Ground Penetrating Radar Stratigraphy and Dynamics of Megaflood Gravel Dunes

Paul A Carling[1], Charles S. Bristow[2], Alexey S. Litvinov[3]

[1]Geography & Environment, University of Southampton, Highfield, Southampton, SO17 1BJ, UK.

[2]Department of Earth & Planetary Sciences, Birkbeck College, University of London, Malet Street, London, WC1E 7HX, UK>

[3]Institute of Monitoring of Climatic and Ecological Systems, Siberian Branch of the Russian Academy of Sciences, 10/3, Academichesky Ave., Tomsk, 634055, Russian Federation.

P.A.Carling@soton.ac.uk

Contact: P.A.Carling@soton.ac.uk

Running Header: GPR facies of gravel dunes

Abstract

Ground-penetrating radar was used to elucidate the stratigraphy of late Pleistocene gravel dunes in the Altai Mountains of southern Siberia that formed when a lake emptied due to ice-dam failure. Survey-lines across dunes had a resolution of decimetres, with depth penetration of 20m. The reflections identify bounding surfaces and radar facies. Two classes of unconformities are identified: (i) an erosional unconformity at the dunes base; (ii) steeply-inclined unconformities that truncate underlying inclined reflections and are down-lapped by overlying inclined reflections within the dunes. Unconformities define six radar facies (RF): RF 1, basal sub-horizontal discordant reflections; RF 2, poorly-defined discordant reflections; RF 3, planar inclined reflections; RF 4, sigmoidal inclined reflections; RF 5, trough fills; RF 6, low-angle inclined reflections. The basal unconformity represents the flood-cut surface, across which the dunes migrated. The inclined unconformities may be interpreted in two ways: (a) Erosional surfaces induced by unsteady-flow within one flood, or (b) Erosional surfaces developed by a series of floods reactivating dunes left stranded by previous floods. Evidence favours the latter model, which is consistent with the occurrence of several dune-forming events within the basin. The broader implications of the study are considered with respect to investigations of megaflood bedforms worldwide.

Several fields of giant gravel dunes have been reported in the Kuray and Chuja Basins of the Altai mountains in southern Siberia (Fig. 1), as well as along the valleys of the Chuja and Katun rivers which lead northward to the Ob river and thence to the Siberian plains (Carling, 1996a; Carling *et al.*, 2002). The conjoined basins were

occupied by a temporal series of ice-dammed lakes during the late Pleistocene (Marine Isotope Stages 2 and 3; Carling *et al.*, 2011) and dunes in the basins were formed when alluvial fan gravels, inundated by the lake, were mobilized as the lake emptied rapidly following the repeated failures of the ice-impoundment (Carling, 1996b). Other gravel dunes were formed in the sediments deposited by the megafloods along the valleys below the former ice-dam (Carling, 1996a; Huggenberger *et al.*, 1998), but these latter dunes are not considered here.

The most extensive and impressive field of dunes lies within the Kuray Basin (Fig. 2). The Kuray dunefield consists of flow-transverse, two-dimensional dunes that are subparallel and developed on a wide plain without lateral confinement that is the northward sloping surface of an older alluvial gravel fan formed by a precursor to the Tetyo river (Figs. 2,3 & 4). Crest lines and toe lines are relatively straight or slightly sinuous with saddles typically every 60 to 120m falling in height to around 1-2m below the intervening lobes (Carling, 1996a). Lee side spurs which might induce strongly three-dimensional curvature to the stratigraphy are absent. The highest (16m) and longest wavelength (192m) dune occur in the west (Fig. 3) and dunes reduce in size to the east; the morphology of each dune usually is asymmetric with steeper lee sides facing down gradient to the east and more gentle stoss slopes facing to the west (Carling, 1996a). Stratigraphic descriptions obtained from hand-dug pits show that the dunes consist of alternating cobble and pebble cross-sets inclined below the angle of repose (Carling, 1996a). These morphological and sedimentological data were used to estimate the local palaeoflow conditions over one dune at a single point-in-time (Carling, 1996b). Recently, a dynamic model of a single draining of the lake, from its maximum high stand until dry (Borhorquez *et al.*, 2015), demonstrated that the lake bed in the region of the Kuray dunefield was subject to sediment-entraining flows that generally traversed the area from west to east, although a recirculating-flow cell developed above the dunefield during part of the draining event.

Figure 1 near here

Figure 2 near here

Figure 3 near here

Not-with-standing the knowledge gained to date, outstanding questions remain. Generally, the stratigraphy of gravel dunes is not well-known (Carling, 1999; Carling *et al.*, 2005) and additional information would contribute to knowledge of gravel

dune dynamics more broadly (Best, 2005). Specifically, an hydraulic interpretation of the detailed stratification within the Kuray dunes could be used to improve lake draining models (Bohorquez *et al.*, 2015) if the nature of the flow field above the dune can be inferred from the stratigraphy. In addition, there is an open question as to whether the Kuray dunes, as seen today, were formed during one lake draining event or are the result of sediment mobilization by several distinct events as the lake filled and drained repeatedly.

Flood bar sediments in the valley downstream of the ice-dam provide dated evidence (C^{14} and infrared stimulated luminescence) for several distinct flood units between 23,350 a ± 400a BP and 20,050a ± 80 a BP. Using these dated horizons and the associated interpretation of the stratigraphic successions, Carling *et al.* (2002) and Carling *et al.*, (2009) argue for at least three major flood events draining the basins during and towards the end of the Sartan glaciation (Last Glacial Maximum). In addition, using cosmogenic isotope dates for the bar-top sediments and ice-rafted boulders, Reuther *et al.* (2006) have argued that the final draining in the Kuray Basin occurred c. 15,900 ± 600a BP. Considering these dates together, there is evidence that the lake in the Kuray Basin drained on four occasions.

Thus, two primary issues arise. Firstly, the basic cross-set dune stratigraphy described in isolated pits (Carling, 1996a) needs further detailed exposition over larger areas. Secondly, there is the question, noted above, as to whether the dunes represent draining by one event or by several events. In the single event scenario, the dunes (as seen today) would have formed on the lake bed by a single draining of the lake to be left as relicts stranded in the landscape. In the multiple event scenario, the relict dunes would be inundated by the lake refilling, as the ice-impoundment closed again. Dunes would then lie unmodified beneath the lake. Subsequent drainings of the lake would remobilize the relict dunes such that they would migrate a short distance in the basin before, once again, being left stranded to give the modern topography.

The above issues can be addressed using ground-penetrating radar (GPR) to define the radar stratigraphy, which in turn can be interpreted with respect to the geometry of stratification induced by dune migration from which the dynamics of the dunes and the associated flows can be inferred.

GROUND PENETRATING RADAR DATA COLLECTION, PROCESSING AND INTERPRETATION

GPR facies of gravel dunes

In 2010 and 2012 GPR data were collected along profiles across the dunes proceeding from East to West in the upstream direction using a PulseEKKO Pro GPR system (Sensors & Software Inc, Ontario, Canada) with 100 MHz antennas. The antennas were arranged in the parallel broadside configuration spaced 1m apart and in 2010 were moved with a step size of 0.25m along tape measures laid on the ground. The time window used for the surveys in 2010 was 260ns but this was increased in 2012 to 400ns because 260 ns was barely sufficient to image the base of 16m high dunes. At the same time the step length was increased to 0.5m and number of stacks was reduced from 64 to 32 to speed up data collection along the 665m long line. A grid of seven profile lines c. 180m long and spaced at 10m intervals was established over the highest dune (Dune 1) in 2010 using compass, measuring tape and fixed ground-pegs. In 2012 a single line c. 180m was established over Dune 1 and extended through four further dunes to the east (Fig. 5). A box-section survey (30m x 30m) also was conducted over the crest of Dune 1 to define the spanwise radar structure as well as the structure in the direction of flow (Fig. 6). All profile lines were surveyed to define the topography using a Geodimeter total station. Approximate georeferencing of profile terminations was accomplished using a Garmin hand-held global positioning system.

Figure 4 near here

Data processing used EKKO42 and EKKO_view2 (Sensors and Software, Mississauga, Ontario, Canada) and includes dewow, AGC gain (max 200), trace to trace average of 2, and down the trace average of 2, followed by migration. Topographic correction was applied using a velocity of 0.135m ns^{-1} derived from curve fitting calibration with diffraction hyperbolae as well as a common mid-point survey collected along the crest of dune 4 (Fig. 5). The latter procedure produced corrected radar sections with reflection events in their true subsurface positions. The migration restored depths and dipping reflections (Fig. 6). Without migration, the cross-sets appeared planar but after migration some were seen to be asymptotic. The box survey shows that the profile orientation was very close to true dip on the W-E profiles and close to strike on the N-S profiles.

Figure 5 near here

The data quality in 2010 was not as good as that obtained in 2012 when data included little or no noise. Nevertheless, the GPR in both years successfully imaged sets of cross-stratification and internal bounding surfaces within the dunes, but in 2010 the erosion surface at the base of the dunes was poorly defined beneath the dune crest because of the inadequate depth penetration in that year. In 2012 the basal contact was well-defined in all cases. Checks of dip and strike orientation were

conducted using the 30x30m box on the crest of the first dune and a 100m GPR profile along the crest of the fourth dune (not illustrated).

GPR Interpretation

The reflections on GPR profiles in sediments are attributed to changes in the complex dielectric properties of layers within the sediments due to changes in sediment fabric including the grain size, grain shape and compaction of sediments (Guillemoteau, 2012), as well as associated changes in porosity and water content (Bristow et al. 2000, Neil and Roberts, 2001, Van Dam, 2001, Neal, 2004). The reflections on the GPR profiles of the giant gravel dunes at Kuray are interpreted to be produced by changes in sediment fabric associated with primary sedimentary structures, specifically sets of cross-stratification produced during dune migration. Cross-cutting relationships where radar reflections are truncated and superposition of reflections provide a relative chronology, younger above older, and truncation of reflections or down-lap indicate stratigraphic gaps due to erosion or non-deposition (e.g. Bristow et al. 2005). Packages of similar radar reflections are termed radar facies (Gawthorpe *et al.*, 1993), and distinct packages of cross-strata can be identified, each defined by the unconformities of the bounding surfaces and the basal unconformity.

RESULTS

Dune 1 is the largest in the dune field (Fig. 3, 4B & 5). As both the 2010 and the 2012 surveys covered parts of this dune, the basic results for both years are presented together in the section '*2010 & 2012 surveys of Dune 1*' below. The section '*2012 surveys of Dunes 2, 3, 4 & 5*' then follows, which is a basic appreciation of the Dunes 2 through 5, which were surveyed in 2012. Finally, the section '*Interpretation of the surveys of all dunes*' is dedicated to the detailed interpretation of the dunefield radar strata and consequent dune dynamics that is consistent with the evidence obtained from all surveyed dunes. Although Figure 6 shows both migrated and unmigrated data, only migrated data are shown in Figures 7 to 12.

2010 & 2012 surveys of Dune 1

Dune 1 rises to 16m high (*H*) with a downflow length (*L*) of 192m (Figs. 7 and 8). The initial stratigraphic interpretation presented below is consistent with the development primarily of foreset bedding due to dune migration in flowing water and this interpretation is amplified in section '*Interpretation of the 2010 & 2012 surveys of all dunes*'.

GPR facies of gravel dunes

Five of the seven surveyed flow-parallel radar transects obtained in 2010 are presented in Fig. 7. Note the ridge is asymmetric with a longer, less-steep stoss slope as opposed to a shorter, steeper lee slope. The radar transect obtained in 2012 (Fig. 8) better defines the erosion surface at the base of the dune than was the case for 2010 and images the alluvial fan strata beneath the erosion surface which have discontinuous sub-horizontal discordant reflections(Radar facies 1). Poorly-defined discordant reflections (Radar facies 2) occur within the upstream end of the dune (Fig 7 line 3; Fig. 8). Steeply-inclined GPR reflections (Radar facies 3) are interpreted to be produced by sets of cross-strata that dip downstream at relatively steep angles between 20° and 30° (Fig. 7 line 4; Fig. 8) with a uniform angle and thickness beneath the stoss side; indicative of steady dune progression. Although several bounding surfaces are evident, in particular, a well-defined bounding surface occurs trending from beneath the crest toward the dune toe in the lee side deposits (Fig. 7; e.g. Lines 1 & 4; Fig. 8). Downflow of this erosional surface cross-sets are coherent but are less-well defined, with more evidence of sigmoidal forms that down-lap and pinch-out (Facies 4), in contrast to the well-defined stoss side examples which tend to be regular sets of planar cross-strata. Whereas typically four to five bounding surfaces can be identified in Fig. 7, in Fig. 8 five well-defined bounding surfaces are evident. In Fig. 8 a set of low-angle reflections (Radar facies 5) occur within the lee side trough. In Fig. 7 low-angle upstream inclined reflections (Radar facies 6) occur on lines 1 and 2.

Figure 6 near here

Figure 7 near here

Figure 8 near here

2012 surveys of Dunes 2, 3, 4 & 5

Dunes 2 to 5 are smaller than Dune 1 with lower heights (*H*) in all cases although the length (*L*) of Dune 2 is similar to that of Dune 1. In all examples (Fig. 9 – 12), Radar facies 1 underlies the duneforms. Radar facies 2 appears clearly in Fig. 9, 10 and 12. The cross-sets (Radar facies 3) in Dune 2 are low-angle (Fig. 9) which may reflect poor development of this dune in the lee of the very steep Dune 1. Radar facies 4 is represented in Figs. 9, 10 and 12 and dominates in Figs. 11 and 12, although it is not as clearly defined in these smaller dunes in contrast to the larger Dune 1. Trough fills are well-developed in all examples (Figs. 9 – 12), whereas Radar facies 6 is absent. Dunes 3 to 5 are better developed, being of greater aspect ratio (*H*/*L*) than Dune 2 with steep cross-set angles (Facies 3 & 4). In all cases the overall interpretation of the radar packages is similar to that presented for Dune 1 with four or five packages

of cross-sets separated by three to four bounding surfaces. However, the relief of dunes 3, 4 and 5 is somewhat subdued owing to trough infilling by fine sediments (coloured green in Figs. 9 to 12).

Figure 9 near here

Figure 10 near here

Figure 11 near here

Figure 12 near here

Interpretation of the surveys of all dunes

The dieletric properties of the reflectors primarily are controlled by lithology and water content. Here the surveys were conducted above the water table, which is several metres below the erosional surface at the top of Facies 1. So reflections largely represent variations in grain-size and porosity; layers of gravel with a fines-infill are poorly sorted and tend to have reduced porosity and tend to be damp in contrast to well-sorted coarse-grained layers and so present strong reflections. In the case of the dune cross-sets these grain-size variations are primary (Carling, 1996a), determined during foreset deposition and are not due to later infiltration or elutriation of fines.

Within all the surveyed dunes two types of unconformity are identified. At the base of the dunes is a sharp but irregular reflection interpreted as a basal erosion surface. The basal deposit (Radar facies 1) below the unconformity consists of stacked, more-or-less sub-horizontal, slightly discordant reflections that are typical of low-energy water-lain deposits (e.g. Leclerc & Hickin, 1997; Roberts *et al.*, 1997). The erosion surface at the base of each of the dunes represents a continuous sub-horizontal and somewhat irregular erosive surface that extends beneath all the dunes, commonly dipping in the downsteam direction. Notably, the basal erosion surface is often higher beneath the dunes than in the troughs (Dunes 1, 2, 4 & 5) which suggests that dune migration was slow, allowing the troughs to actively scour whilst the gravel masses forming the dunes protected the underlying sediments below from scour.

Above the basal erosion surface, the radar stratigraphy within the dunes is dominated by inclined reflections which downlap onto the erosion surface in a downstream direction. The distinct reflections represent cross-sets consisting of alternating layers of open-work gravel interbedded with matrix-filled gravel (Facies 3 & 4); the latter infilled by sand and silt, as observed in open pits (Carling, 1996a;

Huggenberger *et al.*, 1998). Within the inclined reflections there are lower angle inclined reflections that truncate underlying reflections and are in turn downlapped by overlying reflections. These unconformities are interpreted as bounding surfaces, enclosing packages of cross-sets, where there has been a break in deposition before a reshaping of the dunes. The exceptions to this interpretation are the furthest up-stream inclined bounding surfaces in Figs. 7 through 10, which are interpreted as a change from RF 2 to RF 3 (RF 2 to RF 4 in Fig. 12) without reactivation as is explained in the Discussion.

Dunes migrate by building lee side cross-sets from bedload sediment supplied from the stoss sides and, at the end of a flow event, the regularly dipping cross-sets tend to be cut by an erosion surface. The erosion surface is due to scouring and degradation of the dune crest and lee side by falling stage flow as sediment supply to the lee side is reduced (Leeder, 1982, p 89). Remobilization of a dune tends to preserve this erosion surface as a 'reactivation surface' (Collinson, 1970) as regularly dipping cross-sets once again prograde downflow of the erosion surface. Reactivation surfaces thus may either indicate: (i) intermittency in sediment entrainment and deposition over the dunes during a single draining event or; (ii) evidence for a number of repeated events. The packages of cross-sets indicate a vertical stacking of cross-strata within dunes growing in height (Dunes, 1, 2, 4), or possibly losing height on occasion (Dunes 2 and 5), in contrast to just a simple downstream translation of migrating bedforms of uniform height. Each radar package almost certainly represents a change in hydraulic regime (Bristow, 1995) that mediated sediment supply (Allen, 1982; p. 498) and that led to morphological adjustments. The presence of packages consisting of regular tabular cross-beds of approximately equal spacings, especially developed on the stoss sides, implies steady progression of the dunes in steady flow with an intermittent supply of self-similar quantities of gravel to the brinkpoints. Such a supply is readily provided by gravel sheets (Carling, 1999), of similar mass to each foreset, progressing up the stoss sides to the crests and running out down the lee sides (Carling, 1996a). In contrast, beneath the preserved crestal regions, and in the lees, cross-strata is less regular and include beds that may thicken or thin towards the base. Significantly, weakly sigmoidal beds are present generally with a pronounced basal curve representing weakly developed toe-sets. In contrast, the top of sigmoidal bedding often is truncated due to local erosion which observation implies pulses of flow occurred over the crests that were not freighted with sediment. The presence of sigmoidal bedding, the thickening and thinning of cross-strata and the concentration of reactivation surfaces might imply unsteady flow over the crest of the dunes towards the end of a single flood in contrast to the steadier flow that occurred at an earlier time when the stoss side beds were deposited as simple, tabular, cross-sets.

On lines 1 and 2 (Fig. 7) there is an area with a different reflection pattern with a planar high amplitude reflection inclined slightly upstream between 100 m and 120 m on line 1 and 106 m and 124 m on Line 2 (RF 6 blue block, Figure 7). Beneath these low-angle inclined reflections the signal is attenuated which suggests the presence of fine grained sediments such as clay minerals and this mass could be interpreted as a layer of fine-grained infused gravel, possibly a low stage drape deeply-infiltrated into the upstream side of Dune 1, but it is too high up the back of the dune and does not continue across to lines 3, 4 and 5. An alternative explanation for the low-angle high amplitude reflection is the presence of a giant (20m) intraformational clast composed of reworked, fine-grained sediments, possibly a frozen clast of ice rafted debris. The slight upstream dip would be consistent with imbrication of a giant clast and it is reasonable that a large flood would erode blocks of consolidated or frozen sediment (Carling, 2013). The underlying parallel reflection is interpreted as a multiple.

The profiles for Dunes 1 and 2 have no fine fill in the stoss side troughs and there are only thin drapes in the lee sides (Figs. 8 & 9). In contrast, there is up to 5m of fine fill draped into the troughs in the profiles of Dunes 3 to 5. The layering evident in the radar surveys accords with auger and pit surveys in the troughs, which demonstrated loess-like fill above clay, silt or sand deposited on the basal gravel (Carling, 1996a). The lower level of the fill may represent late-flood ponding of fines in the dune troughs. The higher portion of the fill may consist of fines flushed out from the steep dune margins by rainfall onto the dune flanks before vegetation became established, plus a possible a subsequent aeolian component trapped in the troughs (Carling, 1996a).

There are two possible interpretations for the stratigraphy of the dunes.

Single-flood interpretation: The ensemble of cross-bed packages could be a response to a rapidly changing flood hydrograph where the bedform wavelength and height try to adjust to the fluctuating discharge. Taking Dune 1 as an example: the rising limb of the hydrograph is marked by the basal scour and initial dune construction with coherent cross-sets only developing after a less-well defined set of strata are deposited (see box for Line 3; Fig. 7) to form an incipient bed undulation which provides a steep lee side against which well-defined cross-sets can develop. During peak discharge the giant dune is formed and this migrates downstream generating sets of well-defined cross-strata (see box for Lines 4 and 5; Fig. 7). As the discharge declines, and flow depth decreases, the duneform increasingly interacts with the shallowing flow resulting in increased flow variability (Allen, 1982; p 498). Foreset

accretion is constrained by the existing giant dune morphology within variable flow, so dune progression is less regular and bounding surfaces and smaller set packages are preserved beneath the crest and in the lee side of the giant bedforms. Thus, the concentration of reactivation surfaces beneath the crestal region is unlikely to be coincidental, but reflects increasing flow variability as discharge declined and dunes increasingly obstructed the flow depth.

Although the above explanation may suffice, the physical scale of the cross-strata within each reactivated package of cross-sets seems too large to be explained by the passage of low-amplitude bedload sheets (see box for Line 1; Fig. 7) as discharge fluctuated and declined. In addition, the single-flood interpretation does not account for the multiple floods known to have originated in the Kuray Basin.

Multiple flood interpretation: An alternative explanation to bounding surfaces reflecting flow variability within a single-flood, is that the dunes were repeatedly reactivated by a series of separate floods (e.g. Allen, 1982; p. 498; Baker & Bunker, 1985, their fig. 6) such that one (or more) large cross-set packages are produced by each flood. These packages scale with the height of the lee side of the parent bedform that was present at the beginning of each flood. Note that if the area of each package of cross-sets between reactivation surfaces is taken as indicative of the power or duration of the associated flow event, then in all profiles (Figs. 7 & 8) at least one large event is indicated in all cases, as all profiles exhibit one large package, with other smaller packages occurring before or after the main event. Using this interpretation, the four distinct reactivation surfaces in Fig. 8 indicate five flood events whereas reactivation surfaces in Fig. 7 indicate at least four major events. However, there is some uncertainty due to reactivation in the crestal regions giving up to six packages that may represent, as noted above, discrete events or variation in flow power during one or more event.

Figure 13 near here

DISCUSSION

In summary six radar facies (RF) have been identified in these radar profiles together with two significant classes of unconformities – (i) a single near-horizontal basal unconformity and (ii) several steeply-inclined unconformities within the dunes. The facies are: RF 1, basal sub-horizontal discontinuous discordant reflections; RF 2, poorly-defined discordant reflections possibly indicative of incipient bedform development; RF 3, planar inclined reflections dipping steeply (20-30°) downstream; RF 4, sigmoidal inclined reflections dipping steeply downstream; RF 5, trough-shaped fills; RF 6, low-angle upstream-inclined reflections with high attenuation.

The distinction between the sub-horizontal strata (RF 1) below the basal unconformity and the steeply-inclined strata above (RF 3 & 4) indicates that the duneforms are not the final phase of continued aggradation of climbing dunes developed over a long time, as no cross-strata is recorded below the unconformity. Rather the dunes represent one or several dune-forming events that prograded across an erosion surface and excavated material from the troughs to feed dunes downstream. Note that the extent of the excavation is recorded by the height of the basal erosion surface on the stoss-side of the dunes, e.g. dune 2 indicates 5m of incision in the lee of dune 1 as it grew. In addition, all profiles demonstrate relatively broad troughs between dunes; which usually are scoured into the underlying alluvial fan deposits. Thus, the wavelengths of the dune spacings are greater than the dune lengths such that individual dunes are isolated from their immediate neighbours by the scoured troughs. Such spaced bedforms are either indicative of (i) sediment-starvation or (ii) formative flow events that are not of sufficient duration to allow a near-equilibrium dunefield to develop. The latter interpretation is preferred due to arguments presented below.

Taken together the disposition of cross-set configurations is remarkably informative of dune dynamics. In the lower stoss-sides, complexes of convex-up humps (Facies 2) of poorly-defined bedding are evident (Figs. 7, 8, 9, 10 & 12) which are interpreted as small-scale incipient bedforms (Lunt et al., 2004; Okazaki et al., 2015) of insufficient height to allow steep cross-sets to form. This initial style of bedding is replaced by well-defined cross-sets downflow which usually are tabular (RF 3) or occasionally weakly sigmoidal (RF 4; e.g. Fig. 12) indicative of steady dune growth and progression. The bounding surface between RF 2 and the downstream facies represents a change in the style of continued deposition, rather than reactivation, and is defined as an activation surface (AS) as shown in Figure 13 marking the onset of slipface development. Preservation of RF2 and the activation surface at the upstream end of the dunes shows that they have not migrated far downstream, less than one dune length. Although, cross-sets are always below the angle of repose (c. 32^0), the steepest examples (RF 3) usually lie beneath the upper stoss-side. Well-developed cross-strata packages indicate a plentiful sediment supply in a powerful sediment-charged flow. Lower angle sets, pinching-out of cross-beds, and a denser configuration of reactivation surfaces tend to occur beneath the crests and within the lee sides (RF 4) which might be indicative, as noted above, of unsteady flow over the lee sides of dunes during the waning of a single flood that formed the complete dune body. Carling (1996a) noted that the dunes could not have advanced by avalanching of a steep lee side. Rather, low-angle cross-beds are deposited by the passage of gravel sheets that pass over the dune crests and, with momentum, run down the lee sides. The sigmoidal-curvature at the base of otherwise tabular cross-

beds can be due to run-out of gravel sheets travelling down the lee side to just beyond the dune toe (Bristow, 1995; Carling, 1996a). Nevertheless, the angle of cross-sets is always steeper than the angle of the lee sides (Figs. 7 – 12). The presence of relatively steep dunes (L/H < 12) and negligible toe-sets indicates that, despite some planing-down of the lee slope, there was little dune height reduction towards the end of the final flow event. In addition, although the trough-fills (RF 5) exhibit reflections, these do not reflect gravel layers within the fine fill (Figs. 8 - 12). The absence of gravel in trough-fill was confirmed by Carling (1996a) by augering, and was interpreted to indicate that there had been little post-flood modification of the dune profiles. Thus, the stabilization of the steep coarse-gravel dune forms is consistent with rapid flood recession. However, there is further evidence (presented below) that the duneforms are compound and formed by a succession of floods modifying pre-existing dunes on the lake bed.

Although the presence of bounding surfaces within the dunes can be related to unsteady flow during a single draining event, as noted above, recent modelling indicates that this is unlikely. A model of the lake drainage (Bohorquez *et al.*, 2015) for a single draw-down (from maximum water level until completely dry) produces a simple draw-down hydrograph without vagaries in the hydrograph shape. However, at times the model shows that recirculating flow occurred above the Kuray dunefield. Nonetheless, during these periods of variable flow direction, the bed shear stress was too low to move gravel (Bohorquez *et al.*, 2015). Rather, bed-sediment movement occurred towards the end of the drainage for only a few tens of minutes at best. In this respect, it is unlikely that a large dune could grow and exhibit multiple reactivation surfaces in such a short time. It is notable that all the dunes exhibit a broadly similar number of reactivation surfaces of similar distribution throughout the individual bedforms. Further, reactivation surfaces beneath the stoss slopes indicate that the dunes were reactivated during periods of strong, prolonged flow and reactivation <u>was not</u> limited to the waning stage of a flood when the final lee side was developing beneath markedly unsteady flows. Nor, given the short duration of competent flow events is it likely that the final flood effaced all earlier evidence of dunes to leave one set of dunes. Thus, on balance it appears that an initial lake draining was responsible for forming a dunefield and subsequent floods reactivated the dunes on two to three occasions to leave the morphology as it appears today. Consequently, it is concluded that all the dunes surveyed were all activated in similar fashion, not by one flood but by several.

CONCLUSIONS

GPR facies of gravel dunes

The stratigraphy of large gravel dunes formed by late Pleistocene catastrophic emptying of an ice-dammed lake is well-defined by ground-penetrating radar surveys. The dunes are dominated by inclined radar reflections, the basic nature of which is in accordance with the physical cross-sets described in excavated pits. However, a wealth of additional information is revealed in the radar stratigraphy. The dune cross-sets are deposited unconformably above an erosion surface rather than conformably above packages of cross-sets laid down by earlier dunes traversing the area. Thus the bedforms represent one or more short-term flow events across a scoured surface in contrast to prolonged aggrading events.

Steep tabular sets dominate below the stoss sides with an increase in sigmoidal sets, pinch-outs and reactivation surfaces beneath the crests and in the lee sides. In terms of the dynamics of formation, there are two interpretations of the recorded stratigraphy:

(1) The presence of distinct cross-strata packages separated by bounding reactivation surfaces can be interpreted as due to a change from steady flow during the main period of dune construction to more varied flow at the end of a singular flow event that was responsible for constructing the dunefield;

(2) An initial flood constructed a dunefield, but this was then modified by subsequent floods. Subsequent floods reactivated the dunes on several occasions, building several packages of cross-sets in the lee sides of the earlier dunes.

On balance, the second model is preferred as it is consistent with other evidence (obtained further down the route of the flood below the lake) for at least three to four floods sourced from the Kuray Basin (Carling et al., 2002; 2009). The radar stratigraphy, notably the presence of reactivation surfaces and well-defined facies packages, indicates that there were most likely four rapid draw-downs of the lake level.

Such detailed interpretations of bedform stratigraphy might be applied to other megaflood locations, such as the well-studied example of Lake Missoula where a long-running debate has occurred with respect to the magnitude and timing of multiple floods (e.g. Benito & O'Connor, 2003; Clague et al., 2003). Baker & Bunker (1985), for example, noted distinct reactivation surfaces separating cross-set packages within a Missoula flood bar. Thus, the broader implications of the present study lie in the possibility to use the detailed stratigraphy and sedimentology of megaflood bedforms at locations around the world to better inform, and in some

cases constrain, the magnitude, timing and hydrograph behaviour of lake drainage models.

ACKNOWLEDGEMENTS

We are grateful for loan of a PulseEKKO Pro GPR system in respect of Projects 931 and 970 from the NERC Geophysical Equipment Facility. Southampton undergraduate students James Rand and Laura Hayes assisted in field data acquisition. The comments of reviewers Victor Baker and an anonymous review were greatly appreciated in preparing the final manuscript.

REFERENCES

BAKER, V.R., BUNKER, R.C., 1985, Cataclysmic Late Pleistocene flooding from Glacial Lake Missoula: A review. Quaternary Science Reviews, 4; 1-41.

BENITO, G., O'CONNOR, J.E., 2003, Number and size of last-glacial Missoula floods in the Columbia River valley between the Pasco Basin, Washington, and Portland, Oregon. *GSA Bulletin*, 115: 624–638.

BEST, J., 2005, The fluid dynamics of river dunes: A review and some future research directions. Journal of Geophysical Research, 110: F04S02, doi:10.1029/2004JF000218.

BOHORQUEZ, P., CARLING, P.A., HERGET, J., 2015, Dynamic simulation of catastrophic Late Pleistocene glacial-lake drainage, Altai Mountains, central Asia. International Geology Review, in press.

BRISTOW, C., 1995, Facies analysis in the Lower greensand using ground-penetrating radar. Journal of the Geological Society, London, 152: 591-598.

BRISTOW, C.S., CHROSTON, P.N., and BAILEY, S.D., 2000, The structure and development of foredunes on a locally prograding coast: Insights from ground penetrating radar surveys, Norfolk, England, Sedimentology 47: 923-944.

BRISTOW, C. S., LANCASTER, N. & DULLER, G. A. T., 2005, Combining ground penetrating radar surveys and optical dating to determine dune migration in Namibia. Journal of the Geological Society, 162: Part 2, 315-322.

CARLING, P.A. (1996a) Morphology, sedimentology and palaeohydraulic-interpretation of large gravel dunes: Altai Mountains, Siberia. Sedimentology, 43: 647-664.

CARLING, P.A. (1996b) A preliminary palaeohydraulic model applied to Late-Quaternary gravel dunes: Altai Mountains, Siberia. In: Global Continental Changes: The Context of Palaeohydrology. J. Branson, K.J. Gregory and A. Brown (Editors) Geol. Soc. London Spec. Publ. No. 115, 165-179.

CARLING, P.A. (1999) Subaqueous gravel dunes, J. Sedimentary Research, 69: 534-545.

CARLING, P.A., 2013, Freshwater megaflood sedimentation: what can we learn about generic processes? Earth-Science Reviews, 125: 87-113.

CARLING, P.A., KNAAPEN, M., BORODAVKO, P., HERGET, J., KOPTEV, I., HUGGENBERGER, P. & PARNACHEV, S. 2011, Palaeoshorelines of glacial Lake Kuray–Chuja, south-central Siberia: form, sediments and process. J. Geological Society, London, 354, 111-128.

CARLING, P.A. KIRKBRIDE, A.D., PARNACHEV, S., BORODAVKO, P.S. and BERGER, G.W. (2002) Late Quaternary catastrophic flooding in the Altai Mountains of south-central Siberia: a synoptic overview and introduction to flood deposit sedimentology. Spec. Publ. Ass. Sediment, 32, 17-35.

CARLING, P.A., MARTINI, I.P., HERGET, J., BORODAVKO, P., PARNACHEV, S., 2009, Megaflood sedimentary valley fill: Altai Mountains, Siberia. pp. 243-264 In: Megaflooding on Earth and Mars, D.M. Burr, P.A. Carling & V. R. Baker (eds.), Cambridge University Press, Cambridge, 319pp.

CARLING, P.A., RADECKI-PAWLIK, A., WILLIAMS, J.J., RUMBLE, B., MESHKOVA, L., BELL, P., BREAKSPEAR, R., 2005, The morphodynamics and internal structure of intertidal fine-gravel dunes: Hills Flats, Severn Estuary, UK. Sedimentary Geology, 183: 159-179.

CLAGUE, J.J., BARENDREGT, R., ENKIN, R.J., FOIT, F.F., 2003, Paleomagnetic and tephra evidence for tens of Missoula floods in southern Washington. Geology, 31, 247–250.

COLLINSON, J.D., 1970, Bedforms of the Tana River, Norway. Geografiska Annaler, 52A, 31-56.

GAWTHORPE, R.L., COLLIER, R.E.L., ALEXANDER, J., LEEDER, M., BRIDGE, J.S., 1993, Ground penetrating radar: application to sandbody geometry and heterogeneity studies. In: North, C. P. & Prosser, D. J. (eds), Characterization of Fluvial and Aeolian Reservoirs, Geological Society Special Publication, 73: 421-432.

GUILLEMOTEAU, J., BANO, M., DUJARDIN, J-R., 2012, Influence of grainsize, shape and compaction on georadar waves: example of an Aeolian sand. Geophysical Journal International, 190: 1455-1463.

HUGGENBERGER, P., CARLING, P.A., SCOTNEY, T., KIRKBRIDE, A. and PARNACHEV, S.V., 1998, GPR as a tool to elucidate the depositional processes of giant gravel dunes produced by late Pleistocene superflooding, Altai, Siberia. Proceeding of the 7th International Conference on Ground Penetrating Radar, 1: 279-28, May 27-30 1998, University of Kansas, Lawrence, Kansas, USA, Austin, Texas. Radar Systems and Remote Sensing Laboratory, Univ. Kansas. ISBN: 0-936352-16-7.

LECLERC, R.F., HICKIN, E.J., 1997, The internal structure of scrolled floodplain deposits based on ground-penetrating radar, North Thompson River, British Columbia. Geomorphology, 21: 17-25.

LEEDER, M.R., 1982, Sedimentology: Process and Product, Kluwer, 360pp.

LUNT, I.A., BRIDGE, J.S., TYE, R.S., 2004, A quantitative, three-dimensional depositional model of gravelly braided rivers. Sedimentology, 51: 377-414.

NEAL, A., 2004, Ground-penetrating radar and its use in sedimentology: principles, problems and progress. Earth Science Reviews 66: 261-330.

OKAZAKI, H., KWAK, Y., TAMURA, T., 2015, Depositional and erosional architecture of gravelly braid bar formed by a flood in the Abe River, central Japan, inferred from a three-dimensional ground-penetrating radar analysis. Sedimentary Geology, 324: 32-46.

REUTHER, A.U., HERGET, J., IVY-OCHS, S., BORODAVKO, P., KUBIK, P.W., HEINE, K., 2006, Constraining the timing of the most recent cataclysmic flood event from ice-dammed lakes in the Russian Altai Mountains, Siberia, using cosmogenic in situ ^{10}Be, Geology, 34: 913–916.

ROBERTS, M.C., BRAVARD, J.-P., & JOL, H.M., 1997, Radar signatures and structures of an avulsed channel: Rhone River, Aoste, France. Journal of Quaternary Science, 12: 35-42.

VAN DAM, R.L., 2001, Causes of ground penetrating radar reflections in sediments. Unpublished PhD Thesis, Vrije Universiteit, Amsterdam, Netherlands,110pp.

FIGURE CAPTIONS

Figure 1: Location of the Kuray Basin (K) and the Chuja Basin (C) in the Altai mountains of southern Siberia.

Figure 2: Detail of the location of the Kuray dunefield in the Kuray Basin. The main dune field was formed by reworking of gravel fan deposits of the Tetyo River.

Figure 3: Soviet (undated) vertical aerial photograph of the Kuray dunefield (Lat: 50.169124^0 N; Long: 87.921584^0 E). The largest dunes occur in the northwest corner, to the east of the valley of the Tetyo River. The boxed area shows the approximate area of GPR investigations (see Fig. 5). Palaeoflow left to right. Scale ~ 1:30,000; field of view c. 3km.

Figure 4: (A) View along the crest of dune 5 towards the south. (B) View to the southwest across the dunefield. Black arrow points to the crest of the largest dune 1 with the pine forest of the Tetyo River immediately behind (see Fig. 3). Photographs were taken from near the confluence of the Chuja and Tetyo Rivers (see Fig. 2).

Figure 5: Location of GPR and topographic survey lines in 2010 and 2012 at the western extremity of the Kuray dunefield. Palaeoflow approximately left to right (west to east). Image source: Google Earth, 2015.

Figure 6: Radar stratigraphy for the box-section (see Fig. 5 for location) (A) without and (B) with migration correction. The labelled cubes show relative GPR profile location and orientation. The sub-horizontal reflections on the north-south profiles 2 and 4 are consistent with strike sections perpendicular to flow and depositional dip confirming that the east-west dip sections are parallel to flow and depositional dip.

Fig. 7: Interpretation of 2010 five flow-parallel radar profiles (10m apart) for Dune 1. As well as well-defined cross-sets the radar stratigraphy shows reactivation surfaces (red curves) and the unconformable interface (purple curve) with the underlying gravel deposits. Palaeoflow right to left. Inset representative examples of radar facies: Radar facies 1, basal sub-horizontal discontinuous discordant reflections; Radar facies 2, poorly defined discordant reflections within the stoss toes; Radar facies 3, Planar inclined reflections dipping steeply (20-30°) downstream; Radar facies 4, sigmoid inclined reflections dipping downstream; Radar facies 6, low-angle inclined reflections with high attenuation. Radar facies 5 is not present in this image. See text for details.

Figure 8: (A) 2012 Radar reflections for Dune 1; (B) Interpretation of the 2012 radar facies for Dune 1. As well as well-defined cross-sets (e.g. blue curves) the radar stratigraphy shows bounding surfaces (red curves) and the unconformable interface

(purple curve) with the underlying gravel deposits (yellow tone). Palaeoflow right to left. See text for details.

Figure 9: (A) 2012 Radar reflections for Dune 2; (B) Interpretation of the 2012 radar facies for Dune 2. The well-defined cross-sets (e.g. blue curves) are of low-angle. The radar stratigraphy shows bounding surfaces (red curves) and the unconformable interface (purple curve) with the underlying gravel deposits. Palaeoflow right to left. See text for details.

Figure 10: (A) 2012 Radar reflections for Dune 3; (B) Interpretation of the 2012 radar facies for Dune 3. The well-defined cross-sets (e.g. blue curves) are relatively steep in contrast to Dune 2. The radar stratigraphy shows boundimg surfaces (red curves) and the unconformable interface (purple curve) with the underlying gravel deposits. Later fill in the lee-side trough (green) has subdued the modern topography. Palaeoflow right to left. See text for details.

Figure 11: (A) 2012 Radar reflections for Dune 4; (B) Interpretation of the 2012 radar facies for Dune 4. The well-defined cross-sets (e.g. blue curves) are relatively steep. The radar stratigraphy shows bounding surfaces (red curves) and the unconformable interface (purple curve) with the underlying gravel deposits. Later fill in the dune troughs (green) has subdued the modern topography. Palaeoflow right to left. See text for details.

Figure 12: (A) 2012 Radar reflections for Dune 5; (B) Interpretation of the 2012 radar facies for Dune 5. The well-defined cross-sets (e.g. blue curves) are of variable steepness. The radar stratigraphy shows bounding surfaces (red curves) and the unconformable interface (purple curve) with the underlying gravel deposits. Later fill in the stoss-side trough (green) has subdued the modern topography. The air wave is due to the tree visible at the end of the GPR profile in Fig. 5. Palaeoflow right to left. See text for details.

Figure 13: Cartoon interpretation of four packages (1 – 4) of cross-strata separated by bounding reactivation surfaces above a basal unconformity, down-lap and truncation of reflections are represented by small arrows. Package 1 consists of irregular stratification, in the upflow position, deposited as an incipient bedform against which steep cross-sets were deposited once steady flow and the bedform were established. Activation surface (AS) represents the bounding surface between Radar facies 2 and Radar facies 3 and activation of the dune slipface. RS1 is a reactivation surface associated with a subsequent event that represents cutting of the lee-side of package 2. Reactivation surface 2 (RS2) represents cutting of the lee-side of package 3. Packages 2, 3 and 4 represents renewed deposition of cross-sets

in the lee of the bedform stranded after three repeated floods to give a history due to four floods in total.