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The Tekapo Glacier, New Zealand, during the Last Glacial Maximum: An active temperate glacier influenced by intermittent surge activity

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ABSTRACT

Quaternary glaciations have created impressive landform assemblages that can be used to understand palaeo-glacier extent, character and behaviour, and hence past global and local glacier forcings. However, in the southern hemisphere and especially in New Zealand, the Quaternary glacial landform record is relatively poorly investigated with regard to glaciological properties. In this study, a 1 m digital elevation model (DEM) was generated from airborne LiDAR data and supplemented with aerial imagery and field observations to analyse the exceptionally well-preserved glacial geomorphology surrounding Lake Tekapo, New Zealand. We describe a rich suite of Last Glacial Maximum (LGM) and recessional ice-marginal, subglacial, supraglacial, glaciofluvial and glaciolacustrine landform assemblages. These represent two landsystems comprising i) fluted till surfaces with low-relief push moraine ridges; and ii) crevasse-squeeze ridges, 'zig-zag' eskers and attenuated lineations. The former landsystem records the behaviour of an active temperate glacier and the latter landsystem, which is superimposed upon and inset within the former, strongly suggests intermittent surge phases. The two landsystem signatures indicate a sequential change in ice-marginal dynamics during recession that was likely to have been partially non-climatically driven. Overall, we present the first evidence of surge-type glacier behaviour in New Zealand.

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1. Introduction

Glacial landforms are a fundamental source of information regarding the extent, character and behaviour of former glaciers. They can provide a valuable record of palaeoglaciological characteristics including the spatial and temporal evolution of ice flow configurations, meltwater drainage patterns, basal thermal regimes, internal dynamics and ice-marginal recession (e.g. Dyke and Prest, 1987; Boulton and Clark, 1990a, 1990b; Kleman and Borgström, 1996; Clark, 1997; Kleman et al., 1997, 2006; Svendsen et al., 2004; Stokes et al., 2012, 2015; Storrar et al., 2014a, 2014b; Newton and Huuse, 2017; Patton et al., 2017; Margold et al., 2018). More specifically, the application of modern analogue glacial landsystems models (cf. Evans, 2003, 2013 and references therein) has facilitated detailed reconstructions of the former ice dynamics of the Laurentide (Dyke and Prest, 1987; Evans et al., 1999, 2008, 2014; Clark et al., 2000; Dyke et al., 2002; Stokes et al., 2012), Innuitian (England, 1999; Ó Cofaigh et al., 2000; England et al., 2006), Fennoscandian (Sollid et al., 1973; Kleman et al., 1997; Winsborrow et al., 2010), Patagonian (Benn and Clapperton, 2000; Glasser and Jansson, 2005; Glasser et al., 2008; Darvill et al., 2017; Ponce et al., 2019), and the British-Irish ice sheets (Evans et al., 2009a; Greenwood and Clark, 2009a, 2009b; Hughes et al., 2010, 2014; Clark et al., 2012, 2018a), as well as smaller ice masses such as mountain and plateau icefields and their outlets (e.g. Evans et al., 2002; Rea and Evans, 2003; Bickerdike et al., 2016, 2018a, 2018b and references therein), and the northern and southern forelands of the European Alps (Reuther et al., 2011; Ellwanger et al., 2011; Reitner et al., 2016; Monegato et al., 2017).

Such palaeo-glacier reconstructions are often used as the basis for understanding global and local forcings on glaciers and hence past climate. The propensity of these climate-focused studies in the northern hemisphere contrasts with the relatively poor coverage and resolution of studies of palaeo-glacier dynamics in the southern hemisphere. Studies in Patagonia on glacier dynamics inferred from the glacial geomorphological record, have been limited to a few selected sites (e.g. Lovell et al., 2012). Such studies across the Southern Alps of New Zealand include Porter (1975), Suggate (1990), and Bacon et al. (2001). The scarcity of this research is perhaps surprising given the identification within individual valleys of extensive and repeated glaciations of the Southern Alps throughout the Quaternary (Gage and Suggate, 1958; Speight, 1963; Gage, 1985; Clapperton, 1990; Suggate, 1990; Pillans, 1991;

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Newnham et al., 1999; Barrell, 2011; Barrell et al., 2011; Shulmeister et al., 2019) that left an exceptionally well-preserved set of landforms and sediments (e.g. Hart, 1996; Mager and Fitzsimons, 2007; Shulmeister et al., 2009; Evans et al., 2010c, 2013; Barrell et al., 2011; Putnam et al., 2013a, 2013b; Cook et al., 2014; Borsellino et al., 2017; Sutherland et al., 2019). However, recent work has also considered the Southern Alps glaciations regionally and numerical modelling of ice extent driven by a palaeo-climate has been compared to the landform record (e.g. Golledge et al., 2012; Doughty et al., 2015). In a contrasting workflow, James et al. (2019) have recently refined the regional landform record and used it to generate glaciological reconstructions of ice across the Southern Alps during the LGM.

The majority of palaeo-glaciological research in New Zealand has focused on constraining the timing of glacial fluctuations (e.g. Suggate, 1990; Preusser et al., 2005; Shulmeister et al., 2005, 2010a, 2010b, 2018; Suggate and Almond, 2005; Schaefer et al., 2009; Kaplan et al., 2010, 2013; Putnam et al., 2010a, 2010b; Putnam et al., 2013a, 2013b; Kelley et al., 2014; Doughty et al., 2015; Rother et al., 2015; Koffman et al., 2017; Thackray et al., 2017; Barrell et al., 2019). Much less attention has been given to landform identification or to the detailed nature of landform-sediment assemblages (Hedding et al., 2018). However, palaeo-climatic reconstructions rely, crucially, on accurate landform identification and interpretation in terms of process-form regimes before dating programmes are initiated. In particular, work to address this problem in New Zealand has been carried out by Kirkbride and Winkler (2012), Reznichenko et al. (2012, 2016), and Winkler (2018). Therefore besides providing an understanding of past glacier extent, character and behaviour, more systematic geomorphological mapping will facilitate targeted efficient field campaigns and a better understanding of past glacier-climate interactions in the Southern Alps.

2. Study site and geomorphological setting

Relatively well-studied glacier forelands on South Island, New Zealand, cluster in the upper Mackenzie basin (Fig. 1a) and include the LGM/Late Glacial landform studies in the Pukaki Valley (Speight, 1963; Hart, 1996; Schaefer et al., 2006, 2015; Mager and Fitzsimons, 2007; Putnam et al., 2010a, 2010b; Evans et al., 2013; Barrell, 2014; Barrell and Read, 2014; Kelley et al., 2014; Doughty et al., 2015) and those around Lake Oahu (Putnam et al., 2013b; Webb, 2009). These studies have collectively shown that glacial retreat since the Last Glacial Maximum (LGM; 26.5–19 ka; P.U. Clark et al., 2009) was complex, and chronological investigations indicate that rapid and sustained glacier recession began c. 18 ka and that subsequently, the glaciers lost up to 40% of their length within no more than c. 1000 yrs. (Putnam et al., 2013b; Schaefer et al., 2015). The other large valley in the upper Mackenzie basin is that containing Lake Tekapo (Fig. 1), which, in contrast, has received very little attention regarding its glacial geomorphology. The aim of this study is therefore to present in detail the glacial geomorphology of the Tekapo Valley in order to characterize the nature and behaviour of its former outlet glacier throughout the LGM and during the subsequent deglaciation, using a landsystems approach (cf. Evans, 2003) in an effort to better inform the regional palaeo-glaciological reconstructions.

The northeastern part of the Mackenzie basin (Fig. 1a) is a region characterized by high mountains (>3000 m a.s.l) and deep troughs. Lake Tekapo is the second-largest of three sub-parallel lakes running

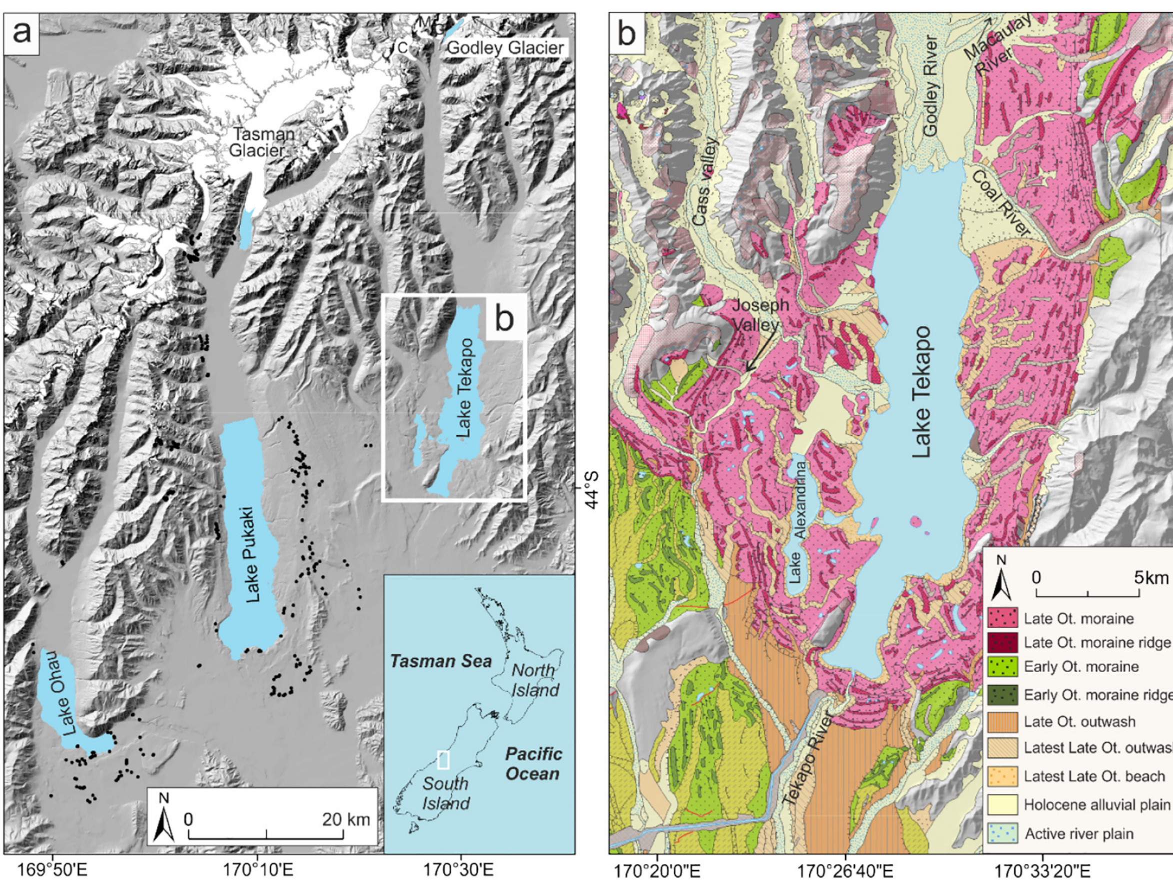

Fig. 1. Study area. a). Location of Lake Tekapo in the Mackenzie basin, South Island, New Zealand. Contemporary glacier outlines (white) from Global Land Ice Measurements from Space (GLIMS) mapped onto hillshaded 8 m DEM. The location of surface exposure ages derived from ^{10}Be terrestrial cosmogenic nuclide (TCN) dating (Schaefer et al., 2006, 2009, 2015; Putnam et al., 2010a, 2010b, 2013b; Kaplan et al., 2010, 2013; Kelley et al., 2014; Doughty et al., 2015) are indicated by small black circles. Note the paucity of quantitative ages in the Tekapo catchment. The Tasman and Godley glaciers are labelled. The Classen, Maud, and Grey glaciers are denoted on the map by C, M, and G respectively. b). Geochronological surfaces around Lake Tekapo as mapped by Barrell et al. (2011). Ot. in legend = Otiran (the latest Pleistocene glaciation in New Zealand).

north-south along the northern edge of the basin (Fig. 1a, b). The other lakes in the Mackenzie basin, Lake Pukaki and Lake Ohau, are situated 30 km and 60 km southwest of Lake Tekapo respectively. Lake Tekapo lies in a long, narrow, ice-scoured bedrock depression, partially blocked by moraine and outwash deposits (Gage, 1975; Sutherland et al., 2019). Lake Tekapo has a maximum depth of 120 m (Mountjoy et al., 2018), is 27 km long and covers an area of 83 km^2. Sedimentation into Lake Tekapo is dominated by the Godley River, which forms an extensive delta in the northern third of the lake (Pickrill and Irwin, 1983). Approximately 52% of the catchment is drained by the Godley, Macaulay and Coal Rivers, all of which drain glacier-fed catchments on the eastern slopes of the Southern Alps. The original outflow from Lake Tekapo was located at the southern end of the lake and into the Tekapo River. However, outflow and lake levels have been controlled artificially by the Tekapo Power Scheme since 1953 (Mountjoy et al., 2018). The present-day lake level is held due to hydroelectric damming at 707 m a.s.l., which is 10 m higher than its natural level.

2.1. Glacial chronology of the Mackenzie basin

The valleys in the Mackenzie basin show geological and geomorphological evidence of multiple glaciations during the Pliocene-Pleistocene defined by moraine belts of the lateral and terminus margins of the LGM outlet glaciers. The Late Otiran Glaciation was the last major glacial event in the Southern Alps (Table 1). Along with associated outwash plains, the moraine belts form a nested succession of glaciogenic landforms (Table 1), named from oldest to youngest as the Wolds, Balmoral, Mt. John, and Tekapo formations (Gair, 1967; Cox and Barrell, 2007; Barrell et al., 2011). Free of the topographic complexities that tend to confound the preservation of older moraines in Westland and central Canterbury, the Mackenzie basin moraine sequences illustrate a key feature of New Zealand glacial succession; that ice was more extensive during previous glaciations and the expansion of glaciers beyond confining valleys has allowed substantial preservation of older deposits (Barrell et al., 2011). Farther down-valley from each glacier trough are sets of Early Otiran and older moraines, best preserved between Lake Pukaki and Lake Tekapo. The outer sector of these complexes (Wolds Formation) comprises discontinuously preserved remnants of subdued moraines and outwash surfaces, whilst increasingly well-preserved moraines and outwash surfaces lie progressively closer to the lake basins (e.g. Mt. John and Tekapo Formations).

Four of New Zealand's largest modern glaciers lie within the Tekapo catchment; the Maud, Grey, Classen and Godley glaciers (Fig. 1a). During Pleistocene glaciations, these glaciers, together with ice from numerous other tributaries, combined to form a single, large valley glacier that extended out from the Southern Alps to the north-western edge of the inter-montane Mackenzie basin. This glacier has sometimes been referred to as the 'former' or 'expanded' Late Otiran Godley Glacier (Speight, 1963; Porter, 1975). However, in this study, we refer to it as the 'Tekapo Glacier' after the lake that occupies part of the trough of the former glacier (Maizels, 1989; Barrell et al., 2011). The Godley Glacier now lies 35 km upstream of Lake Tekapo (Irwin, 1978) and is presently the eighth largest glacier in New Zealand. It has retreated >8 km since the end of the Little Ice Age (LIA), 1900 CE (Chinn, 1996; Winkler, 2004), which is the farthest retreat of any Southern Alps glacier throughout the 20th Century (Chinn et al., 2012; Pelto, 2017).

The LGM ice configuration across the Southern Alps has been suggested by Barrell et al. (2011), Golledge et al. (2012), and most recently by James et al. (2019). During the LGM, the Tekapo Glacier constructed a suite of landforms, recognised in the literature as sequences of inset latero-frontal moraines and associated outwash fans (Barrell et al., 2011). The surficial and bedrock geology has been mapped by Cox and Barrell (2007), and the extensive glacial geomorphological map of the central South Island by Barrell et al. (2011) encompasses Lake Tekapo and its surroundings (Fig. 1b). However, their map mainly focusses on chronological classification of land surfaces, with little information on the geomorphological properties from which process-form regimes for glaciogenic features and assemblages can be deciphered. In the adjacent Pukaki Valley, Schaefer et al. (2006, 2015), Putnam et al. (2010a, 2010b), Kelley et al. (2014), and Doughty et al. (2015) combined geomorphological mapping based upon Barrell et al. (2011) and cosmogenic nuclide exposure ages to present a glacial chronology of LGM and Late Glacial age. Based on landforms located at roughly the same longitudinal positions and elevations, we therefore interpret that the landforms mapped in this study approximately mark the culmination of the last major advance of the Tekapo Glacier and the subsequent retreat that documents the end of the Late Otiran Glaciation.

2.2. Palaeo-climate of New Zealand during the LGM

Warmer interstadials occurred during the generally cold LGM in New Zealand, with mean annual air temperatures similar to, or up to 3.7 °C cooler than present (Marra et al., 2006; Alloway et al., 2007; Woodward and Shulmeister, 2007). Palaeo-precipitation in New Zealand during the LGM is poorly recorded by terrestrial proxies. Nonetheless, growth rate variability and deviations in isotopic composition of speleothems from cave systems in South Island generally indicate drier than present conditions punctuated by transient wetter episodes (Whittaker et al., 2011). Thus, the LGM in New Zealand and the southwest Pacific was marked by changes in regional atmospheric character, perhaps reflecting zonal circulation shifts (e.g. latitudinal migration or

Table 1
Glacial stratigraphy of the Mackenzie basin. Interglacial phases are shaded in grey.

<table>
<tr><th colspan="3">Assigned Geological Period</th><th colspan="2">Glacial Formation</th><th>Assigned Marine Isotope Stage (MIS)</th><th>Approximate Age (yrs ago)</th></tr>
<tr><td colspan="3"></td><td>Mackenzie basin</td><td>Southern Alps (undifferenciated)</td><td colspan="2"></td></tr>
<tr><td colspan="3" rowspan="2">Holocene</td><td>Ben Ohau</td><td></td><td rowspan="2">1</td><td>0 – 11,7000</td></tr>
<tr><td>Birch Hill</td><td>Late Glacial</td><td>11,7000 – 14,5000</td></tr>
<tr><td rowspan="3">Otira (last glaciation)</td><td rowspan="2">Late Otira</td><td>Latest Late Otiran</td><td>Tekapo</td><td>Latest Last Glacial Maximum</td><td rowspan="2">2</td><td>16,000 - 18,000</td></tr>
<tr><td></td><td>Mt John</td><td>Last Glacial Maximum</td><td>18,000 - 24,000</td></tr>
<tr><td colspan="2">Early Otira</td><td>Balmoral 2</td><td></td><td>4</td><td>59,000 – 71,000</td></tr>
<tr><td colspan="3">Kahinu (last interglacial)</td><td></td><td></td><td>5</td><td></td></tr>
<tr><td colspan="3">Waimea</td><td>Balmoral 1</td><td></td><td>6</td><td>128,000 – 186,000</td></tr>
<tr><td colspan="3">Karoro Interglacial</td><td></td><td></td><td>7</td><td>190,000 – 244,000</td></tr>
<tr><td colspan="3"></td><td>Wolds</td><td></td><td>8</td><td></td></tr>
</table>

(Adapted from Barrell et al. (2011).)

dynamic changes in the strength of westerlies (Rojas et al., 2009; McGlone et al., 2010; James et al., 2019).

Rother and Shulmeister (2006) showed that temperature at modern mean equilibrium line altitudes (ELAs) in New Zealand are much warmer than those at an equivalent latitude in the European Alps where snowfall is less. Indeed, mean ELA temperatures in the Southern Alps are above freezing but despite the relatively high temperatures, extensive Pleistocene glaciation occurred. To maintain positive mass balance and support ice growth, modern and Pleistocene glaciers in New Zealand are, and were, fueled with exceptionally high snow volumes supplied by westerly airflow.

3. Methods

A combination of high resolution topographic data, aerial photography and extensive fieldwork was used to identify, map and interpret glacial landforms around Lake Tekapo. Over 357 km^2 of airborne LiDAR data was acquired by a fixed wing aircraft in December 2016 by Canterbury Regional Council and has been released by Land Information New Zealand (LINZ) under Creative Commons Attribution 4.0 New Zealand (CC BY 4.0). In this study the LiDAR-based point cloud data (3.44 points per m^2) were meshed into a 1 m gridded digital elevation model (DEM) in CloudCompare 2.10 (CloudCompare, 2016). Solar-relief-shaded visualisations (315° and 45° azimuth) were constructed from the DEM, primarily to provide topographic context in areas of complex relief. These elevation data far surpass the resolution of all other (NZ) national topographic datasets (8 m) and are ideal for detailed geomorphological mapping and especially for the detection of subtle (low-relief, metre-scale) landforms.

As a complement to the high-resolution elevation data, 0.4 m resolution aerial photography was also used to map landforms. The aerial imagery and orthophotographs were captured from airborne sensors and cameras, sourced from the LINZ Data Service and licenced by the Canterbury Aerial Imagery (CAI) consortium, for re-use under the Creative Commons Attribution 4.0 international licence. Orthophotographs for the Canterbury region were acquired in the austral summer 2013–2014.

Fieldwork was completed in December 2017 and December 2018 and was used to ground-truth mapping completed to that date, to inform future mapping and to seek out and make visual examination of sediment exposures wherever available.

As recently highlighted by Chandler et al. (2018) there are numerous different approaches on how to compile glacial geomorphological maps ranging from very trivial to highly sophisticated techniques. In this study, geomorphological criteria to identify ice-marginal and subglacial landforms from high resolution topography was guided in part from Table 1 in Carrivick et al. (2017a) and in part from Tables 1 and 2 in Perkins and Brennand (2015). The identification of glaciolacustrine landforms was guided by Table 2 in Sutherland et al. (2019). Interpretations and subsequent landform classifications were based upon known process-form relationships in recently deglaciated terrains.

Our geomorphological mapping was completed using ESRI ArcMap 10.4 (© ESRI) software and all files were projected in the New Zealand Transverse Mercator 2000 (NZTM_2000) grid system. Mapping was driven by visual analysis of landform position, size, shape, surface composition and association (proximity to other landforms). This analysis was achieved via on-screen digitization of each individual landform visible in the topography and aerial imagery datasets. Additionally, elevation profiles and individual contours were constructed using the software extension 3D Analyst to assist with identifying landforms and to promote the semi-automated delineation of some of them, such as prominent palaeo-lake shorelines (cf. Carrivick et al., 2017a, 2017b).

4. Results

The glacial landform assemblages in the Tekapo Valley are mapped in Fig. 2. Many landforms are situated within clusters and are not easily individually distinguishable and so the map is designed to be viewed at A0 size (see .pdf file available as online Supplementary data). Fourteen glacial landform types have been recorded on the geomorphological map (Fig. 2) and they are divided into five assemblages: (i) subglacial, (ii) supraglacial, (iii) ice-marginal, (iv) glaciofluvial, and (v) glaciolacustrine. Lakes, rivers, and alluvial features were also mapped to provide some geological and topographic context. In the following we provide an overview of their distribution and describe and interpret the glacial landform assemblages before discussing their palaeo-glaciological significance.

4.1. Subglacial landforms

4.1.1. Drumlins

Conspicuous, streamlined mounds that range from 130 m to 1200 m in length, from 60 m to 380 m wide and from 2 m to 40 m in height appear in distinct clusters in the central western part of the Tekapo Valley (Fig. 2). In total, 138 of these landforms are identifiable across the Tekapo basin. They are extensively fluted and are often draped/overprinted by other landforms such as push moraine ridges, crevasse-squeeze ridges and eskers (Fig. 3B, D). Their surficial composition is that of well-rounded pebbles and cobbles (Fig. 4C). Generally, the features are more subdued, rounded and smoothed in the east and get progressively more elongate towards the northwest. Standing water often occupies the depressions between the individual landforms which accentuates their morphology. Their shapes are non-uniform and two distinct morphological types are identified across the study area. The first, mainly concentrated in between Lake Tekapo and Lake Alexandrina (Figs. 3A, B and 4A, B), tend to be transverse asymmetrical or parabolic (Shaw, 1983) in the sense that they taper up-ice to form triangular, barchan-type shapes. Here, their orientation varies as they occur in a splayed fan distribution. The second morphological type is identified between the northwestern side of Lake Alexandrina and the Joseph Valley (Fig. 3C, D) where they display elongate, highly attenuated forms. Here, the features are long (>500 m) and thin (<50 m wide), with well-defined crestlines (Fig. 3C, D). The majority are spindle-shaped and taper towards each end. Their morphology is clearly linear and they have a distinct northeast-southwest orientation.

We interpret these streamlined mounds as individual drumlins in view of their similar morphologies and elongated oval shapes (e.g. Krüger and Thomsen, 1984; Boulton, 1987; Maclachlan and Eyles, 2013; Eisank et al., 2014; Jónsson et al., 2014). We classify the Múlajökull drumlin field, the only known active drumlin field (Johnson et al., 2010; Jónsson et al., 2014; Benediktsson et al., 2016), in central Iceland as a unique analogue to the drumlins mapped in between Lake Tekapo and Lake Alexandrina. The drumlins at Múlajökull also occur in a splayed fan distribution, representing an arc of 180°. Such landforms are considered rare in alpine regions as they can be easily reworked by post-glacial erosion. However, drumlins have been identified on the alpine forelands of the European Alps, such as those around Lake Constance (Ellwanger, 1992), and lone drumlins or drumlins in small groups have been described in the Turtmann Valley, Switzerland by van der Meer (1983) and van der Meer and Tatenhove (1992). Drumlins have also been identified in the Clarée Valley in the French Alps by Cossart et al. (2012) and interpreted to represent temperate bed conditions.

The precise genesis of drumlins is contentious (cf. Shaw, 1983, 1989; Krüger and Thomsen, 1984; Piotrowski, 1987; Shaw and Sharpe, 1987; Shaw et al., 1989; Clark, 1994; Benediktsson et al., 2016; Eyles et al., 2016; Schomacker et al., 2018), but it is widely agreed that they are subglacially streamlined landforms related to the deformation and/or erosion of a soft-substrate by fast-flowing glacier ice (Boulton, 1987; Stokes et al., 2011). The drumlinised terrain mapped was likely to have formed by subglacial deformation during advances of warm, wet-based ice (Clapperton, 1989; Benn and Clapperton, 2000; Greenwood and Clark, 2008; C.D. Clark et al., 2009). Drumlins form

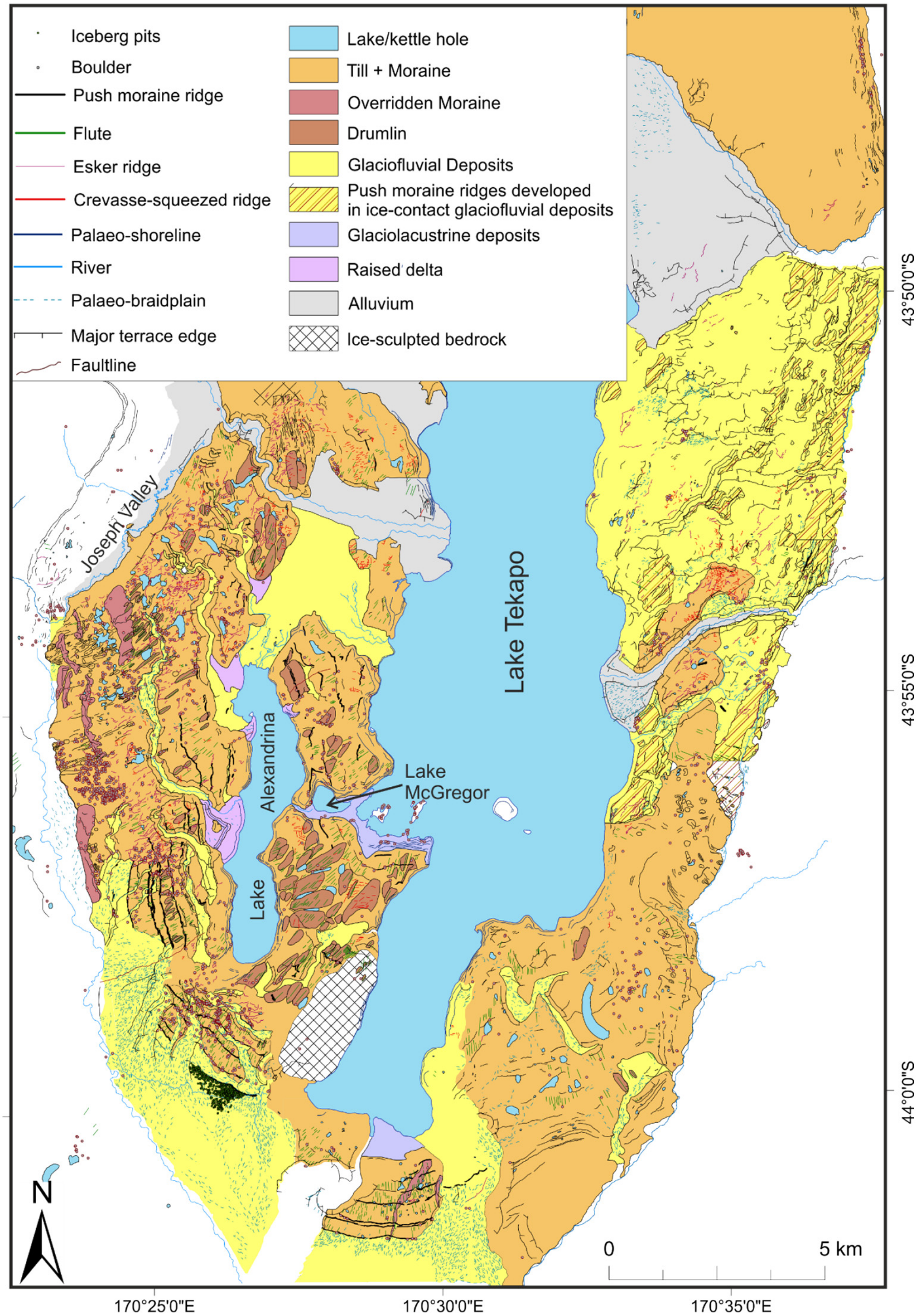

Fig. 2. Geomorphological map of the Tekapo Valley based on 1 m DEM, 0.4 m aerial imagery and widespread field observations. For a full resolution version of the map see online Supplementary data. The full resolution version is designed for viewing at A0 size.

parallel to local ice-flow (Jónsson et al., 2014; Clark et al., 2018b) and the direction of flow is identified by observation of the steeper stoss (up-ice) and tapered lee (down-ice) form. In the case of the Tekapo Valley drumlins, former ice flow was predominantly towards the southwest. The splayed fan distribution of drumlins around Lake Alexandrina (Figs. 3A, B; 4A) reveals that ice flowed in a radial pattern, thus representing a piedmont glacier lobe.

We mapped the drumlins as polygon outlines along their lower break-of-slope in order to show their planform because it more accurately reflects their morphometry and allows calculation of bedform elongation, which is an important proxy for ice velocity (cf. Stokes and Clark, 1999, 2001, 2002; King et al., 2009). Stokes and Clark (2002) demonstrate that long subglacial bedforms such as attenuated drumlins (length:width ratios ≥10:1) are indicative of former areas of fast-flowing ice. In this study, we quantified individual lineation elongation ratios, of which approximately 40% are ≥10:1 and hence infer that the lineations were formed by persistent and relatively fast ice-flow (cf. Boulton, 1987; King et al., 2009). We suggest that the swath of attenuated bedforms that occur in isolation northwest of Lake Alexandrina (Fig. 3C, D), as well as their parallel concordance and the abrupt lateral margins of this zone, are similar to those in areas of former rapidly flowing ice (cf. Stokes and Clark, 1999, 2001; Evans et al., 2008, 2014;

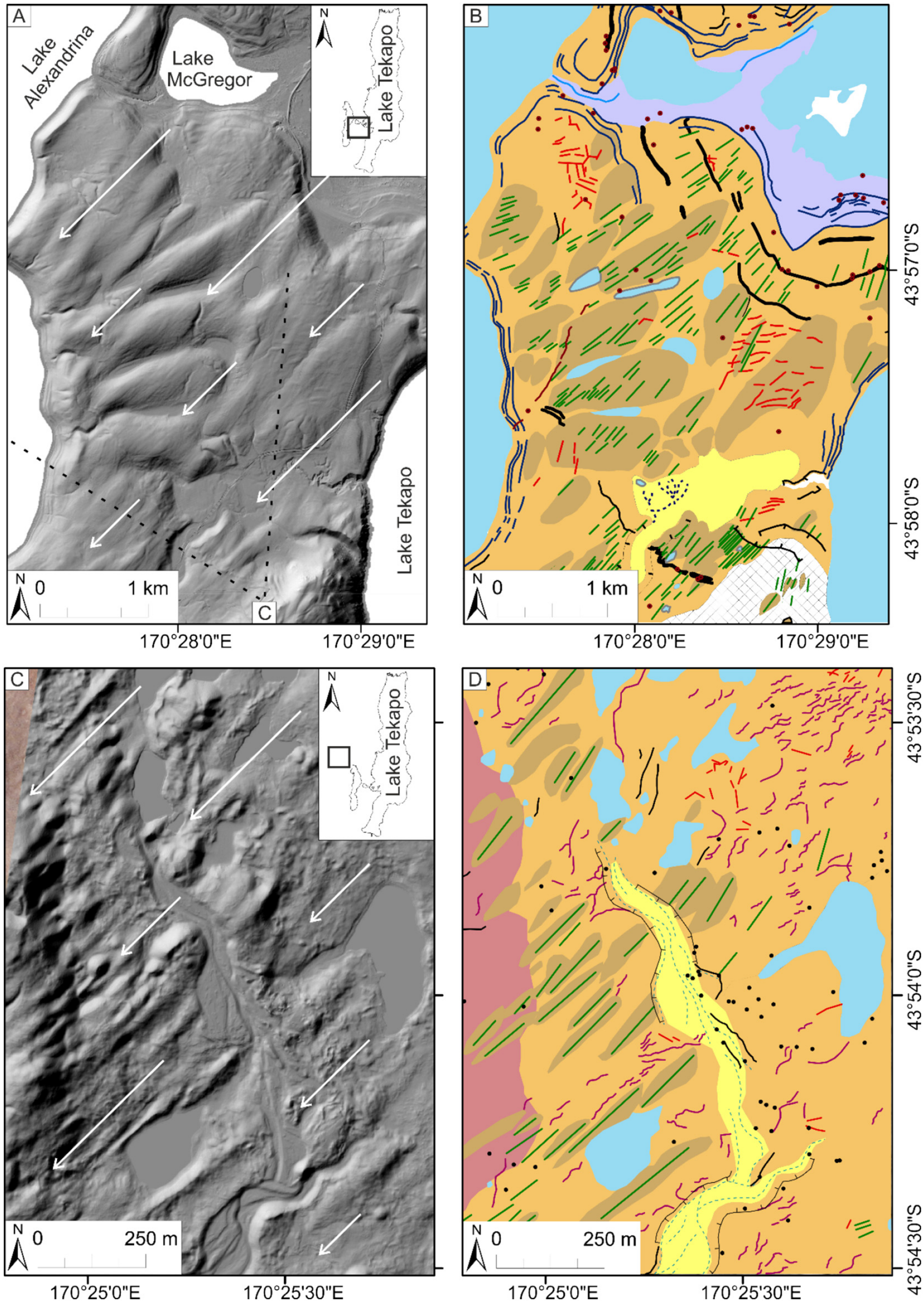

Fig. 3. A. Drumlin terrain in between Lake Tekapo and Lake Alexandrina identified from hillshaded 1 m DEM. B. Mapped interpretation of panel A. Legend is as given in Fig. 2. C. Spindle-shaped drumlins northwest of Lake Alexandrina identified from hillshaded 1 m DEM. D. Mapped interpretation of panel C. Legend is as given in Fig. 2. Note the differences in drumlin morphology and elongation between panels A, B and C, D of this figure. Former ice flow direction is delineated by white arrows on panel A and C.

Lovell et al., 2012), in this case within an active temperate piedmont glacier lobe (cf. Boulton, 1987; Hart, 1999; Evans and Twigg, 2002; Kovanen and Slaymaker, 2004; Evans et al., 2018a).

4.1.2. Flutings

Linear, parallel, positive-relief landforms displaying high directional conformity are widespread throughout the whole study area (Fig. 2). The narrow, streamlined ridges range in length from 30 m to 230 m, but are generally shorter (<80 m) in the southern part of the valley. They are <5 m wide and <1 m high. The majority of the lineations are subdued but closely-spaced and interspersed between moraines ridges (Fig. 5). They are broadly orientated north-northeast to south-southwest. In total, 1294 of these landforms are identifiable across the Tekapo basin.

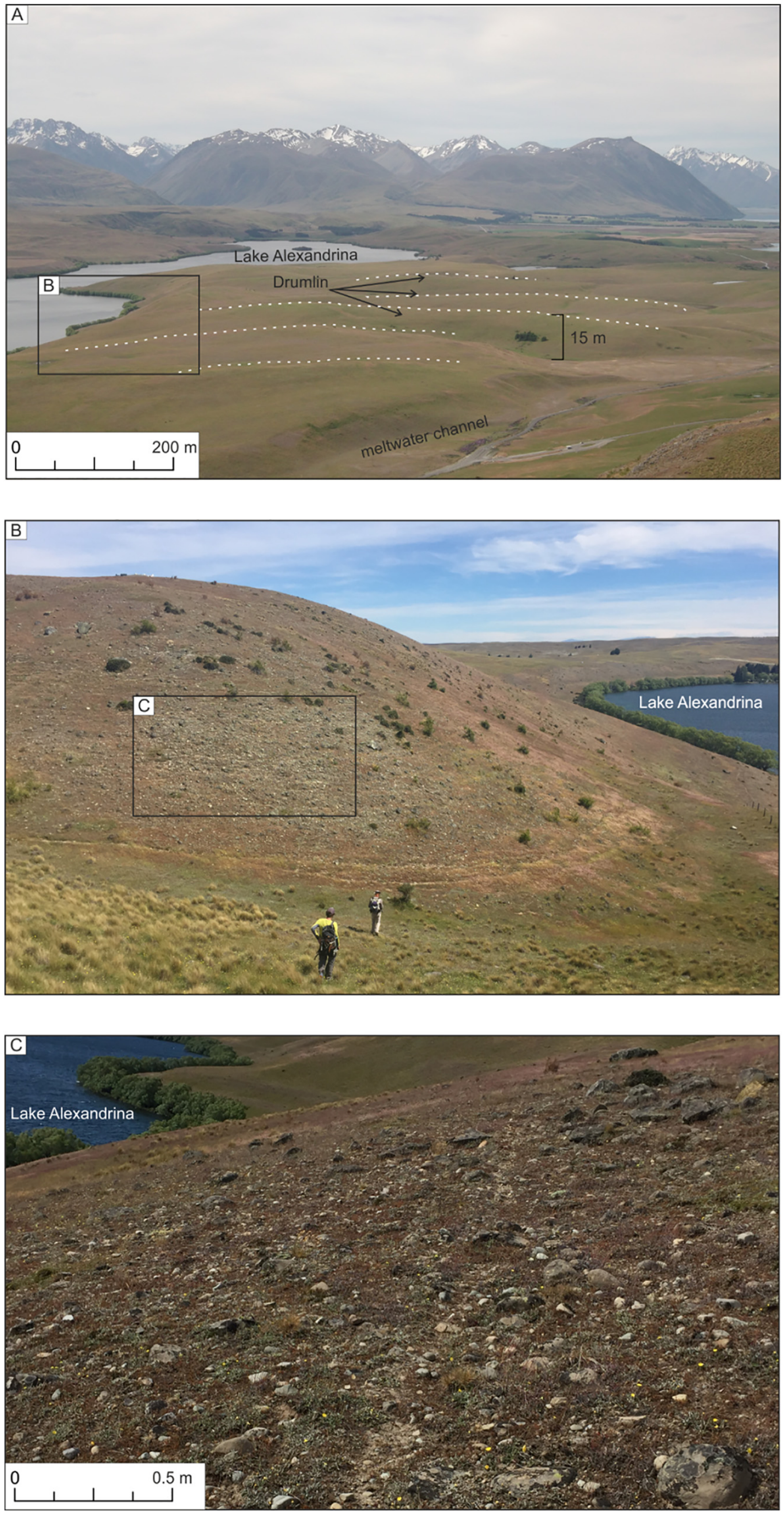

Fig. 4. A. Ground photograph of drumlins, view shown in Fig. 3A. Ground photograph of parabolic-type drumlin in between Lake Alexandrina and Lake Tekapo. View shown in panel A. Note people for scale. C. Surficial deposits showing rounded clasts of drumlin shown in panel B.

We interpret these linear, parallel, positive-relief landforms as glacial flutes (cf. Boulton, 1976; Rose, 1989; Gordon et al., 1992; Clark, 1993; Benn, 1994; Hart, 1998; Ely et al., 2017; Ives and Iverson, 2019). Flutings around Breiðamerkurjökull (Evans et al., 1999; Evans and Twigg, 2002) can be considered a modern analogue for those mapped in the Tekapo Valley due to their similarities in dimensions and

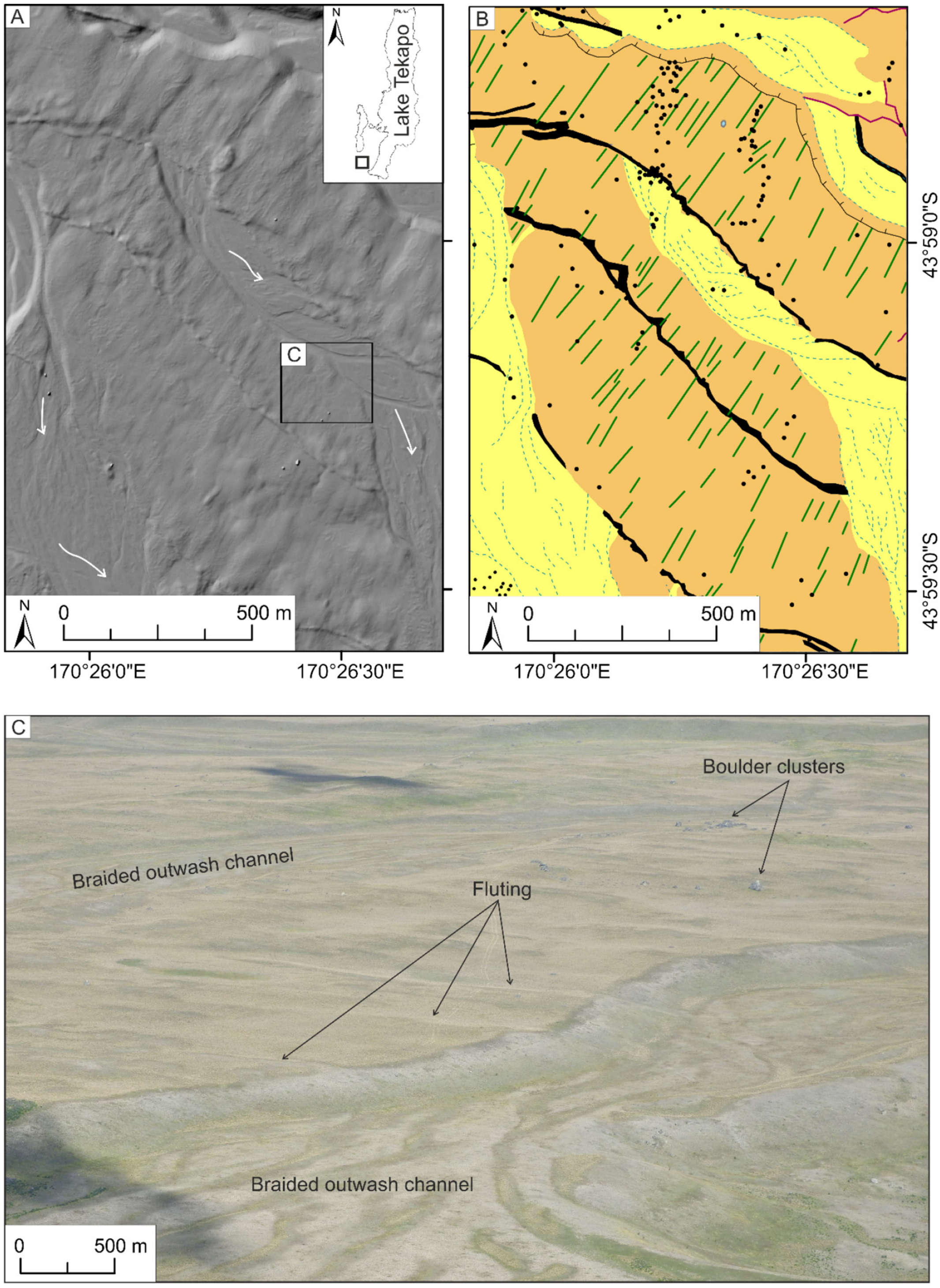

Fig. 5. A. Flutings immediately south of Lake Alexandrina identified from hillshaded 1 m DEM. Note areas of moraine dissected by outwash pathways (white arrows). B. Mapped interpretation of panel A. Legend is as given in Fig. 2. C. Ground photograph of prominent fluting from recessional push moraines ridges. View shown in panel A.

morphometrics. Their spatial distribution represents the principal ice-discharge pathways along the major north-south trending Tekapo Valley axis. The flutings between each moraine ridge are slightly offset from one another and this offset is probably due to small ice flow directional changes at the glacier terminus from year to year (cf. Evans and Twigg, 2002; Evans et al., 2010a, 2017a, 2018b). Nevertheless, some larger flutings can be traced through several push moraines and hence indicate that ice flow direction remained constant between subsequent years (Evans and Twigg, 2002; Chandler et al., 2016a, 2016b, 2016c; Evans et al., 2018a). Overall, the flutings around Lake Tekapo show that the glacier terminus expanded out in a radial flow pattern across the southern part of the lowland Tekapo basin to form a piedmont lobe. Flutings that occur on lowland glacier forelands are a product of the interaction of wet-based ice with a subglacial deforming layer (Eyles et al., 2015) to produce either: i) cavity infill down-flow of lodged stoss boulders (Benn, 1995; Benn and Evans, 1996); or ii) the ploughing of grooves by boulders protruding from the ice base (cf. Boulton, 1979, 1982; Nelson et al., 2005; Kjær et al., 2006); or iii) till squeezing into grooves carved into the glacier sole as the ice slides over lodged boulders (cf. Boulton, 1976; Evans and Rea, 2003). The close association of flutings and recessional push moraines in this study suggests that their formation may be intimately linked, as has previously been suggested at Icelandic glaciers (cf Boulton, 1976; Boulton and Hindmarsh, 1987; Benn, 1994; Evans and Twigg, 2002; Evans, 2003; Chandler et al., 2016b, 2016c). However, the prominence of these subglacial forms does not persist through long periods of surface weathering and erosion. Due to their more subdued morphology, flutings generally have a low potential for preservation unless the ice retreat was continuous without any intermitted re-advances or seasonal advances.

4.1.3. Geometric ridge networks (crevasse-squeeze ridges)

We identify linear, discontinuous ridges that form cross-cutting relationships to one-another. These landforms are subtle. They tend to be of low relief and can be identified only from the hillshaded 1 m DEM. The majority in the Tekapo Valley are short (a few tens of metres) but the longest reach >200 m in length. They are generally <10 m wide and rise to ~2 m above the surrounding terrain. They are an order of magnitude narrower and much more irregular in orientation than any of the adjacent flutes. The conjugate ridges are predominantly orientated transverse or oblique to former ice-flow direction and are arranged in a rectilinear network. Their distribution is widespread throughout the study area and they are most extensive in the central part of both sides of the Tekapo Valley. The mapping reveals they are located in clusters and form dense geometric ridge networks (Fig. 2). Their cross-cutting relationships are easily identified (Fig. 6) and they form in close association with eskers and nearby outwash corridors. In total, we identified and mapped 434 discontinuous, conjugate paired ridges around Lake Tekapo.

Short, linear ridges that cross-cut one-another are classified descriptively in glacial geomorphological mapping programmes as geometric ridge networks (sensu Raedecke, 1978; Bennett et al., 1996) and have been long-regarded as the remnants of crevasse-squeeze ridges (Sharp, 1985b, 1988; Rea and Evans, 2011). Crevasse-squeeze ridges have a low preservation potential and minor sediment mounds, or hummocks that occur in close association might likely reflect other, now collapsed, crevasse-squeeze ridges (e.g. Evans et al., 2016). Crevasse-squeeze ridges have traditionally been regarded as diagnostic of glacier surging in a landsystem sense (cf. Sharp, 1985a, 1985b, 1988; Evans and Rea, 1999, 2003; Evans et al., 2007; Schomacker et al., 2014; Ingólfsson et al., 2016). More recently, however, their occurrence on freshly deglaciated active temperate glacier forelands has been interpreted as indicative of radial crevasse-development in non-surging snouts, because their ridge orientations mimic and continue into the extended limbs of extreme 'sawtooth' push moraines (Evans and Twigg, 2002; Evans et al., 2016, 2017a, 2017b). Where they occur in narrow corridors on palaeo-ice stream trunks, such landforms have also been related to ice stream shutdown (Ó Cofaigh et al., 2010; Evans et al., 2016) and stagnation (Andreassen et al., 2014; Klages et al., 2015; Kurjanski et al., 2019).

It is not possible to classify the genetic origin of such landforms from remote sensing alone since it is the sedimentological pre-conditions that support the interpretation of surge-diagnostic crevasse-squeeze ridges. However, since there were no sedimentary exposures of these landforms observed in the field, we consider their morphological characteristics and in particular, their cross-cutting relationships to be comparable with those around modern surging glacier margins such as Brúarjökull and Eyjabakkajokull in Iceland (Evans et al., 2007; Kjær et al., 2008) where the ridges can be traced from their forelands and into crevasse systems of their snouts. The presence of crevasse-squeeze ridges in the Tekapo Valley, and their cross-cutting relationships however, indicates that the glacier margin was highly fractured by crevasses whose pattern is more diagnostic of a surge origin than radial crevasse infilling (cf. Sharp, 1985a, 1985b; Evans et al., 2007, 2016, 2017a; Rea and Evans, 2011; Lovell et al., 2015). The complex, reticulate networks typical of crevasse-squeeze ridges in the Tekapo Valley were therefore likely produced by till squeezing upwards into open basal crevasses and likely formed where there was soft, readily deformable sediment at the ice-bed interface, either during or immediately after a surge (Sharp, 1985a, 1985b; Evans and Rea, 1999, 2003; Benediktsson et al., 2009; Rea and Evans, 2011; Evans et al., 2016; Evans, 2018;

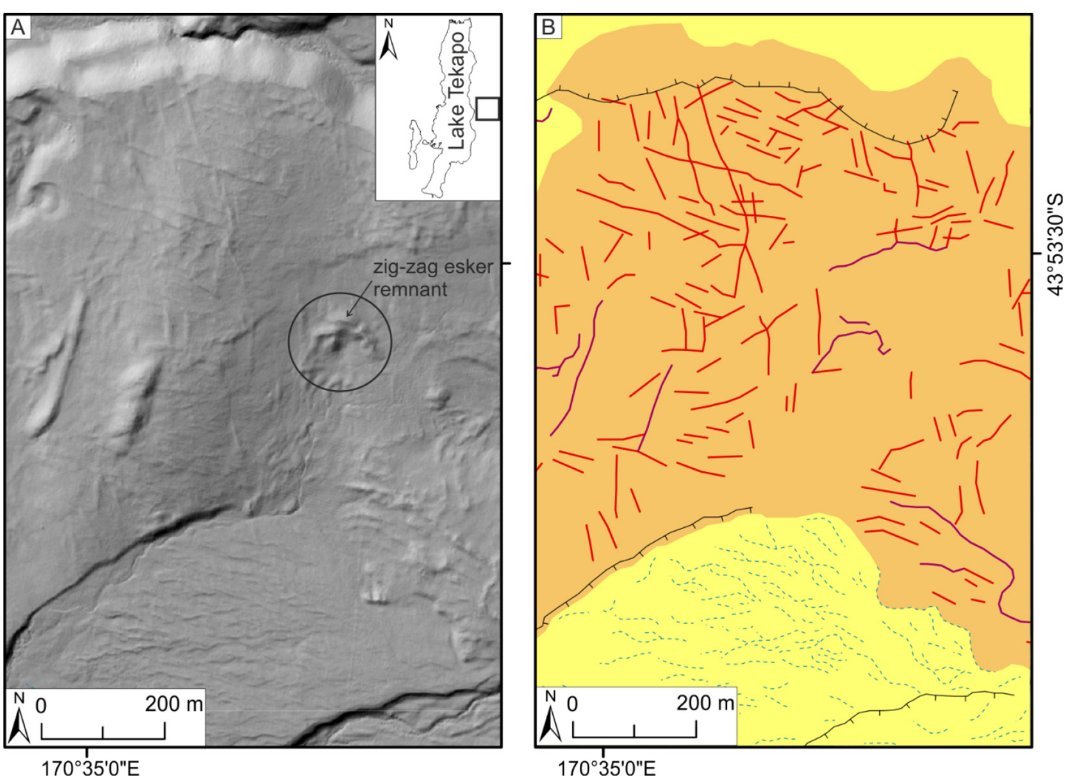

Fig. 6. A. Crevasse-squeeze ridges identified on the eastern valley side of Lake Tekapo from the hillshaded 1 m DEM. Note their cross-cutting relationships. Possible 'zig-zag' esker (Section 4.4.4) remnant circled. B. Mapped interpretation of panel A. Legend shown in Fig. 2.

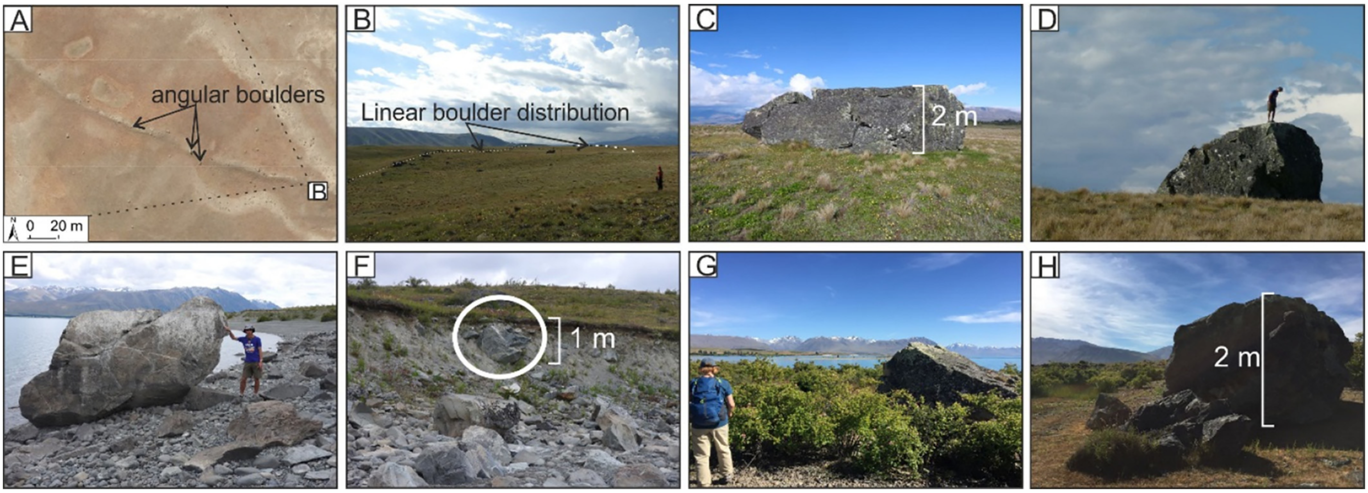

Fig. 7. A. Large angular boulders on push moraine ridge identified from 0.4 m aerial imagery. B. Ground photograph of boulders identified in panel A. Dashed white line indicates their linear distribution. View shown in panel A. Note person for scale. C. Ground photograph of boulder on moraine ridge west of Lake Alexandrina. D. Ground photograph of boulder on moraine ridge west of Lake Alexandrina. Note person for scale. E. Ground photograph of angular boulder on eastern shores of Lake Tekapo. Note person for scale. F. Angular boulder (circled) being exhumed from cliff face. G. Ground photograph of angular boulder on southern shores of Lake Tekapo. Note person for scale. H. Ground photograph of angular boulder on southern shores of Lake Tekapo.

Farnsworth et al., 2016). Darvill et al. (2017) noted the presence of elongated, closely-spaced drumlins and suites of landforms including possible crevasse-squeeze ridges that were are indicative of some re-advances linked to surge activity during the last glacial cycle in Patagonia, a region comparable to the Southern Alps of New Zealand. These and the examples documented around Lake Viedma (Ponce et al., 2019) are the only occurrences of Crevasse-squeeze ridges in Patagonia reported thus far in the literature.

4.2. Ice-marginal landforms

4.2.1. Moraine complexes and deposits

This unit mainly comprises undulating topography within which distinctive ridges occur, but with some larger composite overridden moraines (Section 4.2.3). The western side of Lake Tekapo is characterized by subdued surfaces that comprise areas of low, arcuate changes in relief (>1 km wide) and contain flutings, esker ridges, drumlins and crevasse-squeeze ridges. The surfaces are commonly dissected by meltwater channels and associated outwash deposits. Small lakes/kettle holes are also distributed across the terrain. Although there were no sediment sections exposed in the field, the mapped unit commonly features large scattered, angular boulders (e.g. Fig. 7), often occurring in clusters. We digitized individual boulders identified from 0.4 m aerial imagery (Fig. 2, Fig. 7A) and note a linearity to their distribution along moraine ridges (Section 4.2.2) and large clusters of scattered boulders in the southwest of the study area (Section 4.3.1).

We interpret smooth, gently undulating mounds and depressions as moraine deposits (cf. Krüger, 1994; Evans et al., 1999, 2009a, 2017a, 2018a; Evans and Twigg, 2002; Evans and Orton, 2015). The linear boulder clusters are interpreted as moraine ridges and thereby mark the extent of ice-marginal deposition, in places associated with more substantial push moraines (Section 4.2.2; Fig. 7A). The occurrence of these angular boulders (Fig. 7) at former ice-marginal locations indicates that the Tekapo Glacier efficiently transported large volumes of englacial and supraglacial debris derived from extra-glacial sources in the mountainous catchment, delivering the material as push and dump moraines on the lowlands (e.g. Boulton and Eyles, 1979; Eyles, 1979; Evans et al., 2017b). Given the active geomorphological environment of the Southern Alps in New Zealand, boulder clusters could also be indicative of former supraglacial debris concentrations (such as medial moraines) or, mass movements taking place onto the glacier surface (Shulmeister et al., 2009).

4.2.2. Push moraine ridges

Arcuate, sharp-crested, closely-spaced, typically asymmetrical ridges of sediment are identified in the south and western parts of the study area (Fig. 2). The ridges have cross-profiles with shorter, steeper ice-distal slopes and longer, gently sloping ice-proximal surface slopes (Fig. 8C). They are generally low relief (1 to 2 m high) and range from 15 to 30 m wide at the crest (Fig. 8C). Crest-to-crest (or longitudinal) spacing between individual ridges ranges from 60 m to 440 m. In some areas, most notably west of Lake Alexandrina, the sharp crests appear to have been smoothed by the action of water and in many places the ridge sequences have been partially dissected by glaciofluvial activity (Fig. 8A, B). The ridges are most-often discontinuous, consisting of a number of smaller individual fragments which form part of longer chains. They are most prominent and more continuous in the southwest (Fig. 2). These chains occur as single-crested, cross-valley ridges throughout southern third of the study area. The ridges are separated by areas of faint flutings, and the flutings extend onto the ice-proximal slopes of the ridges in places (Fig. 8A, B).

We interpret these arcuate asymmetrical ridges as recessional push moraines. The geometrical characteristics of the ridges are typical of push moraines formed in modern temperate glacier settings where each moraine ridge represents the thickening of sub-marginal till wedges advected to the ice margin by subglacial deformation (Price, 1970; Evans and Twigg, 2002; Evans and Hiemstra, 2005; Leysinger Vieli and Gudmundsson, 2010; Evans et al., 2018b). Recessional push moraines are a characteristic signature of the active temperate glacial landsystem (Price, 1969; Krüger, 1995; Evans and Twigg, 2002; Chandler et al., 2016a, 2016b, 2016c; Evans et al., 2016, 2017a, 2017b, 2018a). Annual ice-marginal fluctuations are manifest in the form of annual (push/squeeze) moraines. They are often linked to annual re-advances through seasonally-driven ice-marginal processes (Sharp, 1984; Boulton, 1986; Krüger, 1995; Bradwell, 2004; Beedle et al., 2009; Chandler et al., 2016a, 2016b, 2016c). Annual moraine formation occurs at the ice-margin during a period when forward ice-front movement exceeds the negligible ablation during the winter (Boulton, 1986; Lukas, 2012; Bradwell et al., 2013). However, unlike those in the adjacent Pukaki Valley (Evans et al., 2013), the push moraines around

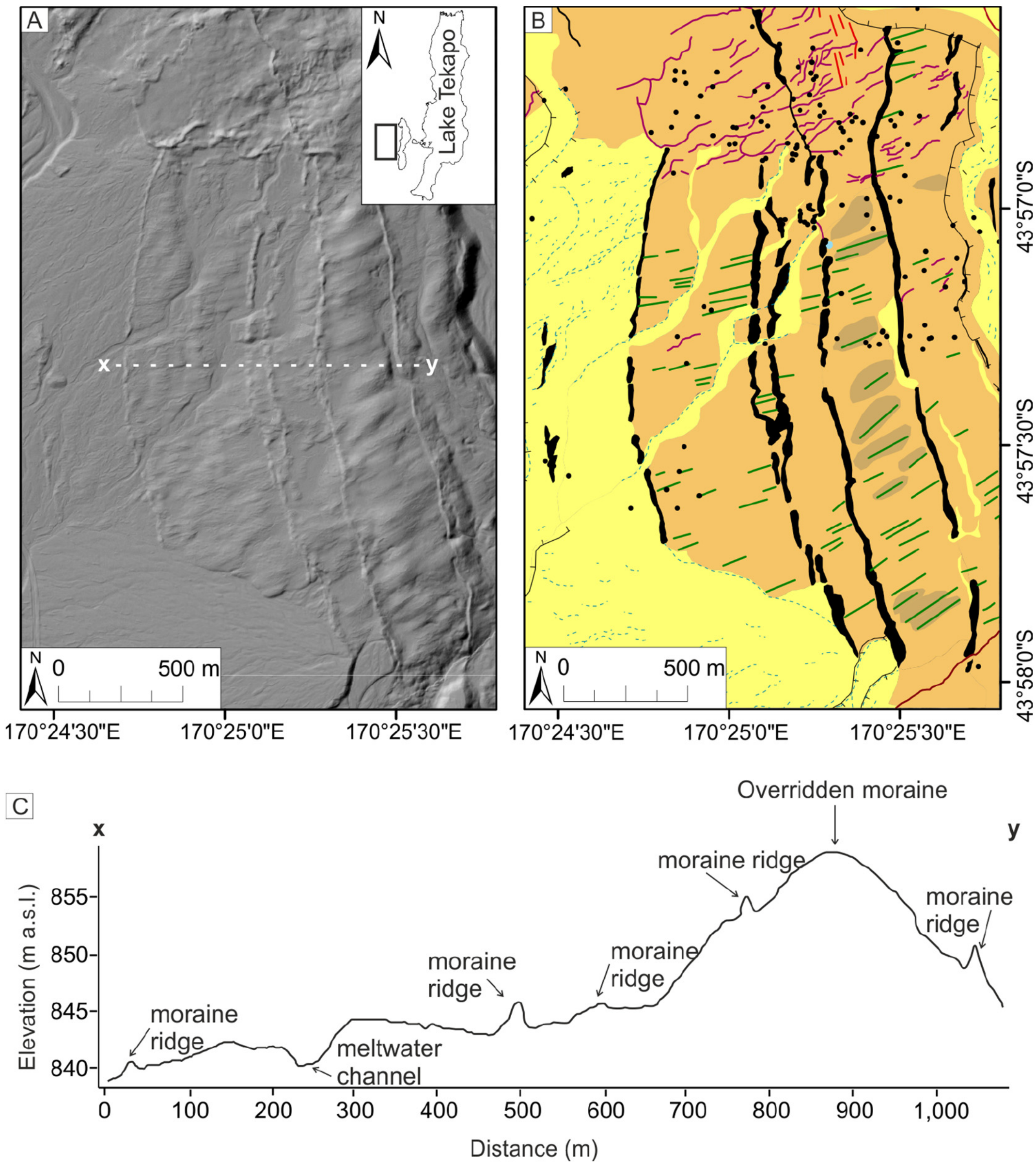

Fig. 8. A. Series of inset frontal push moraine ridges immediately west of Lake Alexandrina identified from hillshaded 1 m DEM. Note glaciofluvial corridors between moraines ridges and the overridden moraine. B. Mapped interpretation of panel A. Legend is as given in Fig. 2. Note small-scale drumlins mapped. C. Cross-profile from 1 m DEM showing five distinct moraine ridges ~2 m high. Note also the overridden end moraine (labelled). Transect from x to y shown in panel A.

Lake Tekapo do not display the distinctive crenulated or 'sawtooth' planform geometries suggesting that the terminus shape was much more regular and not as heavily longitudinally crevassed as those described in relation to 'sawtooth' push moraines (e.g. Matthews et al., 1979).

The ridges therefore demarcate the limits of the former Tekapo Glacier terminus and provide the spatial extent and pattern of the receding ice margin. The regional distribution of these moraine ridges reflects a general pattern of northwards glacier retreat but also of ice-margin recession that was initially upslope but then down an adverse slope (e.g. Fig. 8C). The arcuate ridges are nestled around the main depression of Lake Tekapo, marking the point at which the glacier was flowing up the adverse slopes of the overdeepening onto higher relief areas (Barr and Lovell, 2014) and the existence of the former piedmont glacier lobe. There is, however, no terminal moraine to mark the maximum position of this lobe. The concentric arcs of moraines, where innermost ridges are lower in absolute elevation that outer ridges, reflects downwasting of the ice surface and thus thinning of the Tekapo Glacier. That thinning was more pronounced than the retreat of the terminus position up-valley, at least in the early stages of deglaciation when the ice was still filling the southern part of the Tekapo basin.

4.2.3. Overridden end moraines

This landform occurs in long 'belts' of minor hills, often drumlinised and their length can be traced for several kilometers (Fig. 2). Their broad crestlines range from 180 m to 550 m wide and trend parallel to the push moraine ridges, perpendicular to the ice-flow direction. They are high-amplitude features, with a steep proximal face, rising to ~50 m above the surrounding terrain, which is extensively adorned with attenuated drumlins and flutings. They have a hummocky surface topography that is superimposed by lower-amplitude push moraine ridges and flutings. These features occur at the southern outlet of Lake Tekapo and west of Lake Alexandrina (Fig. 9A, B), intimately linked with the hummocky moraine complex (Section 4.3.1).

We interpret these high-amplitude belts as large composite overridden end moraine ridges that parallel former ice-margins, similar to

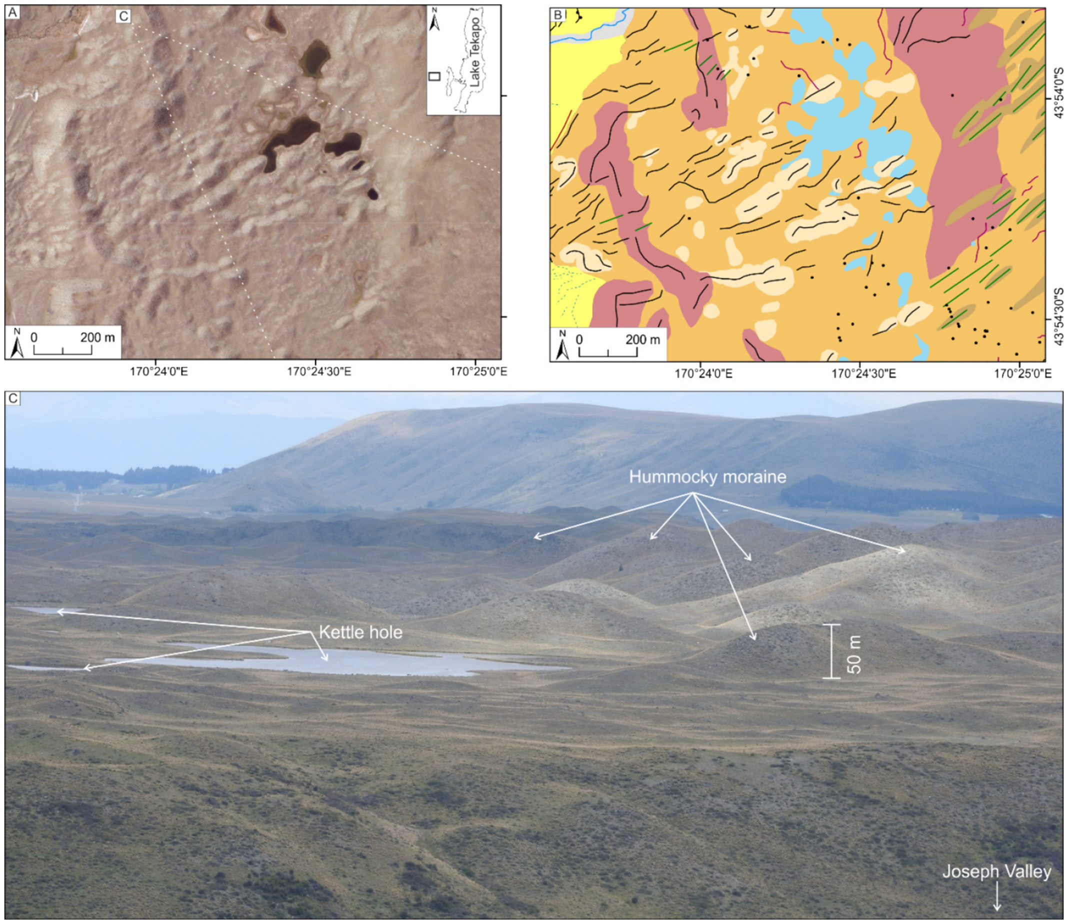

Fig. 9. A. Discontinuous hummocky ridges northwest of Lake Alexandrina identified from 0.4 m aerial imagery. B. Mapped interpretation of panel A. Legend is as given in Fig. 2. C. Ground photograph of hummocky moraine ridges and kettle holes. View shown in panel A.

those at Bruarjokull, Iceland, described by Kjær et al. (2008) especially as they occur in association with patches of hummocky dead-ice terrain (Section 4.3.1). Their superimposition by fluted subglacial till, as well as their surface topography and cross-cutting relationship with the hummocky moraine, lead us to interpret them as having been overridden by glacier ice. This interpretation of ice-overriding is supported by sedimentological evidence of glaciotectonized lacustrine sediments (Section 4.5.3; Sutherland et al., 2019) which serve as another example of overridden material. Overridden moraines are indicative of seasonal oscillations of a warm-bedded piedmont lobe with a subglacial deforming layer (Price, 1970; Boulton, 1986; Evans and Twigg, 2002) which is associated with an active temperate landsystem. The flutings and drumlins have been constructed over the hummocky topography and appear fragmented in the imagery. The fragments exhibit a weak barchanoid form, suggestive of overriding by active ice. This indicates that the moraine arcs were partially overrun by a re-advance of the western Tekapo Glacier margin before complete melt-out of buried ice.

As displayed in Fig. 8C, the overridden moraines are very prominent features when compared to the younger landforms superimposed on them. This discrepancy suggests that they had been formed during a major previous advance, perhaps from the Balmoral or the Mt. John advances (Table 1). Overridden end moraines may also be intra-zonal remnants of older surges Evans and Rea (2003), such as observed in inner zones of glacier surges into open marine waters and fjords from Svalbard (Ottesen and Dowdeswell, 2006; Ottesen et al., 2008; Ottesen et al., 2017). They are also associated with streamlined landforms, such as those that occur on the western side of Lake Alexandrina. The presence of overridden moraines in our study area indicates that the terrain surface relief results from numerous landform generations. Schomacker et al. (2014) attributed the presence of overridden moraines to the existence of multiple surges of different magnitudes.

4.3. Supraglacial landforms

4.3.1. Hummocky moraine with discontinuous linear ridges

There is a conspicuous landform assemblage in the western part of the map area which trends from the northeast to the southwest as a narrow, 2 km long, arcuate band, lying between Lake Alexandrina and the Joseph Valley (Fig. 9). This zone is defined by accumulations of high-amplitude, closely-spaced, circular to semi-circular rounded hummocks interspersed between long (200 to 500 m) discontinuous linear ridges 5 to 20 m high. The crest lines of these ridges are less well-defined than the push moraine ridges described above (Section 4.2.2), and the boundaries of individual ridges are difficult to delimit. The ridge crests are spaced evenly at ~100 m apart. Towards the southern part of this zone, shorter ridges are connected to form longer, more distinct, undulating chains. This area grades into that of the scattered, angular boulders that occur in clusters (e.g. Fig. 7). The large-scale hummocks are predominantly irregular in-between the linear ridges, giving a chaotic appearance with occasional crude linearity.

This mapped unit is extensively interspersed with numerous small lakes, ponds and hollows, northwest of Lake Alexandrina (Fig. 2). Here, lakes of varying size, depth and degree of inundation, occur in clusters and occupy depressions in the surface of the hummocky topography. The majority of the lakes are rounded, oval, or slightly oblong, with smooth concave profiles, steep sides and without any contemporary surface stream flows. The smallest lakes are <10 m in diameter whereas others reach between 50 and 100 m in diameter. The lakes are orientated along a distinct northeast to southwest trending axis, following the arc formed by the strip of hummocky topography. In the area directly west of Lake Alexandrina, dry lake beds contain piles of large, angular boulders within them (Fig. 7).

The term hummocky moraine is used as a descriptive term for irregular, formerly ice-cored morainic topography with high variation in relief (Benn, 1992; Krüger and Kjær, 2000; Kjær and Krüger, 2001; Schomacker, 2008; Benn and Evans, 2010). This assemblage, restricted to the western edge of the Tekapo basin, could therefore represent a broad moraine complex of formerly ice-cored hummocky terrain where the lakes represent melt-out hollows. We interpret the lakes and dried lake beds to be kettle holes, formed by the burial of large blocks of residual ice which subsequently collapsed as the ice melted. Individual kettle holes up to 100 m in diameter are large enough to suggest that extensive areas of the glacier terminus became buried by glaciofluvial sediment, whereupon large bodies of otherwise actively receding glacier ice were subject to localized stagnation (cf. Price, 1969, 1971; Fleisher, 1986; Evans and Twigg, 2002; Carrivick, 2005; Carrivick and Twigg, 2005; Evans et al., 2009a, 2017a, 2017b, 2018a). These landforms bear a similar resemblance in pattern and dimensions to the continuous hummocky ridges and terrain mapped by Bendle et al. (2017) in central Patagonia, interpreted as marking the former ice marginal zone or zone of stagnant ice.

Hummocky moraine is typically associated with deposition of supraglacial debris (Boulton, 1972; Benn, 1992; Kjær and Krüger, 2001; Schomacker, 2008; Darvill et al., 2017), although the transport pathways of debris resulting in this terrain are highly variable (Evans, 2009, and references therein). We infer that the disorganized nature of the landform assemblage here indicates periods of localized ice stagnation and downwasting during overall recession of the ice lobe, leaving behind areas of buried ice and resulting in topographic inversion of the terrain (Clayton, 1964; Boulton, 1972; Etzelmüller et al., 1996; Kjær and Krüger, 2001; Schomacker, 2008). More specifically, organization of hummocky terrain in arcuate bands and the occurrence of minor transverse moraine ridges has been interpreted as a product of incremental stagnation (sensu Eyles, 1979; Bennett and Evans, 2012) in other lowland settings (e.g. Attig et al., 1989; Ham and Attig, 1996; Clayton et al., 2001; Dyke and Evans, 2003; Evans et al., 2014). A similar process-form regime could also apply in the Tekapo Valley if the glacier was episodically carrying large englacial and supraglacial debris loads (Eyles, 1979; Kirkbride and Warren, 1999; Kirkbride, 2000; Deline, 2005; Bennett and Evans, 2012; Darvill et al., 2017).

4.4. Glaciofluvial landforms

4.4.1. Meltwater channels

Straight, sinuous or meandering channels that begin and end abruptly are abundant across the Tekapo basin and vary in size. Many channel lengths reach >5 km and channel widths range from 20 m to 300 m. None of them contain any contemporary surface water drainage. The channels are deeply incised into the surrounding terrain and also cut through moraine crests by ~ 20 m to 100 m. There are a series of channels aligned parallel to each other on the western side of Lake Alexandrina where they trend obliquely across local slopes. The larger channels all have sharply defined terrace edges and are fed by smaller tributaries. There is also an extensive network of longer, but shallower channels towards the southwest of the study area that dissect the hummocky terrain at right angles. Palaeo-channels have been mapped as linear features plotted long their centerline.

All of these channels are considered to be of glacial meltwater origin and we interpret the incised channels as products of glaciofluvial erosion (Dyke, 1993; Greenwood et al., 2007). The glacier hydrological system experiences constant change due to variations in glacier volume and position and hence the meltwater channels consequently changed their course accordingly. They have since adjusted to the contemporary hydrological drainage system in the Tekapo Valley. Glaciofluvial erosion is manifest in a variety of lateral and proglacial marginal meltwater channels at a range of scales throughout the study area and thereby record the recession of ice from upland surfaces (cf. Dyke, 1993; Greenwood et al., 2007; Syverson and Mickelson, 2009; Livingstone et al., 2010; Darvill et al., 2014). They probably record the early stages of glacier marginal drainage before the main linear outwash corridors became established. Proglacial meltwater channels also would have issued from frontal moraine systems and are orientated parallel to the former ice margin. Meltwater channels are abundant across the valley floor, typically flowing between and accentuating moraine ridges. However, the moraines have been cut through in several places as the drainage reworked downslope towards the Tekapo River, resulting in the moraines becoming increasingly intersected and eroded towards the unconfined outwash plains as a result of glaciofluvial erosion.

Additionally, four major spillways have been identified, west of Lake Alexandrina. These spillways have head cuts that correspond in elevation to the palaeo-shorelines in the Joseph Valley (Section 4.5.1). The spillways lie sub-parallel to the transverse moraine ridges, conforming to the lobate planform displayed by the ice-marginal landform record. They extend across the majority of the moraine sequence, reaching up to 500 m wide and 60 m deep. Here, a clear northwest-southeast direction of meltwater flow is evident.

4.4.2. Ice-contact outwash plains and corridors

Large, relatively flat areas without major topographic interruptions are identified in distal locations to prominent moraine crests. The southern extent of the Tekapo Valley is dominated by such plains, many of which are distinctly fan-shaped. They form expansive deposits composed of sub-angular to rounded sands, gravels and cobbles. The plains have a smooth appearance from a distance but are imprinted with complex abandoned braided channel networks and many exhibit distinctive terrace levels. Prominent terraces and breaks of slope have been mapped on to the outwash surfaces where they are present (Fig. 2). Their surfaces grade gently downslope (c. 5°) from former ice margins, and can be traced tens of kilometers down-valley. These glaciofluvial deposits have been interpreted as outwash plains and fans that record the evolution of the proglacial drainage system, reflecting changes in glacier margin position, ice-marginal topography, and meltwater discharge (cf. Price, 1969; Thompson and Jones, 1986; Marren, 2005). The outwash plains are closely associated with moraine ridges and complexes and are useful in determining former ice-frontal positions. The outwash plains and fans either prograde from the frontal moraine complexes to form extensive outwash plains, where meltwater flow was unimpeded by moraine ridges and the fans were able to coalesce, or they occur within ice-margin parallel corridors, topographically controlled by the arcuate moraine complexes (cf. Price, 1969; Price and Howarth, 1970; Krüger, 1994; Evans and Twigg, 2002; Marren, 2002; Evans et al., 2009a, b, 2016, 2017a, b, 2018a; Evans and Orton, 2015).

Extensive outwash plains emanating from the downstream end of Lake Tekapo provide evidence of a large proglacial aggradational fan at the toe of the former Tekapo Glacier. The glacier terminus appears to have been buried by the ice-contact outwash head because no terminal moraine is evident (cf. Kirkbride, 2000; Benn et al., 2003). Additionally, we note the subdued nature of the moraines, where they are volumetrically small in comparison to their outwash fans (cf. Alexander et al., 2014). Minor push moraines that occur on the terraced ice-contact face of the outwash head document contemporaneous glaciofluvial sediment aggradation and push moraine construction in sand and gravel. At present, the Tasman Glacier is the closest modern analogue to the LGM Tekapo Glacier because there, a large outwash head similarly indicates prolonged proglacial aggradation distal to a relatively stable terminus position (Kirkbride and Warren, 1999; Kirkbride, 2000). This outwash head damming process has been described in numerous locations across New Zealand, such as on the West Coast of South Island where the deposits are dominated by outwash processes (e.g. Evans

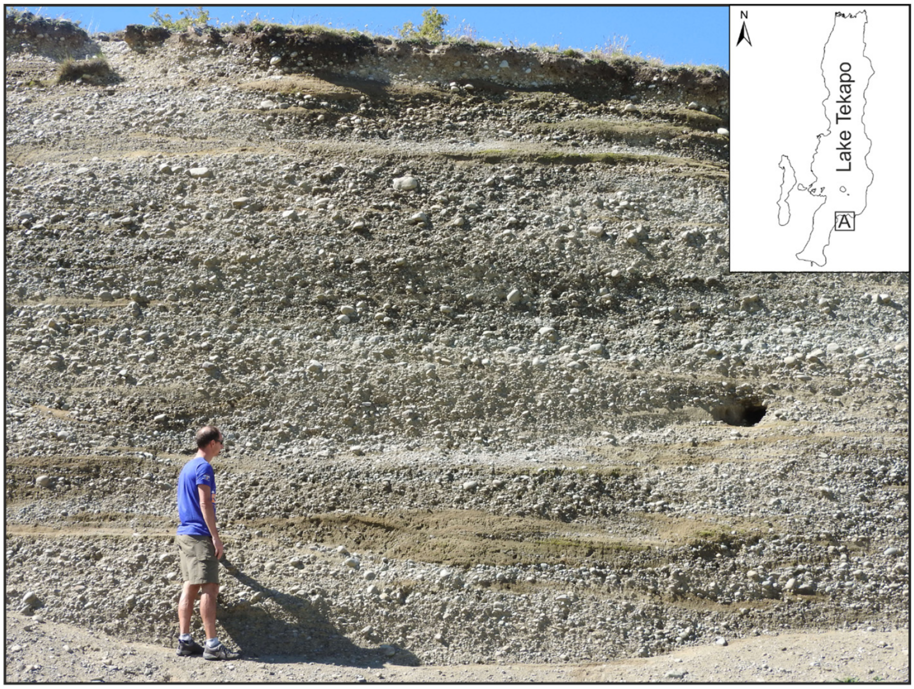

Fig. 10. A Sediment exposure displaying cross-stratified, horizontal to planar-bedded cobble to pebble gravels and trough to planar bedded sands with gravel lags.

et al., 2010c; Shulmeister, 2017), and are also similar to that reported in modern Icelandic glacier forelands (Evans and Orton, 2015).

A 10 m high sediment exposure through outwash deposits exists along Lilybank Road (Fig. 10; 43.96602 S 170.51950 E). The exposure consists of coarse, clast-supported, very well-rounded cobbles and pebbles suggesting a predominantly high-energy source. The sediments are arranged in horizontal beds with lenses of fine sand indicative of variable discharges. We interpret these sediments to have been glaciofluvially deposited as gravel sheets with waning discharge sandy bedforms in a wide outwash channel characterized by longitudinal and bank-attached bars (Boothroyd and Ashley, 1975; Miall, 1977a, 1977b).

Immediately south of Lake Alexandrina, the apex of the outwash head is locally pitted (Fig. 11). The pitted, or collapsed surfaces are manifest as small, near circular pits, approximately 5–10 m wide. The pits are concentrated within 250 m south of the outwash head, yet they extend laterally along the ridge for >1 km. The pits decrease in size down valley and occur in trains. In the field, the pits can be seen to be delineated not only by subtle topography but also by a change in vegetation, which probably indicates a change in sedimentology that is characteristic of iceberg melt-out hollows and obstacle marks as identified in modern jökulhlaup-fed sandar settings (Russell, 1993; Fay, 2002; Russell et al., 2005a, 2005b). The occurrence of pits that decrease in size and are organized into trains at the ice-contact face of an outwash head indicates that jökulhlaup discharges operated around the glacier snout (cf. Russell et al., 2005a, 2005b, 2006; Waller et al., 2018).

4.4.3. Push moraines developed in ice-contact glaciofluvial deposits

Long and narrow alternating sequences of flat-topped glaciofluvial surfaces and parallel ridges (Fig. 12) occur at perched elevations on the eastern side of the Tekapo basin (Fig. 2), limited only to this side of the valley. The zone within which these parallel ridges and glacifluvial surfaces occur is widest towards the north where it has an average width of approximately 5 km. The flat-topped surfaces range from a few metres wide to >100 m wide and have a gentle slope (typically <5°) that trends from north to south. The sequences are dissected by palaeo-subglacial meltwater channels that run from east to west.

We interpret these glaciofluvial deposits on the eastern side of the Tekapo basin to have been focused within extensive flights of kame terraces, where sediment-laden meltwater was constrained between the valley wall and the lateral margin of the glacier (Gage and Suggate, 1958; Soons, 1963; Bitinas et al., 2004; Benn and Evans, 2014). Hence, kame terraces reflect the position of the ice margin and are therefore ideal for palaeo-glaciological reconstructions (Fredin and Hättestrand, 2002). Kame terraces within the formerly glaciated valleys of New Zealand are common and their preservation is spectacular (e.g. Borsellino et al., 2017; Shulmeister et al., 2018). Where glacier marginal oscillations have impinged on the kame terrace edges, the glaciofluvial deposits have been locally reworked into small push moraines. Therefore, we map the alternating sequences of parallel ridges as areas of push moraines developed in ice-contact glaciofluvial deposits. Push moraines have a particular mode of formation and consist in almost all cases of the pre-existing (e.g. proglacial) sediment that the glacier is overriding. Good modern analogues are the moraines formed recently at Franz Josef Glacier and Fox Glacier on the West Coast of New Zealand, where up to c. 2000 CE, and again between 2005 and 2009, the glaciers each experienced short but strong terminus advances. Their terminal moraines consist mainly of glaciofluvial sediment (Carrivick and Rushmer, 2009) simply because they overrode their valley outwash. It is most likely that the Tekapo Glacier similarly experienced short but strong minor advances during the LGM. Likewise, in a comparable setting to Lake Tekapo at the LGM, sediments within push moraines of Gornergletscher, an alpine valley glacier of the Swiss Alps, are composed of proglacial outwash since outwash fans formed prior to the piedmont glacier advance as well as during their retreat (Lukas, 2012).

4.4.4. Eskers

Prominent ridges that lie within areas of glaciofluvial sediment are differentiated from push moraine ridges due to their highly sinuous planform. These ridges range from ~100 m to >1 km in length (Fig. 13) and some have a 'zig-zag' shaped appearance (Figs. 6A, 14A, B; cf. Knudsen, 1995; Evans and Rea, 1999, 2003; Evans et al., 2007). They are often discontinuous, but fragments aligned in the same orientation can be traced in chains as single or bifurcating ridges for many

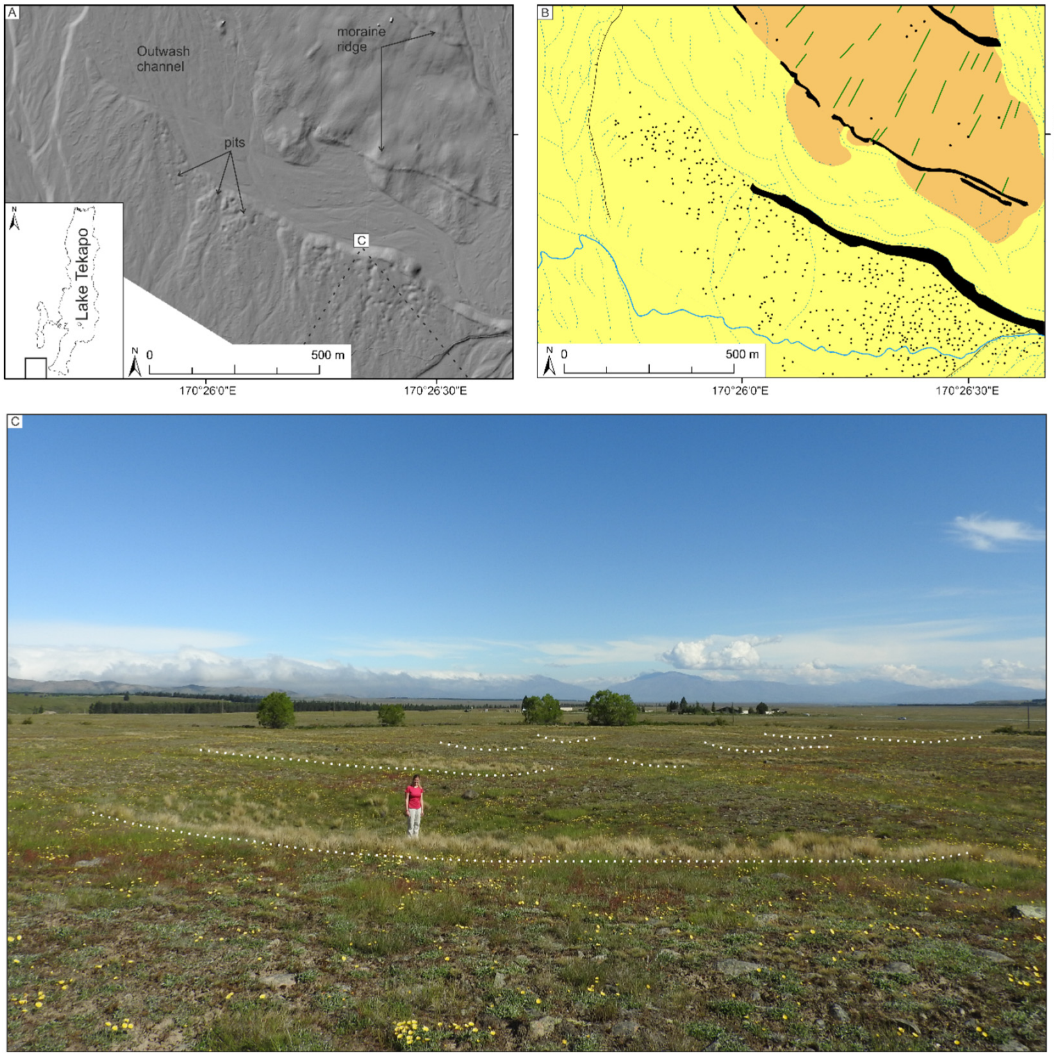

Fig. 11. A. Pitted outwash plain along the margin of the outwash fan-head identified from hillshaded 1 m DEM. B. Mapped interpretation of panel A. Legend shown in Fig. 2. C. Ground photograph of isolated pits, delineated by change in vegetation view shown in panel A. Note person for scale standing in pit.

kilometers. The more prominent ridges are long with a high sinuosity; they are typically 5 to 15 m wide and rise to 10 m in height above the surrounding terrain. They have an oblique orientation relative to former ice-flow direction. The majority occur as isolated ridges with few or no tributaries (e.g. Fig. 13E, F, H), whilst others are arranged in a dendritic pattern with many of the constituent ridges forming a dense network of short (subtle) segments (e.g. Fig. 13A, B, C, D). Their crestlines are smooth, rounded and undulating. The majority of ridges trend northeast-southwest. On the western wide of Lake Tekapo, their crestlines trace into kettle holes and also into meltwater channels. Whilst no sedimentary sections within these ridges could be found during our fieldwork, surficial sediments consisted of very well-rounded cobbles and gravels (Fig. 13F, H).

These sinuous ridges composed of glaciofluvial deposits are interpreted as eskers deposited in meltwater conduits within or beneath glacier ice via ice-walled and ice-roofed channels (e.g. Warren and Ashley, 1994; Carrivick and Russell, 2007; Storrar et al., 2014a, 2014b, 2015). Eskers represent significant subglacial/englacial drainage pathways and so glacier recession reveals the courses of these pathways in the form of esker ridges and networks. Fragments of eskers can be joined together so that they collectively represent the spatial and temporal development of the same meltwater tunnel during glacier recession. Eskers are highly unstable and ephemeral as landform elements (Price, 1966, 1969; Howarth, 1971; Storrar et al., 2015) and hence their traces gradually become fragmented by the progression of ice melt-out to form semi-continuous ridges. Eskers could therefore also be mistaken for isolated kame mounds, especially on the eastern valley side of the Tekapo basin. Their orientation indicates a southwesterly subglacial drainage pattern. Given that the eskers are draped over other subglacial landforms such as flutes and crevasse-squeeze ridges, and that some feed into kettle holes, it is likely that they have melted out from englacial tunnels. Well-preserved 'zig-zag' eskers occur at several locations in the Tekapo Valley (Figs. 6; 14A, B), which is significant because they have been linked to crevasse infilling by heavily sediment-laden high meltwater discharges during glacier surges (Evans and Rea, 1999, 2003; Evans et al., 2007). A zig-zag esker is ice-cored when first formed and hence they eventually melt out to form discontinuous mounds of gravel or short fragments.

4.5. Glaciolacustrine landforms

4.5.1. Palaeo-lake Shorelines

Distinct continuous or near-continuous linear features with a break of slope at a uniform altitude that closely parallel the current active shorelines of Lake Alexandrina, Lake McGregor and Lake Tekapo have been previously mapped by Sutherland et al. (2019). The best examples occur around Lake Alexandrina where a prominent stack of three almost continuous terraces have been mapped at elevations of 721 m, 726 m,

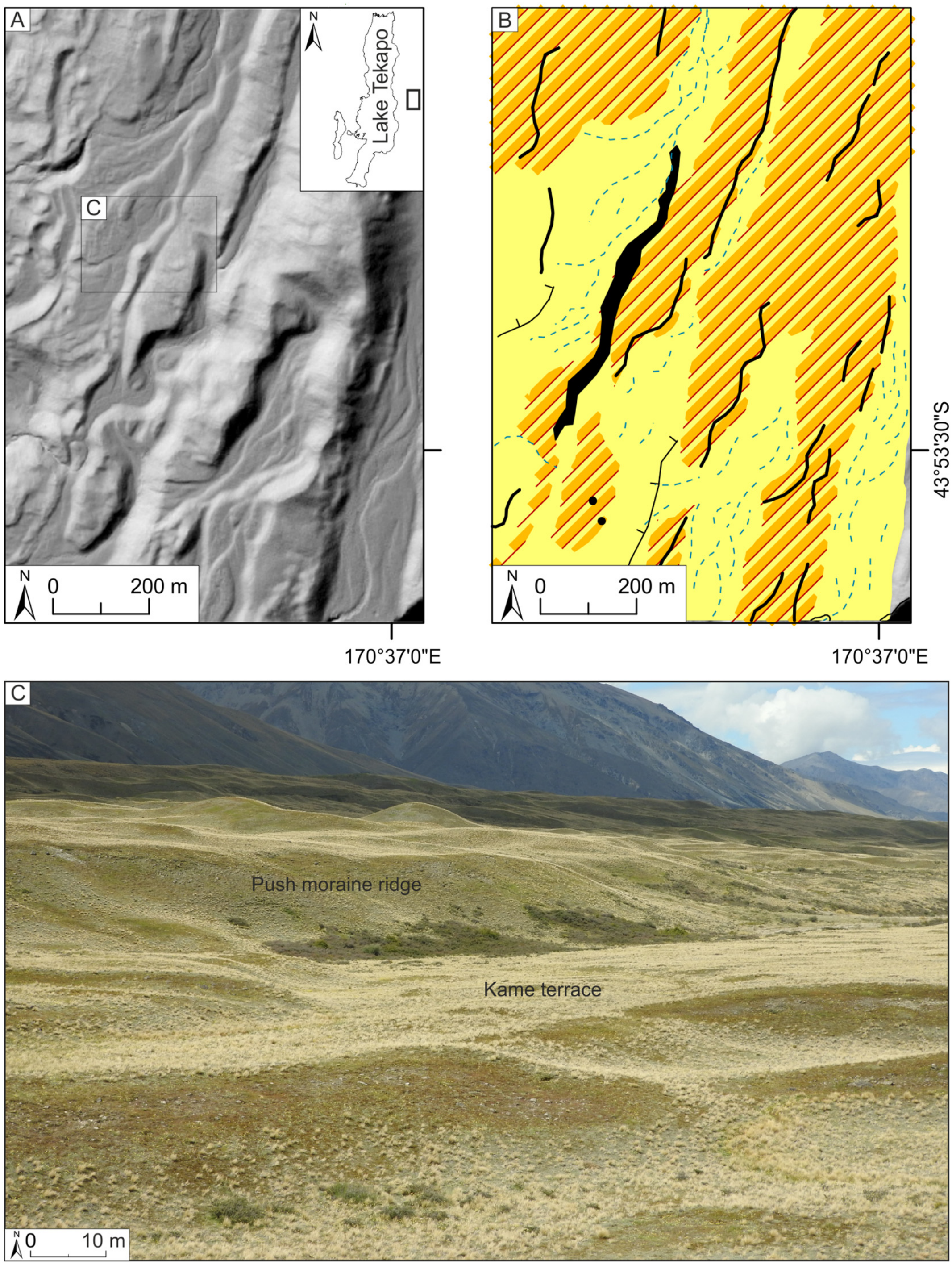

Fig. 12. A. Inset series of kame terraces with push moraines on their ice proximal edges. B. Mapped interpretation of panel A. Legend is as given in Fig. 2. C. Ground photograph of kame terrace and push moraine sequences looking towards the southeast.

and 740 m a.s.l respectively (Fig. 15A, B, C). These former shorelines are closely nested within each other around Lake Alexandrina but other examples are found on the eastern side of Lake Tekapo. Some are located as much as 350 m away from the present-day shorelines. From field observations, they are bare of vegetation and are eroded into unconsolidated sediments (interpreted herein as glacial diamict). Three of these terraces were also observed in the field and from aerial imagery on both sides of the Joseph Valley at 820 m, 830 m, and 848 m a.s.l (Fig. 15D) but are very subtle and difficult to identify from the DEM. They are not laterally extensive, with a maximum length of ~200 m.

These terraces are interpreted as wave-cut palaeo-lake shorelines. The highest 740 m shoreline around Lake Tekapo can be traced continuously from a topset-foreset junction of a relict delta on the western side of Lake Alexandrina (Fig. 16). Former lakes are delimited by a series of abandoned shorelines surrounding the present day lakes Alexandrina, McGregor, and Tekapo. The stack of shorelines evidences progressively lower lake levels, and those were likely the product of Lake Tekapo water level lowering as its outlet was incised. The occurrence of localized palaeo-lacustrine evidence leads us to infer that these shorelines related to a once-larger proglacial Lake Tekapo that existed at various times in the over-deepenings in front of the former ice lobe (Sutherland et al., 2019). LGM lake volume was not calculated because it is difficult to constrain both the location and thickness of the ice margin during lake expansion. Lake levels will also have varied

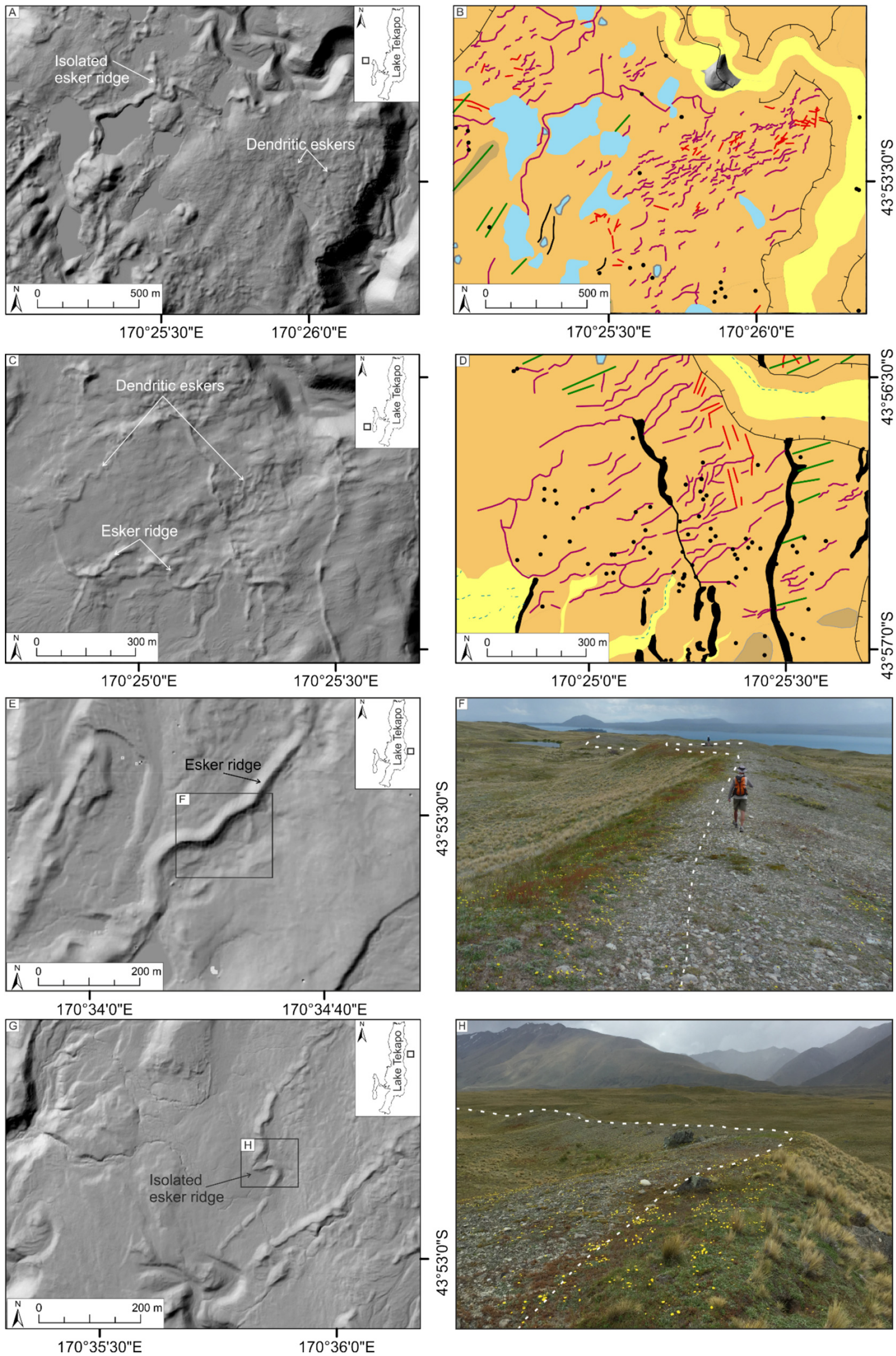

Fig. 13. A. Esker ridge between kettle holes and dendritic esker network situated northeast of Lake Alexandrina, identified from hillshaded 1 m DEM. B. Mapped interpretation of panel A. Legend is as given in Fig. 2. C. Dendritic esker networks situated west of Lake Alexandrina, identified from hillshaded 1 m DEM. D. Mapped interpretation of panel C. Legend is as given in Fig. 2. E. Isolated esker ridge on the eastern side of Lake Tekapo, identified from hillshaded 1 m DEM. F. Ground photograph of esker ridge in panel E. Note person for scale. Crestline is shown as white dashed line. G. Isolated esker ridge on the northeastern side of Lake Tekapo, identified from hillshaded 1 m DEM. H. Ground photograph of esker ridge in panel G. Crestline is shown as white dashed line.

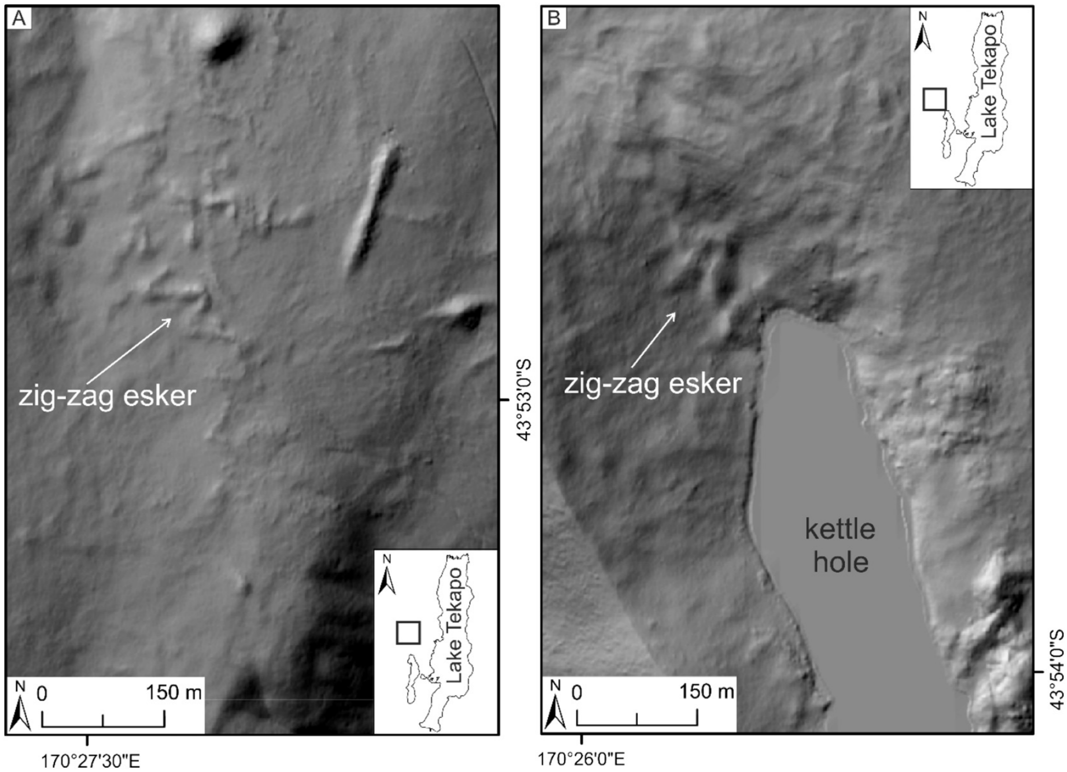

Fig. 14. A. 'Zig-zag' esker network manifest as a discontinuous chain of ridge fragments. Note the adjacent flutings and crevasse-squeeze ridges onto which the esker has been draped B. 'Zig-zag' esker ridge feeding into kettle hole on northwest side of Lake Alexandrina.

through time according to the position of the glacier snout in the valley. The ice-marginal truncation of the mapped shorelines suggests that they formed in front of the ice lobe (ice-contact), probably during recession into their respective topographic basin/depositional overdeepening impounded by moraine and outwash fan-head deposits (Sutherland et al., 2019).

The shorelines identified in the Joseph Valley do not lie within the extent of the former enlarged Lake Tekapo and so we relate these

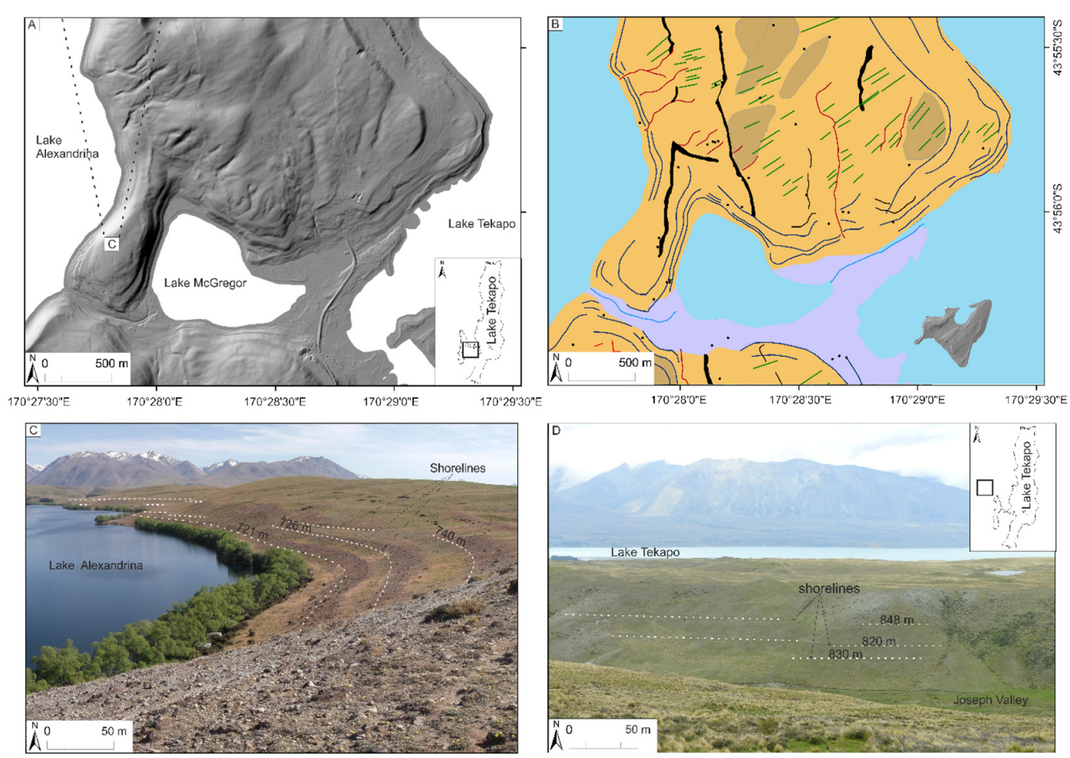

Fig. 15. A. Series of three, relatively continuous palaeo-shorelines between Lakes Alexandrina, McGregor and Tekapo (at 721 m, 726 m, and 740 m a.s.l respectively) identified from hillshaded 1 m DEM. B. Mapped interpretation of panel A. Legend is as given in Fig. 2. C. Ground photograph of palaeo-shorelines on eastern margin of Lake Alexandrina. View shown in panel A. Elevations of palaeo-lake levels labelled D. Ground photograph of palaeo-shorelines in the Joseph Valley. Elevations of palaeo-lake levels labelled.

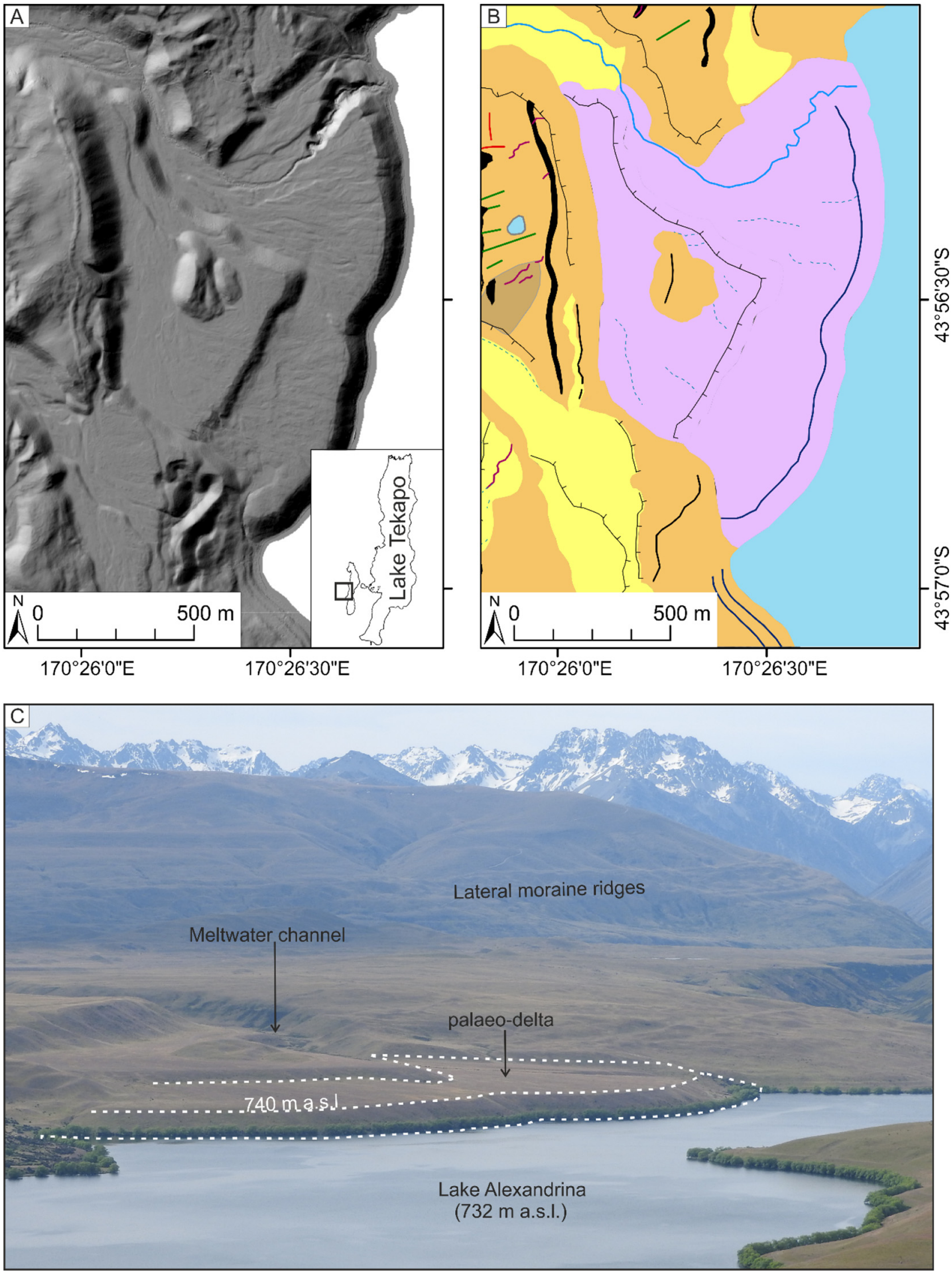

Fig. 16. A. Palaeo-delta on the central western margin of Lake Alexandrina identified from hillshaded 1 m DEM. B. Mapped interpretation of panel A. Legend is as given in Fig. 2. C. Ground photograph of palaeo-delta identified from panels A and B.

shorelines to an isolated ice-contact proglacial lake in the Joseph Valley, formed upon retreat of the tributary ice lobe that issued from the Cass Valley. The shoreline elevations have been linked to meltwater spillways (Section 4.4.1).

4.5.2. Raised deltas

Low gradient (<5°) sediment accumulations with deeply incised lake-facing scarp slopes are formed at the mouths of tributary streams and perched above the western margins of Lake Alexandrina at 770 m and 740 m a.s.l. (Fig. 16; Sutherland et al., 2019). We note a marked absence of these features around Lake Tekapo.

They are interpreted as relict lacustrine Gilbert-type deltas (e.g. Ashley, 2002; Slaymaker, 2011; Dietrich et al., 2017), which formed when proglacial meltwater streams entered a lake after draining from a retreating ice margin. Narrow wave-cut shorelines are present on some of the delta fronts and their incised, telescoping morphology provides evidence of lake water level lowering in several steps. Due to the spatially restricted location of the deltas on the western side of Lake Alexandrina, we interpret them to have formed via the meltwater

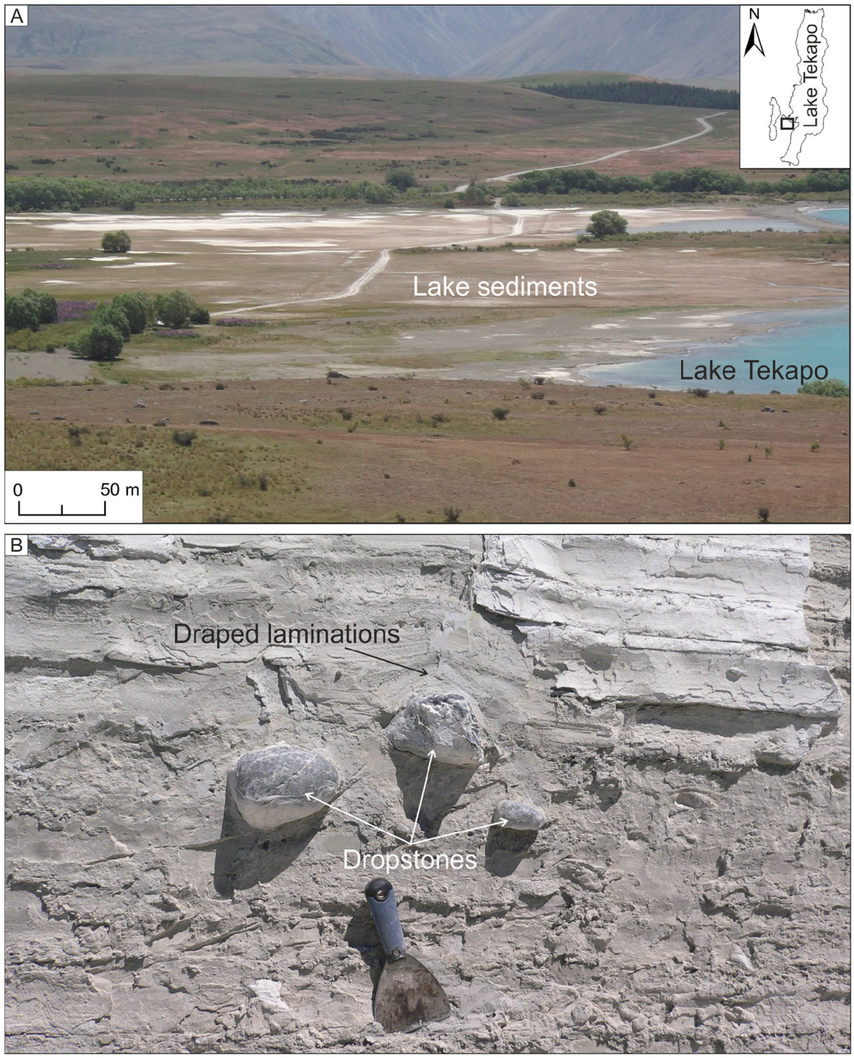

Fig. 17. A. Ground photograph of relict lake sediments in between Lake Tekapo and Lake Alexandrina. B. Dropstones with impact structures and draped laminations in varved lacustrine sediments exposed on western cliffs of Lake Tekapo.

spillways from the Joseph Valley, suggesting a lake in the Joseph Valley and an enlarged Lake Tekapo were contemporaneous.

4.5.3. Glaciolacustrine deposits

Broad, flat accumulations of fine-grained glaciolacustrine sediment are present in between Lake Alexandrina and Lake Tekapo, south of Lake McGregor (Figs. 2, 17A). The terrain has a distinctive white colouration on aerial imagery and appears flat on the hillshaded DEM, distinct from adjacent deposits (e.g. moraine). We interpret this landform to have been deposited in a former lake embayment. This is consistent with the presence of relict lacustrine sediments exposed along the southwestern shores of Lake Tekapo (43°56′54″ S 170°30′01″ E; Upton and Osterberg, 2007; Sutherland et al., 2019). The sediments are exposed above the present lake level where the cliffs are ~20 m high and they extend laterally for >2 km. Pickrill and Irwin (1983) first noted the presence of varves (rhythmites) that were later interpreted by Upton and Osterberg (2007) to have been deposited soon after the last glaciation when lake levels were considerably high than they are today. Field investigations confirms the presence of dropstones (Fig. 17B), which indicates the presence of calving icebergs and hence the lake was in contact with the ice margin. Internal structures comprise glaciotectonised outwash and lacustrine sediments (Sutherland et al., 2019). Similar deposits of varved glacial silts have been noted in the Rakaia valley (Speight, 1926). The glaciolacustrine deposits and former shorelines bear a resemblance to those mapped around the present day fjords and proglacial lakes of the Cordillera Darwin Icefield by Izagirre et al. (2018).

5. Discussion: glacial landsystems model

We identify two landsystem signatures in the Tekapo Valley. Together, the landforms and their assemblages described above record a transition in glacier character and behaviour from that of an active temperate piedmont lobe to that of intermittent surge-type activity, particularly by the western margin of the Tekapo Glacier. The identification of these two landsystems is based upon: (i) fluted till surfaces and associated low-relief push moraines, as well as ice-push modified ice-contact outwash (kame terraces), that record active temperate recession (e.g. Evans and Orton, 2015; Chandler et al., 2016a, 2016b, 2016c; Evans et al., 2016, 2017a, 2017b, 2018a; Darvill et al., 2017) with localized ice-stagnation or incremental stagnation (Price, 1969; Evans and Twigg, 2002; Bennett and Evans, 2012) and (ii) landforms that are diagnostic of occasional surging glacier activity, specifically crevasse-squeeze ridges and ‘zig-zag’ eskers (Evans and Rea, 1999, 2003), which are superimposed on, and inset within, the geomorphology of the more widespread active temperate landsystem.

5.1. Active temperate landsystems signature

The landform assemblages that border Lake Tekapo are consistent with those created by an active temperate glacier lobe emerging from mountainous terrain (Evans and Twigg, 2002; Bennett and Evans, 2012). The arcuate assemblage of inset minor push moraine ridges and associated flutings and drumlins in the Tekapo Valley, in addition to locally ice-pushed kame terraces and linear outwash controlled by overridden moraine arcs, are all diagnostic of an active temperate landsystem (Evans, 2003; Evans et al., 2009a, 2017a, 2017b, 2018b). The assemblages of small recessional push moraines adorned with subdued flutings are indicative of seasonal oscillations of a warm-bedded piedmont lobe with a subglacial deforming layer (Price, 1970; Boulton, 1986; Evans and Twigg, 2002; Evans, 2018). Flutings and ice-marginal wedges of subglacial tills (that constitute push moraine ridges) document warm-based subglacial traction processes (Boulton, 1986; Evans and Twigg, 2002; Evans and Hiemstra, 2005; Evans et al., 2010a, 2010b, 2018b; Evans, 2018). Eskers are also well developed in active temperate glacial systems. Given the fact that glaciofluvial process-form regimes dominate most glacier forelands in New Zealand (Shulmeister et al., 2010a, 2010b; Evans et al., 2010a, 2010b, 2010c, 2013), it might be reasonable to suspect that eskers have been formed elsewhere in the Southern Alps but have hitherto not been recognised due to the lack of high-resolution imagery similar to that employed here. Deep incisions through moraine ridges are associated with corridors of terraced outwash fans. These products of proglacial meltwater deposits are typical in most glaciated terrains, and they are most common on the forelands of active temperate piedmont glaciers (Price, 1969; Price and Howarth, 1970; Gustavson and Boothroyd, 1987; Evans and Twigg, 2002; Marren, 2002; Evans et al., 2009a, 2018a).

High-relief hummocky moraine has long been interpreted as recording stagnation from the ice margin and deposition from the surface of dead-ice downwasting in situ (Eyles, 1979; Ham and Attig, 1996; Bennett and Evans, 2012). The abundance of meltwater channels and kettle holes in this area is also characteristic of topography formed in a final stagnant ice phase (Eyles, 1979; Bennett and Evans, 2012; Evans et al., 2014). Eyles (1979) first proposed a model of localized incremental stagnation whereby large volumes of debris-covered ice became detached from the retreating glacier snout during active recession, resulting in the formation, stagnation and detachment of terminal ice-cored moraine complexes. This is exemplified in the Tekapo Valley by the closely-spaced hummocks and depressions concentrated in the southwestern part of the study area that are indicative of a stagnant glacial regime, recording periodic switches from transport-dominant to ablation-dominant conditions over time, or incremental stagnation (Bennett and Evans, 2012).

Over recent years, conceptual models have been developed for different types of glacier systems mainly based on studies in Icelandic glaciers. In particular, Iceland is the key location where process-form models are being advanced for modern active temperate glaciers. The similarity between the landform assemblages surrounding Lake Tekapo and those associated with active temperate glaciers supports the assertion that the Tekapo Glacier predominantly operated as an active temperate glacier lobe. The record of systematic, active ice-marginal retreat clearly reflects climatic control of ice recession. This suggests that the ice was sensitive to regional climate variability, with active re-advances during overall retreat of the ice margin. An active temperate landsystem has also been ascribed to the neighbouring valleys surrounding Lake Pukaki (Evans et al., 2013) and Lake Ohau (Webb, 2009).

5.2. Surge signature

Crevasse-squeeze ridges have not previously been identified in other glaciated valleys in New Zealand, possibly due to a lack of high-resolution imagery similar to that employed in this study. Recognition of these crevasse-squeeze ridges, especially within cross-cutting networks, is particularly notable because they are diagnostic of a surging glacier process-form regime (Evans and Rea, 1999, 2003; Evans et al., 2007, 2016; Benediktsson et al., 2015; Farnsworth et al., 2016; Lønne, 2016; Lovell et al., 2018). Glacier surges have not previously been recognised in New Zealand in either ancient or modern landsystems. Other landforms that we have identified in the Tekapo Valley such as long flutes and hummocky moraine have also been linked to glacier surges (cf. Sharp, 1985a, 1985b, 1988; Evans and Rea, 1999, 2003; Evans et al., 2007, 2009b, 2010a, 2010b, 2010c; Schomacker and Kjær, 2007; Benediktsson et al., 2008, 2010; Jónsson et al., 2014; Schomacker et al., 2014). However, these long flutes and this hummocky moraine cannot be regarded independently as diagnostic surge indicators because they can also be produced by non-surging glacier margins, especially where debris supply rates are high, as they are in the central Southern Alps (Shulmeister et al., 2009). The occurrence of crevasse-squeeze ridges that are not demonstrably of marginal crevasse/sawtooth push moraine origin (Evans et al., 2016, 2017a) but instead comprise complex reticulate ridge networks typical of surging glacier termini, strongly suggests that intermittent surge activity punctuated the overall active temperate glacier recession. Thus, the Tekapo Valley is an example of landsystem superimposition. Modern examples of such landsystem superimposition have been reported from several modern Icelandic glacier forelands, the most prescient example being that of Breiðamerkurjökull (Boulton et al., 2001; Evans and Twigg, 2002), where intermittent surging of the glaciers' eastern margin is recorded by the development of thrust moraines, crevasse-squeeze ridges, and long flutings over the more dominant active temperate landsystem imprint. Surges in this case have been historically documented and observed to be of small magnitude and hence the landforms developed as a consequence are largely subtle. It is entirely feasible therefore that the crevasse-squeeze ridge imprint in the Tekapo Valley similarly records minor surge events potentially driven by intermittent phases of increased meltwater storage and discharge.

Related to such meltwater discharges in modern surging glaciers are 'zig-zag' eskers (Knudsen's (1995) "concertina eskers"; cf. Evans and Rea, 1999, 2003), which because of their supraglacial/englacial genesis have extremely low preservation potential, but which unequivocally appear on the former Tekapo Glacier foreland (e.g. Figs. 6, 14A, B). We acknowledge that the preservation of landforms diagnostic of surging, such as 'zig-zag' eskers and crevasse-squeeze ridges is known to be very poor (Rea and Evans, 2011; Evans et al., 2016). This poor preservation potential is due to fluvial re-working of glacial deposits, which in New Zealand is pervasive (Shulmeister et al., 2019). Primary subglacial sediments (i.e. subglacial traction tills) have a very low chance of preservation in this setting, their absence often linked to having been washed out by meltwater incision and outwash development. Yet subglacial landforms, although subtle, are present in abundance in the Tekapo Valley. We argue that these landforms appear so well-preserved at Tekapo (existing for ~18,000 yrs) because of little postglacial modification by fluvial activity, which is partly because the glacier retreated down a retrograde bedslope (Sutherland et al., 2019) thereby allowing meltwater to pond, instead of re-working the landforms and partly because of semi-arid conditions during the Holocene.

The abundance of sinuous eskers is also conspicuous. They probably document later stage drainage by meltwater moving though englacial and subglacial pathways because it is likely that substantial quantities of water were stored behind the surge front and that this water was discharged periodically during and at the termination of the surge. Elevated basal water pressures would have forced the water to drain via englacial or supraglacial drainage systems, which would have rapidly exploited the extensive network of crevasses created during the surge.

Drumlins have been described in association with surging glaciers, although they are not considered as particularly typical of surging (Hart, 1995; Kjær et al., 2008; Waller et al., 2008; Johnson et al., 2010; Schomacker et al., 2014). Jónsson et al. (2014) hypothesize that drumlins, which have only experienced a few surges, are shorter and wide with low relief, whilst those that have been attributed to more surges are longer and narrower with her relief. This hypothesis could account for the wide differences in morphology observed in the drumlinised terrain when comparing the eastern and western sides of Lake Alexandrina (e.g. Fig. 3A, C), although drumlin evolution under an active temperate regime is more likely, especially as the landsystem signature in the Tekapo Valley is predominantly of that type (cf. Boulton, 1987).

Certainly, the well-defined trunk of highly elongate, streamlined glacial lineations mapped as drumlins in this study have the characteristic shape and dimensions of rapid ice-flow landforms. Approximately 40% of the highly attenuated bedforms mapped as drumlins have elongation ratios ≥10:1 (ranging up to 11:1), which is similar to bedforms recently imaged beneath ice streams in West Antarctica (King et al., 2009). There is spatial variation in lineation elongation (cf. Lovell et al., 2012) with lineations clearly showing a lateral transition in both length and elongation ratio, with the longest and most elongate lineations focused within a narrow zone (Fig. 3C, D). The orientation of lineations exhibits a high degree of parallel concordance and there are no examples of cross-cutting lineations, suggesting they formed relatively rapidly (isochronously), with minimal modification or re-moulding during subsequent deglaciation (cf. Clark, 1999; Lovell et al., 2012). Their convergent flowlines appear orientated from the Cass Valley into the main trunk of the Tekapo ice, and based on their shape and dimensions and the abrupt lateral margin of the drumlinized zone, we interpret this as a zone of rapid ice-flow in a piedmont lobe (cf. Hart, 1999; Kovanen and Slaymaker, 2004; Lovell et al., 2012; Carrivick et al., 2017b). This interpretation is supported by megaflute elongation ratios of up to 15:1, which in Patagonia have been attributed to be produced by fast-flow/surge glacier behaviour by Ponce et al. (2019).

In contrast to the large surge-type glacier clusters widely known for several mountain regions in the northern hemisphere, the presence of surging glaciers in the southern hemisphere is less well known. Whilst surge-type glacier behaviour has not previously been recognised in New Zealand, similar geomorphological suites of landforms in other southern hemisphere locations, such as the central Andes of Argentina and Chile (Falaschi et al., 2018) and southernmost Patagonia, have previously been hypothesized to represent transient rapid ice flow (Benn and Clapperton, 2000; Lovell et al., 2012). There is evidence of surge-like fast flow within outlet glaciers of the former Patagonian Ice Sheet, possibly associated with proglacial lake formation at ice margins (Glasser and Jansson, 2005; Darvill et al., 2017; Ponce et al., 2019).

5.3. Palaeo-glaciological implications

Since surging has not previously been recognised in glaciers in the Southern Alps of New Zealand, it is worthwhile to briefly consider what surges are and why they might occur in some glaciers but not in others. Glacier surges are cyclical short-term flow instabilities, typically within a glacier system (Meier and Post, 1969), not directly related to external triggers and widely agreed to occur in response to internally driven conditions in basal conditions (Meier and Post, 1969; Sharp, 1988). Surging glaciers exhibit a wide range of observed characteristics and behaviours, encompassing land-terminating and tidewater glaciers, cirque glaciers, valley glaciers and ice streams, as well as both temperate and polythermal regimes (Murray et al., 2003). In response to this diversity, a wide variety of potential surge mechanisms has been proposed, focusing on basal processes and includes cycles of thermal or hydrological conditions (Kamb et al., 1985; Clarke et al., 1984; Kamb, 1987; Van Pelt and Oerlemans, 2012), or instabilities in the deforming bed (Clarke et al., 1984). Surge-type glaciers are common in smaller, polythermal glaciers mainly concentrated in Alaska, Arctic Canada, Greenland, Iceland, Svalbard, Novaya Zemlya, as well as the Caucasus, Karakorum, Pamir, ad Tian Shan mountain ranges (Sevestre and Benn, 2015; Ingólfsson et al., 2016). Several studies have been undertaken to compare surge-type and non-surge type glaciers within individual clusters, aiming to identify possible topographic and geological controls on surging behaviour (Post, 1969; Clarke et al., 1984; Hamilton and Dowdeswell, 1996; Barrand and Murray, 2006). Such studies have identified a number of statistically significant correlations within clusters, but they have not provided definitive answers as to why some glaciers in some regions are prone to surging whereas others are apparently not. Sevestre and Benn (2015) predicted a climatically optimal surge zone where the highest densities of surge-type glaciers within an optimal climatic envelope were bounded by temperature and precipitation thresholds. In their model, New Zealand was one of few glacierized regions where their model predicts low probabilities of surge-type glacier behaviour likely due to high mean annual temperatures.

Possible triggers for surges elsewhere have been suggested including the reorganization of the basal hydrological system (Kamb et al., 1985), switching of the thermal regime (Sevestre and Benn, 2015), as well as sediment deformation on the glacier bed (Clarke et al., 1984; Björnsson, 1998). The location of surge-type glaciers has been associated with the presence of sedimentary basins, leading to the suggestion that saturated soft-sediments at high pore water pressure are likely to facilitate rapid motion (Anandakrishnan et al., 1998). Our field investigations in the Tekapo Valley confirm the presence of soft, potentially deformable sediments in the area which overlie sequences of glaciofluvial and glaciolacustrine sediments. However, Stokes and Clark (2003) conclude that whilst bed properties are likely to be influential in determining the occurrence and vigour of surging glaciers, widespread soft-bed geology is not an essential requirement for those without topographic control.

Considering these generally accepted climatic, glaciological and trigger conditions for glacier surges, we speculate that a combination of a source of soft, erodible sedimentary rocks and sediments, the presence of an ice-contact lake, and the topographical setting of the Tekapo Valley at the LGM all probably combined to produce surge-activity of the Tekapo Glacier. In support of this assertion are case studies by: (i) Kehew et al. (2005), who presented evidence of fast-flow of the Lake Michigan lobe from the Laurentide Ice Sheet as a consequence of a combination of the topographic setting, weak and low permeability basal sediment and the existence of proglacial lakes that enhanced the development of high subglacial pore pressures; (ii) Evans et al. (2008), who proposed that recession of the southwestern margin of the Laurentide Ice Sheet resulted in the damming of progressively more extensive and deeper proglacial lakes, which facilitated surging and ice streaming through ice-margin decoupling and drawdown; and (iii) Ponce et al. (2019), who speculated that the presence of proglacial Lake Viedma, southern Patagonian ice field, propagated fast ice-flow through a series of thermos-mechanical feedbacks involving ice flow and temperature. We rule out switching of the thermal regime as a potential mechanism for surge-type behaviour in the Tekapo Glacier, because our findings (Section 5.1) indicate that the glacier was already temperate at the LGM.

Furthermore, the development of (a very large) ice-contact lake in the Tekapo Valley following the LGM would have prompted ice calving and rapid terminus retreat. Probable ensuing collapse of the ice margin would have promoted preservation of the surge-diagnostic landforms such as crevasse-squeeze ridges and 'zig-zag' eskers. This proglacial lake may also have temporarily interrupted the primary climatic driver of glacial activity, indicating that some changes in ice-marginal dynamics during recession were non-climatically driven. Switches from non-surging to surging behaviour could therefore have resulted in response to destabilization of the glacier margin where it contacted proglacial lakes (e.g. Stokes and Clark, 2004) or where changes in the subglacial drainage system were triggered by meltwater storage and release during deglaciation (e.g. Fatland and Lingle, 2002).

This Tekapo case study adds to the growing body of evidence for palaeo-surge/'surge-like' activity identified across the globe in many formerly glaciated regions, both onshore (e.g. Evans et al., 1999, 2014; Lovell et al., 2012; Gribenski et al., 2016; Darvill et al., 2017; Delaney et al., 2018; Ponce et al., 2019) and offshore (e.g. Ó Cofaigh et al., 2002; Andreassen et al., 2014; Klages et al., 2015; Kurjanski et al., 2019), as well as ice stream shutdown and stagnation (Clayton et al., 1985; Patterson, 1997; Jennings, 2006; Ó Cofaigh et al., 2010, 2013).

6. Conclusion

Detailed geomorphological analysis has enabled recognition of fourteen main glacial landforms in the Tekapo Valley (Fig. 2): drumlins, flutings, crevasse-squeeze ridges, moraine complexes and deposits; push moraine ridges, overridden end moraines, hummocky terrain, meltwater channels, ice-contact outwash plains, push moraines developed in ice-contact glaciofluvial deposits, eskers, palaeo-shorelines, palaeo-deltas, and glaciolacustrine deposits. Together, these landforms and their assemblages characterize two glacial landsystems that record a transition in glacier behaviour from that of an active temperate piedmont lobe to that of intermittent surge-type behaviour. The geomorphological evidence of surging glacier activity comprises the juxtaposed coincidence of subglacial crevasse-squeeze ridges, 'zig-zag' eskers and attenuated lineations (such as elongate drumlins and long flutes). The characteristics and distribution of the mapped glacial landforms in this study have similarities with the conceptual models of surge-type landsystems (Evans and Rea, 1999, 2003; Schomacker et al., 2014), and particularly with those in comparable mountain environments such as the landform assemblages identified in the eastern sector of Lake Viedma in Patagonia (Ponce et al., 2019). Glacier surges have never been documented in New Zealand either in modern processes or in the Quaternary record, and whilst they have been documented elsewhere in the southern hemisphere (Glasser and Jansson, 2005; Lovell et al., 2012; Darvill et al., 2017), there is a paucity of examples of glacier landsystems with a surging signature in the Quaternary record.

The existence of palaeo-surge activity in the Tekapo Valley opens up the possibility of the occurrence of similar intermittent behaviour in other LGM New Zealand active temperate glaciers, mainly in those with similar geological and topographical characteristics to the Tekapo Glacier. At Tekapo we consider that it is a combination of the topographic setting, soft-erodible sediments and the presence of an ice-contact proglacial lake that could have contributed to the development of intermittent surge activity. Our mapping allows us to assess and understand the major drivers and influences on glacial landform development and we have interpreted those for each landform type. However, other than the conceptual discussion presented herein, further analytical, perhaps numerical, work is required to determine whether the pattern and style of glaciation in the Tekapo Valley during and immediately after the LGM predominantly reflects external (i.e. climatological) controls or internal (i.e. glaciodynamic) controls. With regard to the rates of change in the glacial landsystem, there is an absence of absolute dating in the Tekapo Valley which contrasts with other Mackenzie basin valleys, Pukaki and Ohau, where intense geochronological campaigns have primarily focussed to constrain the timing of the last glacial ice retreat (Fig. 1a). Quantitative ages are needed in the Tekapo region to constrain the timing and rates of glacier changes and also to infer regional spatiotemporal patterns of deglaciation across the Mackenzie basin. However, any sampling strategy to obtain geochronological dates needs to be mindful that parts of the landsystem in the Tekapo Valley represent non-climatically forced glacier fluctuations, especially surges.

A full-resolution version of the glacial geomorphology of the Tekapo Valley, South Island, New Zealand (Fig. 2) is presented as an interactive PDF file. Supplementary data to this article can be found online at https://doi.org/10.1016/j.geomorph.2019.07.008.

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