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Late-Holocene changes in character and behaviour of land-terminating glaciers on James Ross Island, Antarctica

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ABSTRACT. Virtually no information is available on the response of land-terminating Antarctic Peninsula glaciers to climate change on a centennial timescale. This paper analyses the topography, geomorphology and sedimentology of prominent moraines on James Ross Island, Antarctica, to determine geometric changes and to interpret glacier behaviour. The moraines are very likely due to a late-Holocene phase of advance and featured (1) shearing and thrusting within the snout, (2) shearing and deformation of basal sediment, (3) more supraglacial debris than at present and (4) short distances of sediment transport. Retreat of $\sim$100 m and thinning of 15–20 m has produced a loss of 0.1 km^3 of ice. The pattern of surface lowering is asymmetric. These geometrical changes are suggested most simply to be due to a net negative mass balance caused by a drier climate. Comparisons of the moraines with the current glaciological surface structure of the glaciers permits speculation of a transition from a polythermal to a cold-based thermal regime. Small land-terminating glaciers in the northern Antarctic Peninsula region could be cooling despite a warming climate.

INTRODUCTION, RATIONALE AND AIMS

The mean annual air temperature in the Antarctic Peninsula region has increased by $\sim$3.7°C over the last century (Houghton and others, 2001; Morris and Vaughan, 2003; Vaughan and others, 2003; Turner and others, 2005). This warming rate is approximately six times greater than the global mean (Vaughan and others, 2003), making the Antarctic Peninsula among the most rapidly changing environments and landscapes on Earth.

There is a spatial and temporal complexity in climate change across the Antarctic Peninsula (e.g. Turner and others, 2005), and this coupled with a range of types of glaciers makes for a highly varied glaciological response (e.g. Davies and others, 2012). Previous studies of the effects of contemporary climate change on Antarctic Peninsula glaciology have focused on ice shelves and tidewater glaciers (e.g. Rott and others, 1996; Vaughan and Doake, 1996; Skvarca and others, 1999; Scambos and others, 2003; Skvarca and De Angelis, 2003; Cook and others, 2005; Glasser and others, 2009; Cook and Vaughan, 2010) and on outlet glacier tributaries to the ice shelves (e.g. De Angelis and Skvarca, 2003; Scambos and others, 2004; Shepherd and Wingham, 2007; Smith and others, 2007; Glasser and others, 2011a). While these types of glacier are the most obviously dynamic parts of the glaciological system, they are ocean-calving and thus have a nonlinear and multifactorial response to climate change (e.g. Joughin and others, 2010) due to responses to changes in ocean currents and sea-surface temperatures (Benn and others, 2007).

In contrast, land-terminating glaciers are relatively neglected in studies of Antarctic Peninsula glaciology. This lack of attention is surprising because (1) the mass balance of land-terminating (non-calving) glaciers is solely dependent on climate and (2) land-terminating glaciers are likely to become more prevalent on the Antarctic Peninsula in the near future as tidewater glaciers retreat onto land (e.g. Glasser and others, 2011a). The first glacier inventory of land-terminating glaciers on the Antarctic Peninsula was by Rabassa and others (1982). This inventory included snowfields, and many of the small glaciers mentioned therein have since disappeared. Skvarca and others (1995) evaluated changes in glacier extent for the period 1975–93, and a subsequent remote-sensing study by Rau and others (2004) showed that of five land-terminating glaciers on the Ulu Peninsula, only one displayed indications of glacier recession between 1975 and 1988. However, in 2002 seventeen of twenty-one land-terminating glaciers were found to be retreating (Rau and others, 2004). Morphological and mass-balance changes for Antarctic Peninsula glaciers are published only from Glaciar Bahía del Diablo on the northern part of Vega Island (Skvarca and others, 2004), but we note that new mass-balance studies have been initiated on Davies Dome and 'Whisky Glacier' on James Ross Island (Fig. 1) in 2006 and 2009, respectively (Nývlt and others, 2010; Engel and others, 2012).

Previous studies of the glaciological effects of climate change on the Antarctic Peninsula have focused on 'micro-' or 'macro-scale' changes. Microscale studies are either at individual glaciers (e.g. Fox and Cziferszky, 2008; Wendt and others, 2010) or regionally over, at most, a few decades (e.g. Shepherd and others, 2003; Scambos and others, 2004; Thomas and others, 2004; Rignot and others, 2008; Pritchard and others, 2009). Macro-scale changes are either continent-wide (e.g. Shepherd and Wingham, 2007; Pritchard and others, 2009), or with consideration of glaciation over millennia. Published ice-core data obtained from the Antarctic Peninsula currently do not extend beyond 1200 a ^{14}C (Mosley-Thompson and Thompson, 2003). This space–time bias, where few mesoscale measurements exist, reflects reliance either on remote-sensing methods for time periods spanning the last three decades (Quincey and Luckman, 2009), or on geological evidence that is dominantly

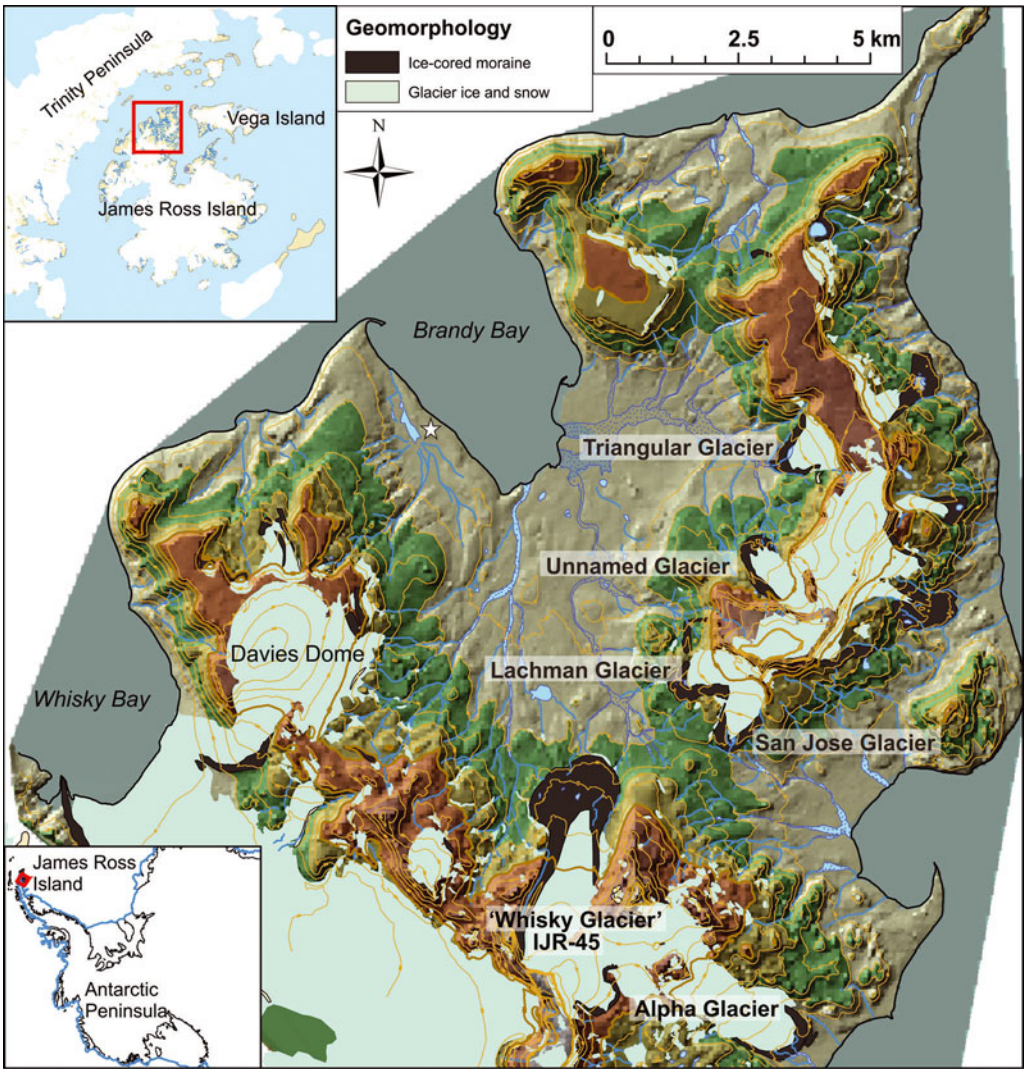

Fig. 1. Location and topography of the Ulu Peninsula, James Ross Island, Antarctic Peninsula. The six glaciers detailed in this study are named in bold; Davies Dome is also mentioned in the text. The white star denotes the location of samples of marine organic material that have been taken to infer a Holocene glacier advance that produced a moraine that runs along the western side of Brandy Bay.

marine-based and thus with a temporal resolution, at best, of several centuries (Sugden and others, 2006; Bentley and others, 2009; Davies and others, 2012).

The overall aim of this paper is to provide the first mesoscale or centennial-scale measurements of glacier character and behaviour on the Antarctic Peninsula. This aim is achieved by (1) holistic geological descriptions of the topography, geomorphology and sedimentology of prominent moraines immediately adjacent to a range of contemporary land-terminating glaciers, (2) interpretation of the character and behaviour of those glaciers while they were at this relatively advanced position and (3) quantification of the geometric changes to these glaciers during the late Holocene.

STUDY SITE AND GLACIAL HISTORY

James Ross Island is located off the northeastern coast of the mainland Antarctic Peninsula. The 600 km^2 Ulu Peninsula on the northern part of James Ross Island (Fig. 1) is one of the largest ice-free areas in Antarctica. Geologically it comprises extensive Cretaceous mud(stone) and sand(stone) beds (e.g. Francis and others, 2006) that are overlain by multiple horizons of subaerial massive basalt and subaqueous–subglacial hyaloclastite breccias.

Holocene glacier fluctuations on the Ulu Peninsula have been inferred from proxy records: datable fossiliferous and organic material within (1) lake sediments (Zale and Karlén, 1989; Ingólfsson and others, 1992; Björck and others, 1996), (2) glacial and marine sediments in coastal sections (Hjort and others, 1997), (3) coastal surfaces identified as moraines (Strelin and others, 2006) and (4) highly weathered glacigenic surfaces (Rabassa, 1983). These studies provide bracketed (uncalibrated radiocarbon) ages of phases of glacier advance and retreat but cannot be used to consider glacier character or behaviour. Overall, the prevailing consensus is that deglaciation from the Last Glacial Maximum had probably occurred (finished) by ~7500 a ^{14}C and that there were readvances of glaciers that culminated at ~4700 a ^{14}C (Zale and Karlén, 1989; Björck and others, 1996; Hjort and others, 1997).

Following the first glacier inventory of James Ross Island glaciers by Rabassa and others (1982), field-based studies of glaciers on the Ulu Peninsula have focused on spatially restricted descriptions of glacial landforms and sediments. Chinn and Dillon (1987) documented the superficial character of a $\sim$1 km^2 debris-covered part of 'Whisky Glacier', Fukui and others (2008) documented the internal and superficial composition of a single $\sim$1 km^2 rock glacier and Lundquist and others (1995) documented a series of moraines and tills within a 10 km^2 area between two rock glaciers. Skvarca and others (1995) identified greatly reduced glacier extent for the period 1975–93, and the subsequent remote-sensing study by Rau and others (2004) showed that of five land-terminating glaciers on the Ulu Peninsula, only one displayed indications of glacier recession between 1975 and 1988.

METHODS

In this paper, we present topographical, geomorphological and sedimentological data obtained in January–March 2011 on the exceptionally well-defined moraines at six land-terminating glaciers on the Ulu Peninsula (Fig. 1): Triangular Glacier, 'Unnamed Glacier', Lachman Glacier, San José Glacier, Alpha Glacier, and 'Whisky Glacier' or IJR-45 (Fig. 1). These glacier names are given on the Czech Geological Survey (CGS) (2009) map. 'Whisky Glacier' (cf. Chinn and Dillon, 1987) is herein referred to as IJR-45 as in the inventory by Rabassa and others (1982), so as not to be confused with the tidewater Whisky Glacier located within Whisky Bay, as named by the British Antarctic Survey (BAS) (2010) (Fig. 1). 'Unnamed Glacier' is a name that we invented for this study. Rabassa (1983) recorded the moraines at two of the glaciers, which Rabassa and others (1982) inventoried as IJR-45 and IJR-47, as the youngest of six distinguishable geostratigraphical units on the Ulu Peninsula.

A differential GPS (dGPS) Leica GPS500 was used in real-time kinematic mode for topographical surveys and to precisely determine the location and elevation of glacier margins, glacier snout positions and moraine crests. The dGPS base receiver first averaged a (static) position for 16 hours and this same base point was used as a reference for all subsequent dGPS 'rover' receiver measurements. Therefore, we consider that our dGPS measurements had an error of 0.1 m horizontally and 0.2 m vertically.

Glaciological structures were mapped from aerial photographs and these interpretations were checked in the field. Glaciological field observations included identification and characterization of present-day ice margin morphology, ice-surface structures such as stratification, foliation and crevasses, and supraglacial debris position and character. Morphological data collected in the field included visual assessment of the position and situation of the moraines relative to the surrounding topography and geomorphology, i.e. the landscape association of the moraines, as well as surface character, characteristic planform and longitudinal profile.

Sedimentological analysis of the moraines provided information on the genesis and depositional history, style and environment (cf. Jones and others, 1999; Hubbard and Glasser, 2005). Field sketches, photography and a handheld GPS were used to log and map the location of exposures and the approximate spatial extent of homogeneous surfaces. All sedimentary observations and measurements included detail of clast lithology, grain-size distribution, matrix content, grading, sorting and texture, thereby complying with standard procedures (Gale and Hoare, 1991; Evans and Benn, 2004). Textural (non-genetic) classifications for poorly sorted sediments were made according to Hambrey and Glasser (2003) as based on Moncrieff (1989). Section exposures of (buried) ice and sediment were sketched and logged as vertical profiles. Ice facies were described according to the revised and unifying scheme proposed by Hubbard and others (2009). Use of proglacial topography, geomorphology and sedimentology to reconstruct glacier dynamic and thermal regimes is now fully established and documented by Hambrey and Glasser (2012).

Our field-based dGPS surveys of modern glacier extent and geometry were supplemented by visual interpretation of BAS aerial photographs, and by digital interrogation of a photogrammetrically derived digital elevation model (DEM) produced by GEODIS (CGS, 2009). The BAS/RN/06 aerial photographs were taken in 2006 with a Leica RC30 at 1 : 25 000 scale. The DEM was stereoplotted from the aerial photographs using digital photogrammetry. The exterior orientations of the aerial photographs were calculated by aerotriangulation and coordinates of ground control points measured using a dual-frequency dGPS. This method produced root-mean-square error (rmse) in the DEM of 0.7 m horizontally and 0.8 m vertically (CGS, 2009).

Late-Holocene glacier extent and ice surface topography were determined for each glacier using standard ArcGIS routines following the method of Glasser and others (2011b; Fig. 2). This method involved interpolating a surface between the highest points on the major moraine crest (Fig. 3), thus avoiding sections of the moraine ridge that have downwasted due to degradation of the ice core. Where a moraine ridge had a double crest, each of similar height, only the innermost ridge was considered in our analysis. Thus a minimum estimate of the height of the moraines was obtained and from this the palaeo-ice surface elevation. Figure 2 illustrates how these palaeo-ice surfaces were analysed to calculate morphometric changes during the late Holocene.

Our reconstructed glacier surfaces were not sensitive to the interpolation routine used: linear, inverse distance weighting, spline, nearest neighbour, or ordinary kriging. A gridcell-by-gridcell analysis produced a mean difference of surface elevation between each of these five interpolation routines of 4 m. This vertical error should be considered to be compounded by the error in the DEM (0.8 m) but nevertheless was considerably less than the mean surface elevation changes computed between the late-Holocene maximum and the present-day position for each glacier, which were 9–23 m (Table 1). The lateral moraines at Triangular, 'Unnamed', Lachman and IJR-45 glaciers extended nearly to the headwall. However, the lateral moraines at San José and Alpha Glaciers extended only part-way up the sides of the glaciers. These latter two glaciers had a more clearly defined accumulation area, and thus for these glaciers our surface interpolations are only robust for the ablation area.

MEASUREMENTS OF MORAINE GEOMORPHOLOGY AND SEDIMENTOLOGY

Prominent moraine ridges immediately in front of Triangular, Unnamed, Lachman, San José, Alpha and IJR-45 glaciers have crests that are typically 20–40 m higher in elevation than the modern glacier surface (Figs 3 and 4). All of the moraines are ice-cored. Multiple crests are evident at most of the glaciers, particularly in the region of the glacier snout, i.e. at lowermost elevations. All of the moraine crests have an undulating or hummocky topography and sides sloping at 30–40° (Figs 3 and 4). The moraines have a complex microtopography with a hummocky ridge crest, shallow surface depressions, some of which contain ponded water, and superficial mass movements of the overlying sediment. This topography and morphology clearly contrasts with the surrounding landscape surfaces (Figs 3 and 4).

Sediment thickness on the moraines appears to be relatively uniform and, where its thickness is demonstrated by an ice scar, apparently varies from ~1 to 2 m. Indeed, the moraines at each of the glaciers in this study have similar sedimentological characteristics. Thus for brevity and clarity we focus our sedimentological descriptions on the moraines at the conjoined and diffluent San José and Lachman Glaciers (Fig. 5) as a representative example. San José and Lachman Glaciers are small cirque glaciers, with steep headwalls of basalt and hyaloclastite cliffs (Figs 4 and 5). The mesa above the glaciers holds a small plateau ice dome from which these glaciers are currently detached (Fig. 4). The contemporary glacier surfaces have no crevasses and few supraglacial streams, with sparse supraglacial debris. Small proglacial lakes and ice-marginal streams have incised through the moraines. Both glaciers currently have pronounced stratification (cf. Hambrey and Lawson, 2000). Stratification is low-angle in the mid-sections of the glaciers, near the assumed equilibrium-line altitude (cf. transient snowline), and becomes progressively more arcuate with distance down-glacier (Fig. 5). Lachman Glacier has an abandoned meltwater channel that is incised through the terminal moraine. It also has moraines at the head of an ice-free valley that trends northwards (Fig. 4).

In detail, the terminal moraine at Lachman Glacier comprises three principal ridges (Fig. 6a). The ridges have steep outward-facing slopes and shallower inward-facing slopes. The innermost ridge is 3 m high, 10 m wide and immediately adjacent to the contemporary glacier snout. This innermost moraine has numerous faults creating a 'staircase pattern' on the moraine surface, with moraine 'blocks' nearer to the contemporary glacier margin ramped up against and 'over' blocks farther from the ice margin. The shallow crests of the top of these ramps are 1–2 m high and together form concentric crescent-shaped arcs. The middle moraine ridge is 10 m wide and 8 m high and has visible ice scars and scattered basalt cobbles. The outermost moraine is the most hummocky and, like the innermost moraine, has numerous faults producing a staircase surface of superimposed smaller ridges up to 2 m high, large ice scars, perched ponds and numerous large perched basalt boulders (Fig. 6a–c).

Sediments on the surface of the moraine range in texture from sandy boulder-gravel to diamicton with numerous angular to well-rounded fine gravel to boulder-sized clasts. Some of these boulders are heavily striated. Frost-shattering of basalt boulders becomes increasingly pronounced towards the outer fringes of the moraine (Fig. 6c). Qualitatively, we observed that there is a clear spatial trend in clast lithology, size and shape; lateral moraines near to headwalls are predominantly basaltic and are clast-supported with proportionally larger (>1.5 m) and very angular clasts, whereas terminal moraines near to the contemporary glacier snout are matrix-supported and contain much smaller (<0.5 m) and more rounded clasts and <50% silts and sands. Sparse supraglacial material, mainly angular basalt boulders, is found on the glacier surface and margins.

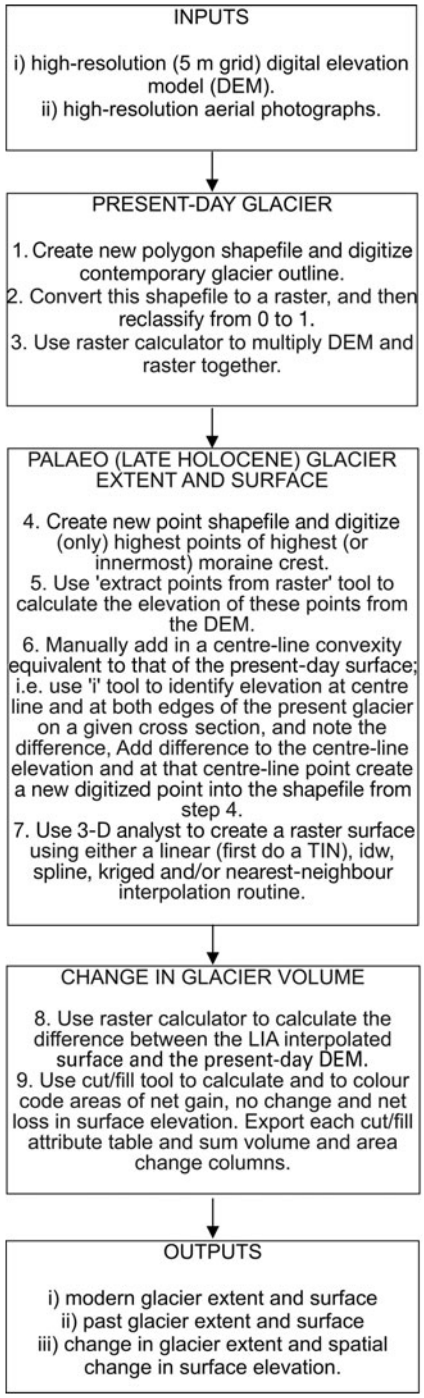

Fig. 2. Method used in this paper for determining previous glacier extent and glacier surface elevation from moraine crests, and the morphometric analysis of changes at each glacier. This method is an adaptation of that developed by Glasser and others (2011b). Different interpolation routines can indicate the sensitivity of reconstructed surfaces to the spatial distribution of input points.

Arcuate hummocky moraines at Lachman Glacier North (Figs 5 and 6a) are up to 30 m high and have steep (40°) distal slopes with abundant loose scree, ice scars up to 2.5 m high and abundant small perched ponds. The moraine complex is covered by a 0.1–0.5 m thick debris layer of silt and sand with a veneer of basalt cobbles. Internally, these moraines comprise basalt boulders and cobbles within a matrix that is derived from the underlying Cretaceous bedrock. Some of the smaller moraine ridges have been cut by contemporary small meltwater streams to reveal

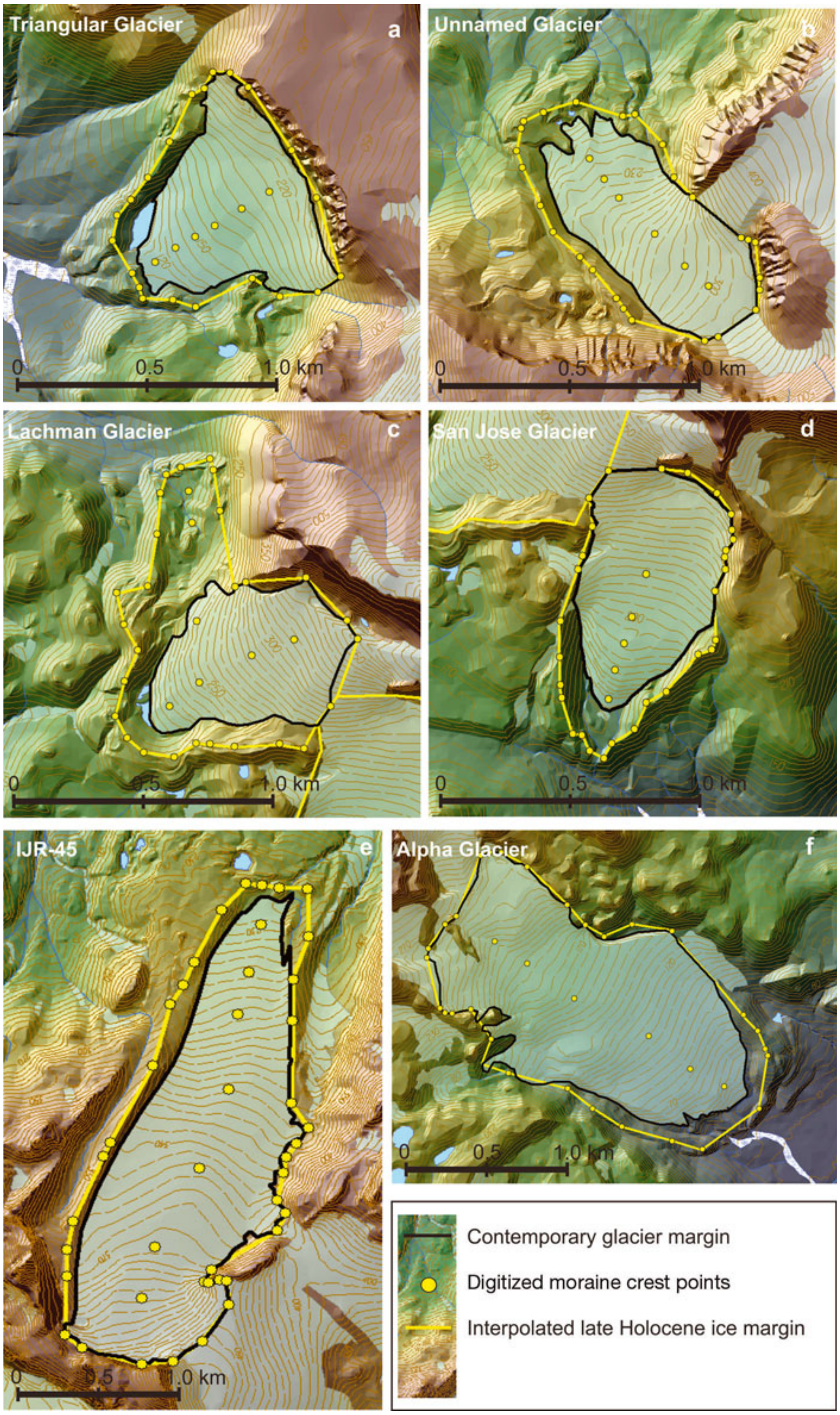

Fig. 3. Data used in the reconstruction of (minimum) late-Holocene glacier extent (line) and points used to provide elevation constraints of the associated ice surface (circles). The contemporary glacier outline and surface contours are displayed for reference and for comparison.

gently folded and occasionally faulted bedrock sand and silty-sand beds (Fig. 7).

Sections 1 and 2 (Fig. 8) at the San José Glacier moraine are excellent examples of back scar with exposed (buried) ice. The ice layers have sharp contacts and can be followed around the corners in three dimensions (Fig. 8). The prominent facies are clean glacier ice, and ice with debris content described as dispersed, laminated and stratified (cf. Hubbard and Sharp, 1995; Hubbard and others, 2009; Table 2). Firstly, there is white, coarse, bubble-rich ice with 0.2 m thick layers separated by both zones of diamicton debris: fine clay to small boulder-sized clasts, dispersed within the ice, and zones of ice with muddy laminations (Fig. 8). Above this is densely laminated ice with debris-rich layers <0.05 m thick incorporating mud and fine gravel clasts (Fig. 8). This densely laminated ice is overlain by clear, bubble-free massive blue ice with dispersed (occasional) cobbles and small boulders (Fig. 8; Table 2). There are several layers 0.15–0.2 m thick, which dip consistently at 40° down-glacier. The stratified ice situated above this

Fig. 4. San José Glacier, which faces southwards, with Lachman Crags behind. The prominent 30 m high ice-cored moraine ridge contrasts topographically, geomorphologically and sedimentologically with the surrounding terrain. The moraine surface comprises subaerial basalt and hyaloclastite clasts from Lachman Crags. The lowermost parts of moraine near the glacier snout also incorporate poorly lithified Cretaceous bedrock.

contains large amounts of debris with interstitial ice, and these layers pinch out downwards (Fig. 8). Layers in this region are contorted and folded, with sheared laminations. The debris-rich ice is overlain by finely laminated debris-rich ice, and then by clear white bubble-rich ice, which pinches out. The whole ice scar is unconformably overlain by 0.7 m thickness of unfrozen, finely bedded, poorly sorted, clast-rich silt and sand, with numerous basalt and sandstone cobbles. These clast-rich beds are overlain by a veneer of cobbles (Fig. 8). The stoss face of the moraine dips at 40°, which matches the angle of dip in the layered ice.

QUANTIFICATION OF GLACIER EXTENT, SURFACE AND VOLUME CHANGES

Land-terminating glaciers on the Ulu Peninsula have decreased in extent and in surface elevation since the deposition of the ice-cored terminal moraines. The decrease in size is a function of both glacier snout recession and ice-margin surface lowering to within lateral moraine ridges (Fig. 3). The magnitude of retreat of the glacier snout position is remarkably similar between glaciers and varies from 75 m at Triangular Glacier to 130 m at San José Glacier (Table 1). There is no significant statistical correlation

Table 1. Summary of late-Holocene properties and changes of the land-terminating glaciers on the Ulu Peninsula. Surface lowering is the mean value for the glacier ablation area. Volume change is the mean of five interpolation routines

Glacier name	Snout elevation	Altitude range	Aspect	Max elevation of moraine crests	Snout recession	Surface lowering	Area change	Volume change
	m a.s.l.	m		m a.s.l.	m	m	km^2 (%)	km^3
Triangular	115	135	SW	230	75	15	0.25 (33)	0.013
Unnamed	195	115	NW	280	100	9	0.18 (23)	0.009
Lachman	186	164	SW	290	120	23	0.44 (46)	0.016
San José	134	156	S	240	130	20	0.16 (22)	0.012
Alpha	24	226	SE	160	115	15	0.25 (12)	0.022
IJR-45	219	231	N	365	90	14	0.41 (15)	0.032

Table 2. Ice facies observed in section 2, San José Glacier moraine, and interpretation of their formation. Facies names and interpretations follow Hubbard and Sharp (1995), Waller and others (2000) and Hubbard and others (2009)

Ice facies	Description	Interpretation
Clean white bubble-rich ice	White ice with coarse to fine bubbles (crystals are difficult to view).	Supraglacial (superimposed) glacier ice formed from compaction of fallen snow (firnification).
Ice with dispersed debris	Dark blue ice with coarse to fine crystals and with occasional diamicton debris (devoid of internal layering).	Clean ice above the basal zone. Advanced compression and firnification.
Ice with laminated debris	Debris-rich laminae intercalated with clean (debris-free) ice. Debris includes clay to fine gravel fractions. Sharp contacts between laminations.	Attenuation of debris by creep when debris-rich facies comes into contact with clean ice.
Ice stratified with debris	Ice intercalated with debris-rich ice; limited interstital ice; frozen (diamicton) debris ranges from fine clay to small boulders.	Net basal adfreezing.

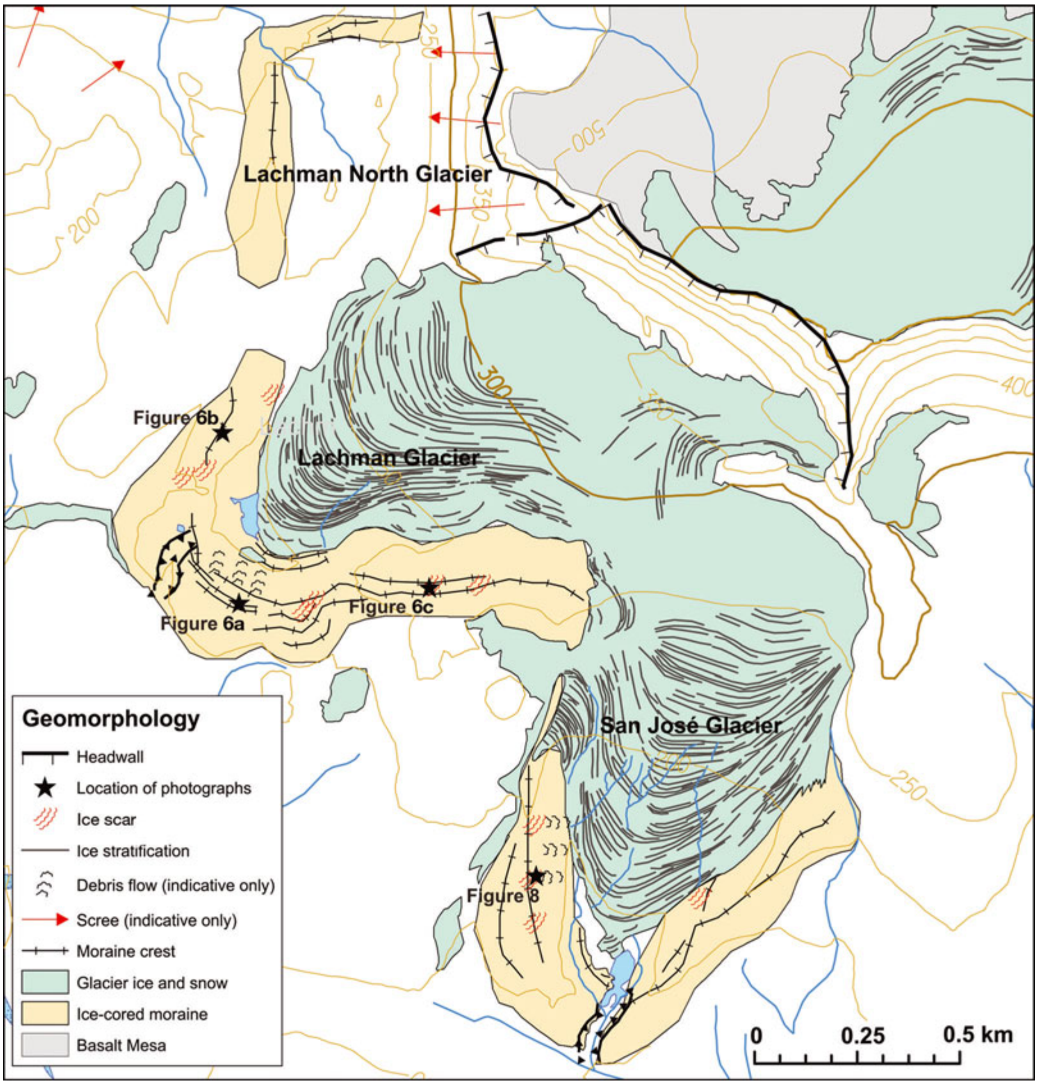

Fig. 5. Geomorphological map of San José and Lachman Glacier areas based on BAS/RN/06 1 : 25 000 scale aerial photographs and on field surveys.

between the snout elevation or the areal extent of a glacier, and the amount of recession or the amount of surface lowering that has occurred. However, generally the glaciers with the largest amounts of snout recession have also experienced the greatest reduction in surface elevations (Table 1).

Surface lowering at each glacier declines in magnitude with increasing altitude up to the maximum elevation of moraine crests (Table 1; Fig. 9). Above the maximum elevation of moraine crests we are less certain of our reconstructed late-Holocene ice surface. Figure 9 also illustrates that all modern glacier surface long profiles comprise a near-linear slope, whereas the reconstructed late-Holocene long profiles have a distinct convexity. All of the glaciers have retreated snout positions and sufficient ice-surface lowering in the vicinity of the snout that there is now an inverse slope and 'space' in front of the contemporary ice margin (Fig. 9). This space is currently occupied by small proglacial lakes and deltas that are fed from the melt of ice-cored moraine, but not glacier ice melt.

Spatial analysis of the pattern of surface lowering reveals asymmetry, which specifically comprises a distinct east–west gradient on each glacier (Fig. 10). There is greater surface lowering at any given altitude on the western part of a glacier than on the east (Fig. 10) and this difference can be up to 20 m, as for 'Unnamed' and Triangular glaciers (Fig. 10). Interestingly, except for Alpha Glacier this spatial pattern is entirely independent of glacier snout elevation and of glacier aspect. The pattern is least pronounced, but still present, on IJR-45 (Fig. 10).

Overall, four of the Ulu Peninsula land-terminating glaciers have lost 0.18–0.25 km^2 in area, and Lachman Glacier and IJR-45 have lost 0.4 km^2. These losses equate to ~25–30% of the late-Holocene surface area for Triangular, Unnamed and San José glaciers, but just 12–15% of the late-Holocene area for IJR-45 and Alpha Glacier. Lachman Glacier has lost 46% of its late-Holocene area, but this anomaly is due to abandonment of a north-trending valley (Figs 3 and 5). IJR-45 and Alpha Glacier are much larger in terms of area and elevation range than the other glaciers. Land-terminating glaciers on the Ulu Peninsula have each decreased in volume during the late Holocene by 0.01–0.03 km^3, to give a combined total ice loss of 0.1 km^3 (Table 1). There is a clear relationship between glacier and volume loss, and there appear to be two groups of glacier by this measure (Fig. 11): Unnamed, Triangular and Lachman glaciers, which all face westwards, have lost less volume for their relative area than San José, Alpha and IJR-45 glaciers.

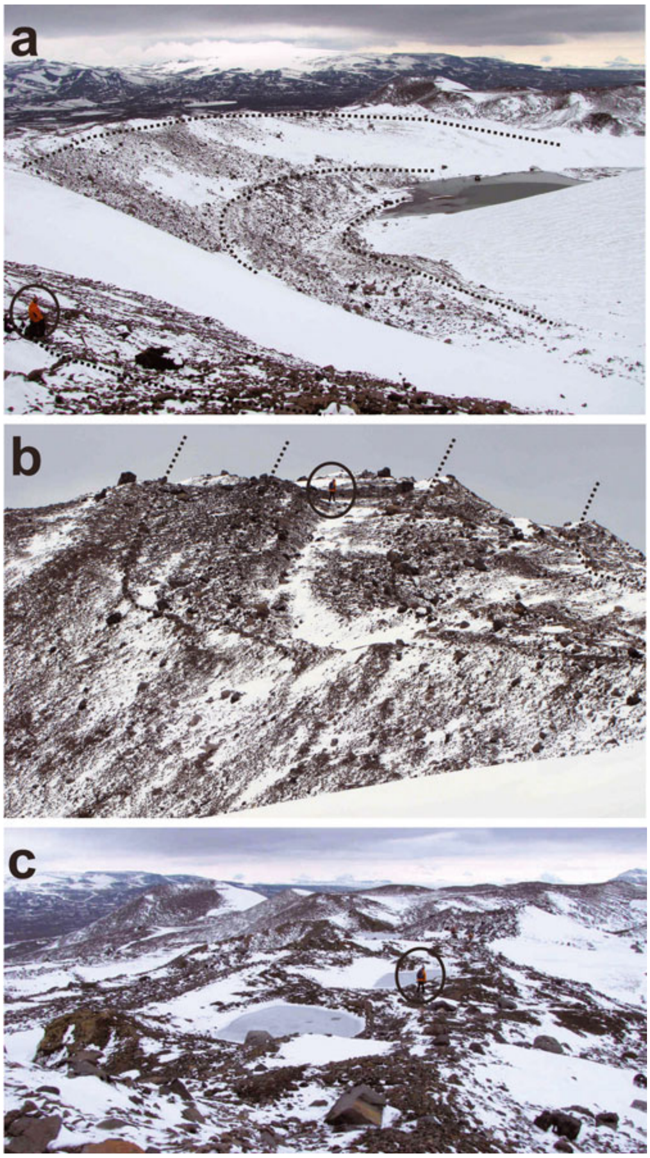

Fig. 6. Annotated photographs illustrating the character of Lachman Glacier moraine: (a) multiple major moraine crests as denoted by dashed lines; (b) superimposed ridges with up to 2 m relief as denoted by dashed lines; and (c) perched ponds on the outer moraine. Note person encircled for scale.

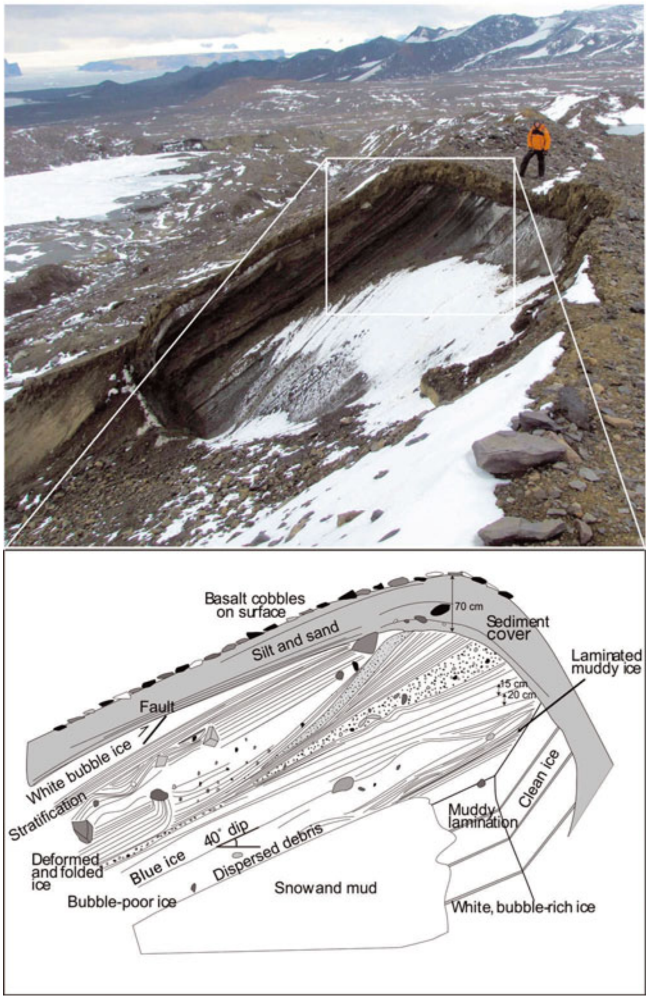

Fig. 8. Photograph and sedimentological log of buried glacier ice at San José Glacier moraine. Note stratification and debris bands in the ice.

Fig. 7. Section of moraine immediately in front of Unnamed Glacier snout, illustrating deformed and faulted silty-sand (Cretaceous) beds. Major horizons are marked with dashed white lines. A likely thrust is indicated by the black arrow.

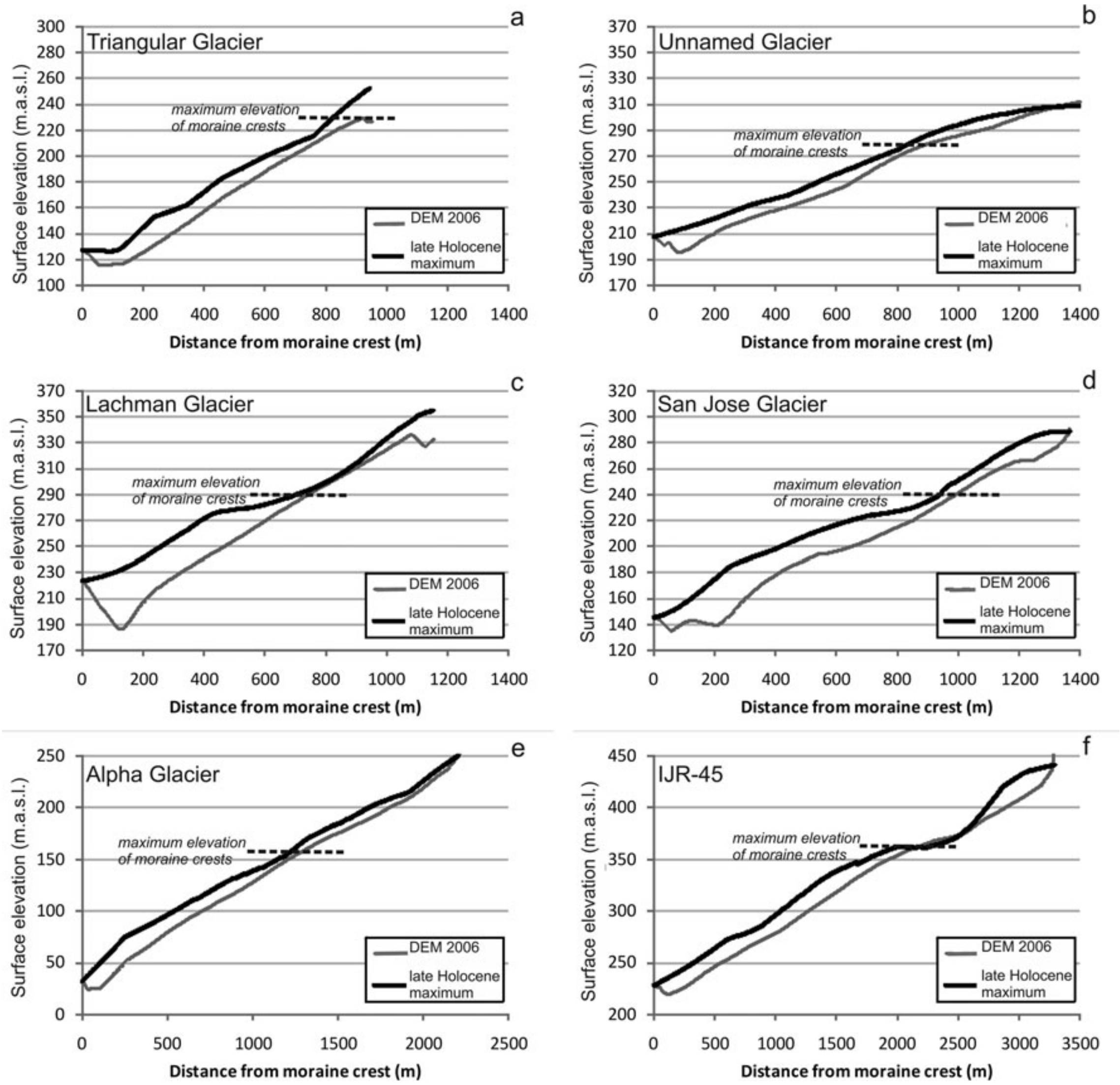

Fig. 9. Surface lowering along the centre line (long profile) of land-terminating glaciers on the Ulu Peninsula. Note the different *X* and *Y* scales used for Alpha Glacier and IJR-45.

INTERPRETATION AND DISCUSSION OF PAST GLACIER BEHAVIOUR

Lateral moraines are composed entirely of boulder-sized angular material near the headwalls. These boulders were thus sourced from local rockfalls and transported supraglacially (Table 3). Upon arrival at the glacier margins these supraglacial boulders effectively buried the glacier ice, thereby producing ice-cored moraine and debris cones. Burial by debris would have been aided by stagnation of the glacier ice either by lateral compression against a pre-existing moraine wall or at lower elevations due to arrival in a zone of net ablation (Benn and Evans, 2010). In contrast, terminal moraines are sedimentologically dominated by a silty-sand matrix and contain striated boulders. This relatively fine-grained sediment matrix must be derived predominantly from Cretaceous bedrock, but also from the matrix of (volcanic) hyaloclastite breccias. Overall this spatial pattern and character of superficial sediment on the lateral and terminal moraines implies very short sediment transport distances (cf. Hambrey and Ehrmann, 2004).

White, bubble-rich ice layers in the ice core have been formed through firnification from frozen snow (Table 2). Such 'superimposed ice' is typical of arctic and subpolar glaciers (Benn and Evans, 2010). We interpret the stratified ice, which has a fine crystal structure and low bubble content with sand, gravel and small boulders, as basal glacier ice formed by adfreezing near the glacier margin (Table 2; cf. Hubbard and Sharp, 1995; Knight and others, 2000; Waller and others, 2000; Cook and others, 2011). The ice facies became interbedded with clear blue basal ice by shearing at the ice margin. This shearing, which implies thrusting, can be seen clearly in section 2 (Fig. 8) producing stacked basal, englacial and proglacial ice (cf. Waller and others, 2000). Debris laminations were formed by the attenuation of debris and interstitial ice at the boundary with cleaner facies ice. Folding of ice laminations occurred through ice creep, particularly around larger obstacles.

Subglacial and englacial thrusting has thus entrained sediment, which subsequently melts out at the surface to bury the ice core and to produce the present-day sediment cover/veneer. Shearing and elevation of basal glacier ice has been rigorously argued to be common within small polythermal glaciers (Hambrey and others, 1999; Glasser and Hambrey, 2003; Hambrey and Glasser, 2012). These processes produce arcuate belts of aligned ridges and superimposed smaller moraine ridges. Orientation of thrusting can be observed in

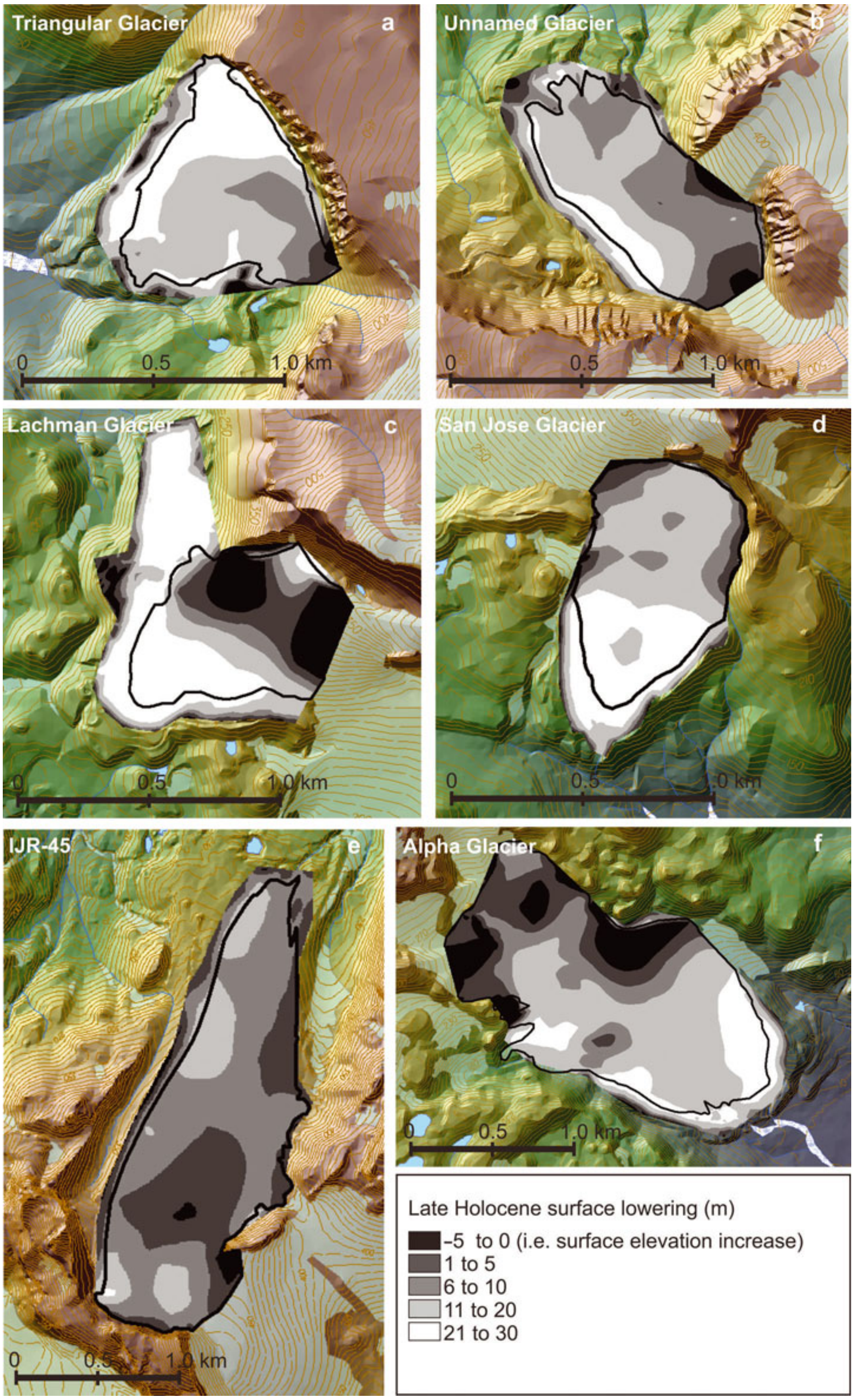

Fig. 10. Spatial variability in surface lowering during the late Holocene at six glaciers on the Ulu Peninsula.

the angle of dip and dip direction in the stratified ice; this closely controls moraine morphology and surface slopes. The thrusting controls not only the angle of the moraine slopes but also the moraine height, width and character. The sediment drape on the moraine mounds has melted out from the debris-laden ice layers. These moraines are particularly prominent because of the presence of this ice and only a thin sedimentary cover; after ice-core meltout there will be little geomorphological expression of them. This perhaps explains the lack of other prominent moraines on the Ulu Peninsula. Sediment redistribution by debris flows and mass movements during meltout (cf. Schomacker and Kjær, 2008) and permafrost reworking further limits the preservation potential of older ice-cored moraines (Ó Cofaigh and others, 2003; Evans, 2009). These processes of englacial sediment meltout producing sediment drapes over ice-cored moraines have been observed in similar glacial and permafrost settings such as Svalbard (Hambrey and others, 1997; Schomacker and Kjær, 2008) and Eugenie Glacier, northern Canada (Evans, 2009).

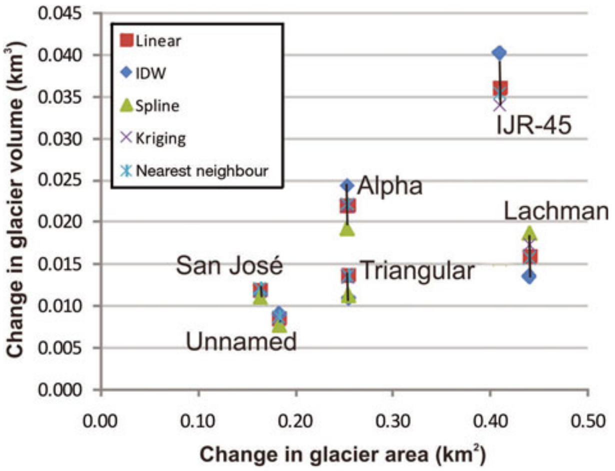

Fig. 11. Relationship between change in surface area and change in volume for six land-terminating glaciers on the Ulu Peninsula. Errors bars are defined by the different interpolation routines which together indicate the sensitivity of a reconstructed surface to the spatial distribution of input points.

We believe that the silty-sand beds exposed near the glacier snouts and exhibiting impressive folding (e.g. Fig. 7) are part of the Cretaceous bedrock of northern James Ross Island. The direction of this deformation, with axial planes parallel to the valley floor slope, and the recumbent style of deformation, together suggest that the sediments were saturated and loaded with a force either in front of the glacier due to pushing and/or beneath the glacier due to overriding by temperate glacier ice (cf. Hart and Boulton, 1991) (Table 3).

The prominent moraines at the land-terminating glaciers on James Ross Island are 'controlled moraines' (Evans, 2009), i.e. the moraine form is controlled by englacial structures. Specifically, moraine form is controlled by the stratification (Fig. 5), which is produced by marginal shear (Hambrey and Lawson, 2000) and by meltout of debris-rich (basal) ice (cf. Ó Cofaigh and others, 2003). We therefore suggest that these moraines record the advance of a polythermal glacier (Table 3) because the englacial and proglacial thrusting and (thick) deformed basal ice sequences are indicative of compressive flow near the glacier snout (e.g. Glasser and Hambrey, 2003; Evans, 2009; Hambrey and Glasser, 2012). It is interesting to note that despite James Ross Island having a similar climate to that of Svalbard, no evidence of surging glaciers (e.g. looped or contorted moraines; cf. Ó Cofaigh and others, 2003; Grant and others, 2009) was observed in this study.

INTERPRETATION AND DISCUSSION OF GLACIER EXTENT AND VOLUME CHANGES

Ice-marginal retreat is greatest at the glacier snout, but the broadly parallel and concentric outlines of former and present glacier extent indicate ice-marginal retreat laterally too (Fig. 3). The magnitude of surface lowering for a centre-line longitudinal profile increases with decreasing altitude and is thus likely to have been driven by air temperature (Table 3). Surface lowering is anomalously low at Unnamed Glacier because that glacier is fed by twin ice lobes from a small ice dome on top of Lachman Crags, whereas the other glaciers have become detached from their plateau accumulation areas.

We interpret the evolution in surface elevation (Fig. 10) to reflect an increasing importance of wind-blown snow for glacier accumulation. Specifically, the enhanced accumulation on eastern glacier sides is most likely to be a product of orographically enhanced accumulation via snowdrifting by the prevailing westerly/southwesterly winds, which are typical of the northern part of James Ross Island (Láska and others, 2011). Thus other factors (e.g. gross precipitation and air temperature) have become relatively less important. Such a change in dominance of controlling factors could be both a cause and a symptom of an evolving glacier mass balance

Table 3. Summary comparison of glaciological character and behaviour of land-terminating glaciers on northern James Ross Island during a late-Holocene advance and at the present day

Method	Glaciological property		Late-Holocene advance	Present day
Measured and calculated	Character; topography and geometry	Areal extent Ice thickness Surface long profile Surface elevation change with distance from snout	Typically 100 m recession of snout Typically 15–20 m surface lowering Convex Uniform rise in elevation across glacier	 Linear Asymmetric (easterly parts are higher)
	Landscaping activity	Sediment sources, entrainment, pathways and deposition	Headwall via rockfall, supraglacially to produce lateral moraines Bed via adfreezing and thrusting, englacially and subglacially to produce debris cover on ice-cored moraines	General absence of contemporary supraglacial debris Meltout of ice-cored moraine
Inferred and speculated	Behaviour	Ice dynamics Net annual mass balance Thermal regime	Shearing and thrusting Positive Polythermal glacier with a temperate core/snout	Passive downwasting Negative Fully cold-based glacier
	Likely dominant control on annual mass balance		Air temperature	Wind-blown snow

and dynamics (Table 3). The fact that the glaciers were more extensive, thicker and with a convex long profile (Fig. 9) demonstrates that they had a positive mass balance (Table 3). In contrast, the present asymmetric surface morphology and linear-slope long profile, and an absence of contemporary surface crevasses or moraines could be taken to imply a modern negative mass balance. Furthermore, the apparent lack of dynamism at the present day is due to glacier stagnation or to a cold-based thermal regime, which has been reported for other present-day James Ross Island glaciers by Hambrey and Glasser (2012).

Overall, comparisons of the geomorphology and sedimentology of the moraines with the current glaciological surface structure of the glaciers permit speculation that a transition from a polythermal to a cold-based thermal regime has occurred. However, since we do not have any data on absolute ice thickness we cannot compute changes in overburden pressure and hence in likely basal ice temperature. This is why we rely on the geomorphological and sedimentological evidence for our inferences of past glacier behaviour (Table 3).

COMPARISON WITH EVIDENCE OF GLACIER EVOLUTION AT OTHER SITES

Advances of both cold-based and polythermal glaciers have occurred during the Quaternary on the Antarctic continent and islands, respectively. Several advances of cold-based ice lobes have been deduced from textural analysis of drift deposits at Upper Taylor Glacier in the Dry Valleys (Marchant and others, 1994). Knowledge of the land-forming result of these Antarctic (cold-based) glacier advances is limited to just a few discrete sites, but they are reviewed by Fitzsimons (2003) and most recently expanded by Hambrey and Fitzsimons (2010). Fitzsimons (1990) established that valley glaciers in the Vestfold Hills with cold-based margins released most debris when basal and englacial debris bands became warped and reached the surface of the glacier, or where debris bands were exposed by ablation of the ice surface. Thus moraine formation at these cold-based ice margins is controlled by the micro-climate at the glacier snout and proceeds as meltout and sublimation tills insulate an underlying ice core (Fitzsimons, 1990). Hambrey and Fitzsimons (2010) describe glaciotectonized ridges and aprons at Wright Lower Glacier in the Dry Valleys and demonstrate a link to glacier structure.

Outside of Antarctica, far more knowledge of the character, behaviour and landforms of polar land-terminating glaciers has been acquired from Svalbard (e.g. Bennett and others, 1996; Glasser and Hambrey, 2001; Lysa and Lønne, 2001; Hambrey and others, 2005) and from the Canadian Arctic (e.g. Ó Cofaigh, 1998). Svalbard glaciers have been noted to have evolved in terms of both dynamics and thermal regime over centennial timescales during the late Holocene (Glasser and Hambrey, 2001; Hambrey and others, 2005), and the same has been suggested for Kårsaglaciären, northern Sweden (Rippin and others, 2011). Those interpretations from Svalbard were drawn from interpretations of structural glaciology, landforms and sediments and from northern Sweden from calculations of overburden pressure and bed shear stress. However, unlike this study they did not include reconstructions of former glacier geometry. It is therefore difficult to say whether our inferences of feedbacks between glacier morphology and climate are unusual. What is more certain is that glacier morphological changes can be due to, or affected by, glacier dynamics processes that produce pressure melting and strain heating (Blatter and Hutter, 1991), as well as due to climatic changes. Therefore while we speculate that ongoing glacier thinning has produced a thermal regime change it must be appreciated that the climatic and glaciological conditions needed for a glacier to change from polythermal to cold-based are debatable.

DISCUSSION ON RATE OF GLACIOLOGICAL CHANGE

There is uncertainty as to the age of the moraines described in this paper. No absolute dates exist for these moraines, not least because of the lack of organic matter within or upon them, and because of the problems and errors associated with dating (Holocene) basaltic surfaces using cosmogenic nuclide techniques. However, the moraines must be much younger than a separate coastal moraine that was initially attributed to an advanced IJR-45 by Rabassa (1983) and that was reappraised to be part of a 'Neoglacial episode' ~4500–4700 a ^{14}C by dating of organic matter within lake sediments (Zale and Karlén, 1989; Björck and others, 1996; Hjort and others, 1997). The location of the samples that yielded these dates is indicated by a white star in Figure 1 on the western shore of Brandy Bay and ~5 km away from the moraines described in this study. We thus consider that the moraines reported in this study are most likely part of a 700–1000 a ^{14}C glacial advance that was broadly synchronous with the early stages of the Little Ice Age, which has been postulated (but undated) for James Ross Island (Strelin and others, 2006), for the western side of the Antarctic Peninsula, albeit from offshore marine sediments such as those in the Palmer Deep (cf. Domack and others, 2001; Hall, 2009), and for the Vestfold Hills area on the Antarctic mainland (Fitzsimons and Colhoun, 1995). That age would produce mean snout retreat rates of 0.17–0.1 m a^{-1}, mean surface lowering of 0.03–0.02 m a^{-1} and mean areal decline of 0.03% a^{-1}, for example.

These 'back-of-the-envelope' calculations of rates of ice mass change are much lower than calculated for the few glaciers on the northern and eastern Antarctic Peninsula with direct mass-balance measurements made over the last three decades. Glaciar Bahía del Diablo decreased in surface elevation by 1.0 m a^{-1} from 1985 to 1998 and by 1.4 m a^{-1} from 1998 to 1999 (Skvarca and De Angelis, 2003). Ongoing work by Nývlt and others (2010) and Engel and others (2012) has demonstrated an annual mean loss of 0.79 m surface lowering of IJR-45 between 1979 and 2006. Therefore, we conclude that either the moraines reported in this study are (much) younger than 700–1000 ^{14}C a BP, or that ice mass loss on the Ulu Peninsula has dramatically accelerated over recent decades.

SUMMARY AND CONCLUSIONS

In summary, combined topographical, geomorphological and sedimentological measurements and observations of prominent moraines on the Ulu Peninsula together describe the first mesoscale changes in the character and behaviour of land-terminating glaciers in the Antarctic Peninsula region (Table 3). The land-terminating glaciers on James Ross Island have retreated (75–120) ± 1.4 m per glacier, and glacier surfaces have lowered (9–23) ± 1.6 m on average since a late-Holocene advance. These areal changes are relatively uniform at individual glaciers and between glaciers, but surface lowering has an east–west pattern that we consider indicates the importance of wind-drifted snow from the west in maintaining accumulation. The late-Holocene moraines are ice-cored and this ice has layering that is inclined at 40° and which is discriminated by laminae and thin beds of bubbly white ice and debris-rich basal ice. This ice-core structure controls the morphology of the moraines, and its composition reflects shearing and thrusting at the glacier snout during its late-Holocene advance. Supraglacial debris is restricted to lateral moraines closest to headwalls, so it seems that sediment transport distances in these glaciers are very short, of the order of a few hundred metres at most. The composition of the moraines and the ice surface reconstructed from these moraines together suggest that glacier mass balance during the late-Holocene advance was most likely controlled by air temperature and that the glaciers were a lot more dynamic than at present (Table 3).

The cause of the late-Holocene negative mass balance is far from clear. The simplest explanation is that it reflects a warming climate, which has been determined for the northern Antarctic Peninsula region (e.g. Björck and others, 1996; Ingólfsson and others, 2003) but is nonetheless surprising since precipitation is also inferred to have increased on James Ross Island in the last few thousand years (Björck and others, 1996). Furthermore, more complex explanations such as a shift in the atmospheric circulation pattern (e.g. Turner and others, 2005) and thus changing wind-blown snow patterns could be important.

Comparison of past glacier dynamics as reconstructed from the geomorphological, sedimentological and geometrical properties of the Ulu Peninsula moraines, with glacier dynamics as inferred from the present morphology and surface character of the Ulu Peninsula glaciers (Table 3), permits speculation that late-Holocene ice-marginal retreat and surface lowering has caused a transition in the glaciers from a polythermal to a cold-based thermal regime. Land-terminating glaciers on the Ulu Peninsula are now apparently cooler despite a warmer climate.

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