

Interpretation

Massive sandstones represent periods of rapid deposition and dewatering, causing soft-sediment deformation, and likely represent flood deposits associated with flow surge (Harms et al., 1975) or bank collapse in a braided river system (Bridge & Lunt, 2004).

4.1.3 | Lithofacies 3: mudstone

Sublithofacies 3.1 interbedded mudstone and sandstone

Within this sublithofacies, beds range from 0.1- to 7-m thick and are generally laterally restricted but can stretch up to tens of metres (Figure 4i). They comprise massive to finely laminated (mm-scale) mudstone (siltstone and claystone grade material) with common interlaminations of fine- to very fine-grained sandstone. Occasional asymmetrical ripples are observed. Sandstone beds are parallel laminated or exhibit climbing ripples, with current ripples and desiccation cracks occurring on bed tops. Locally, a few surfaces containing small, vertical sand-filled features are observed. Reduction spots are common throughout massive mudstones (Figures 4 and 7).

Interpretation

Planar lamination in mudstone indicates settling from suspension in a low-velocity current or in pools of standing water (Miall, 1977; Rust, 1972b), with each lamina representing the suspended load from one flood cycle (Rust, 1972b). Asymmetrical ripples indicate a unidirectional, low-velocity current and may represent deposits of waning floods within overbank deposits (Miall, 1977). Sandstone interbeds form as sheet-like deposits during flooding events (Harms, 1975), whereas mudstones represent deposition in a floodplain complex developing lateral to the main channel system. Desiccation cracks indicate periods of subaerial exposure. The vertical, sand-filled structures indicate bioturbation and have been identified previously as arthropod burrows (e.g. Armstrong & Paterson, 1970; Shillito & Davies, 2017), indicating periodic surface colonization.

4.2 | Facies associations

4.2.1 | Fluvial channel facies association

The fluvial channel facies association dominates the area, accounting for over 90% of the lithologies and comprises mainly conglomerates (lithofacies 1; Figure 5) and sandstones (lithofacies 2; Figure 6). This facies association directly overlies the unconformable Ordovician surface at the base of the LORS (Figure 8). It accounts for over 8,000 m of the 9000-m thick LORS succession and extends across the northern Midland Valley (Figure 9). This facies association is interpreted as pebbly, medium to coarse-grained bedload deposition in a predominantly conglomeratic, braided river system. Due to exposure limitations, lateral accretion is hard to identify in the downstream area of the fluvial system, however, more distal downstream channels may become meandering due to a decrease in sediment load.

The fluvial channel facies association is composed of channel fills and barforms. Two scales of structure can be recognized: 1. Low-relief barforms comprising elements of gravel bar formation such as unit bars, bar tails and bar heads, and 2. Multi-storey channel fills that detail the development of cross-channel bars in a downstream accreting fluvial system (e.g. Bridge, 1993; Lunt et al.., 2004). Trough cross-strata, with individual sets up to 1-m thick, are either intercalated with bigger bedforms and interpreted as channel fills or develop lateral to larger bedforms and represent bar-tail deposits (Bluck, 1979; Bridge, 1993; Lunt et al., 2004). Overall, on a regional scale, the system is subject to a downstream change from pebble-rich to pebble poor deposits (Figure 9).

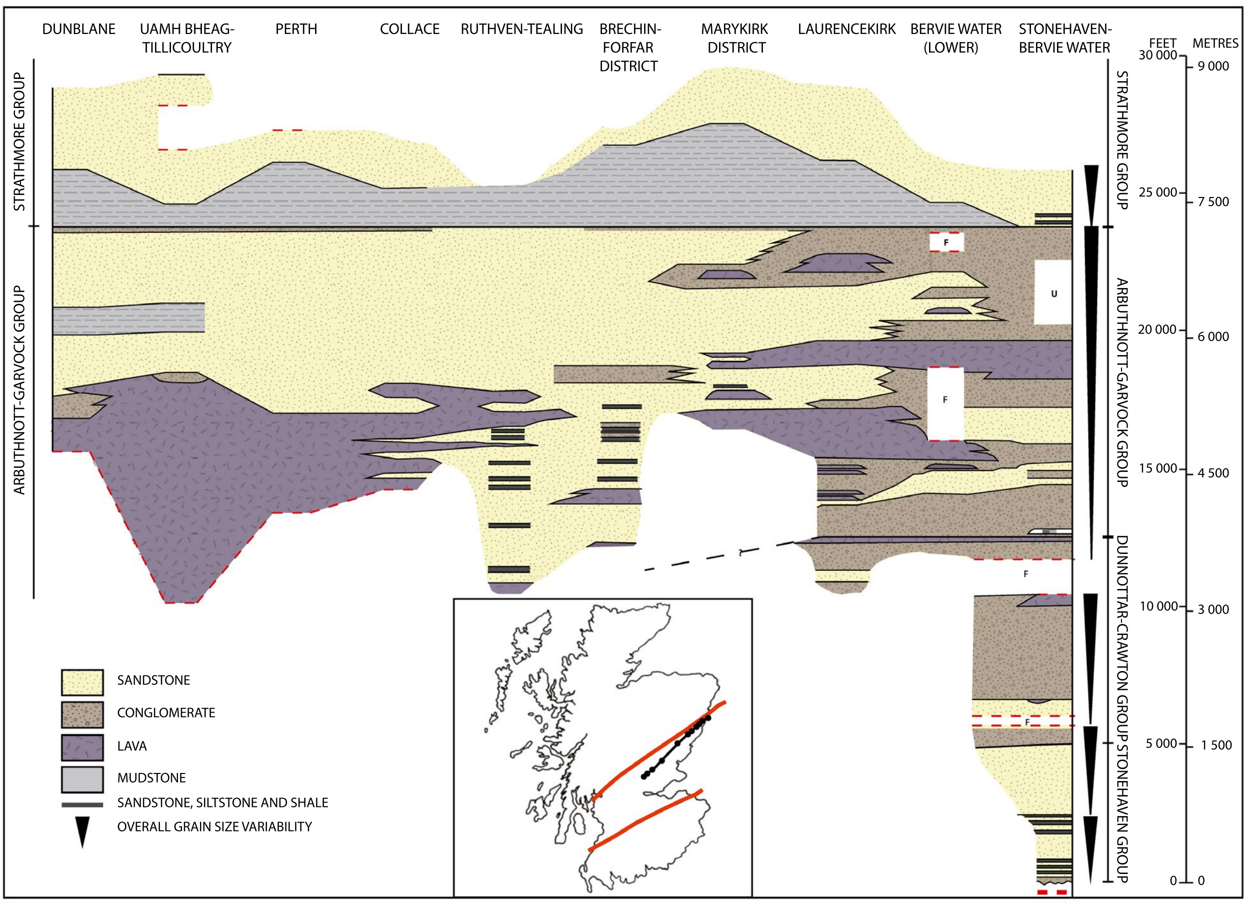

FIGURE 8 LORS lateral lithofacies variation across the NW limb of the Sidlaw Anticline. Based on and modified from Armstrong and Paterson (1970)

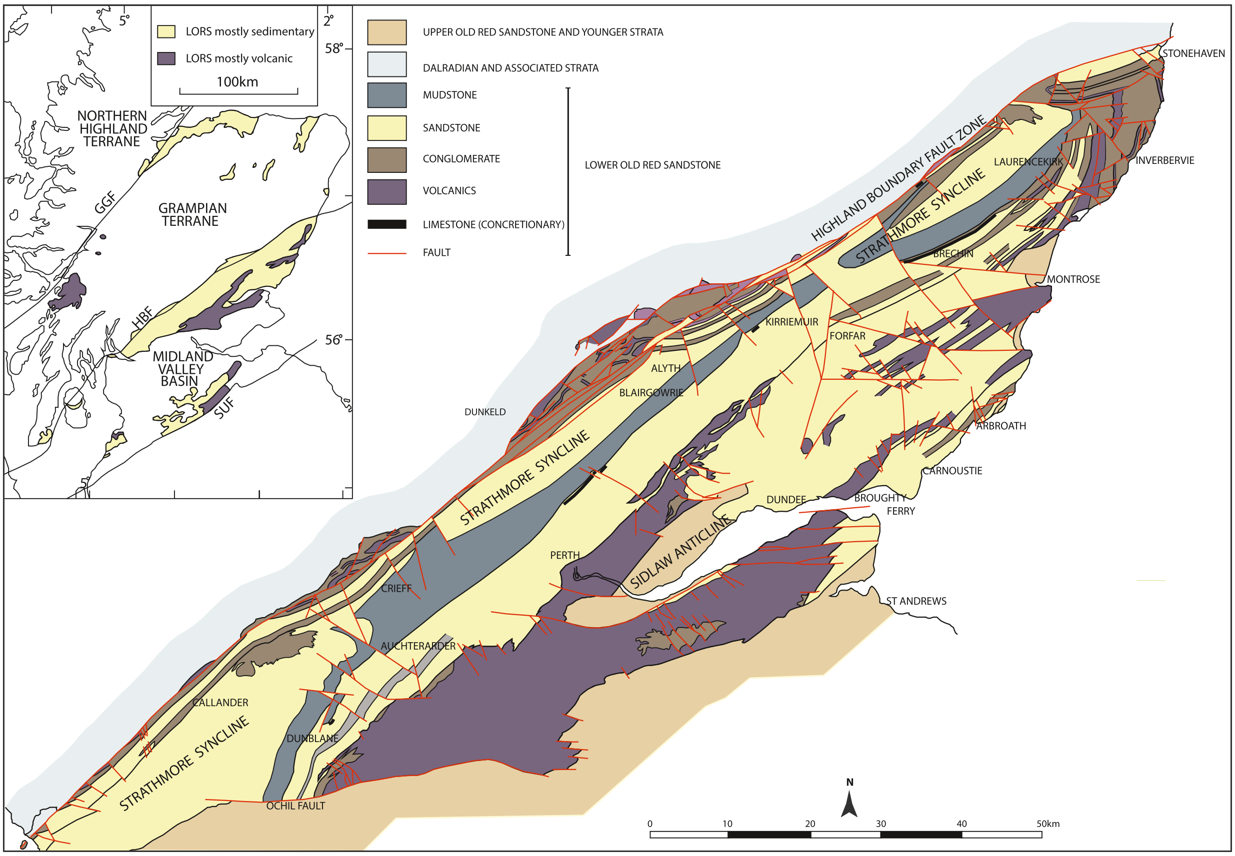

FIGURE 9 LORS lithofacies distribution across the northern MVB. Based on Armstrong and Paterson (1970)

4.2.2 | Floodplain facies association

Floodplain facies typically comprise fine-grained sandstones, claystones and siltstones and exhibit interlamination, cross-lamination and desiccation cracks (Selley, 1988). Burrows occur throughout facies where interbedded mudstone and sandstone occur (Figure 7). Packages of floodplain deposits can either be mud-dominated or interbedded packages of mud and fine sandstone. An increase in sand proportion in units likely indicates increased proximity to the active fluvial channel (Guccione, 1993; Pizzuto, 1987). Floodplain deposits constitute ~10% of the studied LORS, including sections inferred to be floodplain based on previous outcrop description of currently unexposed lithologies (e.g. Carroll, 1995a).

4.3 | Facies distribution across the northern Midland Valley Basin

By combining these field observations and measured stratigraphic sections with the stratigraphic framework established by Armstrong and Paterson (1970), it is possible to correlate lithofacies across the MVB. Correlation is particularly clear along the north west limb of the Sidlaw Anticline where it is possible to trace packages laterally across the axis of the basin (Figure 8). This represents the most complete line of section preserved across the basin following large-scale folding of the strata. Whilst the stratigraphic framework has been documented in the remaining limbs of the Strathmore Syncline and Sidlaw Anticline, it is too fragmentary to be correlated basin-wide (e.g. Armstrong & Paterson, 1970; Browne et al., 2002). Despite the fragmentary nature of the stratigraphical evidence, an overall fining to the SW can be observed, with the majority of coarser-grained lithofacies concentrated towards the NE of the basin. It can also be noted that the succession thins to the SW, with only the Arbuthnott-Garvock and Strathmore groups cropping out to the SW and exposure of the Stonehaven and Dunnottar-Crawton Groups confined to the north east. It should, however, be noted that the true extent of these groups remains unknown due to lack of exposure and post-depositional structural modification.

The distribution of facies associations varies significantly throughout the stratigraphy and across the basin (Figures 8 and 9). The Stonehaven and Dunnottar-Crawton groups predominantly represent fluvial channels with a very small percentage of floodplain facies in the lower part of the Stonehaven Group. Floodplain facies development in the

Stonehaven Group occurs prior to the main development of conglomerate lithofacies in the Dunnottar-Crawton Group. In contrast, at the top of the LORS, the Strathmore Group comprises predominantly floodplain facies, with this facies association becoming more prevalent towards the SW.

4.4 | Palaeocurrent data

The predominant palaeocurrent direction observed in the northern MVB is from NE to SW (Figure 10). In the Stonehaven Group, previous workers have presented conflicting results regarding palaeoflow direction. Gillen and Trewin (1987) described the palaeocurrent direction at the base of the Cowie Formation as towards the east-south-east, and towards the south west in the remainder of the formation. In contrast, Robinson, Rennie, and Oliver (1998) recorded palaeocurrent directions in both the Cowie and Carron formations as having a predominantly north-westerly direction, indicating a source to the south east. However, observation of cross-stratification in the steeply dipping to slightly overturned beds of the Cowie and Carron Formation indicate a dominant palaeoflow direction towards the west throughout, as recorded by Davidson and Hartley (2010) and Hartley and Leleu (2015), indicating an easterly source.

Palaeocurrent data for the overlying Dunnottar-Crawton Group indicate a N–NE source, with palaeocurrent indicators abundant in the thickly bedded conglomerates or cross-bedded sandstone intercalations. While Haughton (1989) records opposing north-westerly and south-easterly palaeoflow in the conglomerates, these were not observed in this study. Conglomerates and sandstones in the Dunnottar Castle Conglomerate clearly exhibit imbrication and trough cross-stratification, indicating a north-easterly source area following correction for the steeply inclined bedding. Evidence of a change in imbrication fabrics from northerly directed in the lower part of the Castle Haven Conglomerate Member to a south easterly direction in the remainder as recorded by Robinson et al. (1998) is also not observed. In the stratigraphically higher Whitehouse Conglomerate transport is towards the south, with little correction required for the shallow dip of bedding.

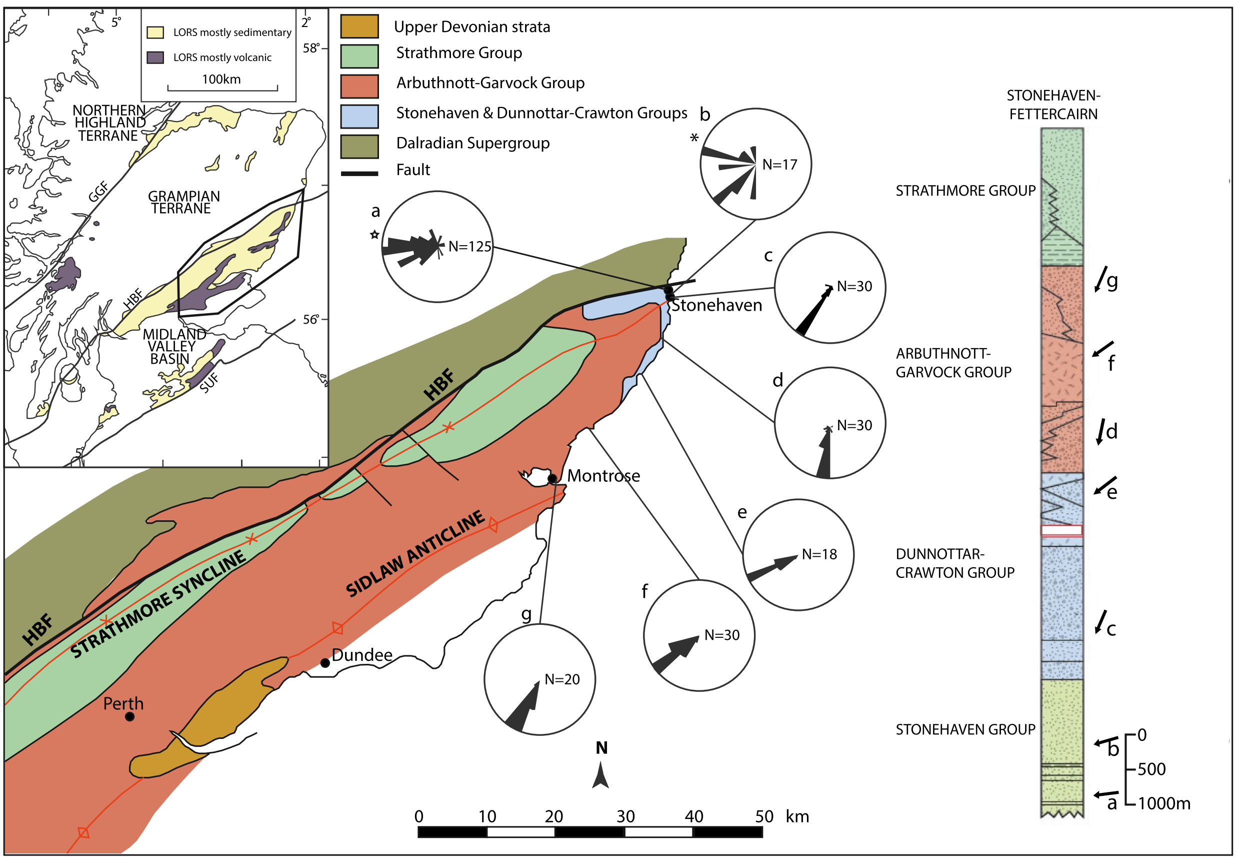

FIGURE 10 Observed palaeocurrents show a predominantly NNE-SSW distribution throughout the stratigraphy. Measurements taken from: (a) cross-stratification in sandstone. Star indicates data from Hartley and Leleu (2015); (b) cross-stratification in sandstone. Asterisk indicates data from Davidson and Hartley (2010); (c) clast imbrication in conglomerate; (d) clast imbrication in conglomerate; (e) clast imbrication in conglomerate and cross-stratification in sandstone; (f) clast imbrication in conglomerate and cross-stratification in sandstone; (g) cross-stratification in sandstone

The Arbuthnott-Garvock Group is characterized by sandstones with south-westerly directed palaeoflow. Palaeocurrent indicators in the Strathmore Group are sparse or absent due to limited exposure, however, Woodcock and Strachan (2000) describes a principally south-west directed flow from a major river system sourced in the NE, constrained to the south by the formation of the Ayr-Ochil-Sidlaw volcanic axis across the MVB.

At a basin-wide scale, Bluck (2000) interpreted large-scale river systems draining the Scandian Orogeny as being able to enter the Midland Valley whenever appropriate accommodation space was available. During deposition of the LORS, Bluck suggested a direct route from the NE which is in keeping with the observed palaeocurrent directions. However, additional palaeocurrents directed towards the north as described by Bluck (1984, 2000) and Haughton (1989) were not observed in this study. Furthermore, McKellar et al. (2020) describe a source to the E–NE based on palaeocurrent and provenance data. On consideration of this and the mean vectors of the palaeocurrents, the dominant direction of palaeoflow across the basin is interpreted as NE - SW.

5 | FACIES ASSOCIATION DISTRIBUTION

The distribution of facies associations in the northern MVB varies greatly throughout the stratigraphy of the LORS as well as across the length of the basin. In the NE, in the stratigraphically oldest units of the Stonehaven and Dunnottar-Crawton groups, the sedimentary succession is principally composed of laterally extensive, highly amalgamated fluvial conglomerates and medium-grained to pebbly sandstones with palaeocurrents indicating flow towards the SW. Limited occurrences of finer-grained floodplain material are observed within the Cowie Formation of the Stonehaven Group, but this group and the overlying Dunnottar-Crawton Group are otherwise entirely composed of fluvial channel deposits. Exposure of these groups is restricted to the NE of the system.

The overlying Arbuthnott-Garvock Group is composed entirely of amalgamated fluvial channel facies association deposits in the NE. This group is, however, much more extensive than the underlying groups and extends across the entire MVB, and towards the SW fluvial channel deposits continue and remain laterally extensive but are separated by distinct packages of floodplain material (Figure 8). There is an evident grain size change within the group with coarse-grained conglomerates in the NE giving way to finer-grained fluvial material across the basin, with floodplain facies associations only developing in the SW. Considering the predominantly NE–SW directed palaeocurrents this grain size variation illustrates a downstream change from coarse- to fine-grained deposits in a river system with a source to the NE. Additional restricted mud and siltstone deposits occasionally occur within the group. These are thought to be the result of impeded drainage due to localized volcanism or syn-sedimentary faulting (Browne et al., 2002), such as that observed at St. Cyrus in the Montrose Volcanic Formation (Hole et al., 2013).

While the previous groups predominantly comprise fluvial channel facies with limited floodplain facies association development, the Strathmore Group is dominated by floodplain deposits. Again, this group extends across the entire basin from NE to SW. Initially dominated by fluvial channel facies sandstones in the NE, the dominant grain size quickly becomes finer, represented by extensive floodplain facies throughout the group and across the basin. With the aforementioned NE-SW palaeocurrent direction this can again be attributed to a downstream decrease in grain size, with the character of floodplain facies being related to proximity to active channels (Guccione, 1993; Pizzuto, 1987), to flood magnitude, or to sediment supply (Moody, Pizzuto, & Meade, 1999; Owen et al., 2015). While the available data suggest floodplain facies, it is difficult to determine the exact palaeogeographic setting. As the original, full extent of the basin is not known, it is unclear whether these facies represent lateral floodplain deposits or terminal facies. When viewed in its entirety, while there is no discernibly significant change in palaeocurrent direction throughout the stratigraphy, there is variation in the proportion of facies associations represented both within the groups and across the basin. The NE of the basin is dominated by the fluvial channel facies association, often very coarse-grained, and conglomerate sublithofacies are prevalent. Floodplain facies account for <10% of the stratigraphy and the fluvial channel deposits are extensive and highly amalgamated. However, when comparing the facies throughout the stratigraphy towards the SW in what can now be described as a downstream direction across the basin, the proportion of floodplain deposits increases through the middle of the basin and becomes the dominant facies association towards the SW, accounting for >60% of the sedimentary rocks. Average grain size is also much-reduced towards the SW, with conglomerate sublithofacies becoming scarce to absent. Overall, the change in dominance from fluvial channel to floodplain facies is indicative of a downstream grain size decrease and reduction in energy as the system becomes more distal to the orogenic front.

6 | DISCUSSION

6.1 | Depositional model

Previous studies by Morton (1979), Wilson (1980) and Bluck (1978) have highlighted the dominant role of fluvial processes in producing the range of sedimentary rocks observed in the LORS of the northern MVB. The emerging model was one of an extended trough, filled laterally by coarse grained alluvial fans and a fine-grained axial braided stream system, with uplift in the region of the present-day North Sea accounting for 70% of the succession (Bluck, 1978). While provenance data suggest a Dalradian source (McKellar et al., 2020), with current exposures of this material to the north of the MVB in the Grampian Terrane, palaeocurrent data are indicative of a source to the east with no evidence of deposition from sources to the north and south. Additionally, the development of alluvial fans would require sufficient topography and a considerable sediment source with appropriate triggering mechanisms to generate the necessary sediment (Blair & McPherson, 2009). Furthermore, the character of the gravelly sedimentary deposits in the MVB is consistent with fluvial rather than alluvial fan deposition, given the formation of longitudinal bars composed of very well rounded pebbles, and thick to crudely horizontally stratified deposits featuring sedimentary features such as clast imbrication and cross-stratification (Blair & McPherson, 1994b; Reinfelds & Nanson, 1993). A fine-grained axial fluvial system is also not indicated, but rather an overall grain size decrease is observed downstream. Additionally, provenance data show little overall variation throughout the stratigraphy, save for occasional increased influence from local volcanic lithologies (McKellar et al., 2020), indicating ongoing rejuvenation of the source area. An increase in sediment maturity is also not observed, with an appreciable quantity of unstable grains throughout the stratigraphy, indicating a lack of major reworking of sediment which may be expected in an axial system influenced by laterally sourced alluvial fans (Santos, Almeida, Godinho, Marconato, & Mountney, 2014).

The laterally fed axial system model was subsequently modified by Haughton and Bluck (1988) who proposed a suite of six alluvial sequences in the northern MVB. They described systems of polycyclic alluvium in large braided streams, input from cryptic flysch sequences located to the north and south of the basin, rapidly aggrading volcanogenic alluvium, sheeted recycled alluvial fanglomerates, fine grained axial alluvium and first-cycle fanglomerates within a series of strike slip sub-basins. As before, palaeocurrent and provenance data do not support supply from sources to the north and south, nor are any indications given of major unconformities or provenance changes which may support the interpretation of amalgamation of a series of sub-basins. This model is also dependent on lateral movement along a suture now concealed by the present-day HBF, as is the fault-controlled onset of sedimentation described by Bluck (2000)..

Decrease in the dominance of channels within the system to the SW, along with a large-scale reduction in grain size was also observed by Bluck (2000), who noted significant reduction in conglomerate clast sizes across the basin. Bluck (2000) attributed this to strike-slip fault-controlled basin formation, with rapid changes between subsidence and uplift resulting in grain size cyclicity generated by increased fault activity towards the east within the HBF zone. These interpretations do not, however, account for the regional NE-SW trends observed such as changes in the dominant facies association type or reduction in grain size. Furthermore, they fail to account for the SW directed palaeoflow and lack of a supply of coarse-grained material from the north. Additionally, syn-sedimentary fault movement indicators on the HBF are absent and there is evidence of an increase in stratal thickness towards the present-day location of the fault, indicating that there was no movement on the HBF during LORS sedimentation (Hartley & Leleu, 2015).

Given that previous models involving strike slip basin formation, alluvial fans sourced from the north and south and/or axial fluvial systems cannot be substantiated an alternative model for deposition must be suggested. This model must take into account the lack of syn-sedimentary fault activity in the area, the NE–SW palaeoflow, downstream grain size reduction and the distribution of the observed facies association across the basin. To account for these factors, we suggest that a Distributive Fluvial System (DFS) model (or fluvial megafan) can best explain the observed characteristics of the LORS fluvial succession in the MVB. DFS are defined by a set number of characteristics. They exhibit channel patterns that radiate away from an apex; channel size and abundance show a downstream decrease; increased floodplain deposits are preserved downstream relative to channel deposits; overall grain size decreases downstream; and proximal areas feature amalgamated channel deposits, whereas distal areas tend more towards smaller, fixed channels with little lateral movement (Cain & Mountney, 2009; Friend, 1978; Friend & Moody-Stuart, 1972; Hartley et al., 2010; Hirst, 1991; Nichols, 1987; Nichols & Fisher, 2007; Owen et al., 2015; Stanistreet & McCarthy, 1993; Weissmann et al., 2010, 2013).

The described down-system trends observed in the MVB are interpreted as being indicative of a large DFS (Figure 11). Overall downstream decrease in grain-size and gravel/sand percentage alongside an increase in the percentage of floodplain facies relative to amalgamated fluvial channel facies is evident. This is consistent with the DFS model proposed by Weissmann et al. (2010, 2013). Additionally, the observed trends across the entire fluvial system, including downstream grain size reduction and increase in floodplain facies, are similar to proximal-to-distal trends reported in other notable DFS, such as the Jurassic Salt Wash DFS (Owen et al., 2015). The MVB DFS is initially dominated by the fluvial channel facies association in the Stonehaven, Dunnottar-Crawton and Arbuthnott-Garvock groups in the east, before increasing dominance of the floodplain facies association in the Strathmore Group. This can again be compared to the Salt Wash DFS which shows this same overall trend, attributed to a downstream change in architectural style from proximal,

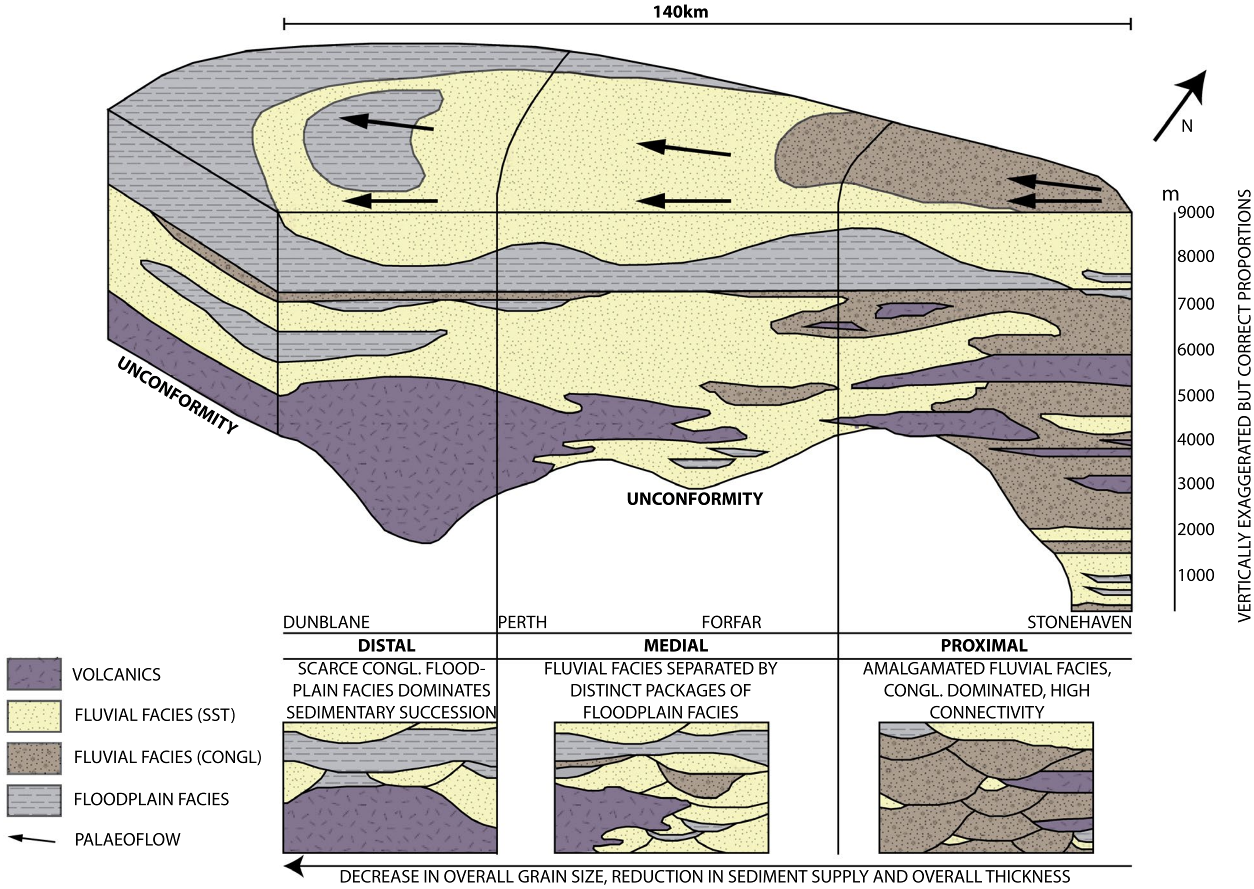

FIGURE 11 A DFS depositional model for the northern MVB LORS. The top surface represents the dominant facies associations throughout the stratigraphy

highly amalgamated fluvial channels to distal regions dominated by floodplain deposits with little to no amalgamation (Owen et al., 2015). The combination of these features indicates a downstream decrease in energy, which is attributed to a combination of channel bifurcation, infiltration, evaporation and flow deceleration as the system radiates away from the apex (Davidson, Hartley, Weissmann, Nichols, & Scuderi, 2013; Hartley et al., 2010; Kelly & Olsen, 1993; Nichols & Fisher, 2007; Owen et al., 2015; Weissmann et al., 2010).

DFS (or megafans) have been recognized in modern sedimentary basins as well as in the rock record at a number of scales, with the largest of these observed in foreland basin settings (Weissmann et al., 2010). Large DFS (or megafans) are considered to be >30 km in length and examples of >700 km are noted in the Andean Foreland; the Pilcomayo and Bermejo DFSs (Hartley et al., 2010). Megafans with thick sedimentary successions on a similar scale to that of the MVB have been observed, such as the c. 7 km Late Eocene-Oligocene Potoco Formation exposed in the Bolivian foreland (Hampton & Horton, 2007), though the preserved length (c. 80 km) suggests that it was somewhat shorter. The c. 200 km long braided Kosi Megafan on the Ganga plain (e.g. Chakraborty, Kar, Ghosh, & Basu, 2010; Weissmann et al., 2010) may be a comparable length to the preserved MVB, though the original extent of the MVB DFS is unknown. Although further examples of similar length are known (e.g. Hartley et al., 2010; Weissmann et al., 2010), the ratio of preserved length (c. 200 km) versus thickness of the MVB DFS makes it somewhat unique.

Two conglomerate members located at the top of the LORS stratigraphy in the Strathmore Group, described by Haughton and Bluck (1988) may represent exceptions to the rest of the main MVB DFS system. These are the Strathfinella Hill and Uamh Bheag Conglomerate Members located at Auchenblae and Callendar, respectively (Figure 1), near the HBF, and represent the youngest part of the Strathmore succession. Both members are suggested to have a northerly provenance (Haughton & Bluck, 1988) and appear to comprise two separate, smaller DFS supplied by sediment from the NW. This again suggests that the HBF was not a basin-bounding fault, and the conglomerates appear to occur independent of the bulk of the MVB LORS sedimentary deposition. They have thus been excluded from the construction of the main DFS model (Figure 11). Mendum (2012) described Highland uplift during the earliest stages of the Acadian Orogeny coinciding with the latter phase of MVB LORS sedimentation, as a result of relative vertical movements on the HBF. As a result, coarse Dalradian detritus was shed into the MVB from the north during the late Lower Devonian. The small, localized DFS at Auchenblae and Callendar may thus have formed as a result of an early Acadian phase of movement on the HBF at the very end of Lower Devonian sedimentation in the northern MVB.

6.2 | Basin evolution and tectonics

Tectonic controls on deposition in the northern MVB are subject to much debate, with Haughton and Bluck (1988) rendering reconciliation of a flanking Caledonian mountain chain incompatible with LORS sedimentation. Deposition into a series of increasingly larger sub-basins is described, through sediment supply from the north and south. The sub-basins are attributed to strike-slip movement along a suture now concealed by the current HBF, within which large, braided antecedent streams were captured. The evidence for a strike-slip regime is taken from what is observed as a SW overlapping relationship between the sub-basins, interpreted to imply progressive opening of the basin floor (Bluck, 1984) alongside descriptions of internal unconformities (Du Toit, 1905; Phillips, Smith, & Carroll, 1997; Robertson, 1987) and soft-sediment deformation (Robertson, 1987). However, deposition across the area of the HBF zone (Tanner, 2008; Figures 1 and 2) imply that the basin was not fault-bounded, as do the lack of syn-sedimentary fault-movement indicators and the increase in stratal thickness towards the present-day location of the HBF (Hartley & Leleu, 2015). As such, the HBF cannot be regarded as a major tectonic control on deposition of the LORS in the northern MVB.

Further evidence of an alternative tectonic setting can be inferred from observation of variations in grain size in the MVB LORS, with several grainsize cycles observed in the LORS of the northern MVB. The cycles, confined to the east of the MVB, comprise a series of five, large-scale (c. 500- to >1000-m thick) coarsening-upward packages from sandstone to conglomerate, or from mudstone to sandstone in the case of the Strathmore Group (Figure 8). Bluck (2000) interpreted the LORS succession as a cycle in itself that originated in a pull-apart basin in a strike-slip setting, prior to fining and maturing petrographically upwards. This petrographic variation was attributed to a decline in tectonic influence following the initiation of the cycle by faulting, but this assessment does not account for the cycles of coarsening upwards grain size (Figure 8) nor the consistent supply of conglomeratic material observed in this study therefore the pull-apart basin model again cannot be substantiated.

Considering the frequency, location and length of the cycles, the origin of the grain size cycles may be attributed to pulsatory active tectonic periods, separated by relatively stable periods (Blair & Bilodeau, 1988). Autocyclic mechanisms, eustasy and climatic variation may locally have a minor impact on grain size (Beerbower, 1964), however, the duration of the cycles—approximately 4–12 Ma—is significantly longer than might be expected from climatic change (Parrish & Barron, 1986). Additionally, while the climate was seasonally wet and variable, no evidence exists to suggest any significant climate changes during the period of deposition at a large enough scale to be a controlling factor. Although no constraints can be given on truncation or erosion in the succession, these have not been indicated by any previous workers and no major unconformities are observed. Alongside their restriction to the NE of the basin and proximity to the likely apex of the MVB DFS, LORS grain size cyclicity is more consistent with a pulsatory tectonic origin (Blair & Bilodeau, 1988). Additionally, provenance data (McKellar et al., 2020) do not reflect evidence of increasing petrographical maturity and rather suggest consistent sediment supply from a tectonically rejuvenating source.

Cyclicity similar to that observed in the northern MVB has been described by Flores and Pillmore (1987) who recognized three distinct coarsening upward megasequences within the Cretaceous-Tertiary Raton foreland basin of Colorado and New Mexico. Similarities in the cycles include comparable grain size to that of the MVB, coarsening upward largely from sandstone to conglomerates with significant development of floodplain facies. Additionally, a similar timescale is described where each cycle represents an average of 7 Ma. In the case of the northern MVB, to account for the coarse-grained nature, the increase in grain size to the east and the coarsening upward cycles, sedimentation must have taken place in a basin located immediately adjacent to a constantly rejuvenating source area. This is in keeping with the descriptions of a synorogenic phase within the two-phase model of foreland sedimentation given by Heller, Angevine, Winslow, and Paola (1988), where gravel-rich coarsening upwards cycles are deposited proximal to the tectonic source but diminish away from the load. In this sense, similarities may be drawn between the northern Midland Valley and the modern fluvial megafans of the Chaco Plain, southern Bolivia, described by Horton and DeCelles (2001) in which an overall coarsening and thickening upward trend are observed. Additionally, the Tertiary Camargo Formation 200 km west of the modern megafans exhibits the same trend (Horton & DeCelles, 2001) and is ascribed by DeCelles and Horton (1999) to an eastward-migrating foreland basin system formed by an advancing fold-thrust belt. Furthermore, McLean and Jerzykiewicz (1978) identify three scales of coarsening upwards cycles within the Brazeau-Paskapoo formations of the Coal Valley area of Alberta, Canada: Large-scale first-order cycles reaching hundreds of metres thick which are comparable to the large-scale cycles observed in the northern MVB; Smaller-scale second- and third-order cycles measuring metres to tens of metres, and centimetres to metres thick respectively. First-order cycles are interpreted as being subject to allocyclic tectonic control, whereas the smaller-scale second- and third-order cycles are heavily influenced by autocyclic controls. As is seen in the northern MVB, the allocyclic first-order cycles within the Brazeau-Paskapoo formations occur in conjunction with a constantly rejuvenating source area, the Rocky Mountains, and result from the encroachment of the thrust belt on the foreland basin.

The overall coarse-grained nature of the lithologies in the east of the basin throughout the stratigraphy implies constant renewal of a very proximal source terrain, however, the relative age of the MVB LORS places deposition too early to be a product of post-orogenic collapse. Additionally, no evidence exists to support a deep, basin-bounding fault to the east consistent with the occurrence of the coarsest-grained material. Furthermore, provenance data (McKellar et al., 2020) indicate no major compositional changes in provenance through the LORS stratigraphy or across the basin consistent with strike-slip basin fill. With this in mind, and on consideration of the similarities with the previously discussed basins, northern MVB sedimentation would be best ascribed to a foreland basin setting heavily influenced by tectonic movement and uplift to the east. (B) Tectonic Framework.

To place the described observations within a wider context, it is necessary to consider MVB LORS deposition within the late Early Palaeozoic tectonic framework. During the Early to mid-Silurian (435–425 Ma), collision between Laurentia and Baltica to the east and north of the Midland Valley and Grampian terranes led to the Scandian deformation phase of the Caledonian Orogeny (Coward, 1990). The timing of deposition of the MVB LORS coincides with a period of orogenic uplift in the Northern and Grampian Highlands, alongside emplacement of numerous granite plutons (e.g. Neilson, Kokelaar, & Crowley, 2009; Thirlwall, 1988) and sinistral movement on major terrane bounding faults such as the Great Glen Fault (e.g. Coward, 1990; Stewart, Strachan, Martin, & Holdsworth, 2001). Deposition continued until interruption by the Middle Devonian Acadian event, considered the end of the Caledonian Orogeny (Mendum, 2012). The commencement of deposition in the northern MVB is also roughly coincident with resumed low ^{87}SR-^{86}Sr I-type (Andean-type subduction related) magmatism in the Grampian Highlands, forming some of the suite of Newer Granites (Oliver, 2001). Models have been presented suggesting a sediment source in the Grampian Terrane as a result of relief generated by this granite emplacement (e.g. Oliver et al., 2008). However, while these models may account for the development of basins locally across the Grampian Terrane, they do not provide an adequate source for continuous deposition of 9 km of largely conglomeratic sediment into the northern MVB. Furthermore, palaeocurrent and provenance data (McKellar et al., 2020) describe a source to the east, and grain size trends, particularly clast-supported boulder conglomerates

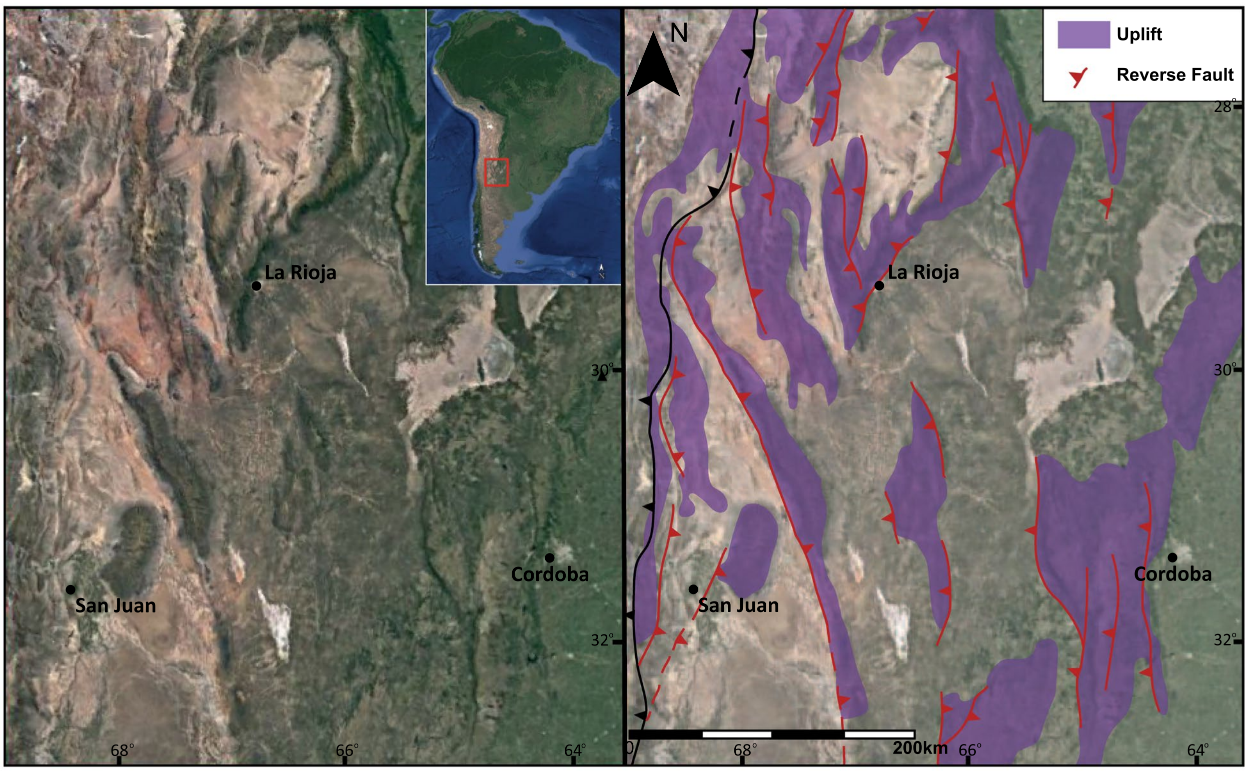

FIGURE 12 Satellite image (left) of the Sierras Pampeanas region, central western Argentina (see inset for location). Image © Google Earth, Landsat/Copernicus. Overlay (right) shows distribution of uplifts and principal faults (red) directly to the east of the Andean Precordillera thrust belt (black). Structural information modified from Jordan and Allmendinger (1986)

indicate close proximity to this source east of the MVB. This implies that deformation associated with the Scandian Orogen created uplift to the east, in the current North Sea area, providing the relief necessary for the extreme grain size observed in the conglomerates and providing a source for the westerly flowing DFS in the MVB.

Whilst it is established that the source of the MVB LORS lay to the east, McKellar et al. (2020) noted that provenance data are not compatible with a Scandinavian source, but rather sediment was supplied from a combination of Dalradian and younger volcanic lithologies. As discussed, generation of sediment of the character and volume of the LORS would require significant uplift of a considerable source area. As the source of the sediment does not lie in the Norwegian Caledonides this poses a problem—a 'missing' mountain range in the Caledonian foreland.

Given the lack of exposure, modern-day analogues may present an opportunity to describe a comparable depositional

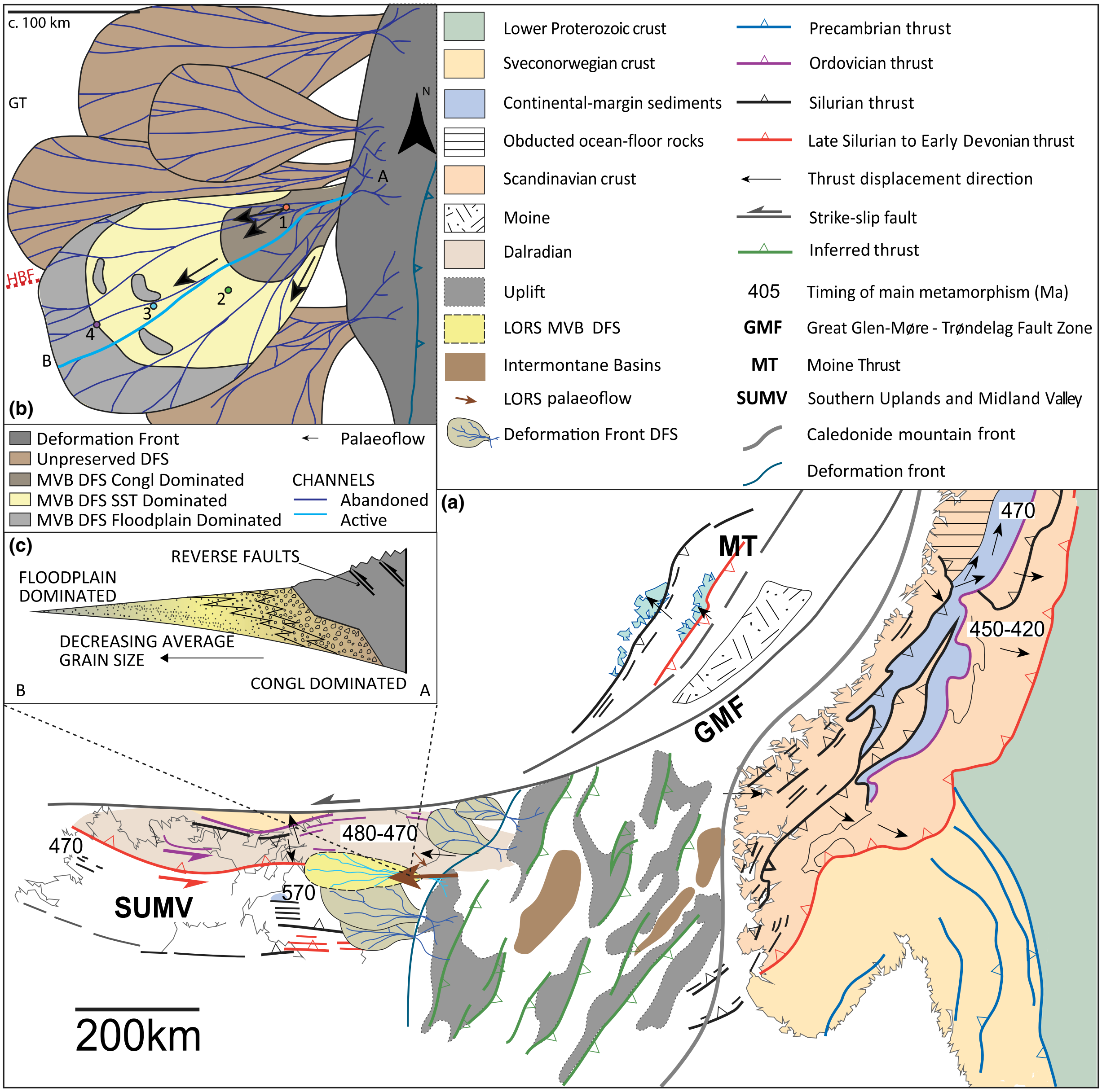

FIGURE 13 (a) Conceptual model of the Caledonide structure of the North Atlantic region during deposition of the northern MVB LORS showing development of large DFS along the mountain front. Inferred zone of uplift is based on simplification of the scale and distribution of fault blocks, uplift and basins within the Sierras Pampeanas, modified to correspond to the Caledonian thrust front. Main palinspastic map of the Caledonides from Coward et al. (2003). (a) Reconstruction of the depositional setting for the northern MVB LORS showing large DFS draining the deformation front to the east, developing in the foreland of the Caledonian thrust front associated with collision of Baltica and Laurentia. (c) Simplified cross section of the northern MVB LORS showing DFS development across the basin and downstream reduction in grain size with pulsatory tectonic influence to the east

setting. Suitable analogues for the Caledonian thrust front exhibiting the necessary thick-skinned basement uplift necessary for generation of the LORS sediment can be found in the Andean foreland. The structural style of the Andes varies along strike (e.g. Allmendinger, Ramos, Jordan, Palma, & Isacks, 1983; Watts, Lamb, Fairhead, & Dewey, 1995), the causes and implications of which are beyond the scope of this paper, and the Andean foreland is thus segmented and variable (Kleya, Monaldib, & Salfityb, 1999). In northwest Argentina the thick-skinned province of the Sierras Pampeanas represents an example of basement uplift within a foreland basin as a result of thick-skinned tectonic inversion (Jordan & Allmendinger, 1986). The Sierras Pampeanas region is a large zone of basement thrusts that forms part of the Andean orogenic zone covering an area 450 by 800 km (Figure 12). Given that compression was ongoing across the Caledonian foreland during the onset of LORS sedimentation, similar structures may have formed along the Caledonian thrust front. Basement inversion in the Caledonian foreland would thus have led to the formation of mountain ranges parallel to the thrust front, as is observed in the Andean foreland. It is possible therefore to infer a mountain range or zone of uplift similar in scale to that of the Sierras Pampeanas in the Caledonian foreland, inverting a northeast extension of the Grampian Terrane as implied by provenance data (McKellar et al., 2020; Figure 13). This uplift would have been a major contributing factor to the sediment supply necessary for the volume and character of the MVB LORS and would satisfy the requirements for the development of the MVB DFS: a source to the east, sufficient relief and proximity to generate the coarse grain size, pulsatory tectonics and continuous regeneration of the source area.

7 | CONCLUSIONS

The sedimentary rocks within the northern MVB can be assigned to two main facies associations, fluvial channel and floodplain. Within these facies associations, nine sublithofacies can be identified—four conglomerates, four sandstone and one mudstone. This study presents an interpretation of these sublithofacies and facies associations, culminating in a new model for deposition of the LORS of the northern MVB, identifying it as an example of a large-scale DFS. The system is observed to exhibit an overall decrease in thickness from east to west, a large-scale reduction in grain size and lower proportions of conglomeratic material. The abundance of amalgamated channel-belt facies associations decreases downstream, with an associated increase in the dominance of the floodplain facies. The architectural style of the deposits shows a significant change through the system, with proximal sections predominantly comprising amalgamated fluvial channel deposits, a medial portion showing a reduction in overall grain size and increase in finer-grained material, and finally the distal regions dominated by floodplain facies, with much reduced amalgamation of fluvial channel deposits. Decrease in downstream energy related to flow expansion as the river enters a sedimentary basin, channel bifurcation and infiltration are interpreted to be key controls in the overall architecture of the system. Evidence of grain size cyclicity observed within the sedimentary rocks has been attributed to a pulsatory tectonic influence and is comparable with that which is observed in several other notable formations deposited in a foreland basin setting.

Assessment of the tectonic setting allows the depositional model to be placed in the wider tectonic setting during the late stages of the Caledonian Orogeny. Previous depositional models relying on strike-slip movement or significant displacement along the HBF both during and prior to basin formation and deposition are rejected on the grounds that there is no evidence for significant post-Ordovician movement on the HBF, nor any indication that this formed the basin margin. Sedimentological observation is instead in favour of the interpretation of a large DFS forming in a foreland basin adjacent to the Caledonian thrust front to the east. Reproduction of this setting has been depicted by the application of a modern-day analogue in the Andean foreland, the Sierras Pampeanas, employing thick-skinned, pulsatory tectonics to infer an ancient mountain range parallel to the Caledonian thrust front. This mountain range and contemporaneous volcanism provided the source for the LORS sediment and the uplift and sediment supply necessary for the formation of a large DFS.

Ultimately, a new model for the tectonic setting and deposition of the northern MVB DFS is presented, consolidating new and existing sedimentological data, thus resolving previous conflictions in the geological history of the northern MVB LORS. Consequently, the LORS deposits fulfil an important role in reconstructing late Caledonian orogenic events in the region and provide a new example of a large foreland basin DFS.

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CONFLICT OF INTEREST

None.

DATA AVAILABILITY STATEMENT

The data that support the findings of this study are available from the corresponding author upon reasonable request.

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