ABSTRACT: Bottomsets are formed in the troughs of migrating bedforms. The flow and sediment dynamics downstream of bedforms differ, and consequently, the thickness, grain size, and internal structure of bottomsets vary greatly. A classification of the controls on bottomset formation is proposed related to two- and three-dimensional processes, flow and sediment unsteadiness, and pre-existing deposits. Structures and deposits created by flow and sediment unsteadiness as well as pre-existing deposits are also considered in relation to changes they can induce in subsequent trough processes. The classification is illustrated by observations from laboratory flumes, the modern Burdekin River (Australia), and the rock record (The Roaches Grit, England, and Hawkesbury Sandstone, Australia). In bottomset examples described in this paper, internal variability, both vertically and laterally, was much greater than that within the dominant foreset component. Complex compound structures were created by the interaction and predominance of different bottomset controls. This, combined with their high preservation potential, makes them a highly useful paleoenvironmental indicator. Bottomset variation (e.g., changes in the abundance of mud laterally) will influence the permeability heterogeneity of fluvial reservoir rocks. Bottomsets formed under relatively steady conditions are likely to be laterally extensive and have similar characteristics over their entire length. They may act as significant barriers to vertical flow. Conversely, bottomsets generated in unsteady regimes tend to be highly variable, potentially creating conduits between cross-bed sets aiding inter-set permeability.

INTRODUCTION

Cross-stratified deposits on all scales, from ripples to deltas, can be divided into topsets, foresets, and bottomsets. This paper focuses on bottomsets formed in the lee of dunes and unit bars, which are often dynamically related to those bedforms (Fig. 1). Cross-stratified sets containing bottomsets are common. They have been described in many depositional environments: fluvial (e.g., Basumallick 1966; Boersma 1967; Boersma et al. 1968; Collinson 1970; Ashworth et al. 2011; Reesink and Bridge 2011), lacustrine (e.g., Gilbert 1885; Jol and Smith 1991), marine (e.g., Terwindt 1971; Sohn et al. 2003; Nielsen and Johannessen 2009), and eolian (e.g., Hunter 1977; Clemmensen and Abrahamsen 1983; Eastwood et al. 2012). The examples presented herein formed at the base of unit bars and dunes migrating in rivers and flumes.

Unit bars, first defined by Smith (1974), are relatively unmodified bars with morphologies determined mainly by depositional processes. Unit bars can develop spontaneously under certain conditions (cf. Callander 1969; Seminara and Tubino 1989; Tubino et al. 1999), or as a result of forcing by channel non-uniformity (cf. Leopold and Wolman 1957; Cant and Walker 1978) or flow and sediment unsteadiness (cf. Jopling 1966; Lunt and Bridge 2007). This contrasts to dunes, which form only through spontaneous development (cf. Kennedy 1963; Richards 1980; Seminara 2010; Vesipa et al. 2014). Internally, high-angle cross-stratification often dominates the foreset component of both unit bars (e.g., McKeen 1957; Jopling 1963; Collinson 1970; Smith 1972, 1974) (Fig. 1A) and dunes (e.g., Rubin 1987; Leclair et al. 1997; Leclair 2002) (Fig. 1B). However, for the former, more complex foreset structures can develop, such as multiple down-climbing sets (e.g., Lunt and Bridge 2007; Reesink and Bridge 2011; often termed compound cross-stratification). Complex unit-bar internal structures develop due to interactions between the host bar and superimposed bedforms (e.g., Reesink and Bridge 2011) or flow unsteadiness (e.g., Jopling 1963, 1966; Lunt and Bridge 2007).

There are more influences on the deposition of bottomsets than with their associated foresets. Consequently, there is a greater variety of possible internal structures (e.g., Fig. 2). Foreset architecture is controlled predominantly by grain-flow and grain-fall processes during bedform migration. The characteristics of bottomsets that form in the troughs downstream of unit bars (Fig. 1A) and dunes (Fig. 1B) are controlled by the flow behavior over and downstream of the bedform, the characteristics of sediment in transport (e.g., grain size, density), and antecedent conditions in the trough. Bottomsets of dunes and unit bars may include: back-flow ripples (e.g., Jopling 1961; Boersma 1967; Martinus and Van den Berg 2011; Herbert et al. 2015) (Figs. 1, 2A), co-flow ripples (e.g., Boersma 1967; Boersma et al. 1968; Martinus and Van den Berg 2011) (Fig. 1A), lower- or upper-flow-regime planar lamination (e.g., Reesink and Bridge 2011), antidunes (e.g., Reesink and Bridge 2011), massive sand (e.g., Rae and Hermansen 2006), muds (e.g., Reesink et al. 2014), and coarse-grained (gravel) layers (e.g., Allen 1982; Kleinhans and van Rijn 2002; Kleinhans 2004; Reesink and Bridge 2011) (Fig. 2D). Bottomset architecture may be particularly complicated if the bedform trough contained transverse flows (e.g., Boersma et al. 1968; Reesink and Bridge 2011), or where tides were reworked before burial (e.g., reworking by low-stage channelization through bar troughs; Reesink and Bridge 2011) or where tides led to changes in flow direction (e.g., Visser 1980; Allen 1981; Van den Berg and Van Gelder 2007; Martinus and Van den Berg 2011; Martinus et al. 2015). As bottomset fabric usually differs greatly...
from that of the associated foreset, they may act as baffles to interstitial flow, influencing oil extraction (Weber 1986; Hartkamp-Bakker and Donselaar 1993) or groundwater contaminant transport (Huysmans and Dassargues 2010).

Under conditions of steady flow and constant sediment transport, bottomsets that extend over the entire set length can develop. However, flow in natural settings is rarely steady for long. Over single discharge events, changing flow conditions may lead to changes in bedform geometry that may (e.g., Gabel 1993) or may not (e.g., Allen 1973) keep pace with discharge change. Changes in flow conditions, sediment transport, and host bedform geometry cause variation in bottomset thickness, extent, composition, and internal structure. In the case of bars that persist over multiple discharge events, bottomset architecture depends on flow unsteadiness over each discharge event and longer-term variation in flow and sediment flux.

Bottomsets have higher preservation potential than the associated foresets because of their low position relative to mean bed level when deposited (Fig. 3). Bottomset thickness, relative to the associated foreset thickness, varies considerably. Initially, bottomset and foreset thicknesses are controlled by the flow, host bedform size, trough scour depth, sediment flux, and antecedent structures (see section below). Because foresets are higher than associated bottomsets (relative to mean bed level), they are more easily eroded, leading to changes in relative set thickness (Fig. 3). In addition, differences in fabric and composition of bottomsets and foresets may influence volume reduction during consolidation and compaction, causing changes in relative thickness. The ratio of foreset to bottomset preservation may be linked to bedform type. For example, Leclair and Bridge (2001) found that sets of cross-stratification were on average 34% of the formative dune height, independent of aggradation rate, suggesting full preservation of bottomsets, but loss of most of the foresets is common.

The degree of preservation influences bottomset abundance within co-sets, with poorer preservation leading to greater dominance of bottomsets (Fig. 3). Relatively thin bottomsets are common (e.g., Collinson 1970; Reesink and Bridge 2011) (Fig. 2A, B, D) whereas examples where bottomsets make up most of a deposit are rarely reported (e.g., Boersma 1967; Martinius and Van den Berg 2011, their Fig. 3.5.11, Fig. 3.5.16). If erosion repeatedly removes foresets, bottomset beds may make up the entirety of a deposit (Martinius and Van den Berg 2011) (Fig. 3F). In this scenario, it would be difficult to identify the nature of the formative bedforms and the paleocurrent direction may be misinterpreted. For
example, the bidirectionality of stacked co-flow and back-flow ripples within bottomsets where the foreset was not preserved could be erroneously attributed to tidal influence.

The high preservation potential of bottomsets and the variety of internal structures within them make them potentially valuable paleoenvironmental indicators, which have largely been overlooked by researchers focusing on more eye-catching foresets. Despite the wide range of potential variation and its ubiquity in fluvial deposits, there has been little reported on bottomset variation within fluvial deposits, although a few have observed it (e.g., Tillman and Ellis 1968; Collinson 1970; Reesink and Bridge 2011).

This paper focuses on improving understanding of bottomsets by categorizing the controls on their development (antecedent, primary 2D, primary 3D, and secondary), describing commonly occurring internal structures, discussing how bottomsets can be identified in the rock record, discussing the role of flow unsteadiness on bottomset variation, and relating the variation to reservoir heterogeneity. The role of different controls on bottomset structure is illustrated with observations from fluvial sandstones including the Carboniferous Roaches Grit, England, and the Triassic Hawkesbury Sandstone, Australia. The sites selected were easily accessible exposures of cross-stratification with associated bottomsets.

Data are also presented from the Burdekin River in Queensland, Australia, where discharge variability plays a major role in bottomset heterogeneity and from flume experiments conducted in the University of East Anglia 10 m × 1 m × 1 m recirculating flume (see Appendix for methodology and setup).

CLASSIFICATION OF BOTTOMSET CONTROLS

Primary Two-Dimensional Controls on Bottomsets

Primary two-dimensional controls on bottomsets are those related directly to the flow and sediment processes occurring over, within, and downstream of flow separation formed by triangular prismatic bedforms (i.e., bedforms with uniform size and flow-parallel shape across their full width) (Fig. 4A; Table 1). Flume experiments suggest that the host bedform height, characteristic grain path-length, sediment flux, mean flow velocity, flow depth, and trough flow deceleration all influence the extent, structure, and relative thickness of bottomset beds (Allen 1965, 1968a, 1982; Jopling 1965a; Hunter 1985; Reesink and Bridge 2009). Allen (1968a, 1982) suggested that bottomset and lower-foreset shape depend on
the suspended-sediment fallout rate in the trough compared to the rate of sediment transport in the trough towards the lee face of the host bedform, which is controlled by the back-flow power. Changes in back-flow power are linked to changes in flow-separation geometry and can be correlated to the relative flow depth at the crest of the host bedform (Allen 1986a, his Fig. 3–10, Fig. 3–11). Allen suggested that the volume of the bottomset relative to the foreset depends on the ratio of characteristic grain path-length to host-bedform height.

Martinius and Van den Berg (2011) suggested zones of bottomset development for dunes on Van den Berg and Van Gelder’s (1993) bedform stability diagram (Fig. 5). This stability diagram uses the modified mobility parameter ($h_0$) calculated from flow depth, depth-averaged flow velocity, sediment and water density, $d_{50}$, and $d_{90}$ (cf. Van den Berg and Van Gelder 1993). Figure 5 includes empirical flume data for dunes (Guy et al. 1966; Costello and Southard 1981; Leclair 2002; Reesink and Bridge 2007). Nearly all the dune data plot in or near to the suggested zone of bottomset development (Fig. 5). However, bottomsets were recorded in only ten runs, with a further six in which the descriptions infer that bottomsets were present (Fig. 5B). This low number could be due to the relatively poor documentation of bottomsets, or to their absence.

Saunderson and Lockett (1983) and Hunter and Kocurek (1986) observed dune bottomset development, but did not state in which runs. In Figure 5B, dunes with bottomsets tend to have a higher $\theta^\prime$ than a large proportion of the dunes without recorded bottomsets (Fig. 5). However, there is overlap between them, suggesting that bottomset formation is influenced by factors not considered in this plot. For unit bars, bottomset formation occurs over a wide range of flow conditions and grain sizes (Jopling 1961, 1965b; Johansson 1963; Allen 1968a; Reesink and Bridge 2007, 2009; Herbert et al. 2015).

Baas and Best (2008) demonstrated that the depth-averaged, volumetric suspended-clay concentration ($C$) can enhance (0.05 $C < 2.67\%$ in their experiments), attenuate (2.67 $C < 8.47\%$), or inhibit (10.45 $C < 12.9\%$) turbulence downstream of bedforms. Turbulence intensity and structure influence suspended-sediment fallout in the trough, trough bedform development (cf. Herbert et al. 2015), and the rate of scour at the reattachment point. Suspended-sediment concentration also affects bottomsets by influencing bedform development within the trough. Wan (1985) showed that as little as 1% volume of suspended bentonite could decrease equilibrium dune height by up to 66%. Baas et al. (2011) found that ripple height and wavelength increased at lower suspended-clay concentrations, whilst higher concentrations suppressed bedform development. Suspended-clay concentration controls the size and position of bedform stability fields (Lowe 1988; Baas et al. 2016).

Bottomsets can form without deposition from suspension, by reworking pre-existing sediment in the trough either by scour at the flow-separation reattachment point (Allen 1968b) or by turbulent packets within the recirculating eddy (cf. Herbert et al. 2015). The role of scour in bottomset development has been linked to host bedform type and antecedent bed characteristics. Allen (1982) suggests that flow separation scour is often an important driver in dune bottomset development. This is due to sand-dominated dunes nearly always migrating over a deposit of similar composition. Thus, underlying sandy sediment can be easily scoured away at the flow-separation reattachment point. If some of this sediment enters the flow-separation zone, it can contribute to bottomset development. In
contrast to dunes, unit bars have been recorded migrating over beds with greatly different composition (e.g., cohesive mud as in Ashworth et al. 2011). Where unit bars migrate over a more resistant underlying deposit (e.g., consolidated mud or gravel layer), scour is likely to play a minimal role in bottomset development. Reesink and Bridge (2011) suggested that turbulence and shear stresses generated by flow separation in the lee of unit bars may often be relatively weak compared to that of dunes, generating less scour in the trough. However, this is unlikely to apply to all unit bars, inasmuch as some types (e.g., bars formed at confluences; Ashmore and Parker 1983; Ashmore 1993) can be associated with active scour zones. Where turbulence in the flow separation zone winnows finer sediment, a coarse-grade (gravel) lag may form that can be preserved within a bottomset (Allen 1982; Kleinhans and van Rijn 2002; Kleinhans 2004).

Under steady conditions, bottomset variation can be dominated by the generation of back-flow ripples driven by flow separation in the lee of the migrating host bedform (e.g., Jopling 1961; Basumallick 1966; Boersma 1967; Herbert et al. 2015) (Fig. 1). Preservation of back-flow ripples results in a periodic thickening and thinning of the bottomset if ripple geometry is preserved at the top of the bottomset (Figs. 1B, 2A). This thickness variation is greater if ripples climbed up the foreset before burial (e.g., Boersma 1967; Herbert et al. 2015).

Under unsteady conditions, bottomset characteristics alter downstream as the conditions of flow and sediment change. In deposits of the Arkansas River, Oklahoma, Tillman and Ellis (1968) recorded a change from tangential to angular foresets, accompanied by the development of back-flow ripples. They attributed this to a decline in flow velocity. Steel and Thompson (1983) observed downstream fining and ripple development in bottomsets of cross-stratified sandstones in the Chester Formation (Staffordshire, UK), which they attributed to falling-stage flow.

Tides cause unsteadiness in many coastal rivers, and the change in flow depth, speed, and direction cause changes in bottomset character. Tidally influenced bottomsets can contain mud related to slack-water periods (e.g., Visser 1980; Allen 1981) or zones of turbidity maximum (e.g., Martinius et al. 2015). Where tidal processes dominate, ripple cross-lamination is related to flood and ebb flows (e.g., De Mowbray and Visser 1984; Van den Berg and Van Gelder 2007). In more transitional environments, where tides are weaker, ripple cross-lamination may have greater irregularity (e.g., Van den Berg and Van Gelder 2007). Bottomset thickness also varies with changes in tidal strength (e.g., Van den Berg and Van Gelder 2007; Martinius et al. 2015).

**Primary Three-Dimensional Controls on Bottomsets**

Primary three-dimensional controls on bottomsets (Table 1) are those that result from the 3D nature of the bedforms, flow, and sediment transport in natural environments (Fig. 4B). Flow three-dimensionality caused by 3D bedforms, obliquity between bedforms and the channel, and channel shape impact bottomset deposition. Bedforms are rarely triangular prisms (i.e.,
having the same size and flow-parallel shape along their full crest length. As bedform height and shape vary across the flow, curved lower erosion surfaces form (cf. Rubin 1987; e.g., trough cross-stratification) and any bottomset will be curved when viewed perpendicular to the mean flow direction.


**Table 1.** Examples of potential bottomset structures resulting from primary, secondary, and antecedent controls as well as feedback structures generated by secondary and antecedent structures influencing subsequent sedimentary processes.

<table>
<thead>
<tr>
<th>Bottomset and Trough Deposit Structures</th>
<th>Driver(s)</th>
<th>Example(s)</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Primary 2D Controlled:</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>• Back-flow ripples</td>
<td>Lee-side flow separation.</td>
<td>Jopling (1961); Fig. 2A</td>
</tr>
<tr>
<td>• Co-flow ripples</td>
<td>Flow downstream of flow-separation eddy and high sediment flux.</td>
<td>Martinus and Van den Berg (2011)</td>
</tr>
<tr>
<td>• Gravel lag at the base of a bottomset</td>
<td>Winnowing at reattachment zone.</td>
<td>Allen (1982); Fig. 16C</td>
</tr>
<tr>
<td>• Mud drapes</td>
<td>Suspended-sediment deposition.</td>
<td>Fig. 19C, E</td>
</tr>
<tr>
<td>• Tidal structures</td>
<td>Changes in flow conditions related to flood and ebb flows.</td>
<td>Martinus and Van den Berg (2011)</td>
</tr>
<tr>
<td><strong>Primary 3D controlled:</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>• Obliquely oriented ripples</td>
<td>Transverse flow along host bedform trough.</td>
<td>Boersma et al. (1968), Collinson (1970)</td>
</tr>
<tr>
<td>• Locally variably oriented ripples</td>
<td>The development of ripple fans by secondary currents.</td>
<td>Allen (1968a, 1982); Fig. 6</td>
</tr>
<tr>
<td>• Spur</td>
<td>Secondary currents within the host bedform trough.</td>
<td>Williams (1971), Smith (1972)</td>
</tr>
<tr>
<td><strong>Secondary controlled:</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>• Mud drapes</td>
<td>Suspended-sediment deposition during low stage.</td>
<td>Fig. 10</td>
</tr>
<tr>
<td>• Vegetation and roots</td>
<td>Subaerial exposure of bar trough during low stage.</td>
<td>Fig. 10</td>
</tr>
<tr>
<td>• Bioturbation</td>
<td>Biological activity during low stage.</td>
<td>Fig. 10C</td>
</tr>
<tr>
<td>• Dunes, antidunes, and upper planar bed</td>
<td>Channelization of host bedform trough during low stage.</td>
<td>Reesink and Bridge (2011); Fig. 11B</td>
</tr>
<tr>
<td>• Bubble sand</td>
<td>Subaerial exposure during low stage.</td>
<td>Reesink and Bridge (2011)</td>
</tr>
<tr>
<td>• Wave-generated structures</td>
<td>Wave activity during low stage.</td>
<td>Collinson (1970)</td>
</tr>
<tr>
<td><strong>Antecedent controlled:</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>• Vegetation and roots</td>
<td>Subaerial exposure of channel bed during low stage.</td>
<td>Figs. 9, 10</td>
</tr>
<tr>
<td>• Gravel layer or extensive cohesive muds.</td>
<td>Pre-existing deposits laid down in an unsteady fluvial regime.</td>
<td>Figs. 9, 11A</td>
</tr>
<tr>
<td>• Pre-existing scour hollows in the trough zone</td>
<td>Pre-existing geometry developed in an unsteady fluvial regime.</td>
<td>Harms and Fahnestock (1965)</td>
</tr>
<tr>
<td><strong>Feedback controlled:</strong></td>
<td></td>
<td></td>
</tr>
<tr>
<td>• Vegetation-induced scour or structures</td>
<td>Flow processes around secondary or antecedent vegetation.</td>
<td>Fig. 11C, E</td>
</tr>
<tr>
<td>• Infilling of scour hollows in the trough zone</td>
<td>Antecedent scour trapping sediment during discharge events.</td>
<td>Harms and Fahnestock (1965)</td>
</tr>
</tbody>
</table>

**Fig. 5.**—Dune flume experiments plotted on the bedform stability diagram of Van den Berg and Van Gelder (1993). **A** Dune experiments in which no bottomsets were noted. **B** Dune experiments with bottomsets (red circles) and possible bottomsets (blue diamonds) based on experimental descriptions. The modified mobility parameter, $\theta'$, is calculated from flow depth, depth-averaged flow velocity, sediment and water density, median grain size, and the 90th percentile of grain-size distribution (cf. Van den Berg and Van Gelder 1993).
This flow structure generates flow cells that can induce the formation of ripple fans and spurs (Allen 1968a, 1982; Herbert et al. 2015). Ripple fans consist of concentrically aligned ripples formed within and downstream of flow separation (Fig. 6). Spurs are elevated ridges that extend downstream into dune and bar troughs (Fig. 7A) and can form in association with ripple fans (Allen 1968a, 1982). Williams (1971) described the internal structure of a spur within a dune (he termed “large-scale ripple”) that he trenched perpendicular to flow direction in the ephemeral Wooldridge River, South Australia. The spur’s ridge shape was preserved, resulting in a significant displacement (relative thinning) of the foresets above it, and thickening of the bottomset into the ridge. The spur contained gently inclined cross-stratification (cf. Williams 1971, his Plate VI B). The internal structure of spurs depends on whether they form from deposition at flow cell seams (Fig. 7B) or from scour at the center of flow cells (Fig. 7C). For the latter, internal structure may relate to antecedent processes and be unrelated to the flow in the trough.

There may be a component of flow along the trough of relatively large bedforms whilst they are migrating. This is common at channel bends (e.g., Boersma et al. 1968), confluences (e.g., Bridge 1993), and with lobate and obliquely orientated bedforms (e.g., Smith 1972). Along-trough flow alters bottomset structure, creating ripple trains that migrate obliquely or at right angles to the migration direction of the host bedform (e.g., Boersma et al. 1968; Collinson 1970; Reesink and Bridge 2011).

Along-trough flow can cause spur migration (e.g., Smith 1972; McCabe 1977). Recent flume research suggests that spur migration can also be induced directly by an obliquely oriented bedform crest line. This leads to secondary currents on one side of the spur dominating, inducing migration of spurs along the avalanche face of a bedform (Swanson et al. 2017). The internal structure of migrating spurs includes undulatory bottomset bedding (e.g., McCabe 1977) and high-angle cross-stratified sets which climb up part of the bar lee (Smith 1972, his Fig. 10C).

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Bedforms vary greatly in height and lee-face orientation along their length. This variation alters the length and strength of the flow separation eddy and the strength of any along-trough flows. Variation in stream flow velocity along the crest of a bedform influences the structure and composition along any trough sediments via localized changes in backflow velocity and in the sediment flux reaching the trough (amount and grain size). This is likely to be important for bank-attached bars where flow close to the bank can differ greatly from that away from the bank. For lobate bedforms, local flow conditions along different parts of the lee face create laterally restricted zones with differing trough characteristics (Fig. 4B). For example, variations in the grain size and bedforms (type, size, and orientation) along troughs of lobate unit bars in the South Saskatchewan River, Canada, were recorded by Ashworth et al. (2011, their Fig. 3a).

Secondary Controls on Bottomsets

Secondary controls (Table 1) alter bottomset characteristics during periods of host-bedform inactivity (Fig. 4C). Upon bedform reactivation, changes in trough characteristics induced by these controls can become buried and incorporated into the bottomset. During low river stage, bars may become inactive and stall, and in some cases become subaerially exposed (e.g., Collinson 1970; Ashworth et al. 2011; Reesink and Bridge 2011). Water may persist in the troughs, forming a low-stage channel.

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Low-stage flow can form bedforms (ripples, dunes, or antidunes) that migrate along the trough (cf. Reesink and Bridge 2011). Alternatively, water in a trough may be static or slowly flowing, depositing sediment from suspension (e.g., Sambrook Smith et al. 2006; Reesink and Bridge 2011; Lunt et al. 2013). Wind-generated waves can rework sediment in troughs at low stage, altering back-flow ripples to symmetrical profiles or forming new wave ripples (Collinson 1970; Steel and Thompson 1983). Subaerial exposure allows eolian reworking of fine-grained trough sediment, forming eolian ripples. Mud in troughs may desiccate and crack. Periods of host bedform inactivity allow bioturbation and vegetation growth (e.g., Fielding et al. 1997; Nakayama et al. 2002). In unsteady fluvial regimes, bar elevation is an important control on trough inundation, subaerial exposure, and vegetation growth.

**Antecedent Controls on Bottomsets**

Antecedent controls (Table 1) are those that create deposits or structures, which are incorporated into bottomsets but which were formed before host bedform development or migration into the area under consideration (Fig. 4D). Thus, sediment and structures in the lee of a bedform may have formed by some process not related to the processes and conditions that generated the bedform.

One situation where antecedence is important is where scour (generated by the flow structure over bedforms) cuts down into sediment that predates the conditions that formed the bedforms, thus generating a new sediment source in the trough. If antecedent deposits are easily scoured, the incorporation of antecedent sediment into a bottomset may greatly influence its final character. Another is where bedforms migrate into an area onto a sedimentary surface of a different character, which is then incorporated into the base of the bottomset.

Antecedent erosion can control trough geometries and thus influence subsequent bottomset-forming processes. Harms et al. (1963) and Harms and Fahnestock (1965) observed antecedent scour hollows downstream of bars in the Red River, Louisiana, and the Rio Grande, Texas, respectively. Similar scour features were later observed by Williams and Rust (1969) in the Donjek River, Canada. Localized depressions, such as these, will alter the geometry of the bottomset lower bounding surface if not infilled before the bedform reaches it. Bar-top hollows (cf. Best et al. 2006), which can develop on the stoss of compound bars, are likely to have the same effect on the bottomsets of any superimposed dunes or unit bars.

In rivers with very variable discharge, the grain size transported in successive events may differ. Thus, gravel may be deposited in one event and sand in a subsequent event. During the sand-depositing event, gravel may have minimal mobility. Sandy dunes and unit bars may form with the antecedent gravel exposed in their troughs. Sand deposited in the trough may infiltrate the gravel, appearing to incorporate pebbles or cobbles into the base of the bottomset (i.e., the gravel part of the bottomset is antecedent, deposited before the formation and growth of the new bedform). This process leads to the formation of layers that appear to be “gravel lags,” independent of scour (examples of this are identified in the bottomset-examples section below). Conversely, if conditions before the formation of a host bedform generated a consolidated mud drape, this could also be incorporated into the bottomset of a unit bar.

Some processes that are considered secondary when they alter the trough zone can be considered antecedent if they create structures downstream of the bedform trough that are eventually incorporated into the bottomset (possibly after a number of discharge events) as the bedform migrates. In some cases, it may be difficult to determine if a bottomset characteristic is of secondary or antecedent origin, and it may be a combination. Examples include the development of mud drapes or vegetation during low stage which extend far downstream of the host bedform trough (see bottomset-examples section below).

**Feedbacks on Subsequent Processes Generated by Secondary and Antecedent Controls**

Structures are formed by secondary and antecedent controls. Before their burial by the migrating host bedform, these structures can influence subsequent primary and secondary processes (Table 1). One key example is how trough deposits influence water flow through the bedform.
FIG. 9.—Satellite photographs over a five-year span at the Burdekin River near the Inkerman Bridge. Photographs taken: A, E) in February 2016; B) in December 2011; C) in December 2012; D) in January 2014. Satellite images courtesy of the DigitalGlobe Foundation.
FIG. 10.—Deposits downstream of a 2-m-high, bank-attached bar formed along the southern margin of the Burdekin River channel at the Burdekin River field site. A) Overview photogrammetric reconstruction (using Agisoft Photoscan) of the bar trough constructed from photographs collected using a Phantom 2 unmanned aerial vehicle. Bar-trough deposits contained: B) relatively thick (ca. 40 mm) mud with desiccation cracks; C) consolidated sand and mud with burrow holes (denoted with white arrows); D) relatively thin mud (ca. 10 mm) with localized desiccation cracks; E) consolidated sand and mud with cracks. The locations of Parts B to E are noted in Part A. The pen is 0.14 m long.
FIG. 11.—Sedimentary structures found in the troughs of unit bars formed along the southern margin of the Burdekin River channel at the Burdekin River field site. A) Gravel layer, in front of a 0.3-m-high unit bar. B) Dune (90 mm high) which migrated perpendicular to the associated unit bar; notebook 205 mm long. C) Sediment tail; shovel 0.6 m long. D) Sediment tails close to a 0.4-m-high unit bar avalanche face. E) Trench, dug parallel to the local trough flow, through a sediment tail. Solid curved lines in Parts A, B, and D denote bar crests, the dashed line in Part B denotes a dune crest.
and then decline almost as quickly (Alexander et al. 1999; Amos et al. 2017).

Vertical flow exiting the bed in a trough can displace the reattachment zone of hyporheic flow. Pressure variation over a bedform induces flow into and out of the bed (advective pumping; Elliott and Brooks 1997a). Higher velocities and altered flow paths due to bedforms and structures can greatly alter subsequent flow processes (Fielding et al. 1997). Submerged vegetation in a trough may decrease mean flow velocity and alter flow near the bed. Large aquatic vegetation in a trough can alter hyporheic flow and thus the flow structure in the trough. For example, a laterally extensive mud drape in the trough formed at low stage could act as an impermeable barrier to hyporheic flow in subsequent higher-stage events.

Other feedbacks also exist. Bedforms and structures developed during low-stage trough channelization alter trough geometry. This influences trough processes until the bedforms and structures are washed out or buried by the reactivated host bedform. Scour depressions in the trough alter local flow and depositional processes, and can infill with fine-grained deposits (e.g., in the Rio Grande, Texas, as described by Harms and Fahnestock 1965). Vegetation that has grown on an exposed bed can be submerged during either high-stage (reactivating the bar) or low-stage (trough channelization) events, leading to localized scour, obstacle marks (Nakayama et al. 2002; Rodrigues et al. 2007), and bar growth (Fielding et al. 1997). Submerged vegetation in a trough may decrease mean flow velocity and alter flow near the bed. Extensive mud deposition in a trough is likely to decrease bed roughness. Antecedent mud drapes or gravel layers may limit the contribution of scour to bottomset development (desiccation and consolidation of mud makes it more difficult to erode when resubmerged, e.g., Sundborg 1956; Postma 1967; Dade et al. 1992). A high content of clays or extracellular polymeric substances (EPS) influences bedform dimensions and development times (Baas et al. 2013; Malarkey et al. 2015; Parsons et al. 2016). This will be the case in troughs if sediment is derived from pre-existing clay or EPS-rich trough sediments.

The wide range of potential feedbacks suggest that secondary and antecedent processes can greatly alter subsequent flow processes downstream of bedforms, independently of changes in bedform geometry and discharge. These feedbacks, by altering subsequent primary or secondary controls on bottomset development, further drive bottomset variation.

EXAMPLES OF BOTTOMSET VARIATION IN MODERN RIVERS AND ANCIENT FLUVIAL DEPOSITS

**Bottomset Formation in the Modern Burdekin River**

The Burdekin River, north Queensland, Australia (130,000 km² catchment), is dominated by short-duration, large-discharge events separated by long periods of minimal flow with great inter-annual variability. Discharge events, caused by monsoon rain or tropical cyclones, rise rapidly to a peak up to three orders of magnitude larger than base flow and then decline almost as quickly (Alexander et al. 1999; Amos et al. 2004).

The riverbed is gravelly coarse sand, with drapes of finer sand and mud sometimes forming in topographic lows. For most of the year, most of the bed of the mid and lower reaches is subaerially exposed, with the flow limited to the lowest parts. Bars, plane beds, dunes, and antidisites are observed on the exposed bed (e.g., Fielding and Alexander 1997; Alexander and Fielding 1997). Bed exposure allows vegetation growth that subsequently influences sedimentary processes (Fielding et al. 1997; Nakayama et al. 2002). The amount and size of vegetation is controlled by discharge variability, with greater amounts growing following successive years of low discharge.

The field site, 17 km from the river mouth at Inkerman Bridge (19° 38' 8.58" S, 147° 24' 14.40" E; Figs. 8A, 9A) is at the downstream end of a nearly straight reach 10 km long. Over this reach, the channel varies from 500 to 1500 m wide and is 800 m wide at the bridge. In March 2012, peak discharge recorded at the Clare gauging station was 17,000 m³ s⁻¹ (Fig. 8). From then until September 2016 only four small discharge events occurred (between 1000 m³ s⁻¹ and 5000 m³ s⁻¹), all much less than bank full. Thus, bedforms observed in topographically high areas during field campaigns in August 2015 and 2016 had been inactive since 2012. Bedforms at lower elevations, near the base-flow channel, were formed or modified by recent, smaller discharge events.

Some of the unit bars studied were reactivated during the February 2016 (2250 m³ s⁻¹ peak) discharge event (Fig. 8B). The height of bar avalanche deposits ranged from 0.3 m to 2 m, and crest lines extended hundreds of meters across the exposed riverbed. One of the bars examined was bank attached. Trenches, located using GPS, were dug to reveal unit-bar internal structure. Satellite imagery was used to identify channel-bed changes since March 2012. Trough deposits of exposed bars were examined to determine the likely degree of bottomset variability that will be generated when the bars are next reactivated. Trough deposits downstream of bars showed secondary and antecedent variability.

The largest bar examined (2 m high, 220 m long, bank attached) was last active in March 2012 (Fig. 9), when its well-defined avalanche face formed. Subsequent smaller discharge events inundated the trough downstream of the bar, but did not reactivate the avalanche face (apart from a small section close to the base-flow channel), resulting in a number of secondary controlled changes in trough characteristics whilst the bar has stalled. In December 2012, the surficial sediment in the trough consisted mostly of medium sand, slightly finer than the coarse sand and granules that dominate the avalanche face. Small discharge events in 2012 reworked the northern margin of the bar, depositing mud close to the low-flow channel (Fig. 8C). Over the subsequent four years more fine-grained sediment was deposited in the trough, with the zone expanding southward (Fig. 9D, E, represented by the darkening trough bed). By 2016, fine sediment covered the entire trough and vegetation had developed (Fig. 9E). The vegetation density was highest and mud thickest near to the low-flow channel (Fig. 10A, Location B; Fig. 10B, 40 mm of desiccated mud blankets the 2012 medium sand). The mud layer thinned with distance from the low-flow stream and with increasing elevation. On topographic highs in the trough (e.g., Fig. 10A, Location C; ca. 0.3 m above the surrounding bed) the top 10 mm of the sediment was well consolidated muddy sand with 5 mm circular burrows (Fig. 10C). In lows (e.g., Fig. 10A, Location D) a 10 mm partially desiccation-cracked mud overlay the 2012 medium sand (Fig. 10D). Close to the channel bank (Fig. 10A, Location E) vegetation was sparser and the upper layer of the sand was muddy and well consolidated (Fig. 10E). In some areas, the bed had been modified by off-road vehicles.

Smaller, topographically lower, unit bars were reactivated during the relatively small discharge events, migrating up to tens of meters in each event. The troughs downstream of many small unit bars contained antecedent rounded gravel (Fig. 11A; clasts up to 0.1 m in diameter) in flow-parallel streaks (cf. Alexander and Fielding 1997). These gravel streaks were probably deposited in March 2012, not moving in the subsequent smaller flows. Instead, the static gravel streaks were reactivated during the February 2016 (14.40 m³ s⁻¹) discharge event (Fig. 8B). The height of bar avalanche faces ranged from 0.3 m to 2 m, and crest lines extended hundreds of meters across the exposed riverbed. One of the bars examined was bank attached. Trenches, located using GPS, were dug to reveal unit-bar internal structure. Satellite imagery was used to identify channel-bed changes since March 2012. Trough deposits of exposed bars were examined to determine the likely degree of bottomset variability that will be generated when the bars are next reactivated. Trough deposits downstream of bars showed secondary and antecedent variability.

The February 2016 discharge event (see Fig. 8B) was channelized along the troughs of some small unit bars, generating secondary controlled...
structures. Dunes formed and migrated at right angles to unit-bar avalanche faces (Fig. 11B). Vegetation had grown within and downstream of exposed bar troughs by February 2016. Subsequent feedbacks between the vegetation and water flow created obstacle marks with teardrop-shaped sediment tails (cf. Nakayama et al. 2002; Rodrigues et al. 2007) which ranged from 0.2 m to > 10 m long (Fig. 11C–E). The tails were finer grained than the surrounding trough deposits and contained mud either distributed throughout the upstream facing head of the mound or deposited as a layer at its base (Fig. 11E). As observed by Nakayama et al. (2002), mud helped harden the surface of the tail. Some mounds had a horseshoe-shaped scour upstream. Dunes, vegetation, and sediment tails increased variability in the structure, thickness, and grain size of deposits in the trough. If these unit bars are reactivated by a subsequent discharge event, then the trough variability observed will likely be incorporated into the bottomset component of the bars.

Figure 12 shows two 1.5 m trenches dug ca. 7 m apart into a lobate-unit-bar avalanche face that migrated and stalled during the small discharge events post 2012. The bottomset and internal structure of the unit bar differ markedly. A single set of planar cross-stratification (Fig. 12A) with a poorly sorted bottomset of mud, sand, and gravel (diameter < 100 mm; Fig. 12C) was exposed in the first trench, whilst in the second (Fig. 12B) the bar deposit had a more complex internal structure of multiple down-climbing cross-stratified sets. This more complex forest structure likely developed as a result of localized periodic washout of the unit-bar lee driven by reattachment scour of superimposed dunes when the bar was active (cf. Reesink and Bridge 2007, 2009). In the second trench, close to the bar lee, two bottomset deposits were exposed. The first developed along the base of the unit bar, and the second formed along the basal surface of a large down-climbing set (Fig. 12D). In contrast to the first trench, both bottomsets were better sorted, containing medium to coarse sand. The variation in bottomset structure observed in this bar is likely a result of a combination of lateral changes in trough flow processes when the bar was active as evidenced by the changing internal bar structure (primary 3D control), local mud deposition possibly laid down after the bar stalled (secondary control), cobbles present on the channel bed before bar formation (antecedent control), and localized secondary and antecedent muds and gravels altering primary trough processes during the February 2016 discharge event (feedback-induced variability).

**Bottomsets in the Roaches Grit**

The Roaches Grit, a 15–90-m-thick deposit in the North Staffordshire Basin, UK, consists of coarse-grained feldspathic fluvial channel bodies and finer interchannel deposits that formed in a paralic setting near the Namurian equator (Jones 1980; Golonka 2002). The fluvial deposits overlie firm seilstones and mudstones, some of which are turbidites, formed in a delta slope setting (Jones 1980). The North Staffordshire Basin formed in the Late Devonian to early Carboniferous by north–south extension, and was in a phase of thermal subsidence by the Namurian (Leeder 1982, 1988; Waters et al. 2008). This, along with rapid glacio-eustatic sea-level variation (Jones and Chisholm 1997; Jones 2014) controlled the sedimentary architecture.

The coarse-grained sandstones of the Roaches Grit are exposed in natural escarpments at the Roaches, UK (53° 09’ 23.79” N, 1° 59’ 28.77” W; Fig. 13A). Here, paleocurrents are predominantly towards the northwest (Jones 1979, his Fig. 4). Jones (1980) records 11 channel fills in erosional contact, one being at least 600 m wide. They have been interpreted as forming in low-sinuosity, 1-km-wide channels in delta-top settings (McCabe 1977; Jones 1979, 1980; Jones and McCabe 1980).

The fluvial sandstone bodies consist of three major lithofacies: faintly laminated coarse sandstone (up to 10 m thick), “giant” cross-stratification (sets up to 20 m thick), and overlying those, co-sets of thinner tabular cross-stratification. The tabular sets, described below, are 0.5–2 m thick and are tens to hundreds of meters long, arranged in co-sets 10–35 m thick (Jones 1980). These tabular co-sets are interpreted by McCabe (1977) and Jones (1979) as “sand wave” deposits that formed on bar tops (the bars forming the “giant” cross-bed sets) and in shallow channel areas. Jones and Chisholm (1997) suggested that they were formed by migrating bars.

A co-set of tabular cross-stratification, 50 m long and up to 5 m high, is exposed at the Roaches (53° 10’ 10.04” N, 1° 59’ 58.53” W; Figs. 14, 15). Sedimentary structures become less visible towards the north end. All of the sets are < 3 m thick. Some sets wedge out whilst others can be followed over the entire exposure length. The number of stacked sets vary from 3 to 8 across the exposure. Only the uppermost set has tangential foresets over the full length of the exposure; others generally contain angular foresets, though these locally become more tangential over short distances (< 10 m). Within individual sets, adjacent cross-strata vary from fine to medium sand to fine gravel. The thickness of individual cross-strata vary from < 5 mm (cross-lamina) to > 50 mm (cross-bed), with the thicker cross-strata tending to be coarser grained. The exposure is tilted towards the east; the strike and dip of the bedding and exposure face at each logging point was used to calculate true thicknesses.

Bottomsets occur at the bases of most sets (Figs. 14, 15). However, at the center of the exposure the three lowest sets lack them (Figs. 14A, 15A). Bottomset thickness ranges from 0 to 180 mm. They vary from massive sandstone to back-flow ripple cross-laminated sandstone. The grain size varies from fine sand to rarer medium gravel. The bottomsets in many cases are of coarser grade than the finer grained cross-strata in the associated foresets. Towards the south of the exposure (P1 in Figs. 14, 15E, F) the thickness and internal structure of individual bottomsets change little (< 33% thickness decrease over ca. 10 m); however, both thickness and internal structure differ between adjacent sets. Towards the middle of the exposure, ca. 20 m north of P1 (P2 in Fig. 14), bottomset variation is greater. Bottomset thickness tends to decrease along individual sets, one by as much as 79% over ca. 10 m, and the internal structure varies. For example, one bottomset changes from back-flow ripple cross-lamination interfingered with foreset cross-stratum to massive sandstone. A second set has poor foreset preservation leading to only the bottomset being preserved over ca. 5 m (Fig. 15C). A little farther to the north, (zone P3 in Fig. 14) the bottomset thickness varies locally over short distances (e.g., 330% increase in thickness over < 0.15 m in Fig. 16A). Here bottomsets thin and thicken in the paleocurrent direction, some locally wedging out and reappearing farther down paleoflow (Fig. 14). Over the exposure, bottomset thickness also varies over short distances where back-flow bedforms interfinger with foreset cross-stratum (varies from 105 to 140 mm in Fig. 16B). Localized changes in bottomset thickness are often accompanied by changes in the preserved thickness of the overlying foresets (e.g., Fig. 16A). A local thickening of a bottomset thins the overlying foresets, as the total bottomset and foreset thickness remains relatively consistent. Towards the north end of the exposure, one bottomset has a basal pebble lag (Fig. 16C).

Most of the changes in back-flow ripples, back-flow ripple cross-lamination, and bottomset thickness at the site were probably caused by changes in primary 2D and 3D controls. These controls might be distinguished if the orientation of ripples could be interpreted, but in this case, this was not possible because of the weathered character of the exposure. The development of bottomsets only in the upper part of the co-set suggests a change in flow or sediment transport over the deposition of the co-set. This could have resulted from an increase in the strength of reattachment scour induced by relatively long-term moderate increases in flow velocity (primary 2D control), with sediment liberated from the bed contributing to bottomset development. The similarity in grain size between the foreset and bottomset deposits suggests that the above proposal is the more likely driver of this transition than an increase in the suspended-sediment flux to the trough zone. The localized increase in bottomset thickness shown in Figure 16A may be due to the development...
(or migration) of a spur (primary 3D control) or obstacle mark with a sediment tail (secondary control) in the trough zone, locally thickening the bottomset (see Fig. 16A schematics). Alternatively, the bottomset may have thickened as a result of a rapid increase in sediment flux to the trough zone (primary 2D control). Localized development and preservation of back-flow ripples (Fig. 16B) is possibly due to increases in back-flow velocity (and the magnitude of associated turbulent packets) in the flow separation zone (see Fig. 16B schematics), possibly driven by short-term increases in

Fig. 12.—A, B) Photogrammetric reconstructions of two trenches dug into a unit bar avalanche face at the Burdekin River field site; trenches were dug ca. 7 m apart. C, D) Photographs of the bottomsets shown in Parts A and B respectively. White dashed lines in Parts C and D denote bottomset bounding surfaces. The pen in Part C is 0.14 m long.
flow velocity (primary 2D control; cf. Herbert et al. 2015). The development of a localized thin gravel lag (Fig. 16C) could suggest increases in reattachment scour also induced by short-term increases in flow velocity (primary 2D control, see Figure 16C schematics). Alternatively, it could relate to an increase in gravel content in the underlying set (anecdotal feedback control).

Bottomsets in the Hawkesbury Sandstone

The Middle Triassic Hawkesbury Sandstone of the Sydney Basin (Australia) was deposited around 60° S of the equator (Golonka 2002) in a braided-river system (Conaghan and Jones 1975; Conaghan 1980; Rust and Jones 1987). Component channels were generally 5–10 m deep, reaching up to 20 m at confluences. Channel-scale architectural elements ranged from about 0.5 km to 3.2 km wide (Miall and Jones 2003). Paleocurrents were towards the northeast (Conaghan 1980; Rust and Jones 1987). Conolly and Ferm (1971) and Miall and Jones (2003) suggested that there was some tidal influence on deposition.

Coastal cliffs and road cuttings expose three major lithofacies: stratified sandstone, massive sandstone, and mudstone. The stratified sandstone is dominated by planar cross-stratification (sets 0.1–7.5 m thick, Rust and Jones 1987) interbedded with minor amounts of trough cross-stratification, ripple cross-lamination, and massive sheet sandstones. Some cross-stratified sets contain reactivation surfaces and bottomsets of back-flow ripples (Ward 1972) and mud drapes. The planar cross-stratification has been attributed to “sand wave” migration (Conaghan and Jones 1975; Conaghan 1980) or dune migration (Miall and Jones 2003) and larger sets to confluence-bar formation (Rust and Jones 1987). The massive sandstone lithofacies has been interpreted as bank failure deposits (Rust and Jones 1987) or upper-stage plane-bed deposits (Conaghan 1980), whilst the mudstone has been interpreted as overbank flood deposits (Rust and Jones 1987; Conaghan 1980).

Exposures containing bottomsets were examined at four sites around Sydney (Fig. 13B). A 19 m × 3 m exposure in the Kamay Botany Bay National Park on the Kurnell Peninsula (Site HS1; Fig. 13B; 34° 00' 46.93° S, 151° 13' 50.39° E) consists of a sandstone co-set of up to seven planar cross-stratified sets, some of which terminate within the exposure (Figs. 17, 18). Fine to medium sand predominates, with some mud drapes toward the base of cross-strata or within bottomsets. In the lower sets, foreset cross-strata are angular, but higher sets tend to have cross-strata that contact bottomsets tangentially (Fig. 17). Between individual cross-strata, there is less variation in both thickness and grain size than observed in the Roaches Grit.

The bottomsets vary in thickness between sets and laterally within sets, ranging from 0 to 0.28 m (Fig. 17A–C). They appear massive and are generally thinnest where foresets wedge out. Their thickness does not vary rapidly over short distances (e.g., a quadrupling in bottomset thickness over a few hundred millimeters, as in Fig. 16A), instead varying gradually along sets. The lowest foreset is 0.91 to 1.11 m thick whilst the associated bottomset is 130 to 280 mm thick over 19 m. There are multiple thin (< 1 mm) undulating mud drapes locally in the upper 30% of the bottomset and along the lower portions of the forest cross-strata (Figs. 17D, 18D). At a few locations, forest cross-stratification extends into the bottomset and their dip reduces (Fig. 17E).

The gradual changes in the basal bottomset thickness at this site may represent a change in suspended-sediment deposition into the trough (primary 2D control) during host bedform migration. The thin mud drapes in parts of the fine-grained bottomset suggest calm trough conditions.

At Site HS2 (Fig. 13B), a 14 m × 1.5 m section of a longer cliff face at Cronulla (34° 03' 25.83° S, 151° 09' 26.30° E) consists of a co-set of planar cross-stratification (Fig. 19) overlain by trough-cross-stratification and massive sandstone deposits. The planar cross-stratified sets extend over 10 to 30 m before wedging out. Their cross-strata range from angular to tangential, and some vary within individual sets. Cross-strata grain size (fine to medium sand) and thickness (most < 10 mm) vary less than in the Roaches Grit exposure. Coarse sand and granules occur in some overlying massive or trough-cross-stratified sets.

Only the two planar cross-stratified sets have bottomsets (Fig. 19A, B). The thickness of the bottomset in the lowest of these increases from 35 to 145 mm over 1.5 m (a 314% increase, P4 in Fig. 19A). This
Fig. 14.—A, B) Schematics of sedimentary structures preserved within the exposure examined at the Roaches Grit. Bottomsets are highlighted in red. C) Photogrammetric model of exposure face denoting the locations of schematics A and B. Rose diagram denotes foreset dip direction in eight cross-stratified sets measured over the exposure.
FIG. 15.—Measured sedimentary logs of the exposure examined at The Roaches. Bottomsets are highlighted in light red. The position of each log is identified in Figure 14A and B. Solid lines between logs denote the continuation or wedging out of sets; the dashed lines note where the sets have been partially eroded away. The zigzag lines on bottomset deposits denote their relatively poor sorting. For illustrative purposes, cross-strata in the logs were drawn dipping to the right; see Figure 14 for information on true dip direction.
Fig. 16.—Roaches Grit photographs of A) a bottomset, third set up, that increases in thickness from ca. 14 mm to 60 mm over < 0.15 m; B) a bottomset that varies in thickness (105 to 140 mm) due to climbing back-flow bedforms; C) a pebble lag (grains up to 10 mm in diameter). White dashed lines denote bottomset bounding surfaces. White lines in Part B denote the bounding surfaces of climbing cross-stratification within the bottomset. White arrows in Part C denote pebbles. Each photograph has two associated schematics; these denote how a change in fluvial conditions over time (from $t_1$ to $t_2$) may have generated these laterally restricted bottomset structures.
FIG. 17.—A, B) Schematics of sedimentary structures preserved within the Hawkesbury Sandstone exposure examined at Site HS1. Bottomsets are highlighted in red. Dashed lines denote likely bottomset boundaries where contact with the foresets is poorly defined. Dotted lines denote reactivation surfaces. C) Photogrammetric model of exposure face denoting the locations of schematics A and B. D) Thin mud drapes (< 1 mm) distributed throughout upper parts of the bottomset (pen is 0.14 m long). E) Infrequent cross-stratum that extended into the bottomset (denoted with black arrows). The locations of Parts D and E is noted in Part C. Rose diagram denotes foreset dip direction in five cross-stratified sets measured over the exposure.
increase is accompanied by a decrease in preserved overlying foreset thickness. The thickness and abundance of mud drapes in the bottomset varies with bottomset thickness. They are relatively thick (1–3 mm; Fig. 19C) at the boundary with the foreset, where the bottomset is relatively thin (ca. 30 mm). Where the bottomset is thicker (ca. 150 mm, Fig. 19D, E), mud drapes are localized, thinner, and are more evenly distributed over the entire bottomset thickness (Fig. 19E). Gentle undulations along parts of the foreset–bottomset boundary could be small back-flow ripples (cf. Herbert et al. 2015), and some thin mud drapes highlight possible bounding surfaces of faint ripple cross-lamination.

There is greater bottomset variation both along and between sets in this section compared to the exposure at Site HS1. Rapid changes in bottomset thickness occur with changes in cross-strata shape, suggesting a rapid and large change in primary 2D controls (e.g., changes in flow or sediment transport). The lack of bottomsets in the upper part of the co-set suggests that there was also long-term variation in flow or sediment conditions, such as a decrease in sediment flux to the trough. The relatively thick mud drapes (Fig. 19C) were not associated with reactivation surfaces and formed at the top of the bottomset, and occasionally extended a short way up the cross-strata. It is likely that they were deposited in a relatively weak flow-separation zone (primary 2D control). The decrease in the abundance of mud drapes close to the top of the bottomset after the bottomset thickens may point to an increase in turbulence in this zone hindering mud deposition close to the migrating lee or a decrease in marine influence (less flocculation near site). Mud instead deposited farther out in the trough, draping ripples. Such variability could be induced either in an unsteady purely fluvial or tidally influenced fluvial regime.

At Site HS3 (Fig. 13B), a cliff section at Cape Banks, Sydney (34° 00' 04.1" S, 151° 14' 57.7" E, Figs. 20, 21) consists of a 2-m-thick, fine- to medium-grained co-set of trough cross-stratification with bottomsets overlain by a set of coarse-grained cross-stratification ca. 6 m thick (Fig. 20D). Two cliff sections at 110° from each other cut the lower cross-stratified co-set approximately parallel (Fig. 20) and perpendicular (Fig. 21) to the migration direction of the formative bedforms. The cross-strata are slightly tangential close to their base; grain size variation across them is minimal.

Bottomsets occur throughout the co-set of trough cross-stratification and are mostly massive with thickness varying gradually along the sets (Fig.
FIG. 19.—

A) Schematic of sedimentary structures preserved within the Hawkesbury Sandstone exposure examined at Cronulla (Site HS2). Bottomsets are highlighted in red. Dotted lines denote possible reactivation surfaces. B) Photogrammetric model of a section of the exposure face. C) A 30-mm-thick bottomset featuring a relatively thick (ca. 3 mm) mud drape along the bottomset-foreset boundary. D) The same bottomset as in Part C but here 150 mm thick. E) A 150-mm-thick bottomset containing thin mud drapes distributed throughout. The locations of Parts C and D are noted in Part B. White dashed lines in Parts C to E denote bottomset bounding surfaces. The upper bottomset surfaces in Part C is denoted by the mud drape. Part B shows dip direction in rose diagram denoted in right cross-sections measured over the exposure.
FIG. 20. A) Photogrammetric model of the Hawkesbury Sandstone exposure examined at Cape Banks (Site HS3), along the migration direction of the formative bedforms. B) Schematic of the internal structure, with 2× vertical exaggeration. Bottomsets are highlighted in red. C) Photograph of the lower co-set of cross-stratification. D) Overview photograph of the site. Rose diagram denotes foreset dip direction in six cross-stratified sets measured over the exposure.

BOTTOMSET ARCHITECTURE

Fig. 21.—A) Photogrammetric model of the Hawkesbury Sandstone exposure examined at Cape Banks (Site HS3), across the migration direction of the formative bedforms. B) Schematic of the internal structure, with 2× vertical exaggeration. Bottomsets are highlighted in red. C, D) Photographs of the amalgamation of bottomsets where cross-stratified troughs wedge out. The locations of Parts C and D are noted in Part A. Black dashed lines in Parts C and D denote bottomset bounding surfaces. Pen is 0.14 m long.
20). Faint ripple cross-lamination was observed in a few bottomsets locally. Where troughs are exposed, bottomsets extend along their entire length, thinning as the troughs wedge out. As bottomsets extend farther out than the foreset troughs, they are often incorporated into the bottomsets of overlying troughs (Fig. 21C, D), thickening the bottomset in these areas. The connectivity of bottomsets in this deposit results in the troughs being entirely encased by bottomset deposits (Fig. 21). Major changes in bottomset thickness occur mostly where the bottomsets of two sets combine; this was driven by primary 3D controls related to the three-dimensional geometry of the host bedforms. Where two bottomsets amalgamate, the bounding surface separating them is not visible.

At Site HS4 (Fig. 13B), a 1.5–2.5-m-thick co-set of down-climbing sets (often termed compound cross-stratification) is exposed over a 60 m cliff section at Little Bay, Sydney (33° 58′ 48.98″ S, 151° 15′ 08.00″ E; Fig. 22). The exposure consists mostly of medium to coarse sand with occasional granules. Underlying the down-climbing sets are thinner sets of planar and trough cross-stratification, the former of which contain backflow ripple cross-lamination locally. As with the previous site, the orientation of the cliff face changes along its length, exposing sections cut approximately parallel to perpendicular to the migration direction of the formative bedforms. Unlike previous examples, bottomsets are absent along most of the co-set’s lower bounding surface, locally, however, thin bottomsets occur.

Towards the west, a 4-mm-thick bottomset, highlighted by quartz and iron oxide cement, occurs along the co-set’s lower bounding surface (Fig. 22C). Towards the center of the examined section a < 20 mm thick bottomset of coarse sands and granules is present (Fig. 22D). This contrasts to the mud bottomset (21 mm thick), which is present towards the southeast of the exposure (Fig. 22E), extending ca. 15 m along the base of the co-set, before wedging out.

The range of bottomset structures observed is likely of primary (coarse sands and granules) and secondary (muds) origin. Unlike other sites, some important bottomset characteristics did not fully develop until long after deposition due to the predominance of iron oxide bottomset cement, which could have implications for vertical permeability. Although the bottomset structures are thin, the two most laterally extensive, the cement and the clay, could greatly reduce vertical permeability through the co-set.

## DISCUSSION

### Identifying Bottomsets in the Rock Record

As described in the examples above, bottomsets can vary greatly in thickness, extent, grain size, and internal structure. The wide range of possible characteristics may make it difficult to distinguish some bottomsets from similar deposits formed by different processes. This could be especially true when analyzing core, as beds that may be bottomsets cannot be observed along their length. For example, during waning flow, sandy upper-plane beds may transition into dunes, creating a planar cross-stratified bed underlying dune cross-stratification (e.g., Roe 1987) that in some circumstances may be confused with a foreset–bottomset pair. Careful inspection of a deposit for characteristic bottomset features could aid identification.

Consideration of foreset–bottomset interconnection, vertical structural variation, and lateral structural variation (where available) may prove useful for bottomset identification. Foreset–bottomset interaction includes interfingerling of back-flow bedforms with the foreset (e.g., Figs. 2B, C; 16B), the extension of foreset cross-stratum into the bottomset (e.g., Fig. 17E), mud drapes extending from the foreset to the bottomset (e.g., Fig. 19C) and foreset sediment infilling gaps between pebbles and cobbles in an antecedent gravel layer (e.g., Fig. 12C). Such interaction suggests that the underlying deposit is a bottomset that could have formed concurrently with the foreset deposit (as with primary controlled bottomset structures), formed during periods of host bedform inactivity (as with secondary controlled bottomset structures) or formed before but influenced bedform depositional processes (as with antecedent controlled bottomset structures). Foreset–bottomset interaction is common (e.g., Boersma 1967; Martinius and Van den Berg 2011; Herbert et al. 2015); however, not all bottomsets show such relationships with the related foresets. For example, a bottomset could show little interaction when formed downstream of a rapidly migrating host bedform, as this could limit the time available for any back-flow bedforms to climb up the host bedform lee before burial. Foreset–bottomset interaction may be difficult to identify in core because of the small width over which the structural pattern resulting from such interaction can be observed. It is also of no use for identifying bottomsets where no foresets have been preserved (e.g., Fig. 3F). Where a deposit shows no interaction with the overlying foresets but instead interacts with underlying cross-strata, as found with the planar-stratified deposits examined by Roe (1987), the bed is likely a topset.

Changes in the character of bottomsets vertically (forming compound bottomsets; see section below) are common. These changes form as a result of unsteady flow and sediment conditions locally (in the host bedform trough) over the duration of bottomset deposition, and can occur even when the mean fluvial conditions are steady. Some of these changes are characteristic of bottomsets or are unlikely to be present in deposits formed by alternate processes (e.g., a mud drape within an upper-plane-bed deposit). For example, a well-documented vertical transition within bottomsets is the change from co-flow to back-flow ripple cross lamination (e.g., Boersma 1967; Boersma et al. 1968; Martinius and Van den Berg 2011). Examples of vertical transitions within bottomsets described above include a sandy bottomset overlying a basal gravel lag in the Roaches Grit (Fig. 16C) and mud drape abundance varying up through a sandy bottomset locally in the Hawkesbury Sandstone (Fig. 17D).

Changes across a potential bottomset deposit (when examining an exposure) could also prove useful for interpreting its origin. In the Roaches Grit and Hawkesbury Sandstone, some bottomset deposits can contain highly localized changes in character downstream, which have been interpreted as the result of changes in fluvial conditions over the duration of the host bedform’s existence. Major changes in bottomset character may coincide with changes within the foresets (e.g., a reactivation surface). Additionally, across-stream changes in bottomset character are likely to be common, induced by the development of spurs (Fig. 7) or ripple fans (Fig. 6) in the trough of the host bedform.

Grain-size changes and the dip direction of any contained cross-strata are also worth considering when identifying bottomsets. Many bottomsets identified in the literature are finer grained than the overlying foresets (e.g., Hartkamp-Bakker and Donselaar 1993; Huysmans et al. 2008; Possemiers et al. 2012; Reessink et al. 2014; Herbert et al. 2015) due to their formation from suspended-sediment deposition downstream of the host bedform. However, grain size is an unreliable indicator when considered on its own because primary controlled bottomsets formed through scour or bottomsets of secondary or antecedent origin may not follow this trend. For example, the bottomsets examined in the Roaches Grit had a grain size similar to that of the foreset deposits (Fig. 15), possibly due to the importance of reattachment scour as a source of sediment for bottomset development. Towards the top of cross-stratified bottomsets, the dip direction of the cross-strata will often differ from the overlying foresets (being reversed, oblique, perpendicular, or a mixture of the three). This is due to the formation and preservation of back-flow ripples (e.g., Figs. 1, 2A, 16B), ripple fans (e.g., Fig. 6), or obliquely to perpendicularly oriented ripples formed by currents directed along a host bedform. Where both co-flow and back-flow ripples are preserved (e.g., Boersma 1967; Boersma et al. 1968; Martinius and Van den Berg 2011), the paleoflow orientation will reverse (from downstream to upstream oriented) up the bottomset.
FIG. 22.—A, B) Photogrammetric models of the compound cross-stratified Hawkesbury Sandstone exposure at Little Bay (Site HS4). Parts A and B are oriented at 95° to each other. C) Photograph of bottomset cement that developed towards the west of the exposure. D) A 20-mm-thick bottomset of coarse sand and fine gravel. E) A 21-mm-thick mud bottomset. The locations of Parts C, D, and E are noted in Parts A and B. Rose diagram denotes foreset dip direction in thirteen down-climbing sets measured over the exposure.
Paleoenvironmental Interpretations from Bottomsets and Compound Bottomsets

Bottomsets have the potential to aid in the differentiation of cross-stratification formed by unit bars from that formed by dunes. Both unit bars and dunes can form simple sets of high-angle cross-stratification. Thus, identification of the formative bedform of a cross-stratified set in the rock record can prove difficult. In the literature, planar cross-stratified sets have been attributed to dunes (e.g., Alexander and Gawthorpe 1993; Miall and Jones 2003; Soltan and Mountney 2016) and bars (e.g., Bluck 1986; Steel and Thompson 1983; Roe and Hernemann 1993; Jones and Chisholm 1997; Soltan and Mountney 2016), with only minimal justification supporting the authors’ choice of bedform in some cases. Some authors suggested bedform type based on the relative cross-stratal thickness, with thicker sets interpreted as resulting from bars and thinner sets from dunes (e.g., Bridge and Tye 2000; Soltan and Mountney 2016). Relations between dune and unit-bar set thicknesses (relative to set length), found in a few modern river systems, support this concept (e.g., Bridge and Lunt 2006). Others interpreted bedform type based on grain-size relationships over a set (e.g., Steel and Thompson 1983; Bluck 1986). Some authors avoided assigning a particular bedform type (e.g., Rust and Jones 1987) or use bedform terms with greater ambiguity, such as “sand waves” (e.g., McCabe 1977; Conaghan and Jones 1975; Jones 1979; Conaghan 1980).

One important difference between unit bars and dunes is a unit bar’s ability to persist over great changes in discharge (whereas dunes will tend to be washed out). Riverine unit bars examined in the Burdekin River and in previous studies (e.g., Tillman and Ellis 1968; Collinson 1970; Smith 1974; Wooldridge and Hickin 2005; Reesink and Bridge 2011; Parker et al. 2013) persisted through changes in flow and sediment conditions. As flow and sediment conditions over a unit bar change, so will the character of its bottomset. In addition, changes in fluvial conditions can lead to the development of complex foreset structures in migrating unit bars, such as down-climbing sets. In such bars multiple bottomsets can develop, with potential for one at the base of the bar and additional bottomsets forming at the base of each down-climbing set (e.g., Fig. 12B, D). In contrast, dunes form sporadically under a relatively narrow range of flow and sediment conditions (e.g., Costello 1974; Harms et al. 1975; Southard and Boguschwal 1990; Van den Berg and Van Gelder 1993). Extended time outside these conditions will lead to washout and replacement (e.g., upper plane bed, cf. Bridge and Best 1988). The narrower range of existence conditions lessens their ability (relative to that of unit bars) to exist over great changes in discharge. Thus the bottomsets of dunes, which may be less likely to form than with unit bars if the trend found in Figure 5 extends to the natural environment, may show fewer and/or lower-magnitude changes in character within and between bottomsets than with unit-bar bottomsets. Dune-derived bottomsets are unlikely to contain secondary controlled bottomset structures (Table 1) as the relatively large changes in flow and sediment conditions that form them is likely to result in dune washout. Applying this idea to the exposures documented above, the planar cross-stratified deposits examined at the Roaches (Figs. 14, 15) and at Site HS2 of the Hawkesbury Sandstone (Fig. 19) are probably unit-bar deposits because of the numerous and relatively large changes in bottomset character observed at these sites.

In addition to aiding in the identification of formative bedforms, the wide variation in bottomset character, which can vary both within and between bottomsets, gives them the potential to be valuable indicators of the conditions before and during deposition. Localized changes in character (e.g., changes in bottomset thickness, Fig. 16A; development of back-flow bedforms, Fig. 16B; gravel-lag development, Fig. 16C; or mud-drape development, Fig. 19C) can provide information on changing flow or sediment conditions over the deposition of a single dune or unit bar. Changes over multiple sets can provide information on longer-term changes in fluvial character (see the Roaches Grit section above for an example of this). Some cross-stratified deposits lack bottomsets entirely (e.g., the lower sets in the Roaches Grit exposure, Fig. 15A). This in itself could prove informative; suggesting the sediment flux to the bedform trough was minimal, that scour in the trough zone was minimal, or that any antecedent trough sediment eroded by reattachment scour was not deposited within the flow separation zone. In addition, it also suggests the back flow (and associated turbulence) in the flow separation zone was not strong enough to rework antecedent trough deposits into back-flow bedforms.

Some bottomsets contain a vertically oriented compound structure (upwards through their thickness) that can provide highly detailed paleoenvironmental interpretations. Compound bottomsets contain stacked structures related to more than one primary, secondary, or antecedent control (e.g., mud-draped gravel bottomset, Fig. 12C; basal gravel lag underlying a sandy bottomset, Fig. 16C; sandy bottomset topped with a mud drape, Fig. 19C) or are amalgamations of bottomsets of multiple bedforms (e.g., Figs. 3F, 21). For the former, compound bottomsets can contain a range of structures related to either spatial changes in local conditions (Fig. 23) or temporal changes in fluvial conditions (Fig. 24) over bedform deposition.

Compound bottomsets generated through spatial variation in conditions contain stacked structures formed by local changes in primary controls only (Fig. 23). Preservation of both co-flow and back-flow ripple cross-lamination (e.g., Boersma 1967; Boersma et al. 1968; Martinius and Van den Berg 2011; Fig. 23A) is a well-documented compound bottomset structure of this type. Compound bottomsets generated through spatial variation in conditions may provide details on the source of bottomset sediment. For example, preservation of and co-flow and back-flow ripple deposits suggests reattachment scour was minimal (Martinius et al. 2015; e.g., Fig. 23A). In contrast, the preservation of a gravel lag at the base of a sandy bottomset (e.g., Figs. 16C, 23B) suggests that reattachment scour was much greater, liberating sediment that likely contributed to bottomset development.

Compound bottomsets generated through temporal changes in conditions contain multiple stacked structures that give a timeline of processes during deposition (Fig. 24). If considering a purely depositional, temporally controlled compound bottomset influenced by primary, secondary, and antecedent controls, structures generated earliest (antecedent structures) should be to the base whilst those generated just before burial (primary structures) will be at the top (Fig. 24C). It may be difficult to identify whether some features are of secondary or antecedent origin, and they may be diachronous (e.g., mud ripples). Changes in the foreset architecture may assist the interpretation, because discharge changes during bar migration (triggering secondary bottomset structures) often generate associated reactivation surfaces (e.g., Fig. 24).

Consideration of feedbacks between trough deposits and structures, formed by secondary and antecedent controls, and subsequent sedimentary and flow processes results in a greater number of possible interpretations of the paleoenvironmental conditions. For example, for a temporally controlled compound bottomset containing back-flow ripple deposits overlain by a secondary derived mud drape (Fig. 24B), the lack of back-flow ripples after the drape could suggest a decline in suspended-sediment flux upon bar reactivation. Alternatively, the sediment flux to the trough may have always been low, but the mud drape increased the scour resistance of pre-existing trough sediment, halting reworking of sediment into back-flow ripples in the flow separation zone. Consideration of other factors, such as the grain size of the bottomset in comparison to that of the underlying deposits may assist in this deduction. Where the back-flow ripple sediment is much finer than the underlying sediment, it suggests that the former is much more likely.

Previous classifications of bedforms and resultant sedimentary structures (e.g., McKee and Weir 1953; Allen 1963; Rubin 1987; Ashley 1990) largely ignored development and structure of bottomsets. Rubin’s (1987)
classification, like the classification of bottomset controls above, recognizes the importance of bedform geometry and orientation relative to mean transport directions, as well as changes in bedform morphology and migration over time. Altering any of these factors changes the internal foreset structures generated for two- and three-dimensional bedforms. However, Rubin's (1987) models are purely geometric, and they lack consideration of the complex fluid dynamics and sedimentary processes that create the full range of bottomset structures. The consideration of the controls on bottomsets discussed herein allows for a complete reconstruction of bottomset deposition separate from that of the foreset. Combined with deductions made from the foresets, this will improve interpretations of paleo-flow and paleo-sedimentary processes.

**Bottomsets in Reservoir Sandstones**

Understanding bottomset variation is important in predicting fluid flow through fluvial rocks, especially if the bottomsets are very different in grain size from the foresets (coarser or finer, e.g., Fig. 2D, 12C, 19C, E). Permeability is generally greater horizontally than vertically in fluvial sandstones (e.g., Pryor 1972; Weber 1982; Miall 1988). Bottomsets may prevent or slow vertical flow and restrict lateral flow if individual sets wedge out (Weber 1986). Air-permeameter measurements of cross-stratified sandstone outcrops (Alexander and Gawthorpe 1993; Hartkamp-Bakker and Donselaar 1993) observed that back-flow ripple cross-lamination increased permeability variation. Bottomsets in all these examples were finer grained than the overlying foresets. Less common bottomsets, such as those containing gravel (e.g., Figs. 2D, 12C), are likely to give different permeability patterns. For example, laterally extensive gravel bottomsets may have high horizontal permeabilities. If present in hydrocarbon reservoirs undergoing fluid injection, such deposits could potentially form "thief zones," layers in a reservoir where large amounts of injected fluid are drawn off (cf. Felsenthal and Gangle 1975). Thief-zone formation in fluvial reservoirs has also been linked to the development of cross-stratified open framework gravel in bedform foreset deposits (Lunt et al. 2004; Gershenzon et al. 2015).

Some models of subsurface flow through fluvial rocks (e.g., Scheibe and Freyberg 1995; Mikes et al. 2006) or wettability (e.g., Van Lingen et al. 1996) have included simple bottomsets (e.g., consistent permeability, consistent lateral extent, grain size the same as fine-grained foreset laminae). This ignores changes in bottomset thickness, grain size, lateral extent, and internal structure that occur with natural changes in flow and sediment conditions. In most deposits, bottomsets are unlikely to be represented well by a single permeability value or grain size.

Changes in the structure, thickness, and grain size of bottomsets may increase or decrease vertical permeability relative to that of underlying and

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**Fig. 23.**—Schematics denoting the internal structure and formation of compound bottomsets induced by local spatial variability in conditions. $U$ denotes mean flow, $u_\text{t}$ denotes local trough flow. Black arrows denote flow direction; black dashed arrows denote a previous flow direction. Gray arrows denote sediment transport.

**Fig. 24.**—Schematics denoting the internal structure and formation (over time, $t_1$ to $t_3$) of compound bottomsets induced by temporal fluvial variability. $U$ denotes mean flow, $u_\text{t}$ denotes local trough flow. Black arrows denote flow direction; black dashed arrows denote a previous flow direction. Where such arrows are oriented upwards, it denotes a flow direction into the page. Gray arrows denote sediment transport. Gray dashed arrows denote a change in water level.
A. Primary and secondary compound bottomset

- Water surface
- Flow-separation zone
- Back-flow ripples
- Host bedform reactivated
- Rise in water level
- No trough reworking

B. Primary and secondary compound bottomset

- Water surface
- Flow-separation zone
- Back-flow ripples
- Host bedform reactivated
- Rise in water level

C. Primary, secondary and antecedent compound bottomset

- Water surface
- Flow-separation zone
- Antecedent gravel layer
- Host bedform reactivated
- Suspended sediment
- Rise in water level
- Back-flow ripples

- U → u

- t2
  - Washed-out avalanche face
  - Perpendicularly migrating dunes
  - Drop in water level

- t1
  - Mud deposition
  - Mud drape
overlying sediment. Cross-channel or downstream variation in a bottomset may create preferential flow pathways which would be absent if bottomsets did not vary (Fig. 25). Vertical connectivity may be enhanced by bottomsets thinning and wedging out (e.g., Fig. 14), especially if the bottomset is clay rich (e.g., Huysmans et al. 2008; Huysmans and Dassargues 2010; Possemiers et al. 2012; Reesink et al. 2014). Conversely, a sand-dominated bottomset with periodic mud drapes (e.g., Fig. 19), may decrease vertical permeability locally. Because bottomsets more often act as barriers to flow (e.g., Hartkamp-Bakker and Donselaar 1993; Huysmans et al. 2008; Possemiers et al. 2012) instead of pathways, bottomset variability will generally improve connectivity between sets by creating zones that can act as conduits (e.g., a thinning and/or coarsening bottomset; Fig. 25B).

If a migrating bar experiences a relatively small change in flow or sediment transport, bottomset changes may be minimal (e.g., bottomset thins by $<20\%$), and this will have only a minor impact on reservoir architecture. A bar migrating through greatly varying conditions is likely to feature a greatly variable bottomset (e.g., bottomset stops forming, transition from sand to mud deposition, formation of a compound bottomset) that could greatly enhance or hinder permeability locally. In situations where bottomsets vary only between individual sets, only one laterally extensive poor-permeability bottomset is needed to hinder vertical flow (Fig. 25A).

CONCLUSIONS

Bottomsets have the highest preservation potential of any deposits generated by bedforms, with their thickness often fully preserved, which is not the case with most foresets. Despite this, they are under-researched and often go unreported in both field- and laboratory-based publications. Bottomsets can be more variable in character than foresets, often changing in thickness, grain size, internal structure, and extent laterally. Bottomsets vary within a single set, between sets, between channel reaches, and between channels. Thus, bottomset character varies along and between exposures of ancient sediment. This variability is controlled by antecedent bed conditions as well as unsteadiness in flow and sediment flux during deposition. Classification of the controls on bottomset development helps with understanding their variation and its paleoenvironmental implications.

Primary 2D controls on bottomset development and character are those relating to two-dimensional processes occurring during host-bedform migration. Primary 3D controls are those relating to three-dimensional flow and 3D bedform shape during host-bedform migration. Secondary controls are those occurring in the host bedform trough during periods of host-bedform inactivity. Antecedent controls are those occurring before the migration of a host bedform into an area of interest, generating deposits or structures that are subsequently incorporated into the base of a bottomset. Structures and deposits formed by secondary and antecedent controls can create feedbacks, which alter subsequent bottomset-forming processes.

Many bottomsets have a compound structure, containing deposits formed in response to more than one primary, secondary, or antecedent control, which vary over time. The wide range of possible characteristics of bottomsets and compound bottomsets provides information on host-bedform type as well as the flow and sediment conditions before and during host bedform migration. The range of internal structures and lateral variability of bottomsets and compound bottomsets has implications for understanding reservoir characteristics of cross-stratified sandstone. Bottomsets formed under relatively steady conditions may be laterally extensive and uniform in character, hindering vertical flow. Bottomsets generated in unsteady regimes can be highly variable, with the potential to create conduits between sets aiding inter-set connectivity.

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APPENDIX

Figures 6 and 7A are from experiments conducted in the 10 m x 1 m x 1 m test channel of the University of East Anglia recirculating flume (a schematic of this is presented in Herbert et al. 2015). The schematic in Figure 6 was constructed from a photograph, taken at the end of the tenth run of a twelve run experiment series, of ripple crests formed downstream of an artificial unit bar. The photograph was taken vertically, looking down at the channel bed. The artificial unit bar was constructed using plastic sheeting supported on sand bags and bricks. The bar was 2.91 m long and 0.244 m high (tough to brink, the latter coincident with the bar crest). The lee dipped 29° downstream, and the stoss dipped 5.6° upstream. The bar’s upstream end, brink, and toe were positioned at 3.52 m, 5.98 m, and 6.43 m along the horizontal flume channel respectively. The artificial unit bar had a 20-mm-thick flat layer of well-sorted fine sand (median grain size = 0.175 mm) and 5.98 m along the horizontal flume channel respectively. The artificial unit bar had a 20-mm-thick flat layer of well-sorted fine sand (median grain size = 0.175 mm) and 5.98 m along the horizontal flume channel respectively. The artificial unit bar had a 20-mm-thick flat layer of well-sorted fine sand (median grain size = 0.175 mm) and 5.98 m along the horizontal flume channel respectively. The artificial unit bar had a 20-mm-thick flat layer of well-sorted fine sand (median grain size = 0.175 mm) and 5.98 m along the horizontal flume channel respectively. The artificial unit bar had a 20-mm-thick flat layer of well-sorted fine sand (median grain size = 0.175 mm) and 5.98 m along the horizontal flume channel respectively. The artificial unit bar had a 20-mm-thick flat layer of well-sorted fine sand (median grain size = 0.175 mm) and 5.98 m along the horizontal flume channel respectively. The artificial unit bar had a 20-mm-thick flat layer of well-sorted fine sand (median grain size = 0.175 mm) and 5.98 m along the horizontal flume channel respectively.