Bedforms and sedimentary structures related to supercritical flows in glacigenic settings

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Associate Editor – Dario Ventra

ABSTRACT
Upper-flow-regime bedforms, including upper-stage-plane beds, antidunes, chutes-and-pools and cyclic steps, are ubiquitous in glacigenic depositional environments characterized by abundant meltwater discharge and sediment supply. In this study, the depositional record of Froude near-critical and supercritical flows in glacigenic settings is reviewed, and similarities and differences between different depositional environments are discussed. Upper-flow-regime bedforms may occur in subglacial, subaerial and subaqueous environments, recording deposition by free-surface flows and submerged density flows. Although individual bedform types are generally not indicative of any specific depositional environment, some observed trends are similar to those documented in non-glacigenic settings. Important parameters for bedform evolution that differ between depositional environments include flow confinement, bed slope, aggradation rate and grain size. Cyclic-step deposits are more common in confined settings, like channels or incised valleys, or steep slopes of coarse-grained deltas. Antidune deposits prevail in unconfined settings and on more gentle slopes, like glacifluvial fans, sand-rich delta slopes or subaqueous (ice-contact) fans. At low aggradation rates, only the basal portions of bedforms are preserved, such as scour fills related to the hydraulic-jump zone of cyclic steps or antidune-wave breaking, which are common in glacifluvial systems and during glacial lake-outburst floods and (related) lake-level falls. Higher aggradation rates result in increased preservation potential, possibly leading to the preservation of complete bedforms. Such conditions are met in sediment-laden jökulhlaups and subaqueous proglacial environments characterized by expanding density flows. Coarser-grained sediment leads to steeper bedform profiles and highly scoured facies architectures, while finer-grained deposits display less steep bedform architectures. Such differences are in part related to stronger flows, faster settling of coarse clasts, and more rapid breaking of antidune waves or hydraulic-jump formation over hydraulically rough beds.

Keywords Antidune, cyclic step, chute-and-pool, glacifluvial delta, ice-contact fan, jökulhlaup.
INTRODUCTION

The role of upper-flow-regime processes for deposition in many clastic depositional environments is increasingly recognized. These bedforms are produced at high Froude numbers at near-critical and supercritical flow conditions and include upper-stage-plane beds, antidunes, chutes-and-pools and cyclic steps. The formation of different bedform types is controlled by the (densimetric) Froude number, grain size, sediment concentration and confinement of the formative flow (Van den Berg & Van Gelder, 1993, 1998; Alexander et al., 2001; Kostic, 2011, 2014; Cartigny et al., 2014; Fedele et al., 2016). Upper-flow-regime bedforms have been documented mostly from alluvial (e.g. Blair, 1999; Fielding, 2006; Froude et al., 2017; Carling & Leclair, 2019), coastal and shallow-marine (Nielson et al., 1988; Massari, 1996, 2017; Slootman et al., 2016, 2018, 2019; Vaucher et al., 2018; Di Celma et al., 2020) and deep-marine deposits (Postma et al., 2014; Covault et al., 2017; Lang et al., 2017a; Hage et al., 2018; Ono & Plink-Björklund, 2018).

The formation and preservation of upper-flow-regime bedforms is generally thought to be favoured in settings with high rates of aggradation and rapidly decelerating flows. These conditions are commonly met in meltwater-dominated glacigenic depositional environments, which are characterized by abundant sediment supply and large spatial and temporal variations in flow strength and discharge (Marren, 2005; Cuffey & Paterson, 2010; Dowdeswell et al., 2015), including rare high-magnitude events due to outbursts of stored meltwater (Duller et al., 2008; Carling, 2013). Previous studies have documented different types of upper-flow-regime bedforms in successions from glacigenic depositional environments characterized by abundant meltwater and sediment supply (Brennand, 1994; Hirst et al., 2002; Russell & Arnott, 2003; Kjaer et al., 2004; Hornung et al., 2007; Duller et al., 2008; Winsemann et al., 2009, 2011, 2016, 2018; Girard et al., 2012a,b, 2015; Lang & Winsemann, 2013; Pisarska-Jamroży & Zieliński, 2014; Dietrich et al., 2016, 2017; Lang et al., 2017b).

This study provides a review of the depositional record of upper-flow-regime bedforms in glacigenic depositional environments preserved in successions from Palaeozoic and Quaternary glacial episodes in Earth’s history. The overall objective is to demonstrate that upper-flow-regime bedforms are formed under a broad variety of flow conditions in glacigenic environments, making the latter an ideal setting for the recognition and study of the stratigraphic signatures of supercritical flows. The study aims to: (i) summarize the controlling factors for bedform morphodynamics and the characteristic sedimentary facies in the different sub-environments of glacigenic deposition; (ii) discuss the specific conditions for the development of upper-flow-regime bedforms in glacigenic depositional settings; and (iii) point out the similarities and differences of deposits of upper-flow-regime bedforms to non-glacigenic depositional environments.

Morphodynamics of bedforms related to supercritical flows

Upper-flow-regime bedforms are known to occur within various sub-environments of glacigenic settings, including subglacial, glacifluvial, glacilacustrine and glacimarine environments (Fig. 1). The large variability in documented bedform scales relates to the very different magnitudes of the formative flows, ranging from shallow sheetfloods in glacifluvial systems (Pisarska-Jamroży & Zieliński, 2014) to deep density flows in glacilacustrine and glacimarine systems (Winsemann et al., 2009, 2018; Hirst, 2012; Lang & Winsemann, 2013; Lang et al., 2017b) and glacial lake-outburst floods (Carling et al., 2009; Winsemann et al., 2011, 2016).

To encompass the wide range of glacigenic depositional environments (Fig. 1) where Froude supercritical flows may occur, some terms used throughout this study need a definition. Subaerial depositional environments are settings characterized by processes related to subaerial runoff, i.e. free-surface or open-channel flows in hydrodynamic terms. The term open-channel flow, however, does not necessarily indicate channelization. Subaerial depositional environments characterized by flowing air (i.e. aeolian) are not considered here. Subaqueous depositional environments are settings characterized by gravity-driven density flows propagating at the base of an ambient water column. The density flows considered here generally have an excess density generated by the sediment load and is higher than the density of the ambient water mass (i.e. sediment-gravity flows). Examples for this group of density flows are low-density and high-density turbidity currents, hypopycnal flows or underflows (Mulder & Alexander, 2001; Talling et al., 2012).
The most commonly described bedforms related to near-critical and supercritical flows are antidunes, chutes-and-pools and cyclic steps (Fig. 2) (Van den Berg & Van Gelder, 1993, 1998; Alexander et al., 2001; Kostic, 2011, 2014; Cartigny et al., 2014; Fedele et al., 2016). Supercritical flows are characterized by the dominance of inertial over gravitational forces and are defined by a Froude number larger than unity. For free-surface flows (also termed open-channel flows), the Froude number (Fr) is given by \( F_{r} = \frac{U}{\sqrt{g h}} \), where \( U \) is the depth-averaged flow velocity, \( h \) is the flow depth and \( g \) is the acceleration by gravity. For density flows, the densimetric Froude number \( F_{r}' \) is given by \( F_{r}' = \frac{U}{\sqrt{g' h}} \), where \( g' \) is the reduced acceleration by gravity with \( g' = g \left( \frac{\rho_{f} - \rho_{w}}{\rho_{f}} \right) \), where \( \rho_{f} \) is the density of the flow and \( \rho_{w} \) is the density of the ambient water. Supercritical flows over mobile sediment beds are characterized by in-phase relationships between the morphology of the sediment-flow interface and disturbances of the upper surface of the flow, which in density flows is represented by a density interface internal to the flow (Hand, 1974; Cartigny et al., 2014; Postma & Cartigny, 2014; Fedele et al., 2016). The formation of upper-flow-regime bedforms requires a free upper flow interface and is therefore suppressed in full pipe-flow conditions (Banerjee & McDonald, 1975).
Depending on (densimetric) Froude number and grain size various types of bedforms develop. For equilibrium conditions in free surface flows the bedforms are well-known from flume experiments and observations in rivers, and their occurrence can be predicted in bedform stability diagrams (Fig. 2A and B). For density flows this is not the case, due to the difficulty to produce stable sediment-density flows in flumes over time periods long enough to create an equilibrium with the evolving bedforms. Cartigny & Postma (2016) presented a series of stability diagrams for bedforms produced by 10 m thick sediment density flows with increasing basal sediment concentration. These authors assumed that the stability fields are basically the same as for free-surface flows, except for a plane-bed stage produced by traction carpets in high-density turbidity currents. Fedele et al. (2016) published a new stability diagram based on a very large number of flume experiments with saline density flows (Fig. 2C). Because the flows in these experiments were not charged with suspended sediment, the diagram may only represent the bed morphodynamics of dilute density flows. The upper limit for upper-flow-regime bedform existence is $F_{r} = 2.8$ (Fig. 2C) (Mastbergen & Van Den Berg, 2003). Above this threshold Kelvin–Helmholtz billows developed at the density flow interface with the overlying less dense fluid reach the bed, resulting in sudden dilution and thickening of the density flow by mixing with the overlying fluid, which in turn causes a reduction of the densimetric Froude number.

Consequently, the densimetric Froude number cannot rise much above this threshold (Mastbergen & Van Den Berg, 2003). Upper-flow-regime bedforms are probably more common in density flow deposits, where supercritical conditions are more easily attained than in free-surface flows and high rates of aggradation result in greater preservation potential (Hand, 1974; Covault et al., 2017). However, these observations might be biased by the strong focus of recent studies on density flows and their deposits. The expression of upper-flow-regime bedforms within the depositional record is also strongly controlled by the rate of aggradation and grain-size distribution, which may lead to very different sedimentary structures (Cartigny et al., 2014; Ono et al., 2020). In addition, reduced aggradation rates may result in a low variability of sedimentary structures (Cartigny et al., 2014), and hence reduce the ability to distinguish between bedforms.

Fig. 2. Bedform stability diagrams. (A) Stability fields of free-surface flows plotted for particle diameter versus flow velocity for a flow depth of 0.06 to 0.1 m (modified from Southard & Boguchwal, 1990; Cartigny et al., 2014). (B) Stability fields of free-surface flows plotted for the dimensionless particle diameter versus dimensionless mobility parameter (modified from Van den Berg & Van Gelder, 1998; Cartigny et al., 2014). (C) Stability fields of submerged density flows plotted for particle diameter versus densimetric Froude number (modified from Fedele et al., 2016).
Experiments by Cartigny et al. (2014) have shown that transitions between the stability fields of antidunes, chutes-and-pools and cyclic steps are poorly defined in the grain size versus Froude number parameter space, probably due to strong flow–particle interactions. Bedform stability fields and their transitions could be better defined by means of the dimensionless particle diameter and the mobility parameter (Fig. 2C) (Cartigny et al., 2014). The dimensionless particle diameter ($D^*$) is given by $D^* = D_{50}((\rho_s - \rho)/\rho_f)^{1/2}$, where $D_{50}$ is the median grain size of the bed material, $\rho_s$ is the density of the sediment particles and $v$ is the kinematic viscosity of the flow. The dimensionless mobility parameter ($\Theta^*$) is given by $\Theta^* = \rho_f U^2/((\rho_s - \rho)(C)^2 D_{50})$, where $C$ is the grain-size related Chezy coefficient. The advantage of using the dimensionless particle diameter and the mobility parameter for bedform stability diagrams (Fig. 2B) is that no shift is expected in the boundaries of the stability fields in the diagram in case of changing the temperature, gravity or fluid density (Van Den Berg & Van Gelder, 1993, 1998).

Upper-flow-regime bedforms can be classified as stable versus unstable bedform types (Cartigny et al., 2014; Slootman & Cartigny, 2019). Bedforms may exist as stable patterns as shown in flume experiments with steady flows, while in nature flows are commonly highly unsteady (Alexander et al., 2001; Cartigny et al., 2014; Fedele et al., 2016). A classic example of unsteady and non-uniform flow is the plane bed – stationary antidunes – breaking antidunes cycle described by Middleton (1965). Upper-flow-regime bedforms should always be considered as inherently unstable under natural conditions and can only be regarded as stable for a short interval of time. However, an exception to this is given by cyclic steps, which once formed can remain active and propagate upstream for very long times, if the flow and sediment supply are sustained.

**Antidunes**

Antidunes are characterized by in-phase trains of surface waves that may be stationary or migrate upflow or downflow, resulting in the formation of a range of bedforms with variable morphodynamics. If the wave steepness (wave height/wavelength) exceeds a critical value (0.142 in open-channel flows), wave breaking may occur at the flow surface, triggering the cyclic destruction and regeneration of underlying antidune bedforms (Kennedy, 1963; Middleton, 1965; Cartigny et al., 2014). The wave steepness of antidunes along the upper interfaces of density flows is generally lower than at the upper interface of free-surface flows, leading to the formation of more stable wave trains less susceptible to wave breaking (Fedele et al., 2016). Antidunes are classified according to their migration direction (stationary, upstream-migrating or downstream-migrating), wavelength and non-breaking (stable antidunes) versus breaking (unstable antidunes) dynamics (Gilbert, 1914; Kennedy, 1963; Alexander et al., 2001; Cartigny et al., 2014; Fedele et al., 2016), leading to a range of sedimentary structures. Tabular beds with trains of backsets, subhorizontal, convex-up or sinusoidal stratification or low-angle cross-stratification point to non-breaking antidunes. Lenticular beds with backsets, low-angle foresets or concave-up concentric infill indicate breaking antidunes (Alexander et al., 2001; Fielding, 2006; Duller et al., 2008; Lang & Winseman, 2013; Cartigny et al., 2014). Stationary non-breaking antidunes may deposit sinusoidal strata, matching the in-phase relation between the flow and the bedforms (Middleton, 1965; Brennand, 1994; Duller et al., 2008; Lang & Winseman, 2013). Upstream-migration or downstream-migration of non-breaking antidunes causes an upflow or downflow offset of the wave crests, respectively (Fielding, 2006; Lang & Winseman, 2013). If antidune migration becomes more pronounced the sedimentary structures are characterized by backsets or foresets (Van den Berg & Lang, this issue; Ito, 2010; Cartigny et al., 2014; Fedele et al., 2016; Lang et al., 2017a,b). Furthermore, the expression of antidune bedforms within deposits is also strongly controlled by the rate of aggradation at the original sediment–water interface, where generally reduced aggradation rates may result in low variability of structures (Cartigny et al., 2014), and hence a reduced ability to distinguish the variety of antidunes.

**Chutes-and-pools**

The morphodynamics of chutes-and-pools are characterized by irregularly spaced supercritical flow reaches (the chute) and subcritical flow reaches (the pool) separated by a hydraulic jump. The hydraulic jumps migrate erratically and may spontaneously form and dissipate (Alexander et al., 2001; Cartigny et al., 2014). However, the exact nature of chute-and-pool bedforms is still debated, and it is possible that these mobile-bed configurations simply reflect
transient, pronounced flow instabilities at the transition between antidunes and cyclic steps at intermediate Froude numbers (Fig. 2A and B) (Cartigny et al., 2014; Slootman & Cartigny, 2019). Deposits of chutes-and-pools typically consist of scours infilled by steeply dipping backsets occasionally accompanied by gently dipping foresets, as well as concentric scour fills. These scour fills are commonly associated with local evidence for antidune deposits formed in the supercritical reach of the formative flow (Hand, 1974; Alexander et al., 2001; Fielding, 2006; Lang & Winsemann, 2013; Cartigny et al., 2014; Postma et al., 2020). Such isolated scour fills are also referred to as deposits of isolated or localized hydraulic jumps (Russell & Arnott, 2003; Duller et al., 2008; Lang et al., 2017b), avoiding explicit reference to possible chute-and-pool bedforms, which might not have been fully preserved. Isolated hydraulic jumps may form also in response to changing flow conditions forced by local slope breaks, for example at the toe of deltas (Winsemann et al., 2007b; Massari, 2017; Postma et al., 2020), at the base of steep barrier spits or beach ridges (Nielsen et al., 1988; Di Celma et al., 2020), or at large flow obstacles, such as stranded ice blocks during a jökulhlaup (Russell & Knudsen, 2002; Burke et al., 2010a; Herget et al., 2013).

Cyclic steps
Cyclic steps are formed at high Froude numbers by flows characterized by alternating reaches with supercritical and subcritical conditions bounded by hydraulic jumps that migrate upflow at a constant distance, defining the wavelength of an individual cyclic step. Cyclic-step morphodynamics are dominated by deposition downflow of the hydraulic jumps and erosion upflow of the hydraulic jumps, typically leading to the formation of asymmetrical bedforms (Cartigny et al., 2011, 2014; Ventra et al., 2015; Fedele et al., 2016; Slootman & Cartigny, 2019). Cyclic-step deposits are commonly characterized by laterally stacked scours filled by massive or diffusely graded/stratified sediment and gently dipping backsets (Postma et al., 2009, 2014; Cartigny et al., 2014; Postma & Cartigny, 2014). Due to the similarity of depositional processes in the hydraulic-jump zone, the distinction between cyclic-step and chute-and-pool deposits remains challenging. Criteria for the recognition of cyclic-step deposits include the predominance of asymmetrical backset cross-stratified scour fills and the lateral stacking of scour fills produced by the steady migration of the hydraulic jumps (Cartigny et al., 2014; Lang et al., 2017a; Postma et al., 2020). Superimposed antidunes may form in the supercritical flow reach on the stoss sides of cyclic steps (Van den Berg & Lang, this issue; Kostic, 2014; Zhong et al., 2015; Lang et al., 2017a). If cyclic steps and antidunes form in the same train, the wavelength of the cyclic steps is typically one order of magnitude larger than the wavelength of the associated antidunes (Kostic, 2014).

Further upper-flow-regime bedforms
Further bedforms developed at near-critical to supercritical flow stages are: (i) upper-stage-plane beds; (ii) humpback dunes; and (iii) supercritical dunes (Fig. 2). Upper-stage-plane beds are formed by bedload sheets and very low-amplitude bed waves, and are characterized by thin (commonly millimetre-scale) planar-parallel stratification in the resulting deposits (Allen, 1984; Paola et al., 1989; Best & Bridge, 1992). In density flows, upper-stage-plane beds have been attributed to deposits of dilute turbidity currents (‘$T_{B-1}$’; Talling et al., 2012) and have been reproduced in flume experiments (Fedele et al., 2016; Koller et al., 2019). In contrast to free-surface flows, where they are produced at near-critical Froude numbers, plane beds represent conditions of relatively high densimetric Froude numbers in density flows. Notwithstanding this fundamental difference, flume experiments show the same bedload sheets as in free-surface flows, suggesting similar depositional processes (Fedele et al., 2016; Koller et al., 2019). Humpback dunes display sigmoidal foreset cross-stratification, with best preserved examples showing a clear differentiation between topset, foreset and bottomset laminae. They are generally considered as bedforms transitional between subcritical dunes and antidunes, forming commonly under high aggradation rates (Saunderson & Lockett, 1983; Chakraborty & Bose, 1992; Fielding, 2006; Lang & Winsemann, 2013; Winsemann et al., 2018). Humpback dunes have originally been described from fluvial environments, where they are commonly flattened and washed-out when upper-stage-plane beds or antidunes develop under increasing flow velocity (Saunderson & Lockett, 1983; Fielding, 2006). However, humpback dunes can also develop at subcritical Froude numbers in deep fluvial and estuarine channels for conditions of fine-grained sand and high bed-grain mobility (Røe, 1987; Martinius & Van den Berg, 2011). In
this case, the formation of sigmoidal cross-stratification can be considered as an initial stage in the transition from dunes to upper-plane-bed regime not forced by a high Froude number, but by the suppression of turbulence by a high near-bed vertical gradient in suspended bed material (Bridge & Best, 1988).

Supercritical dunes are asymmetrical, downstream-migrating bedforms observed in supercritical density flows. Internally, they are characterized by sigmoidal or tangential foreset cross-stratification (Fedele et al., 2016; Lang et al., 2017a). In flume experiments, supercritical dunes display clear flow-separation on the lee sides, lack an in-phase relationship with the flow and form at higher densimetric Froude numbers than downstream-migrating antidunes (Fedele et al., 2016). Subcritical dunes did not form in the experiments of Fedele et al. (2016) because the bed-shear stress necessary for dune formation was only attained in supercritical flows.

However, distinguishing between deposits of humpback dunes, supercritical dunes and downstream-migrating antidunes might be nearly impossible as the characteristic sedimentary structure of these three bedform types is sigmoidal foreset cross-stratification (Saunderson & Lockett, 1983; Fielding, 2006; Lang & Winsemann, 2013; Fedele et al., 2016; Lang et al., 2017a,b; Winsemann et al., 2018). The issue is further complicated because sigmoidal cross-stratification may also relate to subcritical (climbing) dunes under high rates of aggradation (Ghienne et al., 2010; Winsemann et al., 2011) and isolated hydraulic-jump bars (Macdonald et al., 2009, 2013).

THE DEPOSITIONAL RECORD OF GLACIGENIC UPPER-FLOW-REGIME BEDFORMS

Deposits of upper-flow-regime bedforms have been documented from a variety of glacigenic depositional environments within Palaeozoic and Quaternary successions (Table 1). Direct observations of upper-flow-regime bedforms in modern glacigenic settings include antidune trains on glaciﬂuvial fans (Gustavson, 1974) and cyclic steps in supraglacial meltwater streams (Izumi et al., 2017). The best example of cyclic steps in a modern submarine setting are from the subaqueous slope of the sand-rich glaciﬂuvial Squamish River delta in western Canada, where bedform morphodynamics, flow triggers and the resulting stratigraphy have been studied in detail from direct observations (Hughes Clarke, 2016; Stacey & Hill, 2016; Hizzett et al., 2018; Hage et al., 2018, 2019; Stacey et al., 2019; Vendettouli et al., 2019). Deposits of antidunes and chutes-and-pools are known from modern jökulhlaup deposits in Alaska and Iceland, where the depositional record can be compared with direct observations and estimations of the flood discharge (Burke et al., 2008, 2010a,b; Duller et al., 2008, 2015).

Pleistocene glacigenic successions occur worldwide in high-latitude and high-altitude settings and are generally very well studied. Detailed documentation of upper-flow-regime bedforms, which were mostly produced by meltwater floods in the vicinity of the ice margin, come from successions related to the Fennoscandinavian (e.g. Hornung et al., 2007; Winsemann et al., 2009, 2011, 2018; Lang & Winsemann, 2013; Pisarska-Jamroży & Zielinski, 2014; Lang et al., 2017b) and British-Irish ice sheets (e.g. Lee et al., 2015; Leszczynska et al., 2017, 2018) in Europe and the Laurentide Ice Sheet in North America (e.g. Brennand, 1994; Russell & Arnott, 2003; Johnsen & Brennand, 2004; Dietrich et al., 2016, 2017).

The Late Palaeozoic Ice Age (Fielding et al., 2010; Montañez & Poulsen, 2013) resulted in the expansion of ice sheets across Gondwana, with a general diachronous trend charting the drift of South Gondwana over the South Pole. A rich archive of glacigenic deposits and associated sand-rich outwash complexes is recorded for example from Devonian basins in Bolivia (Bache et al., 2012), Late Carboniferous to Permian basins of Argentina, Brazil and Uruguay (Vesely et al., 2015; Aquino et al., 2016; Alonso-Muraga et al., 2018; Assine et al., 2018), roughly coeval successions in South Africa (Dietrich & Hofmann, 2019) and Ethiopia (Bussett, 2014), and Permian deposits in Oman (Martin et al., 2012) and Australia (Fielding et al., 2010). With the sole exception of the eastern Karoo Basin, South Africa (Dietrich et al., 2019), there is no specific interpretation of upper-flow-regime bedforms in Late Palaeozoic glacigenic strata. The reason for this is probably neither lack of evidence nor failed preservation, but failed recognition.

Successions related to the Late Ordovician (Hirnantian) glaciations of Gondwana are widespread and well documented from North Africa and the Arabian Peninsula (Le Heron et al.,
Table 1. Overview of well documented field examples of upper-flow-regime bedforms in glacigenic deposits.

| Bedform                  | Grain size      | Geometry | Sedimentary structures                                                                 | Dimensions\(^b\)                                                                 | Location; Age; Database                      | References                |
|-------------------------|-----------------|----------|----------------------------------------------------------------------------------------|----------------------------------------------------------------------------------|---------------------------------------------|---------------------------|
| Subglacial (esker)      |                  |          |                                                                                       |                                                                                  |                                             |                           |
| Antidunes               | Sand to gravel  | Tabular  | Sinusoidal stratification                                                              | d: 1–3 m; λ: 4–10 m; y: 1–3 m; y/λ: 0.25–0.3; θ: 10°                             | Wales (UK); Pleistocene; outcrop            | Lee et al. (2015)          |
| Antidunes               | Pebble to cobble gravel | Tabular  | Sinusoidal stratification                                                              | λ: 12 m; y: 3 m; y/λ: 0.25                                                      | Alaska (USA) and Iceland; Holocene; georadar | Burke et al. (2008, 2010b) |
| Glaci fluvi al          |                  |          |                                                                                       |                                                                                  |                                             |                           |
| Chutes-and-pools        | Pebbly sand     | Lenticular | Scours filled by concave-up backsets                                                   | l: 0.3–1.3 m; d: 0.04–0.3 m; d/l: 0.2; θ: 12–34°                              | Northern Germany; Pleistocene; outcrop      | This study                |
| Antidunes               | Pebbly sand     | Tabular  | Low-angle cross-stratification and subhorizontal stratification                        | d: 0.05–0.2 m; θ: 5–10°                                                      | Northern Germany; Pleistocene; outcrop      | This study                |
| Antidunes               | Pebbly sand     | Tabular  | Low-angle cross-stratification                                                        | d: 0.05 m; θ: <10°                                                            | Iceland; Holocene; outcrop                  | Kjær et al. (2004)        |
| Upper-stage-plane beds  | Pebbly sand     | Tabular  | Planar-parallel stratification                                                        | d: 0.3 m                                                                      | Iceland; Holocene; outcrop                  | Kjær et al. (2004)        |
| Upper-stage-plane beds  | Granule to cobble gravel | Tabular  | Planar-parallel stratification                                                        | d: 0.8–1.2 m                                                                  | Poland; Pleistocene; outcrop               | Pisarska-Jamrózy & Zieliński (2014) |
| Upper-stage-plane beds  | Coarse-grained sand | Tabular  | Planar-parallel stratification                                                        | d: 0.1–1.0 m                                                                  | Poland; Pleistocene; outcrop               | Pisarska-Jamrózy & Zieliński (2014) |
Table 1. (continued)

| Bedform                          | Grain size                  | Geometry     | Sedimentary structures<sup>a</sup>                                                                 | Dimensions<sup>b</sup>                                                                                                           | Location; Age; Database | References                      |
|----------------------------------|-----------------------------|--------------|----------------------------------------------------------------------------------------------------|-------------------------------------------------------------------------------------------------------------------------------------|-------------------------|---------------------------------|
| Jökulhaups/glacial lake-outburst floods |                             |              |                                                                                                    |                                                                                                                                     |                         |                                 |
| Cyclic steps                     | Pebble sand                 | Lenticular   | In-flow: Scours with thin massive basal layer and gently-dipping concave-up backsets. Across-flow: Scours with concave-up concentric infills or subhorizontal stratification | l: 2–13 m; d: 0.1–0.4 m; l/d: 0.04; x: 5–10°                                                                                  | Northern Germany; Pleistocene; outcrop | Lang & Winsemann (2013)          |
| Cyclic steps or antidunes        | ?                           | Tabular      | Sinusoidal stratification; shorter wavelength features are associated with isolated foreset-filled scours | l: 350 m; d: 10 m; λ: 60–90 m; y: 3.8–5.0 m; y/λ: 0.06                                                                             | Northern Germany; Pleistocene; outcrop and georadar | Winsemann et al. (2018); Kostic et al. (2019) |
| Chutes-and-pools                 | Interbedded gravel and sand | Lenticular   | Sigmoidal backsets                                                                                   |                                                                                                                                     | Switzerland; Pleistocene; outcrop | Fiore et al. (2002)             |
| Chutes-and-pools                 | Scoria granule gravel       | Lenticular   | Infill of regular-based scours; base concordant concave-up stratification; onlap on base of scour     | l: up to 60 m; d: 3–4 m; d/l: 0.09–0.11; λ: 10 m; y: 0.2 m; y/λ: 0.02                                                             | Iceland; Holocene; outcrop | Duller et al. (2008)            |
| Local hydraulic jump formed at isolated trough or obstacle | Scoria granule gravel       | Lenticular   | Infill of scours with steep upflow margin; convex-up stratification; onlap on scour base and upflow margin | l: 13 m; d: 3 m; d/l: 0.11                                                                                                       | Iceland; Holocene; outcrop | Duller et al. (2008)            |
| Local hydraulic jump formed at obstacle | Cobble to boulder gravel    | Lenticular   | In-flow: Scour fill with foresets (lee side) and backsets (stoss side). Across-flow: Convex-up scour fill | l: 40–70 m; w: 15–20 m; d: 3–7 m; d/l: 0.1–0.7                                                                               | Iceland; Holocene; outcrop and georadar | Burke et al. (2010a)            |
| Antidunes (net-erosional)        | Pebble gravel               | Waveforms    | Waveforms are eroded into diamicton and draped by up to 2.5 m thick gravel                          | λ: 300 m; y: >10 m; y/λ: 0.03                                                                                                      | Altai Mountains (Russia); Pleistocene; outcrop | Carling et al. (2002, 2009)      |
| Bedform                          | Grain size       | Geometry | Sedimentary structures | Dimensions^b | Location; Age; Database | References               |
|---------------------------------|------------------|----------|------------------------|---------------|------------------------|--------------------------|
| **Antidunes**                   |                  |          |                        |               |                        |                          |
| (net-erosional)                 | Gravel           | Waveforms| Backsets               |               | British Columbia (Canada); Pleistocene; outcrop and georadar | Johnsen & Brennand (2004) |
| Non-breaking antidunes          | Scoria granule gravel | Tabular or lenticular | Concave-up stratification above a gently inclined concave-up basal surface | l: 45 m; d: 0.5–2.0 m; d/l: 0.01–0.09; l: 24–96 m; y: 0.3–1.3 m; y/l: 0.02; x: <10° | Iceland; Holocene; outcrop | Duller et al. (2008) |
| Antidunes                       | Pebby sandstone  | Tabular  | Sinusoidal stratification and subhorizontal stratification | d: 10–20 m; l: 100 m; y: 5 m; y/l: 0.05 | Murzuq Basin (Libya); Upper Ordovician; outcrop | Girard et al. (2015) |
| **Subaqueous ice-contact fans (glacilacustrine/ glacialmarine)** |                  |          |                        |               |                        |                          |
| Chutes-and-pools                | Fine to medium-grained sand | Lenticular | Diffusely graded or cross-stratified scour fills | l: 20 m; d: 3; d/l: 0.15 | Ontario (Canada); Pleistocene; outcrop | Russell et al. (2007) |
| Chutes-and-pools                | Granule to cobble gravel | Lenticular | Scours filled by concave-up backsets | l: 1.7–4.6 m; w: 2–13 m; d: 0.35–0.9; d/l: 0.11–0.26; x: 12–44° | Northern Germany; Pleistocene; outcrop and georadar | Lang et al. (2017b) |
| Chutes-and-pools                | Pebby sand       | Lenticular | *In-flow:* Scours filled by steeply dipping concave-up or planar backsets. *Across-flow:* Concave-up concentric scour fills | l: 0.4–0.8 m; d: 0.2–0.3 m; d/l: 0.4; x: 10–20° | Northern Germany; Pleistocene; outcrop | Lang & Winsemann (2013) |
| Chutes-and-pools                | Pebby sand       | Lenticular | Scours filled by planar or concave-up backsets | l: 0.4–13 m; w: 0.5–1.5; d: 0.08–1.5 m; d/l: 0.1–0.2; x: 10–20° | Northern Germany; Pleistocene; outcrop and georadar | Lang et al. (2017b) |
| Bedform | Grain size | Geometry   | Sedimentary structures $^a$ | Dimensions $^b$ | Location; Age; Database | References                  |
|---------|------------|------------|----------------------------|-----------------|--------------------------|-----------------------------|
| Local hydraulic jump | Pebbly sand | Lenticular | Massive or diffusely graded infill of steep-walled scours | $d$: 3 m | Ontario (Canada); Pleistocene; outcrop | Russell & Arnott (2003) |
| Local hydraulic jump | Granule to boulder gravel | Lenticular | Massive, normally graded or planar cross-stratified infill of scours | $l$: up to 25 m; $d$: 1–3 m; $l/d$: 0.04–0.12; $\alpha$: 10–30° | Northern Germany; Pleistocene; outcrop and georadar | Hornung et al. (2007) Winsemann et al. (2009) |
| Breaking antidunes | Pebble to cobble gravel | Tabular | Pebble to sandy gravel couplets with planar stratification; local gravel clusters | $l$: 8–10 m; $d$: 0.5–2.0 m; $d/l$: 0.06–0.2 | Ontario (Canada); Pleistocene; outcrop | Russell & Arnott (2003) |
| Breaking antidunes | Pebble sand | Tabular or lenticular | In-flow: Subhorizontal stratification or low-angle cross-stratification; scours with convex-up or concave-up foresets or backsets. Across-flow: Concave-up concentric scour fills or subhorizontal stratification | $l$: 0.5–16 m; $w$: 2–13 m; $d$: 0.3–1.5 m; $d/l$: 0.09–0.6 m; $\alpha$: 5–15° | Northern Germany; Pleistocene; outcrop and georadar | Lang et al. (2017b) |
| Breaking antidunes | Granule to cobble gravel | Lenticular or tabular | Subhorizontal stratification or low-angle cross-stratification | $l$: 1.9–9.5 m; $d$: 0.2–1.5 m; $d/l$: 0.1; $\alpha$: <10° | Northern Germany; Pleistocene; outcrop and georadar | Lang et al. (2017b) |
| Non-breaking antidunes | Pebble to cobble gravel | Tabular | In-phase waveforms | $\lambda$: 12 m; $y$: 0.5 m; $y/\lambda$: 0.04 | Ontario (Canada); Pleistocene; outcrop | Brennand (1994) |
| Non-breaking antidunes | Pebbly sand | Tabular | In-phase waveforms; internal diffuse grading | $\lambda$: 5–20 m; $y$: 0.25–1.25 m; $y/\lambda$: 0.05 | Ontario (Canada); Pleistocene; outcrop | Brennand (1994) |
| Non-breaking antidunes | Medium-grained sand | Tabular | Low-angle cross-stratification | $l$: 8–10 m; $d$: 0.4–1.5 m | Ontario (Canada); Pleistocene; outcrop | Russell & Arnott (2003) |
Table 1. (continued)

| Bedform                | Grain size                        | Geometry | Sedimentary structures\(^a\) | Dimensions\(^b\) | Location; Age; Database References |
|------------------------|-----------------------------------|----------|-----------------------------|------------------|-----------------------------------|
| Non-breaking antidunes | Sand and pebbly sand               | Tabular  | In-flow: Sinusoidal stratification. Across-flow: Concave-up or convex-up lenticular and subhorizontally undulating geometries | l: 7–40 m; w: 3–13 m; d: 0.5–1.2 m; l: 1.2–40 m; y: 0.1–0.7 m; y/\(\lambda\): 0.02–0.05; \(\alpha\): <10° | Northern Germany; Pleistocene; outcrop and georadar | Lang & Winsemann (2013); Lang et al. (2017b) |
| Delta slopes (glacilacustrine/glacimarine) | | | | | |
| Cyclic steps\(^c\) | Mud, <5% sand                      | Waveforms| Downslope asymmetrical sinusoidal stratification | l: 3000 m; w: 2000 m; d: 60 m; d/l: 0.02; \(\lambda\): 16–600 m; 0.1–8.6 m; y/\(\lambda\): 0.02; \(\alpha\): 3.8–7.7° (stoss side) 1.7–2.9° (lee side) | Yukon (Canada); Holocene; sub-bottom profiles | Gilbert & Crookshanks (2009) |
| Cyclic steps          | Pebby sand                        | Lenticular| Scours with gently dipping concave-up backsets | d: 1–3 m; \(\lambda\): 10–20 m; y: 1–3 m; y/\(\lambda\): 0.1 | Quebec (Canada); Pleistocene; outcrop | Dietrich et al. (2016) |
| Cyclic steps          | Medium to coarse-grained sand      | Waveforms| Massive or subhorizontally stratified | d: 1–5 m; \(\lambda\): 10–20 m | Quebec (Canada); Pleistocene; outcrop | Dietrich et al. (2017) |
| Cyclic steps          | Pebby sand                        | Lenticular| In-flow: Scours with concave-up or sigmoidal backsets; basal scour fill may be massive or normally graded. Across-flow: Concave-up or convex-up concentric scour-fills | l: 0.7–7 m; w: 1–5 m; d: 0.08–1.6 m; d/l: 0.01–0.1; \(\alpha\): 5–23° | Northern Germany; Pleistocene; outcrop and georadar | Lang et al. (2017b); Winsemann et al. (2018) |
| Cyclic steps          | Pebby sand to pebble gravel       | Lenticular| Scours with gently dipping concave-up backsets | l: 0.5–1.0 m; d: 0.1–0.3 m; d/l: 0.2–0.3 | Northern Germany; Pleistocene; outcrop and georadar | Winsemann et al. (2011, 2018); Kostic et al. (2019) |
Table 1. (continued)

| Bedform            | Grain size       | Geometry          | Sedimentary structures                       | Dimensions\(^b\)                                                                 | Location; Age; Database                  |
|--------------------|------------------|-------------------|-----------------------------------------------|----------------------------------------------------------------------------------|------------------------------------------|
| Antidunes          | Pebble sand      | Lenticular        | Sinusoidal stratification and subhorizontal stratification | d: 0.2–5.0 m; λ: 3 m                                                                | Quebec (Canada); Pleistocene; outcrop    |
| Breaking antidunes | Pebble sand      | Tabular           | Low-angle cross-stratification                | l: 0.5–4.0 m; d: 0.05–0.7 m; α: 5–15°                                             | Dietrich et al., 2017                   |
| Non-breaking       | Medium to coarse-grained sandstone | Tabular | Horizontal to subhorizontal stratification within undulating bedforms | d: 1–10 m; λ: 40 m; y: 0.5–2.0 m; y/λ: 0.01–0.05                                  | Upper Ordovician; outcrop               |
| Non-breaking       | Medium-grained sand | Tabular           | Sinusoidal stratification                     | d: <1 m; λ: 1 m; y: 0.1 m; y/λ: 0.1 m                                               | Girard et al., 2012b                    |
| Glacimarine density flows | Fine-grained sandstone | Tabular           | Sinusoidal stratification                     | d: 2–3 m; λ: 2.0–2.5 m; y: 0.1–0.25 m; y/λ: 0.05–0.1 m                             | Illizi Basin (Algeria); Upper Ordovician; outcrop |
| Non-breaking       | Fine-grained sandstone | Tabular           | Sinusoidal stratification                     | d: 0.5–5.0 m; λ: 1.5–3.5 m; y: 0.1–0.25 m; y/λ: 0.07 m                            | Hirst, 2012                             |

\(^a\)Sedimentary structures are generally described in flow parallel direction only. If across flow descriptions are available, this is highlighted.
\(^b\)Length refers to flow parallel dimension; width refers to flow normal dimension. Only the available dimensions are provided.
\(^c\)The bedforms are referred to as ‘sediment waves’ in the original publication (Gilbert & Crookshanks, 2009). However, the inferred sedimentary process exactly matches cyclic steps.
The dominantly glacifluvial and glacimarine successions indicate depositional environments with abundant meltwater and supply of sand-grade sediment (Le Heron et al., 2009; Bataller et al., 2019). These conditions should favour the formation and preservation of upper-flow-regime bedforms, which indeed have been documented by a range of studies (Hirst et al., 2002; Le Heron et al., 2010, 2013, 2015; Hirst, 2012; Lang et al., 2012; Girard et al., 2012a,b, 2015; Moreau & Joubert, 2016).

Unlike Pleistocene and Palaeozoic records, no upper-flow-regime bedforms have been described from the Neoproterozoic record. Excellent exposures of diamictite successions are known globally (Arnaud et al., 2011). The issue appears to be that the successions are either dominated by thick, stacked diamictites (Le Heron et al., 2014) or alternatively by fine-grained turbidites intercalated with chemical precipitates (Lechte et al., 2018). Although upper-flow-regime bedforms are known from deposits of fine-grained low-density turbidity currents (Fedele et al., 2016; Lang et al., 2017a; Onishi et al., 2018), they have so far not been reported from the abundant slope and deep-marine deposits, which prevail as a general rule in the Cryogenian glacial record (Busfield & Le Heron, 2016; Spence et al., 2016). Ice-proximal successions comparable to the North Gondwana record of the Hirnantian glaciation were yet not described, possibly due to a preservational bias in glacimarine rift basins, favouring the preservation of diamictites over meltwater deposits (Eyles & Januszczak, 2004; Eyles, 2008).

UPPER-FLOW-REGIME BEDFORMS IN DIFFERENT GLACIGENIC DEPOSITIONAL ENVIRONMENTS

Subglacial environments

Subglacial meltwater deposits form in subglacial meltwater conduits and may be preserved in eskers or as the infills of tunnel channels and tunnel valleys (Fig. 1; Table 1). Eskers are typically deposited in meltwater conduits incised upward into the ice, while tunnel channels and tunnel valleys are incised into the substratum. Tunnel channels represent subglacial conduits adjusted to the magnitude of the (bankfull) meltwater discharge, while tunnel valleys are larger-scale features cut during multiple episodes of erosion and deposition. Both eskers and tunnel-valley fills may include ice-marginal and other non-subglacial deposits, and may include non-glacigenic deposits formed after ice-sheet retreat (Banerjee & McDonald, 1975; Brennand, 1994; Russell et al., 2003; Kehew et al., 2012; Lang et al., 2012; Janssen et al., 2013; Ahokangas & Mäkinen, 2014; Steinmetz et al., 2015).

Antidune deposits have been described from successions inferred to have been deposited in subglacial environments, although their formation should be suppressed under full pipe-flow conditions in subglacial conduits (Banerjee & McDonald, 1975). The occurrence of these deposits in subglacial environments can be explained by: (i) the occurrence of high-density, internally stratified underflows within a full conduit; (ii) non-full pipe-flow conditions; and (iii) flows expanding into subglacial cavities or lakes (Brennand, 1994; Brennand & Shaw, 1996; Fisher et al., 2003; Russell et al., 2003; Burke et al., 2008, 2010b; Lee et al., 2015). At the ice margins there is commonly a transition between subglacial tunnel-valley fills and eskers, respectively, and glacifluvial systems at terrestrial ice margins or subaqueous ice-contact fans, if the ice margin terminates in a water body (Gorrell & Shaw, 1991; Brennand, 1994; Russell et al., 2003; Deschamps et al., 2013; Ahokangas & Mäkinen, 2014).

Burke et al. (2008, 2010b) documented antidune strata in the basal part of modern esker deposits in Alaska and Iceland, with wavelengths of up to 12 m and amplitudes of up to 3 m. The eskers are interpreted to have formed during jökulhlaups in subglacial or englacial conduits close to the ice margin (Burke et al., 2008, 2010b). Fiore et al. (2002) observed chute-and-pool deposits in Pleistocene esker deposits in Switzerland. The deposits consist of sigmoidal gravelly and sandy backsets that infill up to 3 m deep and more than 7 m wide troughs. Chutes-and-pools are interpreted to have formed in wider segments of subglacial conduits during jökulhlaup conditions (Fiore et al., 2002).

Subaerial depositional environments

Sedimentary structures related to upper-flow-regime bedforms have commonly been described from subaerial meltwater deposits (Fig. 1A; Table 1). The observed sedimentary structures and bedforms are similar to those observed in non-glacigenic, high-energy, commonly braided fluvial systems (Alexander & Fielding, 1997; Fielding, 2006; Froude et al., 2017; Carling & Leclair, 2019) and alluvial fans (Blair, 1999). Fielding (2006) proposed ‘upper-flow-regime
sheets, lenses and scours fills’ (‘UFR’) as a possible architectural element in fluvial deposits supplementing the widely applied architectural-element scheme of Miall (1985). Upper-flow-regime bedforms are considered as potentially indicative of fluvial sedimentation under strongly seasonal climate, with pronounced variability and discharge peakedness (Alexander & Fielding, 1997; Fielding, 2006; Froude et al., 2017; Carling & Leclair, 2019; Wang & Pinkl-Björklund, 2020). These conditions are met also in glacifluvial systems, where meltwater discharge is subject to strong seasonal fluctuations (Marren, 2005; Cuffey & Paterson, 2010). Even more extreme discharge events in glacifluvial systems are represented by glacial lake-outburst floods or jökulhlaups caused by the release of water stored in proglacial, subglacial or supraglacial lakes. Such cataclysmic events contribute in a large measure to the accumulation of ice-proximal glacifluvial deposits (Pisarska-Jamroży & Zielinski, 2014) and may aggrade thick successions related to supercritical flows (Duller et al., 2008; Lang & Winsemann, 2013; Winsemann et al., 2016).

Glacifluvial deposits
Glacifluvial systems are formed by subaerial meltwater streams, which are generally characterized by braided channel networks (Fig. 1A) (Zielinski & van Loon, 2002, 2003; Blažauskas et al., 2007; Pisarska-Jamroży & Zielinski, 2014). The sedimentology and geomorphology of glacifluvial systems are controlled by the scale of the system, the meltwater and sediment supply, the local hydraulic gradients and lateral confinement. Steep glacifluvial systems commonly form small-scale fan shaped sediment bodies, referred to as sandar (singular: sandur), which may resemble alluvial fans. Gently sloping glacifluvial systems form larger-scale braidplains, which may merge into ice-marginal trunk rivers (Zielinski & van Loon, 2002, 2003; Blažauskas et al., 2007) or feed into glacifluvial deltas (Slomka & Hartman, 2019).

Deposits of upper-stage-plane beds and antidunes are interpreted to form in shallow braided channels and unconfined sheetfloods during high discharge events in the melt season (Figs 3 and 4) (Zielinski & van Loon, 2002, 2003; Krzyszkowski, 2002; Kjær et al., 2004; Pisarska-Jamroży & Zielinski, 2014). Kjær et al. (2004) described gravelly and sandy antidune deposits from an Icelandic sandur. Gravelly antidune deposits display subhorizontal and backset cross-stratification, while sandy antidune deposits display low-angle cross-stratification and are associated with sandy upper-stage-plane beds (Kjær et al., 2004). Pisarska-Jamroży & Zielinski (2014) described several sedimentary facies related to supercritical flows from Pleistocene glacifluvial deposits in Poland. Massive clast-supported gravel is interpreted as representing rapid sedimentation from highly sediment-laden supercritical flows during peak-flood conditions. Horizontally stratified gravel and sand are deposited as upper-stage-plane beds by waning supercritical flows. Low-angle gravelly or sandy foreset cross-stratification points to the flattening of dunes at the transition from subcritical to supercritical flow conditions (Pisarska-Jamroży & Zielinski, 2014). Typical vertical successions pass from planar or low-angle cross-stratified gravel via horizontally stratified gravel into horizontally stratified sand, indicating the formation of longitudinal or transverse bars. Bar vertical accretion decreases the local flow depth and allows for the formation of upper-stage-plane beds (Pisarska-Jamroży & Zielinski, 2014).

Deposits of glacial lake-outburst floods and jökulhlaups. Glacial lake-outburst floods or jökulhlaups are cataclysmic drainage events due to the release of water stored in proglacial, subglacial or supraglacial lakes. The largest glacial lake-outburst events relate to the drainage of ice-dammed lakes along the margins of continental ice-sheets (Baker, 1973; Herget, 2005; Carrivick, 2006; Alho et al., 2010; Carling, 2013; Margold et al., 2018; Lang et al., 2019; Panin et al., 2020). Erosion and deposition by such floods have an enormous geomorphological impact due to the breaching of drainage divides and the rerouting of drainage systems (Baker, 1973; Man-gerud et al., 2004; Gupta et al., 2007, 2017; Lang et al., 2019; Panin et al., 2020). Flow conditions during glacial lake-outburst floods are dominantly supercritical, due to the great flow depths. Supercritical flow conditions will commonly be attained due to local flow constrictions and shallow flows over valley margins or substratum obstacles (Carling et al., 2009; Alho et al., 2010; Bohorquez et al., 2016; Winsemann et al., 2016; Lang et al., 2019; Hansen et al., 2020).

Supercritical flows in outburst flood-related successions may form both net-erosional and net-depositional bedforms (Table 1). Net-erosional waveform structures of antidune origin incised during a major glacial lake-outburst flood were documented by Carling et al. (2002,
In plan-view, the waveforms are straight crested with wavelengths of ca 300 m and amplitudes of up to 20 m. The waveforms are incised into diamicton. The formation of the antidune bed-waves is attributed to flow thinning in the lee of an obstacle (Carling et al., 2002, 2009). Johnsen & Brennand (2004) documented slightly asymmetrical antidune waveforms of similar scale (wavelength 100 to 230 m, amplitude 6 to 14 m) on a glacifluvial delta plain truncated during a lake-drainage event. The preservation of up to 1.5 m thick backset cross-stratified gravel on the stoss-sides of the waveforms suggests a transition from

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degradational to aggradational conditions (Johnsen & Brennand, 2004; Carling et al., 2009).

Deposits of cyclic steps, chutes-and-pools and antidunes have been documented from successions related to glacial lake-outburst floods and jökulhlaups (Figs 3, 5 and 6) (Duller et al., 2008; Burke et al., 2010a,b; Winsemann et al., 2011; Lang & Winsemann, 2013; Le Heron et al., 2013; Girard et al., 2015; Winsemann et al., 2016; Hansen et al., 2020). Successions related to supercritical flows accumulated in the lee of glacitectonic ridges by a Pleistocene glacial lake-outburst flood were studied by Lang & Winsemann (2013) and Winsemann et al. (2016). An up to 15 m thick succession of pebbly sand, passing upward from cyclic-step and chute-and-pool deposits to deposits of antidunes, humpback dunes and three-dimensional dunes (Fig. 5A), was deposited within a channel incised by flows spilling over a glacitectonic ridge (Lang & Winsemann, 2013). Cyclic-step deposits are characterized by laterally and vertically amalgamated scour fills (length 2 to 13 m, depth 0.1 to 0.4 m) that contain low-angle (5 to 10°) backsets (Fig. 5A). Chute-and-pool deposits comprise amalgamated scour fills, which are 0.4 to 0.8 m long, 0.2 to 0.3 m deep and contain steeply dipping (10 to 20°) backsets. Deposits of chutes-and-pools are associated with antidune deposits that are characterized by up to 2.5 m
thick tabular beds of low-angle cross-stratified and subhorizontally stratified pebbly sand (Fig. 5A). The variety of facies is interpreted as representing a single flood event, which dissected the glacitectonic ridge and subsequently accumulated a vertical facies succession related...
to waning flow conditions (Lang & Winsemann, 2013). Deposits of more unconfined, overspilling floodwaters were described by Winsemann et al. (2016) from a denudated glacitectonic ridge (Fig. 5B and C). The up to 10 m thick tabular succession comprises subhorizontally stratified and low-angle cross-stratified medium-grained sand to pebbly sand, indicating probable deposition by antidunes. Isolated scour fills may point to the occurrence of chutes-and-pools. The antidune deposits are associated with deposits of 3D dunes and climbing ripples, pointing to pulsating discharge conditions with rapidly changing flow velocity and depth (Winsemann et al., 2016).

Duller et al. (2008) presented a detailed architectural analysis of Holocene volcaniclastic jökulhlaup deposits from an Icelandic sandur,
comprising antidune and chute-and-pool deposits (Fig. 6). Antidune deposits are characterized by tabular or lenticular concave-up bedsets that are conformably overlying concave-up basal surfaces. Based on outcrop observations, Duller et al. (2008) reconstructed antidune wavelengths...
ranging from 24 to 96 m and amplitudes of up to 1.3 m. Deposits of chutes-and-pools form convex-up bedsets that infill deep, steep-walled scours up to 13 m long and 3 m deep (Fig. 6A and B). Antidunes are interpreted to have been the dominant bed configurations during the waxing and waning stages of the jökulhlaup, while chutes-and-pools formed during peak flood conditions (Duller et al., 2008). Similar jökulhlaup deposits were described by Girard et al. (2015) from an Upper Ordovician glacifluvial delta-plain succession in Libya. Antidune deposits comprise up to 20 m thick, sinusoidally and subhorizontally stratified pebbly sandstone, displaying wavelengths of up to 100 m and amplitudes of up to 5 m (Girard et al., 2015). The distal equivalents of these deposits are deposits of antidunes and climbing dunes in sandstones accumulated at the delta front and hybrid-event beds on the delta slope (Girard et al., 2012a, 2012b).

Scour fills related to isolated hydraulic jumps were described from Icelandic jökulhlaup deposits by Burke et al. (2010a). Scours are 40 to 70 m long, 15 to 20 m wide and 3 to 7 m deep. Internally, the scour fills comprise cobble to boulder gravel, forming foresets on the lee sides and backsets on the stoss sides. Hydraulic jumps are interpreted to have formed upstream of stranded ice blocks that acted as flow obstacles, leading to scouring and deposition around the ice blocks (Burke et al., 2010a).

Subaqueous depositional environments

Bedforms related to supercritical density flows are common in deposits of glacialasturine or glaciomarine subaqueous ice-contact fans, deltas and glaciogenic deep-water fans (Fig. 1B; Table 1). A major distinction between glacialasturine and glaciomarine settings is given by the density difference between freshwater and seawater, affecting the density contrast between a sediment-laden flow and the ambient water and hence the mechanics of sediment transport by density flows. Glaciomarine settings may be more favourable to generate hyperpycnal flows due to the relatively low density of their ambient water (Russell & Arnott, 2003; Winsemann et al., 2009; Lang et al., 2017b), while glaciomarine settings may foster the formation of buoyant plumes (Powell, 1990; Dowdeswell et al., 2015).

Subaqueous ice-contact fans

Subaqueous ice-contact fans (also known as grounding-line fans) are fed by meltwater discharging from englacial or subglacial conduits of grounded ice margins into glacialasturine or glaciomarine basins (Fig. 1B) (Rust & Romanelli, 1975; Powell, 1990; Gorrell & Shaw, 1991; Lenné, 1995; Le Heron et al., 2004; Hornung et al., 2007; Winsemann et al., 2009; Deschamps et al., 2013; Aquino et al., 2016). The meltwater effluent behaves as a jet when entering a standing water body (Powell, 1990). Jets are momentum-driven flows that evolve into gravity-driven flows due to the turbulent entrainment of ambient water, leading to flow expansion and deceleration (Bates, 1953; Launder & Rodi, 1983; Hoyal et al., 2003). Glaciogenic jets with very high discharge and momentum may relate to outbursts of large volumes of meltwater from subglacial or englacial reservoirs (Russell & Arnott, 2003; Hornung et al., 2007; Winsemann et al., 2009; Dowdeswell et al., 2015).

The meltwater jet erodes large-scale scours that laterally transition into proximal fan-deposits. The geometries of the scour and of the fan are controlled by the densimetric Froude number at the outlet, by sediment-grain size and aggradation rate (Lang et al., in review; Powell, 1990; Hoyal et al., 2003). Winsemann et al.
(2009) documented large-scale (1.3 to 3.2 km long, 0.8 to 1.2 km wide, 7 to 25 m deep) scours that radiate from the former meltwater outlet and pass downflow into proximal gravel-rich fan lobes of a Pleistocene glacilacustrine ice-contact fan. The scours closely resemble those formed by experimental jets (Lang et al., in review; Hoyal et al., 2003) and are interpreted to indicate outlet densimetric Froude numbers larger than five (Winsemann et al., 2009). Proximal gravel-rich fan lobes are interpreted as the mouth bars of the glacigenic jet flow and are unconformably overlain by more distal sand-rich lobes deposited by sustained density flows on the lee side of and between the gravelly mouth bars (Hornung et al., 2007; Winsemann et al., 2009; Lang & Winsemann, 2013; Lang et al., 2017b). Deposits of supercritical flow are very common within gravelly and sandy subaqueous ice-contact fan deposits and may be considered characteristic facies for these systems (Figs 3, 7 and 8) (Brennand, 1994; Russell et al., 2003; Hornung et al., 2007; Russell et al., 2007; Winsemann et al., 2009; Lang & Winsemann, 2013; Leszczynska et al., 2017, 2018; Lang et al., 2017b). The large-scale geometry of gravel-rich fan deposits is wedge-shaped, with an overall gentle (3 to 5°) glacierward dip on the proximal side and basinward dip on the distal side (Lønne, 1995, 2001; Hornung et al., 2007; Russell et al., 2007; Winsemann et al., 2009). The most proximal fan deposits are characterized by highly scoured gravel (Fig. 7A to C). The largest and most proximal scour fills commonly display both foreset and backset cross-stratification, interpreted as the progradational infill of the scours or as deposits of local hydraulic jumps (Winsemann et al., 2009; Lang et al., 2017b). Scours are probably formed during an early flow-stage and relate either to the impingement of eddies generated by the expanding meltwater jet or slope failures on the inner margin of the mouth bar (Lang et al., in review; Lang et al., 2017b). Distally, scours become smaller and are associated with more sheet-like deposits, forming vertical and lateral stacks characterized by erosional contacts. Scour fills comprise backset or foreset cross-stratification or concentric concave-up stratification, indicating deposition by chutes-and-pools, isolated hydraulic jumps or breaking antidunes (Fig. 7C and D) (Lang et al., 2017b). Sheet-like deposits consist of low-angle cross-stratified, subhorizontally or sinusoidally stratified gravel formed by breaking or non-breaking antidunes (Brennand, 1994; Russell & Arnott, 2003; Leszczynska et al., 2017, 2018; Lang et al., 2017b). Planar and trough cross-stratified gravel indicates the migration of 2D and 3D gravel dunes under high-energy subcritical flows (Russell & Arnott, 2003; Hornung et al., 2007; Winsemann et al., 2009; Lang et al., 2017b).

Sand-rich fan deposits form gently (<15°) downflow dipping beds (Fig. 3) (Gorrell & Shaw, 1991; Hornung et al., 2007; Winsemann et al., 2007a, 2009; Lang et al., 2017b; Virtasalo et al., 2019). Slopes of subaqueous ice-contact fans may become steeper in basins with high topographic gradients (Lønne, 1995) or where subaqueous ice-contact fans prograde in confined settings (Deschamps et al., 2013). Internally, the sand-rich fan deposits may comprise laterally and vertically stacked lobe elements with a recurring internal architecture probably controlled by autogenic flow morphodynamics (Lang et al., 2017b). Low-angle cross-stratified, subhorizontally or sinusoidally stratified sand and pebbly sand, pointing to deposition by aggrading quasi-stationary antidunes, represent the typical facies association of these sand-rich fan deposits (Figs 3 and 8). These antidune deposits are associated with facies representative of chutes-and-pools and humpback dunes. Chute-and-pool deposits are characterized by scours filled by diffusely stratified or backset cross-stratified pebbly sand and occur both as isolated and as amalgamated stacks. They are commonly associated with laterally extensive erosional surfaces (Fig. 8A and B) (Gorrell & Shaw, 1991; Russell & Arnott, 2003; Russell et al., 2007; Winsemann et al., 2009; Lang & Winsemann, 2013; Leszczynska et al., 2017; Lang et al., 2017b). Lateral and vertical transitions from sigmoidally cross-stratified humpback dunes into sinusoidally stratified antidune deposits are common (Fig. 8D) (Russell et al., 2007; Lang & Winsemann, 2013; Lang et al., 2017b). Distally, thick successions of trough cross-stratified pebbly sand and climbing-ripple cross-laminated sand indicate deposition of dunes and climbing ripples by subcritical flows under high rates of suspension fall-out (Powell, 1990; Russell & Arnott, 2003; Winsemann et al., 2009; Deschamps et al., 2013).

Glaciogenic deltas
Glaciogenic deltas are formed by meltwater discharge at the shorelines of glacilacustrine or glacimarine basins (Fig. 1A). Glaciogenic deltas may be classified based on their position relative to the ice margin and by the gradient and
Fig. 8. Sand-rich subaqueous ice-contact fan deposits (Porta Fan, Middle Pleistocene; northern Germany). (A) Sinusoidally stratified antidune deposits separated by an irregular scour surface. Pebbles and sandy intraclasts occur at the scour surface. Trowel for scale is 28 cm. (B) Concentric scour fill, pointing to an isolated hydraulic jump (chute-and-pool deposit) at the contact between sinusoidally stratified antidune deposits. Trowel for scale is 28 cm. (C) Antidune deposits, comprising sinusoidally stratified fine-grained to medium-grained sand. Note internal erosive truncation (above trowel) and pinch-and swell of laminae. Trowel for scale is 28 cm. (D) Interbedded deposits of chutes-and-pools, antidunes and humpback dunes. Sinusoidally stratified antidune deposits form the lower part of the succession. The concentric scour fill in the left centre indicates an isolated hydraulic jump (chute-and-pool). Sigmoidal cross-stratification (upper right) points to deposition by humpback dunes. (E) Interbedded deposits of chutes-and-pools and antidunes. Note the steep-walled scour filled by three generations of backsets in the centre. Lateral (upflow) fining of backsets points to upflow migration of the hydraulic jump during chute-and-pool formation. (F) Sinusoidally stratified and low-angle cross-stratified sandy antidune deposits, forming packages that are 0.5 to 1.5 m thick and fine upward. Contacts between packages are erosional and commonly overlain by gravel. (G) Structureless inversely graded beds, forming part of antidune deposits. This stratification style indicates deposition from high-density flows.
accommodation space at the receiving basin margin. Ice-contact deltas are deposited directly at the ice margin and are characterized by a short and poorly developed subaerial delta plain, while glaci fluvi al deltas have well-devel oped delta plains and may develop at a considerable distance from the ice margin (Lønne, 1995). Ice-contact deltas are commonly characterized by coarse-grained gravity-flow deposits, including flow tills and resedimented till clasts. During progradation, ice-contact deltas may evolve into glaci fluvi al deltas (Lønne, 1995). High-gradient settings allow for the deposition of Gilbert-type deltas, displaying longitudinal profiles with a distinctive subdivision into topset, foreset and bottomset deposits, while low-gradient settings feature shoal-water deltas dominated by subaqueous mouth-bar complexes (Ashley, 1995; Winsemann et al., 2018). Depositional processes in deltas include tractional flows, sustained and surge-type, high-density and low-density turbidity currents, cohesionless debris flows and suspension fall-out (Nemec, 1990; Mulder & Alexander, 2001; Winsemann et al., 2011, 2018; Gobo et al., 2014, 2015).

Upper-flow-regime bedforms commonly occur within deposits of glaci fluvi al deltas (Fig. 3; Table 1) (Postma et al., 1983; Winsemann et al., 2011, 2018; Girard et al., 2012a,b; Dietrich et al., 2016, 2017; Moreau & Joubert, 2016; Nehyba et al., 2017; Normand eau et al., 2017; Lang et al., 2017b; Hanáček et al., 2018; Leszczynska et al., 2018; Kostic et al., 2019). The successions are very similar to those described from non-glaci genic coarse-grained deltas (e.g. Nemec, 1990; Gobo et al., 2014, 2015; Ventra et al., 2015; Massari, 2017; Kostic et al., 2019; Okazaki et al., 2020; Postma et al., 2020). The formation of upper-flow-regime bedforms is favoured by steep delta slopes of 5 to 35° (Lønne, 1995; Dietrich et al., 2017; Winsemann et al., 2018), by far exceeding the necessary threshold of 0.45 to 0.6° for submerged density flows to attain supercritical conditions (Sequeiros, 2012; Covault et al., 2017). Further parameters leading to the formation of upper-flow-regime bedforms are: (i) the occurrence of high-magnitude outburst floods (Girard et al., 2012a,b); (ii) rapid high-magnitude lake-level falls, triggering incision and bypass of coarse-grained sediment to the delta toe that commonly involve supercritical flows (Winsemann et al., 2011, 2016, 2018); and (iii) flow strength amplified by tidal drawdown (Dietrich et al., 2017).

Glaci fluvi al distributary systems on delta plains may display upper-flow-regime bedforms similar to those described in other glaci fluvi al systems (Fig. 3). Girard et al. (2012a,b) documented subaerial and subaqueous delta-plain deposits that comprise up to 10 m thick packages of sinusoidally and (sub-)horizontally stratified sandstones, partly forming undulating bedforms with wavelengths of up to 40 m, interpreted as the product of antidunes formed during the early waning stages of jökulhlaups. Antidune deposits are commonly associated with metre-scale scours filled by intraformational conglomerates that point to bypass and cut-and-fill processes (Girard et al., 2012a,b), probably related to chutes-and-pools. Dietrich et al. (2017) described antidune deposits from the infill of braided channels in Pleistocene subtidal delta-plain deposits in Canada. Antidune deposits comprise subhorizontally stratified pebbly sand, forming undulating bedforms with wavelengths between 5 m and 10 m (Dietrich et al., 2017).

Base-level fall may lead to the dissection of delta plains and slopes and to the formation of incised valleys (Fig. 9A). In glacial lacustrine settings, very rapid high-magnitude lake-level falls may occur due to the opening of outlet channels (Winsemann et al., 2011, 2018). The initial valley incision may relate to the formation of (erosional) cyclic steps (Strong & Paola, 2008; Winsemann et al., 2011, 2018; Muto et al., 2012; Kostic et al., 2019). Examples of such incised valleys related to lake-level falls include symmetrical to slightly asymmetrical bedforms (wavelength of ca 60 to 90 m) associated with isolated scour fills and interpreted as representing cyclic steps or antidunes (Winsemann et al., 2011, 2018; Kostic et al., 2019). Scours infilled by gravelly backsets occur along erosional surfaces related to forced regression and point to deposition in the hydraulic-jump zone of cyclic steps (Fig. 9A) (Kostic et al., 2019).

Upper-flow-regime bedforms on delta slopes are deposited by both surge-type and sustained turbidity currents. Deposits of surge-type supercritical turbidity currents are characterized by backset cross-stratified gravelly or sandy scour fills, interpreted as deposits of the hydraulic-jump zone of cyclic steps, associated with low-angle cross-stratification or sinusoidal stratification in pebbly sand or sand, pointing to deposition by antidunes (Fig. 9B) (Lang et al., 2017b; Winsemann et al., 2018; Kostic et al., 2019). Scour fills related to hydraulic jumps are commonly widely spaced and isolated, indicating deposition at the lower limit of the cyclic-step
stability field or to aggradation associated with the local development of chutes-and-pools (Fig. 10) (Massari, 2017; Lang et al., 2017a,b). Cyclic-step and antidune deposits commonly form the basal part of fining-upward successions and are overlain by ripple cross-laminated sand (Lang et al., 2017b; Winsemann et al., 2018), pointing to deposition by waning flows (Mulder & Alexander, 2001). Surge-type turbidity currents are probably triggered by frequent small-volume slope failures along the upper delta slope (Hughes Clarke, 2016), favoured by high aggradation rates (Gobo et al., 2014, 2015; Winsemann et al., 2018).

Deposits of sustained supercritical density flows are characterized by laterally extensive trains of scours filled by backset cross-stratified pebbly sand, probably deposited by cyclic steps.

Fig. 9. Upper-flow-regime bedforms in glacifluvial deltas. (A) Deposits of cyclic steps overlie erosional surfaces (red) formed during forced regression. Cyclic-step deposits comprise elongate scours filled by backsets. Prograding delta-foreset deposits downlap the erosional surfaces and cyclic-step deposits (Betheln Delta, northern Germany; Middle Pleistocene; modified from Kostic et al., 2019). (B) Delta-foreset deposits, comprising, isolated gravelly and sandy backset cross-stratified scour fills, which are interbedded with sinusoidally stratified sand. The succession is interpreted as deposits of cyclic steps and superimposed antidunes that were formed by surge-type supercritical density flows. Tool for scale is 1.4 m (Porta Delta, northern Germany; Middle Pleistocene). (C) Delta-foreset deposits, comprising laterally stacked scours filled by backset cross-stratified pebbly sand interpreted as cyclic steps deposited by sustained supercritical density flows (Freden Delta; northern Germany; Middle Pleistocene). (D) Detail of the stacked backset cross-stratified scour fills from Fig. 5C. (E) Isolated backset cross-stratified scour-fills (arrows) related to cyclic steps within foreset packages dominated by climbing-ripple cross-laminated sand (Freden Delta, photograph by courtesy of C. Brandes; modified from Lang et al., 2017b, and Winsemann et al., 2018).
The scours may amalgamate to form laterally extensive downflow-dipping composite erosion surfaces (‘pseudo-foresets’, Dietrich et al., 2016). Backsets in cyclic-step deposits may pass downflow into low-angle cross-stratified or sinu-oidal stratified pebbly sand or sand deposited by antidunes (Fig. 9C) (Dietrich et al., 2016, 2017; Lang et al., 2017b; Winsemann et al., 2018). Some foreset beds consist entirely of antidune deposits (Fig. 10). Cyclic-step and antidune deposits are associated with sandy foreset beds, comprising deposits of humpback dunes, (climbing) dunes and climbing ripples (Dietrich et al., 2016; Lang et al., 2017b; Winsemann et al., 2018). Sustained density flows on steep subaqueous slopes can stay supercritical and form hydraulic jumps at the base of the slopes (Komar, 1971; Clemmensen & Houmark-Nielsen, 1981; Mutti & Normark, 1987; Nielsen et al., 1988; Dabrio et al., 1991; Winsemann et al., 2007b; Covault et al., 2017; Rubi et al., 2018). Hydraulic jumps may also form on the slopes of Gilbert-type deltas, resulting in the formation of local scours filled with backsets or massive deposits, surrounded by or passing upslope into diffusely stratified, low-angle stratified or spaced planar stratified sand or pebbly sand in the upper or middle part of the foreset bed (Clemmensen & Houmark-Nielsen, 1981; Nemec, 1990; Massari, 1996, 2017; Dietrich et al., 2016; Selim, 2019; Okazaki et al., 2020). Hydraulic jumps formed on the upper part of the slope may also trigger the formation of upslope-migrating climbing-ripples on the lower delta slope due to large backflow areas (Clemmensen & Houmark-Nielsen, 1981; Winsemann et al., 2007b, 2018). The slope of these Gilbert-type deltas commonly varies between 5° and 26° and is thus well below the angle of repose, excluding grain-flow avalanches as a cause of the diffuse stratification or spaced planar stratification. Therefore, diffuse stratification and spaced planar stratification are interpreted to represent conditions of high densimetric Froude numbers in sustained density flows (Fedele et al., 2016).

Sustained supercritical density flows are triggered by sediment-laden flows, plunging over the delta brink and evolving into hyperpycnal flows. The formation of hyperpycnal flows is commonly interpreted as related to high-discharge events (Ventra et al., 2015; Carvalho & Veselý, 2017), such as jökulhlaups (Ghiennie et al., 2010; Girard et al., 2012a,b). Deposits pointing to sustained density flows are favoured in settings where the delta plain is bypassed during lake-level lowstand or highstand when
accommodation is low (Lang et al., 2017b; Winsemann et al., 2018).

Downslope, the steeply dipping foreset beds typically pass into gently dipping to flat toeset and bottomset beds. The slope break at the transition may trigger hydraulic jumps in supercritical density flows, leading to the formation of isolated scours filled by backsets or massive deposits, which may be incised into fine-grained delta-foreset or toeset deposits (Fig. 7E) (Winsemann et al., 2007b, 2018; Leszczynska et al., 2018). Increased sedimentation rates downflow of hydraulic jumps may cause the accumulation of thick successions featuring climbing-dune cross-stratification and climbing-ripple cross-lamination in the delta toeset (Winsemann et al., 2011, 2018).

Glacigenic subaqueous fans
Glacigenic subaqueous fans are deposited by glacial density flows and are common in deepwater glacimarine successions, where they extend downdip from subaqueous ice-contact fans or deltas (Fig. 3; Table 1) (Lønne, 1997; Plink-Björklund & Ronnert, 1999; Lajeunesse & Allard, 2002; Le Heron et al., 2010; Hirst, 2012; Lang et al., 2012; Dietrich et al., 2017, 2019; Normandeau et al., 2017).

Successions comprise sheet-like sediment bodies that represent depositional lobes (Lønne, 1997; Plink-Björklund & Ronnert, 1999; Le Heron et al., 2004, 2006; Hirst, 2012; Dietrich et al., 2017) or ribbon-shaped sediment bodies that represent the infills of channels (Hirst et al., 2002; Hirst, 2012; Girard et al., 2012a,b). The most commonly reported upper-stage bedforms from deposits of glacimarine subaqueous fans are antidunes (Fig. 11) (Hirst et al., 2002; Le Heron et al., 2010; Hirst, 2012; Lang et al., 2012; Girard et al., 2012a,b). The sedimentary facies of glacimarine lobe deposits are very similar to those of non-glacigenic unconfined supercritical density flows, whose facies signatures are typically dominated by antidune deposits (Lang et al., 2017a; Postma & Kleverlaan, 2018; West et al., 2019). Antidune deposits comprise sinusoidally stratified fine-grained sandstone with wavelengths from 1.5 to 3.5 m and amplitudes from 0.1 to 0.25 m, forming up to 5 m thick vertical stacks with antidune crests displaying upflow or downflow migration (Fig. 11A to C) (Hirst et al., 2002; Hirst, 2012). Antidune deposits are associated with deposits of chutes-and-pools and humpback dunes. Deposits of chutes-and-pools are characterized by steep-walled scours filled by partly deformed backsets (Fig. 11B). Humpback dunes display sigmoidal cross-stratification and may be laterally and/or vertically transitional to sinusoidally stratified antidune deposits (Fig. 11D) (Hirst, 2012; Le Heron et al., 2013).

DISCUSSION

Characteristic dimensions of bedforms and sedimentary structures

The compilation of the documented field examples (Table 1) provides an overview of the occurrence of upper-stage deposits in different glacigenic depositional environments and of their characteristic dimensions. The formation and preservation of the different bedform types relate to environment-controlled parameters like flow depth, flow velocity and aggradation rate, and to their temporal and spatial evolution. Upper-stage-plane beds are documented from glacial systems, where aggradation rates are comparatively low and bedforms are thus poorly preserved. Deposits of antidunes are documented from all studied glacigenic depositional environments, although their architecture and dimensions are highly variable, depending on the respective setting. Cyclic-step deposits occur in successions from deltas and glacial lake-outburst floods. Chute-and-pool deposits were observed in successions from glacial systems, subaqueous (ice-contact) fans and glacial lake-outburst floods. However, the distinction between deposits of cyclic steps and chutes-and-pools may be ambiguous (cf. Lang & Winsemann, 2013; Postma et al., 2020). Some sedimentary structures interpreted as cyclic-step deposits (Fig. 9B) may also represent deposits of chutes-and-pools, and vice versa (Fig. 10) (Masaari, 2017; Lang et al., 2017a; Okazaki et al., 2020).

Dimensions that are commonly considered characteristic for different bedforms and sedimentary structures can also be linked to flow parameters, including the wavelength and wave steepness of bedforms, and aspect ratio and spacing of scours (Kennedy, 1963; Hand, 1974; Cartigny et al., 2014; Fedele et al., 2016). Wave-length and wave steepness can only be measured from well-preserved deposits, while values of aspect ratio and spacing of scours can also be obtained from less complete successions (Symons et al., 2016). In glacigenic depositional
environments, bedform wavelength and steepness are documented from deposits of cyclic steps, chutes-and-pools and non-breaking antidunes (Table 1). Cyclic-step wavelengths from the slopes of glacigenic deltas vary over three orders of magnitude (Fig. 12A). Shorter wavelengths relate to coarse-grained systems with steep gradients, while longer wavelengths relate to fine-grained systems with gentle gradients. The linkage between grain size and slope and bedform wavelength is in line with other studies of experimental and natural bedforms (Kostic, 2011; Slootman & Cartigny, 2019). If deposits of cyclic steps and antidunes occur in the same depositional system, cyclic-step wavelengths are one order of magnitude larger than antidune wavelengths (Fig. 12A), matching the results of numerical simulations by Kostic (2014) and Kostic et al. (2019).

The wavelength of antidunes and their deposits scales with the flow depth (Kennedy, 1963; Hand, 1974). The variation of antidune

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Fig. 11. Upper-flow-regime bedforms deposited by supercritical density flows in glacimarine settings (North Africa; Upper Ordovician). (A) Sheet-like sandstones that display sinusoidal stratification and low-angle cross-stratification with common internal low-angle truncations, pointing to deposition by antidunes (Jebel Asba region, Kufra Basin, Libya). (B) Thickly bedded sinusoidally stratified sandstone, indicating highly aggrading antidune deposits, is unconformably overlain by foreset and backset cross-stratified sandstone, probably indicating chutes-and-pools. Hammer for scale is 33 cm (Jebel Eghei region, Kufra Basin, Libya; from Le Heron et al., 2014). (C) Sinusoidally stratified sandstone deposited by antidunes (Tassili N’Ajjer region, Illizi Basin, Algeria). (D) Sigmoidally stratified sandstone deposited by humpback dunes, passing upward into sinusoidally stratified antidune deposits. Hammer for scale is 33 cm (Jebel Asba region, Kufra Basin, Libya; from Le Heron et al., 2010).
Different wavelengths over three orders of magnitude (Fig. 12A) can thus be explained by strong variations of the flow velocities and depths between different environments. Glacigenic deltas are rather small systems, but their steep slopes will cause strong acceleration of flows (e.g., Dietrich et al., 2016; Lang et al., 2017b; Winsemann et al., 2018). Relatively shallow fast flows will cause high densimetric Froude numbers that may trigger antidune-wave breaking (Kennedy, 1963), potentially resulting in a transition to cyclic steps (Fedele et al., 2016). Jökulhaups or glacial lake-outburst floods are characterized by high-magnitude, deeper flows (Duller et al., 2008; Alho et al., 2010; Winsemann et al., 2011, 2016; Girard et al., 2012a,b; Lang et al., 2019), leading to the formation of correspondingly long wavelength bedforms (Fig. 12A). Antidune deposits in subaqueous (ice-contact) fans display a wide range of wavelengths (Fig. 12A), indicating that a wide range of flow magnitudes may occur in these depositional systems. Previous field-based studies suggested that longer wavelength bedforms relate to higher-magnitude events (Hirst et al., 2002; Russell & Arnott, 2003; Lang & Winsemann, 2013; Lang et al., 2017b).

The wave steepness (wave height/wavelength) of antidune and cyclic-step deposits is fairly similar in all studied depositional environments (Fig. 12B). Antidune deposits have wave steepness values consistently lower than 0.1, which is below the experimentally determined critical antidune-wave steepness of 0.142 (Kennedy, 1963). If antidune-wave steepness exceeds the critical value, wave breaking and reworking of the bed will occur (Kennedy, 1963; Middleton, 1965). Therefore, it is not surprising that the steepness of preserved bedforms is well below the critical value.

The aspect ratios (depth/length) of scours related to cyclic steps, chutes-and-pools and breaking antidunes show no clear trends or relation to the depositional environments (Fig. 12B). Scours interpreted as formed by cyclic steps tend to display lower and less variable aspect ratios, while other bedform types display a wider range of values.

**Controls on the formation and preservation of upper-flow-regime bedforms**

The documented bedforms and bedform successions from glacigenic settings are very similar to those known from non-glacigenic settings. The great diversity of glacigenic depositional environments allows for the study of bedforms and sedimentary facies deposited under a wide range of flow conditions and sediment supply. The

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differences in the observed sedimentary facies are controlled by various parameters, including: (i) slope; (ii) confined versus unconfined conditions; (iii) sediment concentration and aggradation rate; (iv) dominant grain size; and (v) discharge variations. However, all of these controlling factors relate to parameters that characterize the respective depositional system but are not specific for glacigenic settings. Observations from glacigenic depositional environments can therefore be transferred to non-glacigenic depositional environments, and vice versa. Glaciﬂuvial systems can be compared to other high-energy ephemeral alluvial or fluvial systems (Blair, 1999; Fielding, 2006; Froude et al., 2017; Carling & Leclair, 2019; Wang & Plink-Björklund, 2020). Glacigenic Gilbert-type deltas display the same facies types and depositional architectures as non-glacigenic Gilbert-type or fan deltas (Massari, 2017; Kostic et al., 2019; Okazaki et al., 2020; Postma et al., 2020), as do glacigenic density ﬂows and non-glacigenic density ﬂows (Lang et al., 2017a; Ono & Plink-Björklund, 2018; West et al., 2019). Proximal subaqueous ice-contact fans and ice-contact deltas may represent an exception here, because they only occur in glacigenic settings and their deposition is controlled by speciﬁc ice-marginal conditions. However, the medial to distal reaches of glacigenic subaqueous ice-contact fans are characterized by density ﬂows that present the same dynamics as in nonglacial settings (Lang et al., 2012, in review; Powell, 1990; Hirst, 2012). Glacial lake-outburst-floods may affect all glacigenic depositional environments, commonly leading to very high-magnitude ﬂows and to the formation of exceptionally large bedforms (Duller et al., 2008; Winsemann et al., 2011, 2018; Girard et al., 2012a,b; Carling, 2013).

Bed slope and flow conﬁnement
The bed slope controls ﬂow acceleration by gravity, and thus has a huge impact on bedform evolution. Steep gradients, for example in delta systems, tend to favour the formation of cyclic steps (Figs 7 and 8) (Dietrich et al., 2016; Lang et al., 2017b; Winsemann et al., 2018), while gentler gradients, for example in glaciﬂuvial systems or subaqueous (ice-contact) fans, are more prone to the formation of antidunes (Figs 2 and 6) (Russell & Arnott, 2003; Lang & Winsemann, 2013; Pisarska-Jamroży & Zieliński, 2014). The impacts of slope and flow conﬁnement commonly act in conjunction, and slope breaks commonly also lead to a transition from conﬁned to unconfined ﬂows (Stacey et al., 2019). Flow conﬁnement impacts the run-out distance and competence of ﬂows. Flows that leave a conﬁned conduit are prone to dissipation due to rapid sediment deposition and water entrainment (Cartigny et al., 2011; Hamilton et al., 2015; Stacey et al., 2019). In settings characterized by supercritical ﬂows this leads to downﬂow transitions in bedform type and wavelength, and ultimately to a transition to subcritical ﬂow conditions (Fedele et al., 2016; Normandeau et al., 2016; Stacey et al., 2019). Based on previous ﬁeld examples, Fedele et al. (2016) suggested that cyclic steps are more common in conﬁned settings (i.e. channels), while antidunes prevail in non-conﬁned settings (i.e. lobes). Observations from glacigenic successions support this idea. Cyclic-step deposits are common in more conﬁned settings, such as the base of chutes, channels or incised valleys (Figs 5A and 9A) (Winsemann et al., 2011; Lang & Winsemann, 2013; Winsemann et al., 2018). Studies from modern delta systems suggest that cyclic steps preferentially form in channels on the delta slope (Hage et al., 2018; Stacey et al., 2019). Antidune deposits tend to dominate in unconfined settings, for example subaqueous aprons of ice-contact fans and lobes deposited by glaciﬂuvial density ﬂows (Figs 8 and 11) (Hirst et al., 2002; Hirst, 2012; Lang et al., 2017b).

Sediment concentration and aggradation rate
Physical and numerical ﬂow experiments have shown that the architecture and internal sedimentary structures of bedforms developed under supercritical ﬂows depend on the sediment concentration, grain size and aggradation rate (Cartigny et al., 2013, 2014; Vellinga et al., 2018; Ono et al., 2020). The effects of sediment concentration, grain size and aggradation rate may hamper the unambiguous recognition of bedforms at outcrop, as reﬂected by the broad range of sedimentary facies observed in glacigenic successions (Fig. 13), for example for deposits of cyclic steps and antidunes. Net-erosional bedforms may form during glacial lake-outburst ﬂoods under conditions of high discharge with comparatively low sediment availability or dominant bypass (Carling et al., 2002, 2009; Johnsen & Brennand, 2004). Cyclic steps formed during the incision of valleys by lake-outburst ﬂoods or lake-level falls comprise laterally amalgamated or spaced scour ﬁlls (Lang & Winsemann, 2013; Winsemann et al., 2016, 2018; Kostic et al., 2019).
2019), indicating the preservation of only the basal part of bedforms when aggradation rates are low (Fig. 13). Higher aggradation is documented by cyclic steps deposited on delta slopes, where trains of laterally stacked scour fills bear evidence of upslope migration (Dietrich et al., 2016; Lang et al., 2017b; Winsemann et al., 2018). Repeated surveys of cyclic steps on a modern channelized glacifluvial delta slope suggest that only ca 11% of the bedform stratigraphy is preserved, while exceptionally high aggradation and preservation relate to less common delta-brink failures (Vendettuoli et al., 2019). Gilbert & Crookshanks (2009) showed that preservation of complete cyclic-step bedforms may occur on fine-grained distal delta-slopes. These cyclic-step deposits resemble those deposited in deep marine settings under highly aggradational conditions (Cartigny et al., 2011; Kostic, 2011; Symons et al., 2016; Lang et al., 2017a). Antidunes also produce a range of sedimentary structures under different aggradation rates. Low-angle backsets and shallow scour fills are deposited in glacifluvial systems during the melt season (Fig. 13) (Kjær et al., 2004; Pisarska-Jamroży & Ziebiński, 2014). These antidune deposits indicate rather low aggradation rates, although glacifluvial systems during the melt...
season typically have comparatively high discharge and sediment concentration compared with other fluvial systems (Marren, 2005). In contrast, antidunes deposited by jökulhlaups (Duller et al., 2008; Girard et al., 2015) or subaqueous density flows (Brennand, 1994; Hirst, 2012; Lang & Winsemann, 2013; Lang et al., 2017b; Winsemann et al., 2018) comprise thickly bedded, sinusoidally stratified units (Fig. 13), where antidune waveforms are almost completely preserved. These deposits are similar to antidune successions deposited by highly aggradational turbidity currents (Ito & Saito, 2006; Ito, 2010; Lang et al., 2017a; West et al., 2019).

Different sediment concentrations and aggradation rates produce variable stratification styles. Well-developed internal lamination is the most common stratification style in glacigenic upper-flow-regime bedforms (Figs 4, 5C and 8C to E), indicating grain-by-grain bedload aggradation under turbulent low-density flows. The occurrence of spaced or crude stratification in some beds (Figs 4B, 7A, 7E, 7F and 11B) points to flows with higher sediment concentration (Cartigny et al., 2013). These beds may display normal (Fig. 7A and F) or inverse (Fig. 8G) grading, the latter commonly attributed to traction carpets (cf. Hiscott, 1994; Sohn, 1997). However, antidune deposits may display highly variable stratification styles and many deposits attributed to traction carpets may relate to antidunes (Yagishita, 1994; Lang et al., 2017a; Postma et al., 2020). The occurrence of structureless, diffusely stratified and graded beds (Fig. 6C and D) points to deposition by concentrated flows (Russell & Arnott, 2003; Winsemann et al., 2004, 2007a, 2009; Duller et al., 2010) where the formation of stratification is suppressed by high sediment concentration (Arnott & Hand, 1989; Cartigny et al., 2013). Structureless or normally graded scour fills (Figs 7A, 7D and 9B) are attributed to rapid suspension fall-out in the hydraulic-jump zones of cyclic steps or chutes-and-pools (Postma et al., 2009; Postma & Cartigny, 2014; Lang et al., 2017a,b; Winsemann et al., 2018).

Sediment grain size
Bedform geometry and sedimentary facies in upper-flow-regime deposits are impacted by grain size (Cartigny et al., 2014; Postma & Cartigny, 2014; Ono et al., 2020). However, detailed studies of very fine-grained (i.e. mud-dominated; e.g. Gilbert & Crookshanks, 2009) and very coarse-grained (i.e. gravel-dominated; e.g. Brennand, 1994; Lang et al., 2017b) bedforms are rare. The cyclic-step bedforms and depositional architecture preserved on a muddy glacilacustrine delta slope (Fig. 13) (Gilbert & Crookshanks, 2009) resemble those documented from deep marine settings (Cartigny et al., 2011; Kostic, 2011; Symons et al., 2016; Lang et al., 2017a). Deposits comprising gravelly sediment tend to display steeper scour margins and backsets and a predominance of lenticular bed geometries with intense internal scouring (Fig. 7B to D) (Winsemann et al., 2009; Lang et al., 2017b). Steeper scour margins and backsets can be related to more rapid settling of coarser-grained clasts. The predominance of lenticular scour fills is caused by the stronger flows required to transport coarse-grained sediment and the lower threshold for antidune-wave breaking or hydraulic-jump formation in flows over hydraulically rough gravelly beds (Breakspear, 2008). The extreme end-member of coarse-grained sediment in combination with low aggradation is represented by the formation of transverse ribs. Transverse ribs, consisting of regularly spaced clusters of imbricated cobbles to boulders, have been described from modern glaciﬂuvial systems (Gustavson, 1974) and are interpreted to form beneath antidune waves (Gustavson, 1974) or beneath trains of hydraulic jumps (Allen, 1983; Grant, 1997), similar to cyclic steps. The occurrence of imbricated gravel clusters on laterally continuous erosional surfaces (Winsemann et al., 2009; Lang et al., 2017b) indicates that transverse ribs are the first bedforms to be deposited after extensive scouring. The clustering of the gravel also hinders remobilization, favouring the preservation of these structures (Allen, 1983).

Discharge variations
Both in glacigenic and non-glacigenic depositional environments the formation and preservation of upper-flow-regime bedforms and sedimentary structures are commonly attributed to discharge variations, especially to high-magnitude flood events. Such discharge variations may affect all studied sub-environments of deposition.

In subaerial settings upper-flow-regime bedforms generally record high-discharge events and/or high discharge variations during the melt season (Kjær et al., 2004; Pisarska-Jamroży & Zielinski, 2014), jökulhlaups (Burke et al., 2008, 2010a,b; Duller et al., 2008; Girard et al., 2015) or the catastrophic drainage of proglacial lakes.
Deposits of subaqueous supercritical density flows are also commonly attributed to high-discharge events (Russell & Arnott, 2003; Winsemann et al., 2009; Girard et al., 2012a,b; Lang et al., 2017b). In some field examples, jökulhlaup deposits that include sedimentary structures related to supercritical flows could be laterally traced from delta-plain into delta-slope deposits (Girard et al., 2012a,b). Observations from modern systems demonstrate that supercritical density flows on a delta slope are triggered on a daily basis from low-density river plumes, and high-discharge events are not a prerequisite (Hage et al., 2018; Hizzett et al., 2018; Vendettuoli et al., 2019). Less common high-magnitude events have far longer run-out distances and supply large amounts of sediment that may partly be redistributed by subsequent lower-magnitude flows (Stacey et al., 2019; Vendettuoli et al., 2019). Many glacilacustrine and glacimarine successions deposited by supercritical density flows display coarse-grained sediment and high thickness, which can more easily be explained by aggradation during high-discharge events. Dowdeswell et al. (2015) pointed out that the deposition of large subaqueous ice-contact fans and thick underflow deposits requires higher discharges, flow velocities and sediment concentrations than those measured in modern systems.

Flow parameters during high-discharge events, like flow velocity and inundation depth, can be estimated from the dimensions of the preserved upper-flow-regime bedforms (Duller et al., 2008; Winsemann et al., 2011). Such quantitative estimates are straightforward in subaerial settings, where bedform wavelength can be directly linked to the flow velocity and depth (Duller et al., 2008; Froude et al., 2017). For bedforms deposited by density flows in subaqueous settings such quantitative estimates need to be handled more carefully. In subaqueous density flows, the excess density of the flow compared with the ambient fluid has to be considered when calculating flow parameters from bedform dimensions (Hand, 1974; Fedele et al., 2016). However, the flow density is very hard to estimate from the depositional record. Furthermore, in many cases (especially high-density flows) a calculated flow depth may only be representative for the dense basal layer of a flow (cf. Postma et al., 1988; Postma & Cartigny, 2014).

CONCLUSIONS

Upper-flow-regime bedforms are ubiquitous in glacigenic depositional environments. A review of the existing literature shows that several studies have presented detailed analyses of upper-flow-regime bedforms and sedimentary facies architecture, as well as numerous studies that simply noted the occurrence of deposits attributed to supercritical flows.

Deposits of upper-stage-plane beds, antidunes, chutes-and-pools and cyclic steps are known from all kinds of glacigenic depositional systems, and typically are indicative of ice-proximal settings and/or high-discharge events. The bedforms can develop in subglacial, subaerial and subaqueous environments, recording deposition by both open-channel flows and submerged density flows. In addition to modern and Pleistocene records, upper-flow-regime bedforms are increasingly recognized from Palaeozoic (particularly Late Ordovician, and to a lesser extent Permo-Carboniferous) glacigenic deposits.

This review implies that the bedforms and bedform successions are generally not specific for any glacigenic depositional environment, and that very similar successions are documented from all sub-environments. Lateral and vertical trends in bedform distribution are similar to those in non-glacigenic settings. Important controlling factors for bedform evolution that differ between depositional environments include the degree of flow confinement, gradient of the depositional surface, sediment concentration, grain size and aggradation rate. Deposits of cyclic steps are more common in confined settings or on steeper slopes, for example in channels or incised valleys or along the foresets of glacifluvial deltas. In contrast, antidune deposits dominate in unconfined settings, like glaci fluvi al fans or subaqueous (ice-contact) fans.

Differences in bedform architecture and the resulting sedimentary facies and stratification styles are affected by different sediment concentration, aggradation rate and grain size, as demonstrated by previous physical and
ACKNOWLEDGEMENTS

Open access funding provided by Projekt DEAL. We would like to thank the editorial team (A. Sloatman, M. Cartigny, A. Normandeau, D. Ventra and S. Hubbard) of the Sedimentology special issue for inviting us to contribute this paper. Discussions on bedform morphodynamics with J. Fedele, D. Hoyal and G. Postma helped to sharpen our ideas. T. Hartmann helped with artwork. Special thanks go to the owners of the open-pits for the permission to work on their properties and R. Duller, M. Pisarska-Jamroży and T. Zieliński for sharing outcrop photographs. Constructive reviews by P. Carling, R. Duller and D. Ventra are highly appreciated.

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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Manuscript received 28 May 2019; revision 7 June 2020; revision accepted 16 June 2020

Supporting Information

Additional information may be found in the online version of this article:

Table S1. Settings, locations and references of the outcrop examples.