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Ice-contact proglacial lakes associated with the Last Glacial Maximum across the Southern Alps, New Zealand

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ABSTRACT

Proglacial lakes can affect the stability of mountain glaciers and can partly disengage glacier behaviour from climatic perturbations. However, their role in controlling the onset and progression of deglaciation from the Last Glacial Maximum (LGM) remains poorly understood. This lack of understanding is partly because the evidence required to consistently and robustly identify the location and evolution of ice-contact lakes is not standardised. In this paper we therefore firstly present a new set of criteria for identifying the landform and sedimentary evidence that defines and characterises ice-marginal lakes. Secondly, we then apply these key criteria with the aid of high-resolution topographic mapping to produce the first holistic definition and assessment of major proglacial lake landforms and sediments pertaining to the end of the LGM across South Island, New Zealand. The major findings of this assessment can be grouped to include that: (i) The localised constraints to proglacial lake extent were topography, glacier size and meltwater/sediment fluxes, (ii) Lake damming was initiated by outwash fan-heads that interrupted water and sediment flows down-valley, and (iii) New Zealand LGM lakes were unequivocally in contact with a calving ice margin. These findings will be useful for reconstructing ice dynamics and landscape evolution in this region.

1. Introduction and aims

Proglacial lakes are pervasive globally within contemporary mountain environments. Some are hydropower resources and others represent natural hazards when they suddenly drain, disrupting local communities and regional infrastructure downstream (Carrivick and Tweed, 2016). Ice-contact proglacial lakes are important for determining glacier geometry and behaviour (Carrivick and Tweed, 2013). They play a significant role in controlling proglacial landform evolution and effectively buffer the delivery of meltwater and sediment to proglacial areas and to the oceans (Larsen et al., 2011; Tsutaki et al., 2011; Carrivick and Tweed, 2013; Staines et al., 2015). They can also provide an effective buttress to the ice margin, evidenced by periods of instability following rapid and sudden drainage events (Kirkbride and Warren, 1999; Diolaiuti et al., 2006; Röhl, 2008). Exceptionally large floods from proglacial lakes are known to have affected global circulation and global climate during the Quaternary (Barber et al., 1999; Clark et al., 2001 Teller et al., 2002; Mangerud et al., 2004).

A range of studies have described and quantified the thermo-mechanic feedbacks operating in modern ice-contact environments (Warren and Kirkbride, 1998; Kirkbride and Warren, 1999; Warren et al., 2001) and it is perhaps logical to hypothesize that the same thermo-mechanical feedback mechanisms observed today in modern ice-contact environments were almost certainly active during Quaternary glaciations (Stokes and Clark, 2003). Indeed, proglacial lakes have been recognised as an integral part of the onset and progression of ice sheet deglaciation (Perkins and Brennand, 2015). It is therefore of wide interest to better understand the effects of ice-contact proglacial lakes worldwide on glacier dynamics and on landscape evolution, for example, especially for those ice-marginal lakes that developed at the end of the Last Glacial Maximum (LGM). Of crucial importance to gaining this improved understanding is the need for a set of key criteria to identify, characterise and interpret the landforms and sediments associated with ice-contact proglacial lakes. This improved

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understanding can be further amplified by the exploitation of high-resolution topographic data that are becoming more widely available for many mountain regions of the world.

There are two aims of this study. First, we will develop a set of key criteria to identify landforms and sediments associated with ice-contact proglacial lakes based on a comprehensive review of the literature. Secondly, we will apply these to the central Southern Alps of New Zealand, in order to characterise the physical attributes and spatial distribution of post-LGM ice-contact proglacial lakes in this understudied region. Our analysis is aided by the availability of recently-acquired high-resolution topographic data encompassing most parts of these lake environs. Throughout this paper, we use the term 'LGM lake' to refer to ice-contact proglacial lakes that formed during deglaciation from LGM ice limits.

2. Study location and previous work

2.1. Study location

The interaction of prevailing, moisture-laden westerly winds with the steep topography of the Southern Alps gives rise to conditions favourable for glaciation. Proglacial lakes represent 38% of the total number of all lakes in New Zealand (Hamilton et al., 2013). However, whilst less than 1% of more than 3000 glaciers in New Zealand have developed proglacial lakes (Chinn, 1999) due to rapid recession from Little Ice Age maxima, 33% of the country's perennial ice is contained within these lake-calving glaciers (Chinn, 2001). Chinn et al. (2012) report that more than two thirds of ice volume change in New Zealand (5.96 km^3 water equivalent) over the last three decades is due to surface lowering and terminus calving into expanding proglacial lakes.

New Zealand is one of only three Southern Hemisphere mid-latitude areas, along with Patagonia and Tasmania, which display evidence of extensive Late Pleistocene mountain glaciation (Fitzharris et al., 2007). It has been widely suggested (Thackray et al., 2009; Shulmeister et al., 2010a, b; Rother et al., 2015; Shulmeister et al., 2018a, b) that the retreat of glaciers in the Southern Alps, immediately after the LGM, could have been relatively rapid, not only because of climate forcing but also because of the widespread formation of large proglacial lakes. In this context, it is surprising that LGM lakes across the Southern Alps of New Zealand (Fig. 1) are inconsistently described, not mapped in detail and poorly understood in terms of evolution and timing. In New Zealand, LGM lakes started to form after ~19 ka, i.e. at the end of the global LGM (Clark et al., 2009). The global LGM is defined by Clark et al. (2009) from 26.5 to 19 ka but the local LGM in New Zealand occurred between c. 35–28 ka (Barrell, 2011). At many sites in the Southern Alps, e.g. Lake Hawea (Wyshnytzky, 2013), Lake Ohau (Putnam et al., 2013a), Lake Pukaki (Schaefer et al., 2006), the Rangitata (Rother et al., 2014), and the Rakaia (Shulmeister et al., 2010a,b), there were two stillstands, or re-advances, between 21 and 17 ka, one close to 21 ka and the other at 17–18 ka. During this period, a quasi-continuous, elongated ice field extended nearly 700 km along the Southern Alps, with snowlines around 600–800 m below those of present day (Porter, 1975). Piedmont and large mountain valley glaciers flowed east and west from the Main Divide (Fig. 1). They extended onto the margins of adjoining forelands and fed outlet glaciers that advanced beyond the present coastline along the west coast of South Island (Gellatly et al., 1988; Alloway et al., 2007).

2.2. Previous mapping

The landform and sedimentary evidence for the location of NZ LGM lakes, their geometry and evolution has not been studied in detail. Furthermore, there is a significant lack of reliable age control for the NZ LGM lakes. To date, the most extensive geomorphological map of glacial landforms and deposits in New Zealand is that presented by Barrell et al. (2011). The mapped glacial geomorphological features include ice-sculpted bedrock, moraine ridges, outwash plains, kettle holes, and meltwater channel margins, as well as a range of other landforms. The map provides the most complete representation of glacial geomorphology at the catchment-scale and it emphasizes geochronological surfaces using relative morphological age indicators. However, the mapping was limited to catchments with active glaciers in their headwaters and focussed on the central Southern Alps. Many major valleys, such as those in Central Otago, were not included in the map and in most cases glaciolacustrine evidence was not recorded.

There have been several recent, more localised studies that have contributed landform and sedimentary evidence for NZ LGM lakes, but these studies are mostly unpublished theses (e.g. Cunningham, 1994; Kober, 1999; Stossel, 1999; Pugh, 2008; Webb, 2009; Hyatt, 2010; Armstrong, 2010; Wyshnytzky, 2013; Evans, 2015; Krause, 2017; Thomas, 2018). Furthermore, these investigations concentrated on individual glaciers and the combined landform and sedimentary of evidence of LGM lakes is too sporadic to enable a regional evaluation.

3. Methods

3.1. Criteria for ice-contact proglacial lake identification

Ice-contact proglacial lakes have previously been reconstructed from sedimentary evidence and landforms within the geological record (Teller, 2001; Jansson, 2003; Mangerud et al., 2004; Larsen et al., 2006; Livingstone et al., 2010; Fiore et al., 2011; Murton and Murton, 2012; Perkins and Brennand, 2015; Carrivick et al., 2017). Despite recent attempts to formulate diagnostic criteria based on knowledge of contemporary proglacial lake processes (e.g. Livingstone et al., 2012), there are still no robustly validated criteria for distinguishing their geological signature. In this study, we constructed a new table of geomorphological and sedimentological criteria (Table 1) that in combination are diagnostic of ice-contact proglacial lakes. A diagram to supplement these criteria is shown in Fig. 2.

3.2. Geomorphological mapping

We compiled landform and sedimentary evidence relating to LGM lakes in New Zealand firstly from the published literature, secondly from our own exploitation of recently-acquired high-resolution topographic data, and thirdly from our own field observations. Figures, maps, and descriptions of proglacial lake landforms and deposits were sought from journal articles, scientific reports, and theses, as well as from archives and databases from the Institute for Geological and Nuclear Sciences (GNS) in New Zealand.

The spatial extent and resolution of the topography and aerial photography data used in this study is detailed in Table 2. Recent acquisition of 800 km^2 fine-resolution LiDAR (Light Detection and Ranging) data enabled landforms to be identified, digitised and interpreted for the first time. LiDAR point clouds were meshed into a digital elevation model (DEM) in the open source software CloudCompare v.2.10 (CloudCompare, 2016). Relief-shaded models (315° and 45° azimuth) were constructed from the DEMs, primarily to provide topographic context in areas of complex relief. The horizontal resolution (0.5–2 m) of the DEMs used is exceptional for the detection of subtle palaeo-lake shorelines, and far beyond that

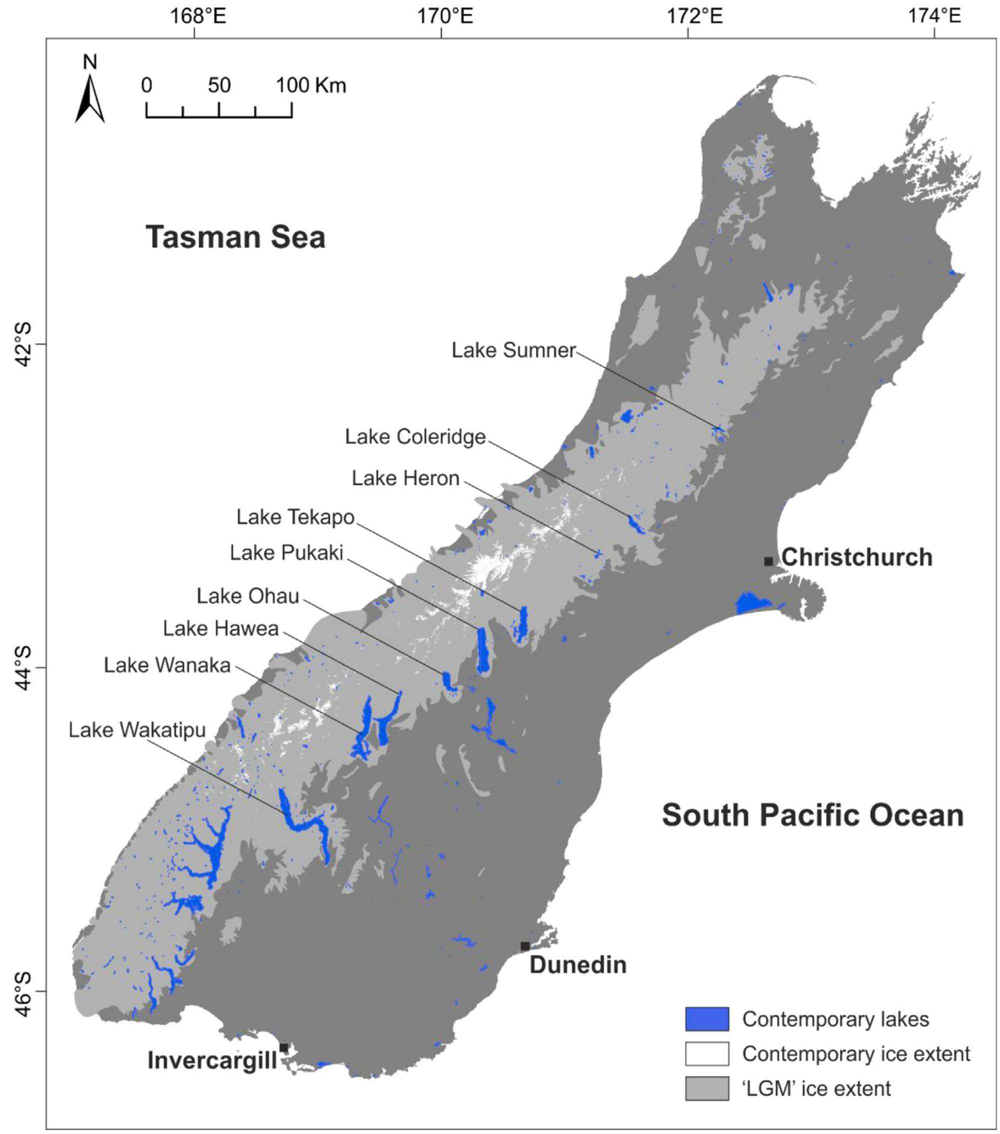

Fig. 1. Glaciological context of modern lakes in New Zealand. Lakes associated with the end of the LGM and discussed in this paper are labelled. Contemporary glacier outlines from Global Land Ice Measurements from Space (GLIMS), contemporary lake outlines from National Institute of Water and Atmospheric Research (NIWA), LGM limits reconstructed from geological evidence (Barrell, 2011). Note that the LGM ice margin was asynchronous.

of the highest resolution elevation data (8 m) publicly available for the whole of New Zealand. In areas where high-resolution topography data were unavailable, aerial photographs were used to map glaciolacustrine landforms and deposits. The aerial imagery was sourced from the Land Information New Zealand (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 Otago and Canterbury regions were taken in summer periods of 2004–2011. Ground-truthing of the mapping was conducted extensively and our field work also sought out sediment exposures at the following sites; Glenorchy (Lake Wakatipu), Beacon Point, Colquhouns Beach, and Glendhu Bay (Lake Wanaka), Dingle Burn and Timaru Creek (Lake Hawea), Lake Alexandrina (Lake Tekapo), Lake Heron, and Lake Coleridge.

Geomorphological maps were produced using the ESRI GIS software ArcMap 10.3.1 and presented in New Zealand Transverse Mercator (NZTM) map projection. Elevation profiles and individual contours were constructed using the software extension 3D Analyst. These spatial analyses permitted a greater number of landforms to be identified and interpreted, and semi-automated identification delineation of prominent LGM lake shorelines, respectively (Table 3).

3.3. Lake bathymetry

Lake bathymetry was provided by NIWA (National Institute for Water and Atmospheric research) mainly in the form of scanned paper maps with water depth contours ranging in interval between 5 m and 100 m. Each map was georeferenced and digitised. The resultant bathymetry for each lake is shown in Figs. 3, 6, 9, 10, 12, and 13 and provided in the Supplementary Information; Fig. 2. These bathymetries were used to calculate individual lake volume for the modern day (Table 3). LGM lake volume was not calculated because, as will be discussed below, it is difficult to constrain both the location and thickness of the ice margin during lake expansion. Lake levels will also have varied though time according to the position of the glacier trunk in the valley.

Table 1
Diagnostic criteria for the identification of palaeo proglacial lakes in the Quaternary record. Letters (a) – (g) correspond to Fig. 2.

Sediment-landform assemblage	Identification characteristics		Process Interpretation	Reconstruction Potential	Limitations	Examples
	Field-based	Remote sensing-based				
Sub-aqueous mass flow deposits (a)	Poorly sorted, weakly bedded, sediment interbedded with planar bedded and ripple cross-laminated sand beds, drape laminations, flame structures and rare rip-up clasts. Convolute lamination, dewatering structures, climbing ripple sequences common.	Homogenous surface texture and colour. Shadowing (break in slope) along former ice-contact slope. Often associated with other ice-marginal deposits (e.g. moraines).	Deposition by a variety of subaqueous flow processes (turbidity currents, mass flows/slumps, traction currents, surge currents – deposition from high velocity expanding jet). Sediment distribution processes are gravity driven for the most part.	Indicates a water-lain ice-proximal environment. Can occur in either proximal (diamictons) or distal (turbidites) settings.		Lawson (1982) Hyatt et al. (2012) Evans et al. (2013) Shulmeister et al. (2018a)
Shoreline (b) Beach (depositional) / Bench or Terrace (erosional)	Near horizontal or cuts in valley side. Often align parallel to modern lake shore. Multiple shorelines at different elevations occur as 'flights' on some hillsides.	Shadowing along former lake-side break in slope of feature. Flat upper surface.	Shorelines are the topographic indent left by wave re-organisation of hillside sediment. Represents initial impact of lake transgression Generally identified in aerial imagery by their photographic lows that result from better drainage and lack of vegetation.	Best approximation of the mean lake level elevation that existed when the shoreline was actively being formed. Inner edge of shoreline is most representative of height at the time of shoreline formation. Used to infer relative timing in between lake levels. Represents a period of stability for the ice margin.	Very faint in areas of low superficial cover (e.g. bedrock) and where narrow (due to minimal shadowing). Sporadically preserved along length of lake, and often discontinuously represented.	Clapperton et al. (1997) Blair (1999) Drake and Bristow (2006) Carrivick et al. (2017)
Grounding Zone Wedge (c)	Diamict 'moraine' material is likely to be deposited on the up-ice side of grounding line fans and ice-contact deltas. Well bedded foreset and bottom-set deposits.	N/A	Formed at the transition from a grounded ice margin to a floating ice margin. Fed by flows entering a lake basin at the base of the water column rather than at the top (as is the case for deltas). An accumulation of sediment builds up at the grounding line when the ice margin remains in a stable position.	They form transverse to ice flow. They represent recessional positions during deglaciation and they can be used to infer ice thickness. The location of the grounding line is important because mass loss is strongly linked to changes in ice margin positions and the grounding line. Change in the grounding line can result in very rapid changes in glacier and ice shelf behaviour.		Powell (1984) Eyles et al. (1985) McCabe and Eyles (1988) Powell and Molnia (1989) Lønne (1995) Powell and Alley (1997)
Sublacustrine (DeGeer) Moraine (d)	A series of separate, subparallel, narrow ridges. Each ridge consists of a till core, capped by a layer of partly rounded boulders. Linear or sinuous crests. Frequently display glaciotectonised structures.	N/A	Deposited beneath the grounded part of an ice sheet that extended into a lake. Formed by coalescing grounding line fans during periods of intermittent glacier margin retreat, as well as by brief ice-margin standstills or minor advances.	The moraine ridges are orientated transverse to ice flow. They are significant features for understanding sedimentary grounding line processes and recession. DeGeer moraines are strictly ice-marginal features.		Barnett and Holdsworth (1974) Larsen et al. (1991) Lindén and Möller (2005) Golledge and Phillips (2008)
Palaeo-Delta (e)	Fan-shaped body of sediment,	Homogenous surface texture and	Formed from deposition of	Formation is characteristic of		Ashley (1975) Gilbert (1885)

Table 1 (*continued*)

Sediment-landform assemblage	Identification characteristics		Process Interpretation	Reconstruction Potential	Limitations	Examples
	Field-based	Remote sensing-based				
	upstream of modern (actively forming) lake deltas. Low gradient (gently sloping) surface. Unbroken, vertical succession of up to 3 distinguishable bedding sets (bottom-set beds, foreset beds, topset beds). Often displays rhythmic fining up sequences. Sharp break in slope and steeply inclined lakeside face.	colour distinct from adjacent terrain. Shadowing (break in slope) along former delta front. Distinct flat, tabular surfaces characterised by steep sides. Presence of numerous channels or gullies.	sediment carried by a river as the flow leaves its mouth and enters slower moving/standing water.	freshwater lakes – used to reconstruct lake water levels. Delta front break in slope approximates (former) glacial lake level Gilbert-type deltas are specifically formed coarse sediments (as opposed to gently sloping muddy deltas).		Corner et al. (1990) Dirszowsky and Desloges (2004) Østrem et al. (2005) Gobo et al. (2014)
Iceberg Grounding Structure (f)	Semi-circular to elongate depressions, enclosed by sharp crested ridges. Often occur in a dense network. Light/dark shadowing distinguishes pits and craters from adjacent ridges.	Light/dark shadowing distinguishes pits and craters from adjacent ridges. Occur in dense networks.	Erosional grooves and constructional ridges (keel and plough marks) are created by intense iceberg scouring, grounding and ploughing events.	Infers an ice-contact environment, indicative of floating terminus break-up (calving). Can be used to infer the number of glacial advances and the direction of those advances.	Possible misidentification as ice-marginal ridges but, unlikely due to differing orientation.	Thomas and Connell (1985) Woodworth-Lynas and Guigné (1990) Eyles et al. (2005)
Below wave base lacustrine deposits (g)	Accumulations of horizontally bedded, well sorted, fine-grained sediment. Often with the presence of massive or rhythmic lamination. Occasional presence of ice-rafted debris (dropstones ***(g)***), rhythmites, climbing ripples and deformation structures within. Found around former ice margins, lake embayments or valley sides.	Distinctive white colouration of terrain on aerial imagery. Distinctly flat surface on DEM.	With distance from the ice-margin, rhythmically laminated bottom sediments develop via a combination of suspended sediment concentrations from input streams, and bottom current transport from underflows, interflows and overflows. Rhythmites reflect variations in sediment supply and depositional conditions.	Represents deposition in a proximal to distal glaciolacustrine environment. Indicative of glacial lake existence and former lake levels. Varves (pairs of layers of clay and silt of contrasting colour and texture) in particular used to establish glacial and glaciolacustrine chronologies.	Underestimation of spatial extent on imagery. Best identified in the field.	Ashley (1975) Smith and Ashley (1985) Benn and Evans (1998) Tiljander et al. (2003)

4. Results

In the following subsections, we provide an overview of the distribution and geometry of nine proglacial lakes situated within parallel valleys east of the Main Divide (Fig. 1). We then describe and critically evaluate existing knowledge for each lake, before presenting new results produced from analysis conducted in this study.

The heavily-skewed distribution of lakes east of the Main Divide (Fig. 1) is at least partially a preservation issue. Ice-contact lakes are recognised in the literature to have been dispersed across the gently sloping terrain beyond (westwards of) many western valley mouths shortly after the LGM (Suggate, 1965; Suggate and Almond, 2005; Thomas, 2018; Supplementary Fig. 1A). With lakes presently persisting at moraine limits (Supplementary Fig. 1A) it is likely that large proglacial lakes would have formed on the Westland piedmont, occupying glacial troughs during ice retreat and thus there would have been more LGM lakes on the western side of the Southern Alps than we see evidence for now. However, because lacustrine deposits are buried they are not discussed in this paper. Other lakes that this review did not investigate further include the glacially carved lakes in Fiordland (Supplementary Fig. 1B) e.g. Lake Manapouri, Lake Te Anau, Lake Monowai, Lake Hauroko, or Lake Poteriteri because the terrain in Fiordland is dominated by densely vegetated bedrock which obscures any glaciolacustrine evidence that could have been preserved (Pickrill, 1976).

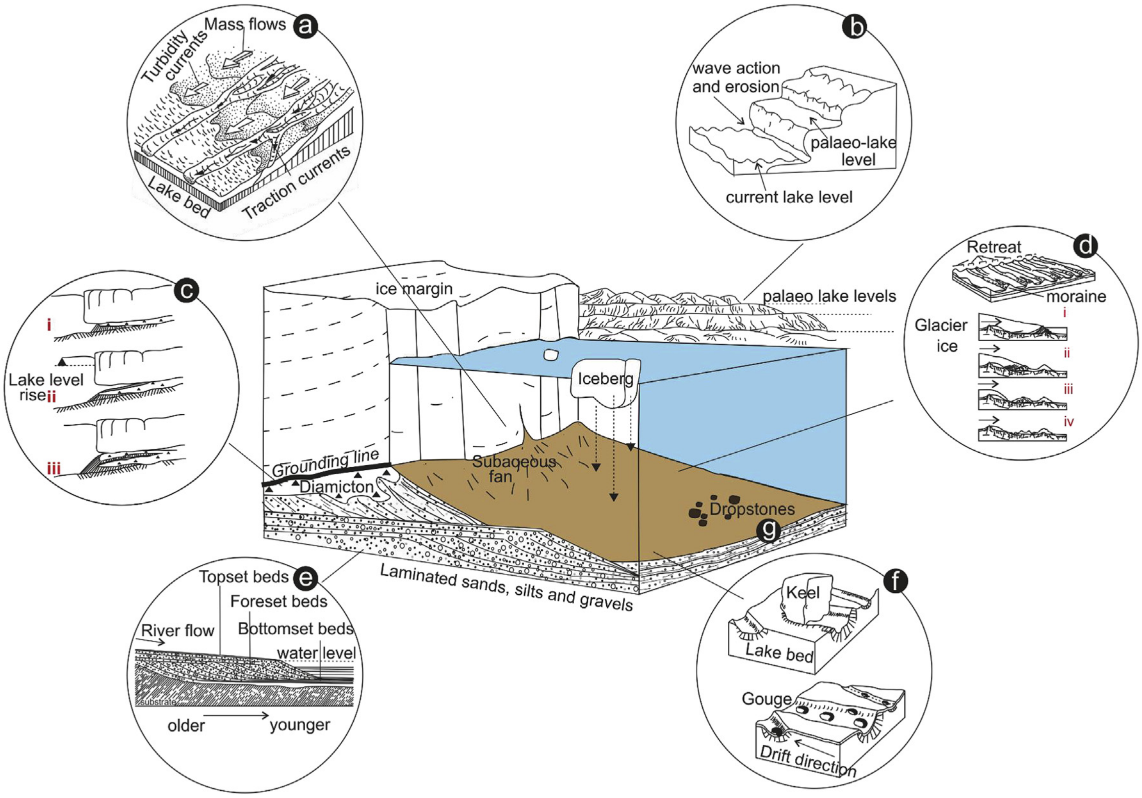

Fig. 2. Schematic of ice-marginal lake and associated sediment-landform assemblages. (a)–(g) correspond to Table 1.

Table 2
A summary of the imagery analysis used to map evidence of proglacial lakes associated with the end of the LGM in New Zealand. QLDC: Queenstown Lakes District Council. LINZ: Land Information New Zealand.

Lake	Extent	Imagery	Horizontal Resolution (m)	Year Acquired	Source	Area (km^2)
Wakatipu	Glenorchy	LiDAR	0.5	2011	QLDC	16
	Queenstown	LiDAR	0.5	2016	QLDC	302
	Kingston	LiDAR	0.5	2011	QLDC	2
Wanaka	Wanaka township	LiDAR	0.5	2011	QLDC	45
Hawea	Hawea township	LiDAR	0.5	2011	QLDC	5
Pukaki	Whole lake buffer	LiDAR	2	2010	Meridian Energy	72
Tekapo	Whole lake buffer	LiDAR	2	2017	LINZ	357
_	Central Otago	Aerial Photographs	0.75	2004–2011	LINZ	–
_	Canterbury	Aerial Photographs	0.4	2013–2014	LINZ	–

4.1. Lake distributions and geometries

None of the following nine proglacial lakes are presently in direct contact with active glaciers. They are all fed by sediment-laden meltwater and snowmelt from alpine tributaries and glacierised high mountain valleys. The majority of the lakes extend down-valley beyond the confines of surrounding steep hillslopes and are partially in a piedmont setting.

An overview of both the modern lake and LGM areal extents are provided in Table 3. The contemporary lake extents do not differ greatly when compared to the LGM, ranging from a 16.5% (Lake Wakatipu) decrease in area to 1.2% (Lake Hawea) decrease in area. However, these figures indicate lake water elevations have fluctuated by many tens of metres, implying large volume changes.

The LGM lakes have conspicuous dams. These dams are composed of a huge thickness of contemporary alluvial outwash boulder-cobble material that has coalesced with subdued ridges of moraine (Gage, 1975, 1985; Irwin, 1975; Soons, 1982) and are known to contribute to the permanence of the lakes (Shulmeister, 2017). Moraines around the ends of the large ice-carved lake basins may be only minor features contributing relatively little to the depth of the lakes, while none of the larger glacial lakes are entirely retained by solid rock (Gage, 1975; Soons, 1982).

The lake bathymetry data (Figs. 3, 6, 9, 10, 12 and 13) reveal that the lakes all have retrograde slopes and they are all remarkably deep (some > 300 m), which is a product of successive multiple Pleistocene glacial erosion episodes forming over-deepenings (Cook and Swift, 2012). The large glacial lakes tend to be steep-sided with flat basin floors. The deepest parts of some of them, for example Lake Wakatipu, are below sea level (Irwin, 1972b). Lakes Wakatipu, Wanaka and Hawea have hummocky lake floors which contain many over-deepened reaches. This hummocky

Table 3
Summary of both present-day and LGM lake geometries for the lakes described and analysed in this paper. Maximum depth and areal extents are taken from the original reference for lake bathymetry survey data. Percentage increase between LGM lake surface elevations and those of the present day are derived from tracing the contour around the highest lake level mapped. Elevation data has been taken from national coverage 8 m DEM.

Modern						LGM		
Lake	Surface Elevation (m a.s.l.)	Surface Area (km^2)	Maximum Depth (m)	Volume (km^3)	Key Reference	Surface Elevation (m a.s.l.)	Surface elevation change (%)	Damming Mechanism
Wakatipu	309	295.4	380	64.2	Irwin (1972a)	360	16.5	Terminal moraine
Wanaka	300	198.9	311	31.92	Irwin (1976)	320	6.6	Outwash fan-head
Hawea	348	151.7	384	24.49	Irwin (1975)	352	1.2	Outwash fan-head
Ohau	520	59.3	129	4.56	Irwin (1970a)	Unknown	Unknown	Outwash fan-head
Pukaki	532	172.8	98	8.9	Irwin (1970b)	520	7.4	Outwash fan-head
Tekapo	710	96.5	120	6.94	Irwin (1973)	765	4.2	Outwash fan-head
Heron	694	6.3	36	0.059	Irwin (1972b)	714	2.9	Terminal moraine
Coleridge	436	46.9	200	3.65	Flain (1970)	447	2.5	Terminal moraine
Sumner	529	13.7	134	1.1	Irwin (1979)	550	4.0	Terminal moraine

bathymetry suggests that those lake floors are probably bedrock. Lakes Ohau, Pukaki and Tekapo have relatively flat floors which suggests substantial sediment infill (Pickrill and Irwin, 1983; Upton and Osterberg, 2007).

4.2. Lake Wakatipu

4.2.1. Existing knowledge

The formation of Lake Wakatipu began with the onset of sedimentation into the basin at 17 100 ± 2600 cal yrs BP, inferred from the Greenstone River Fan (Figs. 3 and 5A; Cook et al., 2013). The number of the Greenstone River fan terraces is similar to the number of palaeo-lake Wakatipu shorelines identified, and indeed on the basis of Optically Stimulated Luminescence (OSL) and exposure dating, which revealed Holocene ages from 10 201 ± 3916 to 784 ± 284 cal yrs BP these dates inspired Cook et al. (2013) to suggest that they correlate to the ages of other palaeo-lake shorelines around Wakatipu. This number and suggested age correlation of the Greenstone River terraces with the Wakatipu shorelines indicates that the lake level remained stable throughout deglaciation and only began to lower due to the downcutting of the Kawarau outlet during the last 10 000yrs.

Cook et al. (2013) noted a prominent shoreline at a height of 351.5 m a.s.l (42 m above the present shoreline). It was assumed by Cook et al. (2013) to represent the maximum height of the lake since its formation at the end of the LGM. However, according to Kober (1999), the level of the Kingston terminal moraine (360 m a.s.l.) represents the highest lake level of Wakatipu. The highest lake level was also interpreted in earlier work by Mutch (1969), and Brodie and Irwin (1970), to correspond to the terminal moraine at Kingston, which first dammed Lake Wakatipu. Sedimentological work carried out by Kober (1999) and geomorphological mapping by Cook et al. (2014), determined that there are at least ten different lake levels present at the Bible Terrace complex, east of Glenorchy (Figs. 3 and 4A, E, Fig. 5C and D). There is also field evidence of a Gilbert-type delta sequence at Glenorchy (Fig. 5B), of which the topset-foreset juncture measured at 352 m a.s.l. Stossel (1999) found evidence of extensive glaciolacustrine silts and sands situated below 351 m that cover the valley floor southwards of Frankton towards Drift Bay (Fig. 5E). Lake Wakatipu is restrained mainly by a c. 305 m high rock rim which is capped by c. 70 m of till (Bayly and Williams, 1973).

4.2.2. New mapping

Our mapping supports the interpretation of Mutch (1969), Brodie and Irwin (1970), and Kober (1999) since the most well-developed and best preserved of all palaeo-shorelines is measured at 360 m a.s.l, which can be traced discontinuously along many parts of the lake basin. For example, a well-preserved sequence of palaeo-lake shorelines is evidenced at Jacks Point on the Frankton arm (Fig. 4C, F), of which the highest is measured at 360 m a.s.l., indicating a drop in lake level by 50 m. Our shoreline mapping shows that as the ice retreated, an enlarged Lake Wakatipu formed a continuous body of water through to Lake Hayes (Queenstown), Diamond Lake in the Glenorchy District, and further south of Kingston (Fig. 3). The evidence reported by Stossel (1999) matches our geomorphological mapping along the flanks of the Remarkables (Figs. 4B and 5E) where the lake sediments grade laterally into small fan-deltas which became isolated from the lake margin when it dropped to its present level.

Our mapping also identified a minimum of ten different lake levels at the Bible Terrace complex, east of Glenorchy (Figs. 3A and 4A, E, 5C). Several lower, but less well-defined intermediate palaeo-lake shorelines are present below the prominent high lake level, e.g. at 342 m, 334 m, and 325 m at Jacks Point (Fig. 4C, F), Bible

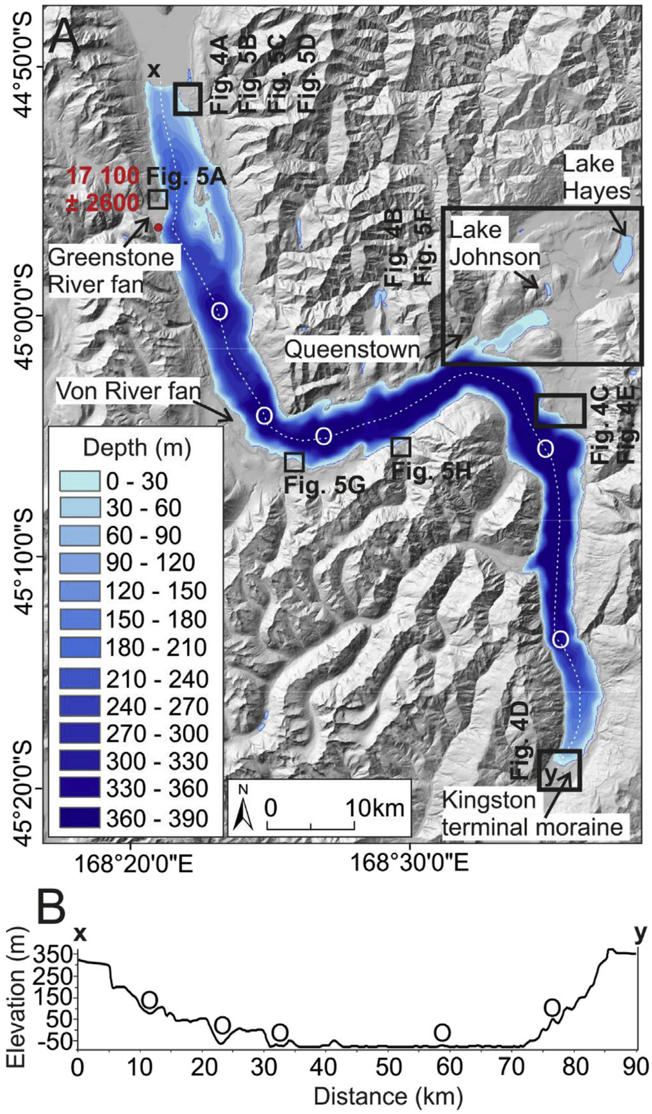

Fig. 3. Regional overview of Lake Wakatipu showing the location of high-resolution topography mapping (Fig. 4) and field sites (Fig. 5). The location of absolute dating, and respective age by Cook et al. (2013) is labelled in red. Rasterised lake bathymetry is displayed and an elevation profile of the glaciolacustrine basin is presented in B, over-deepenings are marked by O. Transects are shown from X–Y. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

Terrace (Fig. 4A, E), and at the southern end of the lake near Kingston (Fig. 4D). Overall, between seven and ten palaeo-lake levels around Lake Wakatipu are observed. There are other assumed former ice-contact features, such as two raised deltas along the Shotover River behind Queenstown Hill, higher than the prominent shoreline in several locations. These raised deltas and palaeo-shorelines are measured at the same elevation of 400–405 m, and suggests that they correlate with one another and also to the surface elevation of Lake Johnson at 400 m a.s.l (Figs. 3 and 5B). Lake Johnson, a glacial rock basin (900 m long, and max depth of 27 m), and associated raised deltas are interpreted to be produced prior to the formation of the LGM lake, possibly formed at the margins of a proto-lake which could have formed in the early part of the recession of the Wakatipu glacier. We have also identified kettle holes or iceberg pits at the southern outlet of the lake at Kingston between the intermediate lake levels (Fig. 4D).

The surface of the major Shotover delta is graded to between 350 and 360 m a.s.l (Figs. 4B and 5F), revealing that the Shotover River must have built the extremely large fan-delta into the enlarged Lake Wakatipu during the high-stand lake level. The delta extends through to the northern end of Lake Hayes, a basin which became impounded by the deposits of the delta as the lake levels fell, and now stands isolated 21 m higher than the present Lake Wakatipu at 315 m a.sl. Several other fan-delta complexes are present along the northern arm of the lake such as the Greenstone River fan and the Von River fan (Fig. 5A, G, H). Subsequent lake lowering has caused the Greenstone River to incise a flight of 10 terraces that step down incrementally towards the level of the modern river floodplain at 354 m, 349 m, 344 m, 341 m, 356 m, 332 m, 327 m, 310 m a.s.l. respectively.

4.2.3. Evolution interpretation

Lake Wakatipu was first dammed by the Kingston terminal moraine. The Mataura outlet at Kingston was abandoned 12 000 yrs ago when Lake Wakatipu became connected to the Kawarau catchment farther to the north (Thomson, 1996) leaving the 360 m a.s.l shoreline behind. The drainage re-routing after 15 kyrs probably occurred as a result of a breach of outwash fan-head sediments previously separating the lake from the Kawarau catchment. The Kawarau River must have been lower in elevation at that time than the river draining the lake at Kingston, and thus became the more favourable outlet for the lake (Park, 1909; Bell, 1982). An enlarged lake with fluctuating levels reflecting subsequent glacier oscillations may have persisted until shortly after c. 10 000 yrs ago, when, through downcutting of the Mataura outlet at Kingston, the Shotover delta became exposed to isolate Lake Hayes.

The highest shoreline (360 m a.s.l) formed over a period of ~5000 yrs while the lake level was stable. As the lake level dropped to the present level, there were episodes of relative lake level stability that allowed several intermediate and less well-defined palaeo-shorelines to be formed at various elevations, for example, at 48 m, 35 m, 27 m and 23 m above the present level (Fig. 4). The decrease of lake levels progressed slowly for the first 15 000 yrs following deglaciation (1.3 mm/yr on average) and from then onwards there has been a step-like lake level drop, most profoundly due to a switch in drainage routing from the Kingston outlet to the Kawarau River near Queenstown. The present-day lake level was likely to have been reached within the last 500 yrs (Cook et al., 2013).

4.3. Lake Wanaka

4.3.1. Existing knowledge

Sedimentary exposures in large aggraded outwash fan-heads and valley fill deposits around Lake Wanaka have been interpreted by Evans (2015) as a mixture of ice-proximal basal, glaciolacustrine and supraglacial facies that have been folded and faulted. This deformation is likely the result of either dewatering under the weight of overlying sediments or slumping due to the melt-out of buried ice. Gilbert-type deltaic facies (of bottom-sets and foresets) have been recorded by Evans (2015) in several of the gullies at Beacon Point and around Glendhu Bay (Fig. 6). The presence of at least two raised beaches were observed in the field (Fig. 8A) and noted by Evans (2015). Sedimentary structures such as asymmetric climbing ripples and recumbent folds were abundant in the laminated sands and silts underlying the raised beaches (Fig. 8B). These are formed in environments characterised by rapid sedimentation from suspension (Table 1g). Deformation of the sediments, noted to be localised by Evans (2015) in other locations is exhibited by flame structures and water escape structures. These could be related to movement of the glacier or to calving at the floating tongue of the

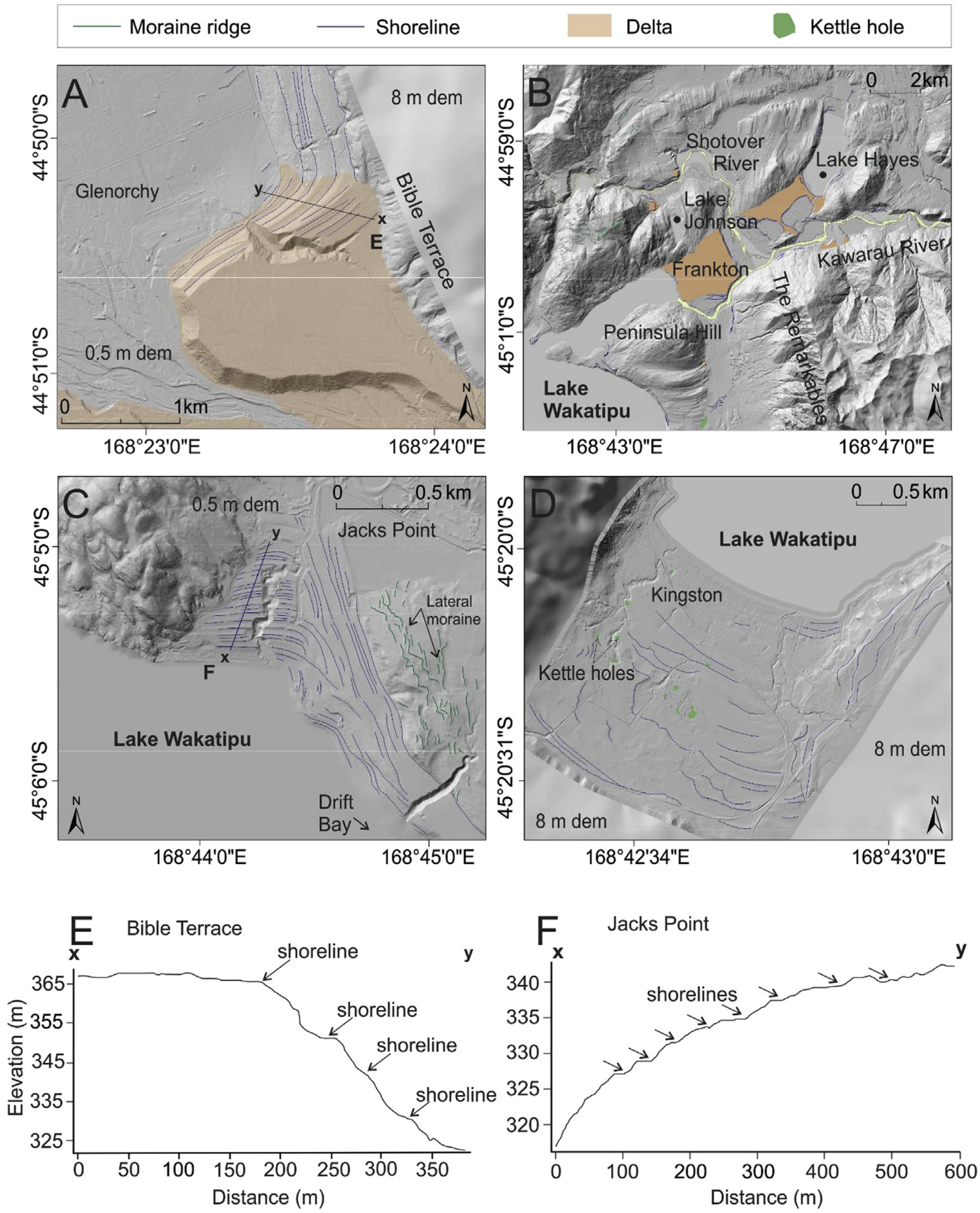

Fig. 4. A. Palaeo-shorelines mapped at Bible Terrace complex, Glenorchy (elevation profile shown in panel **E**). **B.** Shotover Delta at the Frankton arm of Lake Wakatipu **C.** Palaeo-shorelines mapped at Jacks Point **D.** Palaeo-shorelines mapped at Kingston **E**. Cross-profile through palaeo-shorelines at Bible Terrace complex, Glenorcy (shown in panel **A**). **F.** Cross-profile through palaeo-shorelines at Jacks Point. Transects are shown from X–Y. Hillshaded DEM data referenced in Table 2. The shorelines are all well captured by the fine-resolution of the DEMs.

ice margin.

4.3.2. New mapping

Past lake high-stands are evidenced from palaeo-shorelines along Lake Wanaka's southern shores (Fig. 7). Most notably, there is a flight of palaeo-lake shorelines on the eastern side of the lake at Colquhouns beach, and these are visible as concentric sub-parallel ridges that follow the boundary of the modern beach (Fig. 7A, E, and F). The highest shoreline at Colquhoun's Beach is at 314 m a.s.l. with a 14 m elevation drop to the present lake level. A sedimentary exposure through one of these ridges, shown in Fig. 8C, displays evidence of a raised beach (a depositional shoreline feature), 18 m above the present lake shoreline. Additionally, and equally spectacular, are the palaeo-shorelines near Beacon point (Fig. 7B). The highest shoreline, and the most prominent, surrounding Lake Wanaka lies at 320 m a.s.l, or 20 m above the current lake level. Palaeo-shorelines below the prominent high lake level are present at 318 m, 311 m and 309 m. There are raised deltas/perched fans along the north-eastern edge of Lake Wanaka, the first set from Rumbling Burn immediately to the east of Mou Waho island (Figs. 6 and 8E), and the second further north, to the east of Bells Creek and Twin Peaks (Fig. 6).

4.3.3. Evolution interpretation

Evans (2015) constrained the local LGM advance in the Wanaka catchment by the onset of gravel aggradation at ~32 100 ± 570 cal yr B.P with its termination defined by the age of the delta that formed at Glendhu Bay (Fig. 8D) when ice retreated into the Wanaka basin at 29 000 ± 500 cal yr B.P. This ice advance is well-constrained chronologically and well-defined

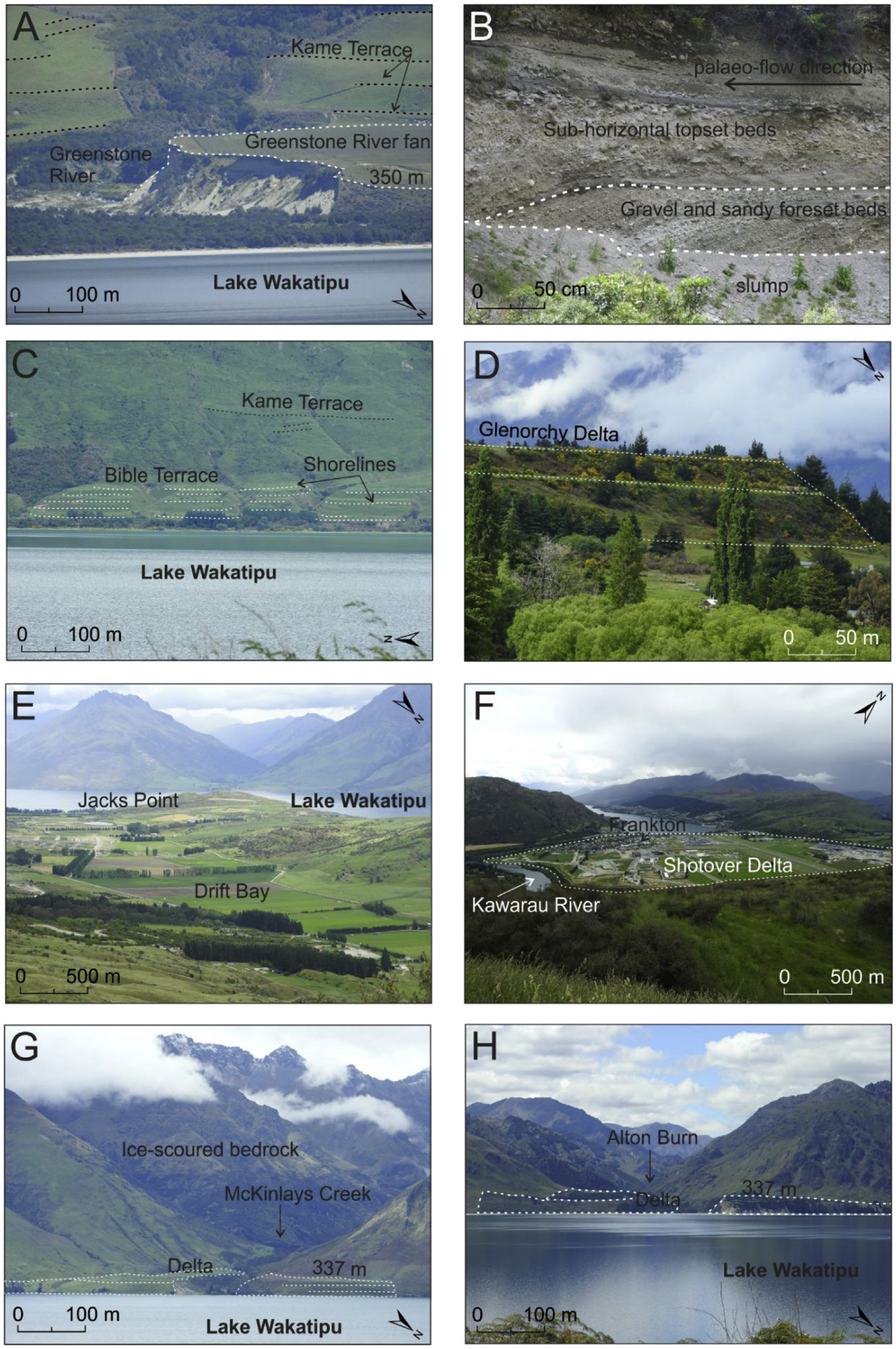

Fig. 5. A. Greenstone River Fan **B.** Sediment exposure through Glenorchy delta **C.** Palaeo-shorelines at Bible Terrace complex, Glenorchy **D.** Gilbert-type Glenorchy Delta **E.** Flat lacustrine sediments of Drift Bay, evidence of an enlarged LGM Lake Wakatipu, from Jacks Point through to Frankton. **F.** Shotover Delta at Frankton (350–360 m a.s.l) **G.** Raised-delta on the north-western valley side (337 m a.s.l) formed during drop in lake level from 360 m a.s.l **H.** Raised-delta on north-western valley side (337 m a.s.l) formed during drop in lake level from 360 m a.s.l.

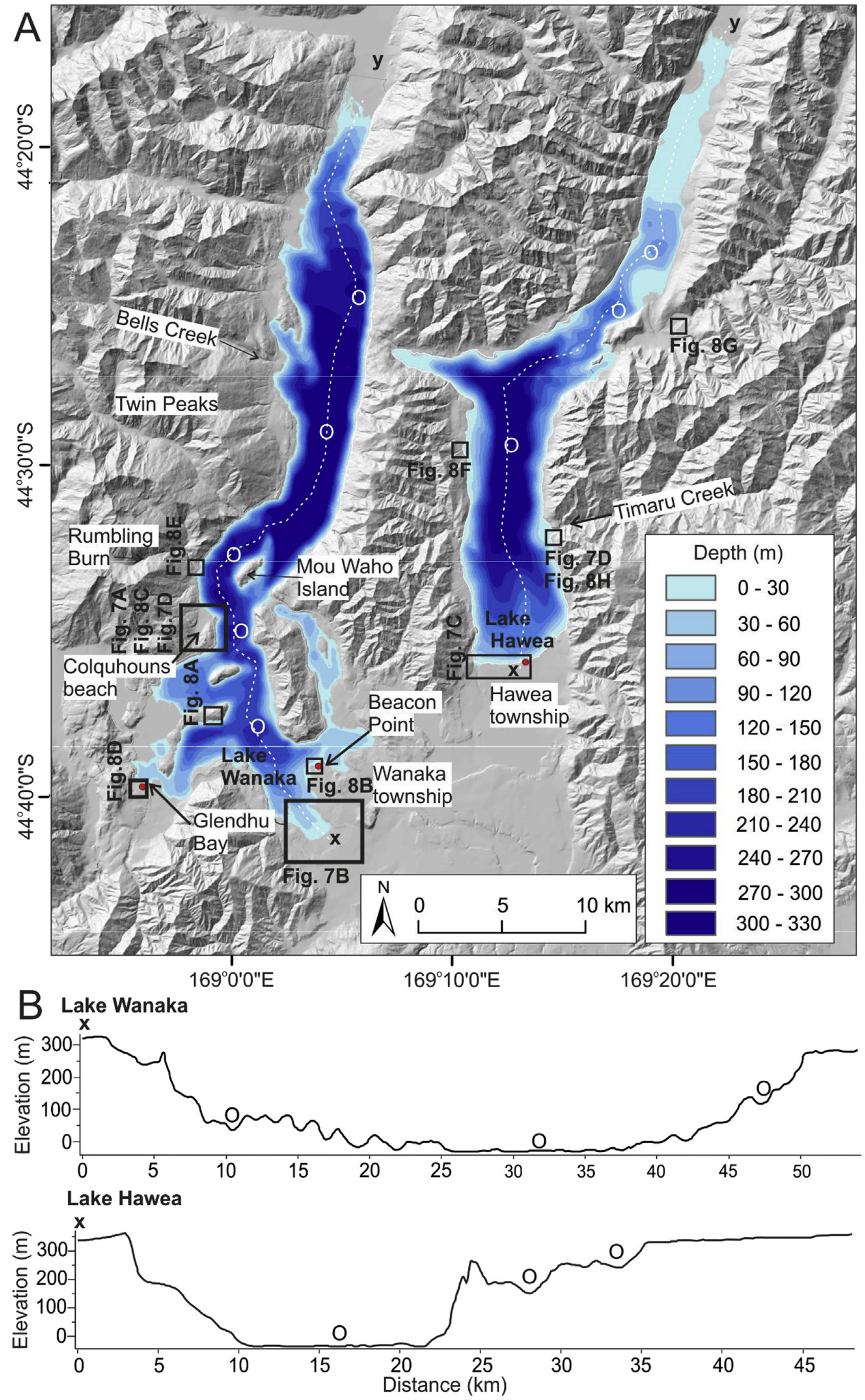

Fig. 6. Regional overview of Lake Wanaka and Lake Hawea showing the location of high-resolution topography mapping (Fig. 7) and field sites (Fig. 8). The locations of absolute dating (Table 4) is labelled in red. Rasterised lake bathymetry is displayed and an elevation profile of the glaciolacustrine basins are presented in B, over-deepenings are marked by O. Transects are shown from X–Y. (For interpretation of the references to colour in this figure legend, the reader is referred to the Web version of this article.)

sedimentologically. Evans (2015), noted good correlation between the Beacon Point and Glendhu Bay sites and thus much of the southern margin of the lake was apparently subject to the same deglacial environment and evolution.

4.4. Lake Hawea

4.4.1. Existing knowledge

The sediments of the recessional moraine, interpreted by Armstrong (2010) and Wyshnytzky (2013), comprise basal outwash gravels which are overlain by deltaic deposits and aggrading and

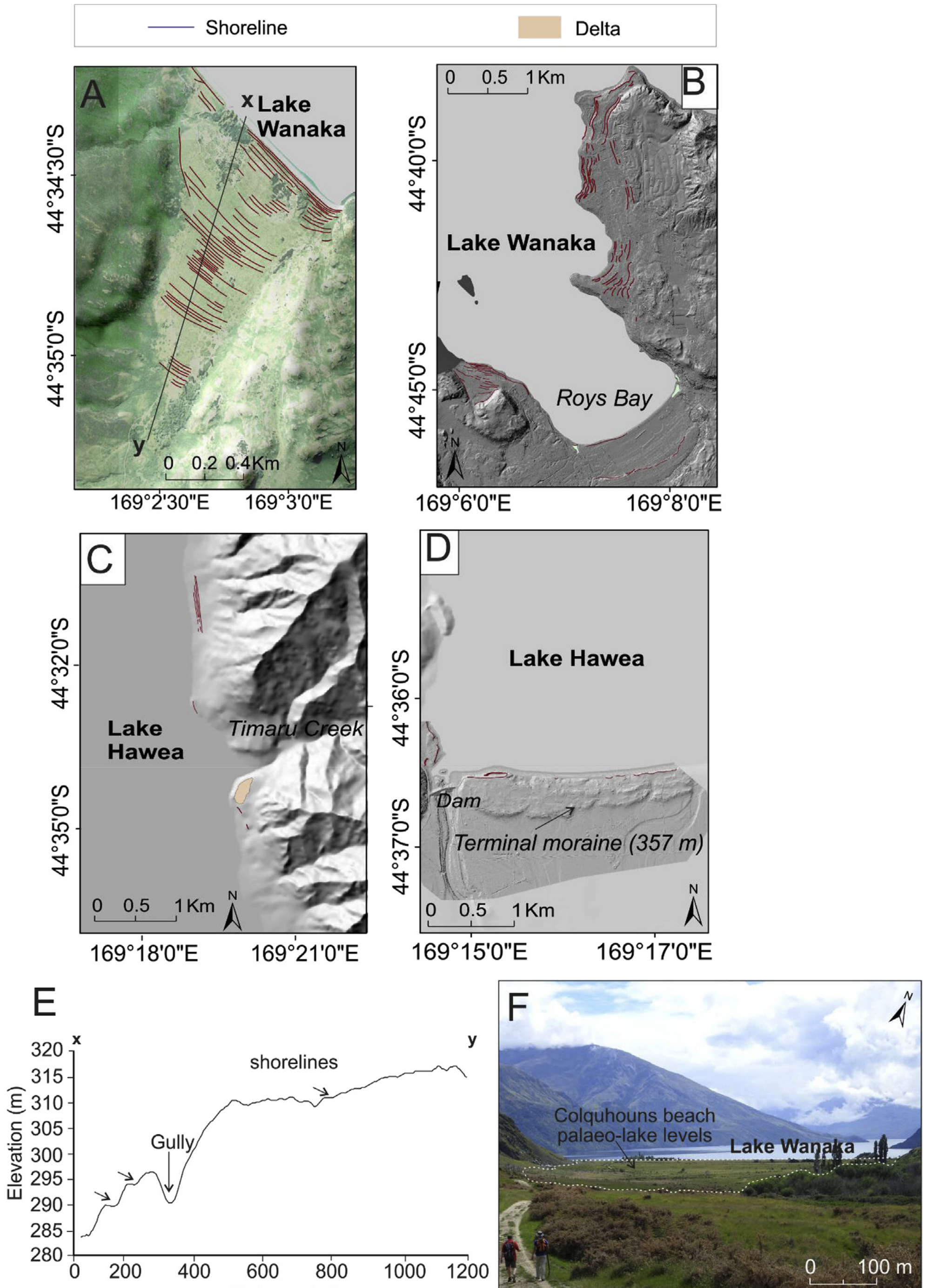

Fig. 7. A. Mapped palaeo-shorelines from 0.4 m aerial imagery at Colquhouns beach, Lake Wanka **B.** Mapped palaeo-shorelines from southern end of Lake Wanaka from 0.5 m DEM. **C.** Timaru Creek Delta and palaeo-shorelines at Lake Hawea **D.** Mapped palaeo-shorelines and the terminal moraine at the southern margin of Lake Hawea **E.** Cross-profile through palaeo-shorelines at Colquhouns beach, Lake Wanaka (transect X–Y, panel A), taken using hand-held GPS. **F.** Field evidence of palaeo-lake shorelines at Colquhouns beach, Lake Wanaka.

prograding sands and gravels that intergrade with horizontally-bedded sands and silts, altogether interpreted to be sub-aqueous ice-contact fans (e.g. Shulmeister et al., 2010a). The fans are overlain by a weakly stratified diamicton and capped by lacustrine silts and sands that are locally deformed due either to dewatering or another ice advance (Armstrong, 2010). McKellar (1960) similarly noted stratified gravels with an absence of sand and silt, with steep depositional slopes, characteristic of a Gilbert-type delta depositing foresets in high energy conditions. No palaeo-lake shorelines at Lake Hawea are recorded in the literature.

4.4.2. New mapping

In 1958, Lake Hawea was artificially raised by 20 m to store more water for increased hydroelectric power generation, and thus probably some pertinent landforms, perhaps especially palaeo-shorelines, are now underwater. Therefore, the majority of

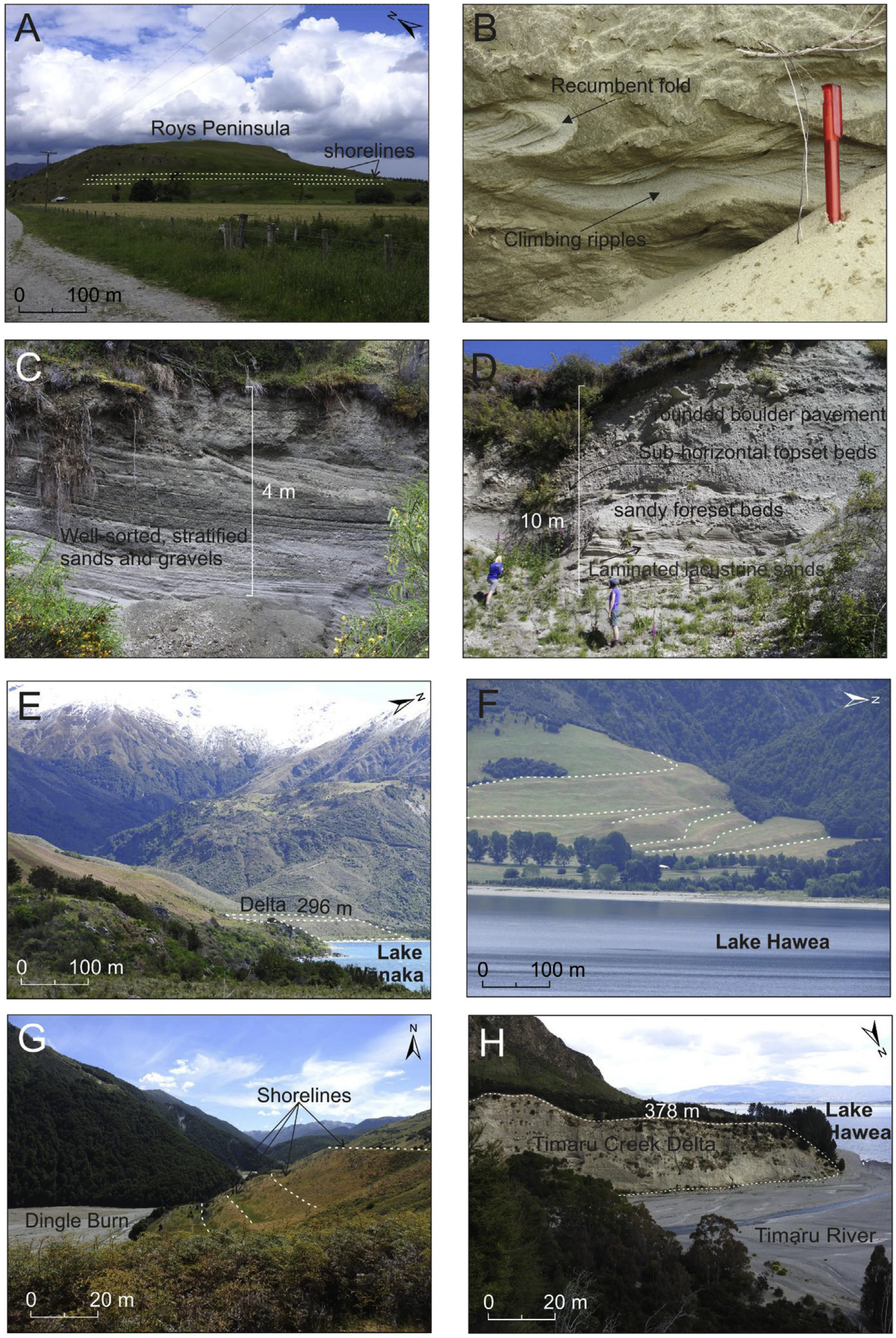

Fig. 8. A. Palaeo-shorelines at Roys Peninsula (south-western shores of Lake Wanaka) **B.** Deformed glaciolacustrine sediments at Glendhu Bay, Lake Wanaka **C.** Well-sorted, clast-supported gravels in 4 m-high sediment exposure at Colquhouns beach, Lake Wanaka. Interpreted as raised-beach. **D.** 10 m-high sediment exposure through Glendhu Bay delta **E.** Rumbling Burn raised delta **F.** Palaeo-shorelines at Mt Burn Creek on the north-western valley side of Lake Hawea **G.** Palaeo-shorelines at Dingle Burn on the north-eastern valley side. **H.** Timaru Creek delta, eastern valley side of Lake Hawea.

glaciolacustrine evidence visible today around Lake Hawea is restricted to a well-defined composite loop of recessional moraine along the southern shores (Fig. 7D). The recessional moraine rises to 9 m above the present shoreline (29 m above that of the original shoreline before artificial damming). The highest shoreline is measured at 352 m a.s.l., or 24 m above that of the original lake level (before artificial damming) near Hawea Township (Fig. 6). Further north, one third of the way up the eastern side of Lake Hawea is Timaru Creek (Figs. 6 and 7C) and the mouth of this valley hosts a perched delta (Fig. 8H). The elevation of for the topset/foreset juncture is at 380 m a.s.l. The dipping gravels beds were interpreted in this study to be delta foresets during field investigation (Fig. 8H).

4.4.3. Evolution interpretation

The outwash gravels and diamictons that comprise the recessional moraine at Lake Hawea are associated with the two most recent ice positions. Specifically, the basal gravels represent

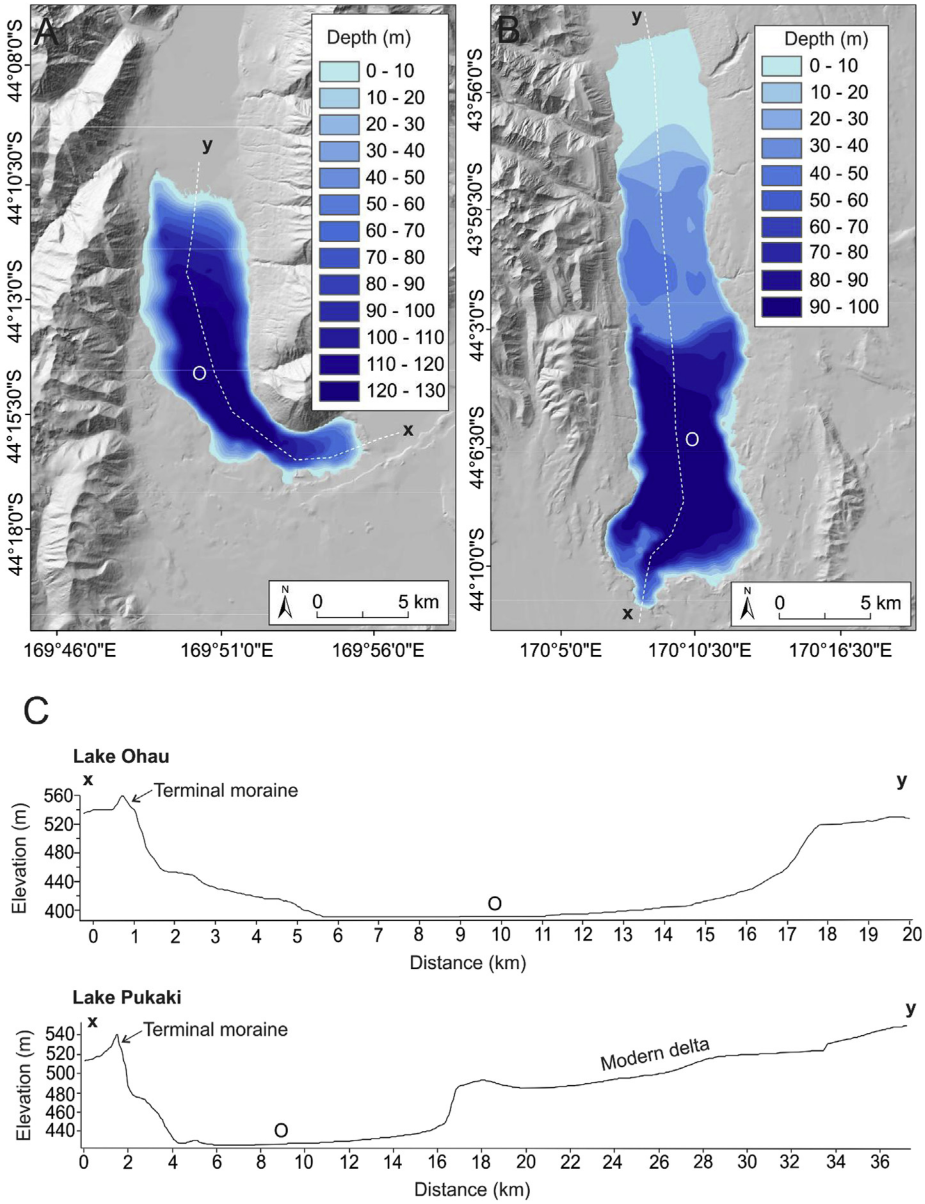

Fig. 9. Regional overview of Lakes Oahu and Pukaki. Note, no mapping of glaciolacustrine deposits or landforms has been conducted at these lakes in this study due to absence of evidence preserved. Rasterised lake bathymetry is displayed and an elevation profile of the glaciolacustrine basins are presented in B, over-deepenings are marked by O.

outwash inferred to have been deposited during retreat from an ice position further down-valley ~32–26 ka, during the local LGM in New Zealand. At < 29 ka, ice re-advanced to deposit the glaciolacustrine sediments. A further ice re-advance occurred >29–18 ka (Wyshnytzky, 2013) and stagnated which enabled the deposition of the diamict when the ice margin was relatively stable, marking the extent of the late Pleistocene Hawea Advance (Thomson, 2009) during the global LGM. Weak stratification suggests reworking in water and therefore an ice-proximal setting as opposed to subglacial deposition. Wyshnytzky (2013) recognises a dynamic ice margin during this period when the recessional moraine was formed. At ~18 ka the glacier began to retreat causing localised ponding between the ice front and the terminal moraine. Weakly bedded silts and sands with small ripples and cobble to boulder-sized dropstones were deposited, interpreted by Armstrong (2010) to represent ice-proximal lacustrine sedimentation. Armstrong also noted the localised extent of ice-contact lake deposits from dropstones with well-preserved impact structures, interpreted to have been deposited from a subaqueous debris flow or via ice-rafted debris. Wyshnytzky (2013) highlighted the dominance of water in all ice-proximal deposits observed in the Hawea moraine exposures. The ice-contact Lake Hawea began to drain when meltwater cut through the moraine/outwash fan-head.

4.5. Lake Ohau

4.5.1. Existing knowledge

Webb (2009) investigated landform development in the Ohau Valley. The geomorphological map includes lateral moraines, trimlines, an abundance of fluvial channels, and proglacial outwash deposits. Lithofacies associations from sediment exposures above the lake were grouped into massive, matrix-supported diamict facies interpreted by Webb (2009) to have formed in a low-energy sub-aqueous environment, either in expanding supraglacial lakes such as those described by Kirkbride (1993), or an ice-marginal lake impounded between the outwash fan-head and retreating ice mass (Mager and Fitzsimons, 2007). The well-preserved terminal moraine complex is composed of glaciolacustrine and ice-marginal diamict. These are capped in places by large volumes of angular supraglacial debris. Clasts within the diamict facies and boulders observed on the lake shore also lack striations or clast faceting. Webb (2009) suggests a lack of contact with the base of the glacier, and inferred a supraglacial origin from valley side material entering the system from rock falls and avalanches, transported to and deposited in the ice-marginal area.

Proglacial lake sedimentation >100 m thick was investigated by seismic imaging by Krause (2017), 35 m of which were interpreted as fluvially derived reworked sediments (rhythmites). The seismic stratigraphic units suggested by Krause (2017) correspond to depositional environments from minimal ice-contact subglacial till, overlain by glaciolacustrine and meltwater discharge deposits. Krause (2017) identified a large amount of highly stratified sediments. This indicates that the majority is most likely rhythmites, corresponding to a low-energy depositional environment (e.g. Carrivick and Russell, 2007) and interpreted by Krause (2017) as transportation via underflows. Massive, clast-supported diamicts were also attributed by Webb (2009) to deposition in a glaciolacustrine environment, formed either as a sediment-density flow or debris flow in an ice-marginal lake. Due to the stratified nature of the deposits, it is possible that they were laid down during a period with a stable ice margin or at a time when the glacial margin had receded enough to no longer be proglacial. Ground Penetrating Radar (GPR) profiles from Webb (2009) confirmed glaciotectonic deformation consistent with either meltout of buried ice or minor compression associated with an oscillating or advancing ice margin.

4.5.2. Evolution interpretation

Terrestrial cosmogenic nuclide (TCN) dating by Putnam et al. (2013a) constrains local LGM glacial recession into the Ohau trough to 17 690 ± 350. By 17 380 ± 510 the Ohau glacier tongue had retreated by as much as 24 km up-valley, and ~40% reduction in the length of the Ohau glacier tongue over approximately 300 yrs, with a net retreat rate of 77 m yr^{-1}. A debris-covered Ohau Glacier abutted against the outwash fan-head which was built up due to prolonged steady-state conditions (Webb, 2009). These outwash sediments, situated >10 m above the modern-day lake level, combined with the terminal moraine complex dammed the outflow of the lake.

4.6. Lake Pukaki

4.6.1. Existing knowledge

Lake Pukaki lies between 34 and 47 m higher than its natural level due to dam construction for hydroelectric water storage in 1979. The natural surface of Lake Pukaki (prior to human control) was measured at an elevation of 484 m a.s.l. Aerial photographs, topographic contour maps and geotechnical reports that pre-date the artificial lake raise report evidence of well-developed palaeo-shorelines and raised beach deposits. These were observed by Bunting (1977) at 520 m a.s.l and have since been submerged. These high-stand LGM lake levels are dated to 16 033 ± 750 cal yrs BP. based on radiocarbon dating of ice-retreat landforms by Moar (1980).

Five main sediment-landform associations in the Pukaki region were first identified by Speight (1963), with classification according to relative age. Speight interpreted the landform associations in the Lake Pukaki area as products of push moraine construction on the ice-contact faces of proglacial outwash fans. Evans et al. (2013) have since identified a total of 7 lithofacies that record pulsed subaqueous grounding line fan progradation, cohesionless debris flows, underflow activity and rhythmite deposition by suspension settling, iceberg rafting and dropstones, as well as pulsed traction current activity. The extent of lacustrine deposits that exist within the moraines that bound the southern shores of Lake Pukaki is significant. Mager and Fitzsimons (2007) concluded that > 50% of the sediment is fine-grained lake sediment with numerous widespread dropstones. Localised disturbance of these deposits by glaciotectonic deformation record minor re-advances by the glacier snout and emplacement of the glaciotectonite. Overall, the Lake Pukaki sediments largely contain glaciolacustrine facies, including deltaic foresets, laminated silts and sands, diamictons and ice-rafted material.

4.6.2. Evolution interpretation

Due to numerous investigations into the Pukaki moraine exposures, differing process-evolution models have evolved for the deglaciation at Lake Pukaki. Sediments near the lake outlet indicate initial formation of the lake between 510 and 520 m a.s.l. with initial lacustrine deposition of ice-proximal, well-bedded sands and localised coarse angular gravel accumulations likely derived from melting icebergs (Barrell and Read, 2014). Mager and Fitzsimons (2007) invoked the damming of lake water by a former latero-frontal moraine arc characterised by a thick ice core, thereby explaining why lake sediments lie at a higher altitude than the older moraine arcs that could have acted as a dam. However, Evans et al. (2013) state evidence that is diagnostic of active temperate glacier recession from a glacially overridden outwash fan-head. They also argue that the glaciolacustrine deposits only lie above the altitude of the outwash fan-head/lateral moraine arc where they have been glaciotectonically compressed.

4.7. Lake Tekapo

4.7.1. Existing knowledge

Speight (1942) described ancient shorelines and noted high level beach deposits near the Lake Tekapo outlet, including an uppermost beach ~ 20 m above the lake level before the construction of the control dam, and another of 'less importance' ~3 m higher. Pickrill and Irwin (1983) also noted the presence of varves (turbidites) in the sediment along the shores of Lake Tekapo. They are visible today in long cliff exposures ~20 m high to the north of Mt John. However, they were later acknowledged by Gage (1985) to be pre-LGM. Ages from Fitzsimons (1997) suggest that the Mt. John and Tekapo surfaces were formed ~18 ka and 15 ka respectively.

4.7.2. New mapping

Palaeo-shorelines are consistently exposed around Lake Tekapo, Lake Alexandrina and Lake McGregor (Fig. 10). The most prominent shorelines occur around Lake Alexandrina (7.2 km long, 27 m deep) where an almost continuous series of at least three palaeo-shorelines have been mapped at elevations of 721 m, 726 m, and 740 m a.s.l respectively (Fig. 10C). From field observations (Fig. 11A and B), the palaeo-shorelines are eroded into glacial diamict. The level that has the highest geomorphological imprint around Lake Tekapo is the 740 m lake. This level can be traced continuously from a topset-foreset juncture of a palaeo-delta on the western side of Lake Alexandrina (Fig. 10B). As the outlet from Lake Tekapo was incised, a series of lake terraces formed at the margins of the falling lake. The presence of three well-developed palaeo-lake shorelines between 740 m and 721 m a.s.l indicates a sequence of lake level changes and a complex lake-level history. The abandonment of the shorelines is presumed to have been caused by lake drainage.

There are also a series of perched, flat-topped delta terraces

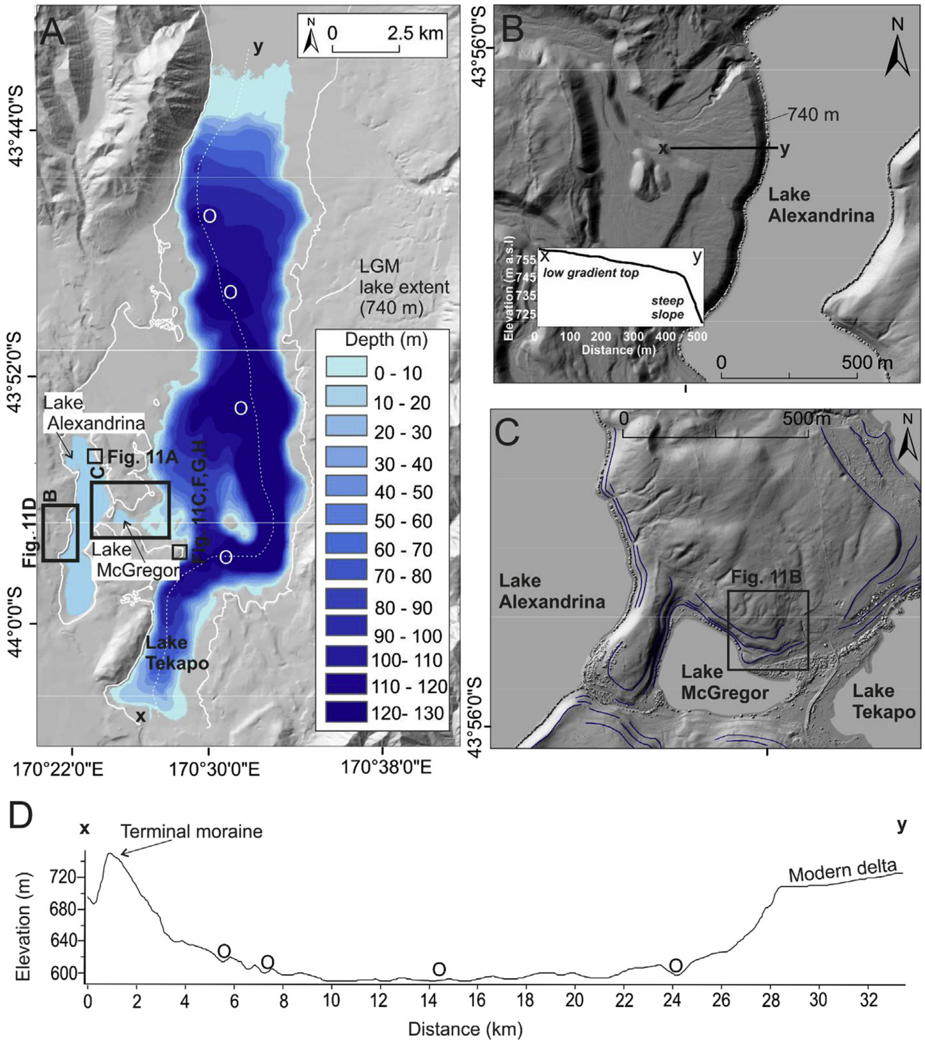

Fig. 10. Regional overview of Lake Tekapo showing the location of high-resolution topography mapping (**B, C**) and field sites (Fig. 11) White contour indicates LGM lake extent at 740 m. Rasterised lake bathymetry is displayed and lake floor elevation profile of the over-deepened glaciolacustrine basin is presented in D, over-deepenings are marked by O. **B.** Impressive palaeo-delta on the western side of Lake Alexandrina, topset-foreset juncture was measured at 740 m a.s.l. **C.** Series of three, relatively continuous palaeo-shorelines mapped between Lakes Alexandrina, McGregor and Tekapo (at 721 m, 726 m, and 740 m a.s.l. respectively) showing they must once have been part of an enlarged lake system.

along the margins of Lake Alexandrina (Fig. 10B and C) at 770 m and 740 m a.s.l. These are interpreted to be Gilbert-type deltas where proglacial meltwater streams have entered the lake (Ashley, 1975; Longhtano, 2008; Bell, 2009; Slaymaker, 2011; Dietrich et al., 2017). These glaciolacustrine deltas were fed by a fluvial network issues from the retreating ice margin (cf. Dietrich et al., 2017; Davies et al., 2018). There are a series of parallel palaeo-channels that feed the raised deltas. The higher delta at 770 m corresponds to the overridden moraine elevation at the southern shore of Lake Tekapo.

4.7.3. Evolution interpretation

Recession of the Tekapo Glacier from its terminal moraine bordering the southern end of the lake enabled the first stage of lake formation. Lake Alexandrina has been previously recorded as a kettle hole, or as a small water table lake which lies in an enclosed hollow within fields of ablation moraine (Viner, 1987). However, there is good correlation between the continuous shorelines mapped around Lake Alexandrina and to those mapped around Lake Tekapo which match in elevation. High-level beach deposits indicate that both Lake Alexandrina and Lake McGregor are remnants of a larger Lake Tekapo that have become disconnected via lake level lowering. Thus, much of the southern margins of the lake were apparently subject to the same deglacial environment and evolution. Well-developed shorelines indicate a stable lake level, and also the prominent deltas needed time to aggrade, indicating the enlarged Lake Tekapo must have been quite long-lived. Water levels in Lake McGregor fluctuate with the controlled water storage levels in Lake Tekapo.

4.8. Lake Heron

4.8.1. Existing knowledge

Lake Heron has a maximum depth of 37 m. Due to the invasion of alluvial fans along its margins, much of the lake is shallow, with 60% being <5 m deep (Bowden et al., 1983). Lake Heron is ponded by a side-valley fan, occupying a depression with a northeast-southwest trending orientation. The lake has been significantly modified by post-glacial alluvium, and its perimeter shape is largely controlled by alluvial fans that splay from the northeast and southwest. Lake Heron also lies in an area where the ice was not thick enough to excavate a deep trough (Pugh, 2008).

Mabin (1984) noted small areas of generally flat ground near the

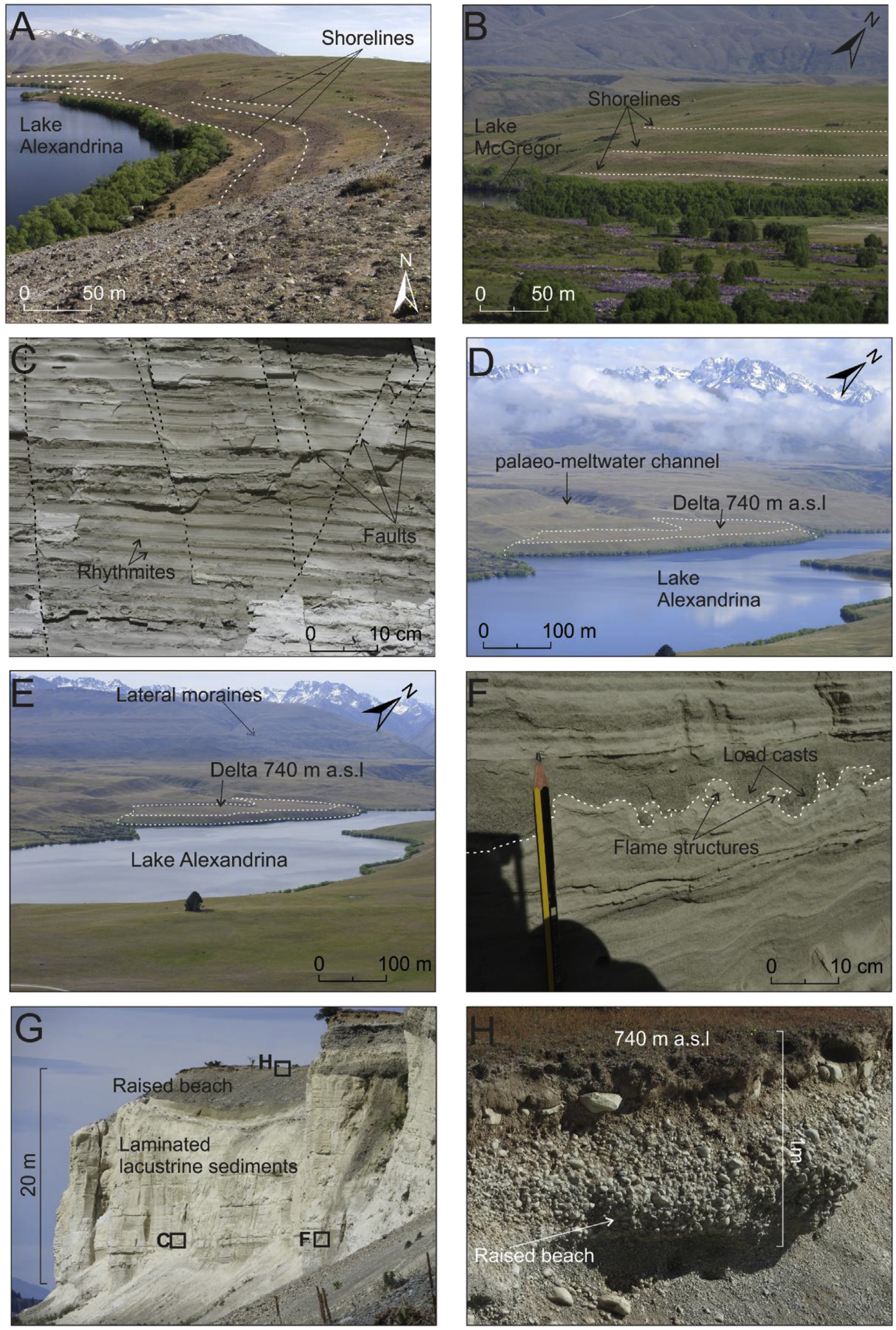

Fig. 11. A. Series of three palaeo-lake shorelines around Lake Alexandrina. **B.** Series of three palaeo-lake shorelines around Lake McGregor **C.** Heavily faulted and fractured rhythmite deposits in >20 m-high cliff exposure on western shores of Lake Tekapo. Interpreted as evidence for glaciotectonic disturbance **D.** Raised delta sets on the western side of Lake Alexandrina (740 m a.s.l) **E.** Raised delta on the western side of Lake Alexandrina **F.** Deformed glaciolacustrine sands and silts in >20 m-high cliff exposure on western shores of Lake Tekapo **G.** > 20 m-high glaciolacustrine rythmites at Tekapo bluffs, southwestern shores of Lake Tekapo. **H.** Very well-sorted, well-rounded cobbles and pebbles interpreted to be a raised beach deposit from the 740 m high lake level.

southern shores of Lake Heron and interpreted them as former lake shorelines (Fig. 12C). Pugh (2008) provides a comprehensive study of the late Quaternary environmental history of the Lake Heron Basin and identified palaeo-shorelines between 700 m and 737 m a.s.l., cut into glacial deposits around the southern shores of the lake. A former lake outlet at the southern end of the lake, with an elevation of 697 m, also indicates that the lake surface was previously higher and that the lake once drained towards the south via the channel currently occupied by Gentleman Smith Stream. Other palaeo-shorelines have been observed a further 9 km up-valley (Speight, 1943). Burrows and Russell (1975) describe lake benches near Prospect Hill, the most prominent at 737 m, and 689 m a.s.l. They are identified as broad, flat surfaces, usually between 5 and 40 m wide and primarily constructed into glacial deposits. Steep scarps up to 2 m high mark the inner and outer margins of the shorelines which represent the abandonment of the land surface by the lake. Raised delta fans at the ends of the meltwater channels on the south-western margin of the lake are further evidence of the

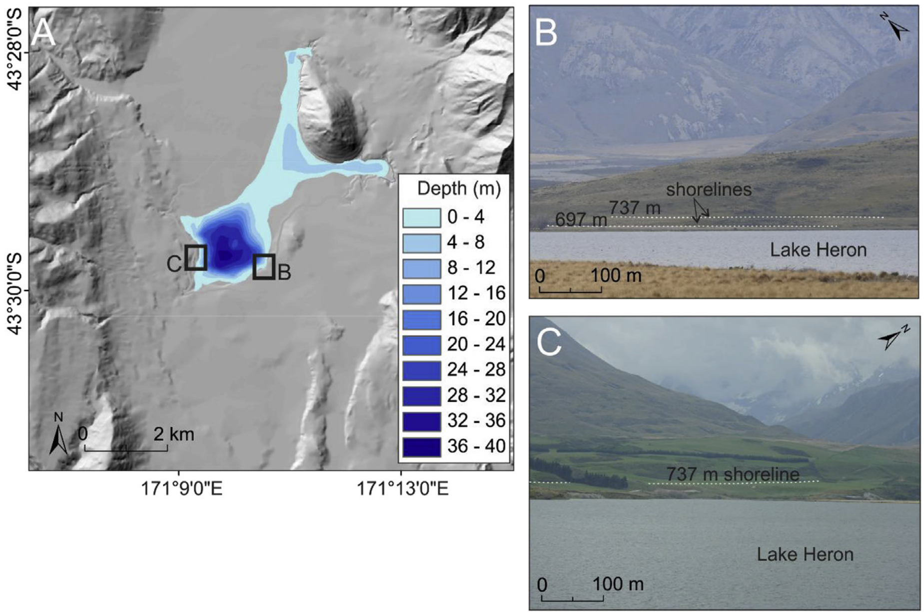

Fig. 12. A. Regional overview of Lake Heron and rasterised bathymetry showing the locations of field photographs. **B.** Field evidence of 697 m and 737 m palaeo-shorelines **C.** Field evidence of 737 m palaeo-shoreline.

existence of a previously much larger lake.

4.8.2. Evolution interpretation

Burrows and Russell (1975) suggested that a re-advance blocked the northern end of the basin, known as the Lake Stream Valley, which filled the basin with meltwater and formed a lake. However, Pugh (2008) states that the lake is more likely to have existed during the retreat phase of the Lake Heron lobe rather than from a re-advance. Degradation terraces occur within the outwash recording fluctuations in lake levels. These terraces are cut into the former alluvial fans (Fig. 12B and C) suggest that the fans have made adjustments according to variations in lake base level. The relationship between glacial deposits and alluvial fans suggest that the fans are largely a post-glacial feature. The Cameron River fan occurs below the height of kame terraces and former lake benches along the western margins of Lake Heron. Its fluvial outwash surface is traceable to the present-day Cameron Glacier, 8 km up-valley. This suggests that rapid fan growth occurred following ice retreat. Pugh (2008) suggested that initial down-wasting of the ice-front was succeeded by the development of a proglacial lake, as indicated by palaeo-shorelines, dammed between the ice margin and an area of hummocky terrain at the southern margin.

4.9. Lake Sumner

The evolution of Lake Sumner is not recorded in the literature. However, palaeo-Shorelines at Lake Sumner have been mapped in this study at the south-eastern edge near the outlet of the lake (Fig. 13E). They do not completely encircle the lake and are only present at the southern extremity between 540 m and 550 m a.s.l.

4.10. Lake Coleridge

The present elevation and size of Lake Coleridge is owed to the impounding alluvium in the plain of the Harper River (Barrell et al., 2011). A vertical exposure of ~15 m observed in the field, reveals a massive, creamy coloured diamicton (Fig. 13B). The faintly stratified and rounded boulders strongly suggest aqueous deposition. The striae on the boulders reveal that ice was proximal. This unit is interpreted as a subaqueous diamicton. At the head of Lake Coleridge there is also a 4.5 m thick exposure of glaciolacustrine sediment. The sequence is capped by large angular boulders and the sediments are interbedded with coarse gravel deposits that have ~10 m lateral extent. Very steep, climbing ripples within the laminated silts and sands indicate high energy and rapid deposition. According to Shulmeister et al. (2010a,b) and Putnam et al. (2013b), ice from the Wilberforce Glacier receded in the area ~18 000 yrs ago.

5. Discussion

Our synthesis and critical review of the hitherto very disparate literature on NZ LGM lakes is combined in this discussion with our own novel detailed identification, mapping, and analysis of landforms and sediments associated with these lakes. That combination produces the first holistic synthesis of the distribution, geometry, landforms and sediments, and evolution of NZ LGM lakes. We discuss each of these properties of the lakes below and highlight improved process-understanding that leads us to comment and make comparisons on modern analogues worldwide.

5.1. Distribution and geometry

The larger LGM lakes in South Island, such as Wakatipu, Wanaka, Hawea, Ohau, Pukaki, Tekapo, and Coleridge were formed in over-deepened troughs. These troughs are eroded into major inter-montane valleys that have fault lines running along their length (Langridge et al., 2016). The presence of the faults helps to explain why many of the troughs, and hence many of the LGM lakes, extend down-valley beyond the confines of mountain hillslopes, i.e. onto a piedmont setting (e.g. Figs. 9 and 10). A piedmont setting is important to note because in confined valleys such as Wakatipu (Fig. 3), erosion of stratified outwash gravel is assisted by high water tables, whereas when a glacier switches to a piedmont lobe beyond the confined valley mouth (such as the lakes in the Mackenzie basin, Ohau, Pukaki and Tekapo), water tables will almost

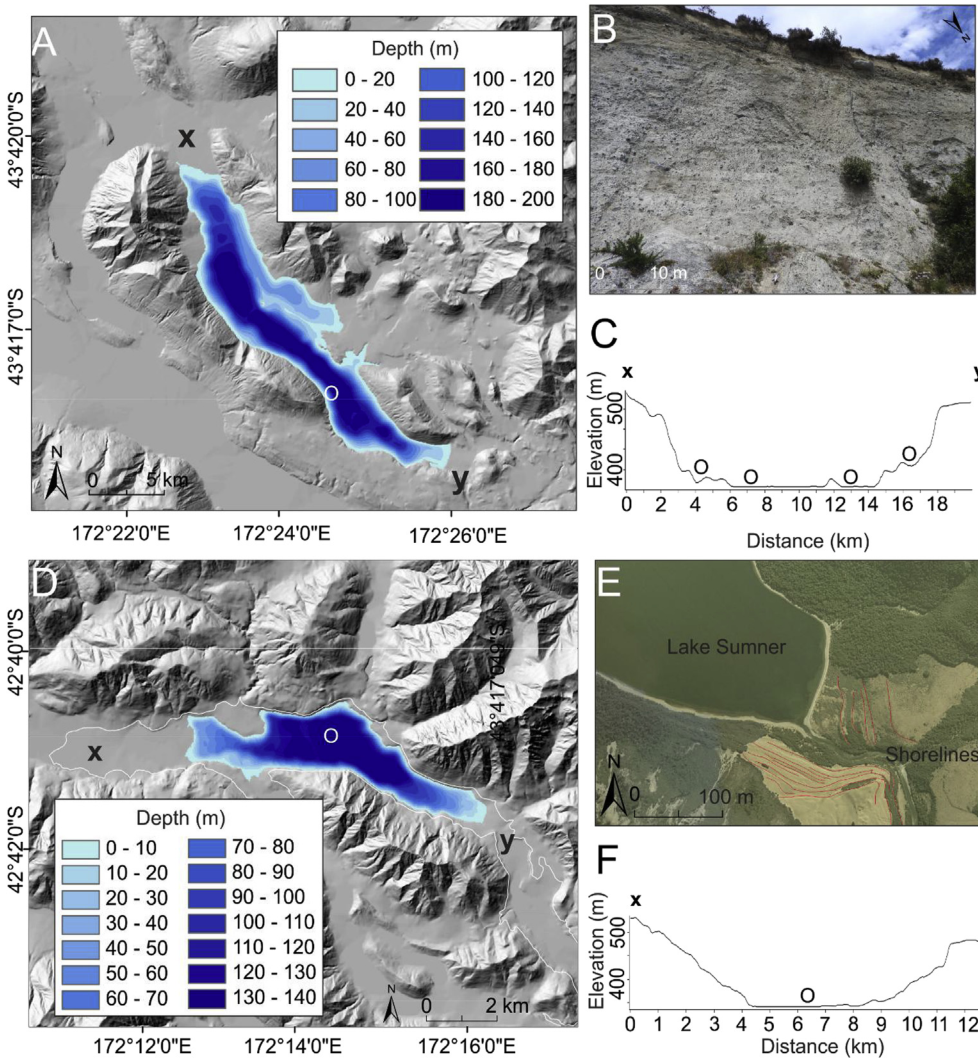

Fig. 13. A. Regional overview of Lake Coleridge with rasterised bathymetry displayed **B.** Subaqueous diamict **C.** Elevation profile of the Lake Coleridge glaciolacustrine basin, over-deepenings are marked by O. **D.** Regional overview of Lake Sumner with rasterised bathymetry displayed **E.** Palaeo-shorelines mapped at the south-eastern end of the lake from 0.4 m aerial imagery (LINZ). **F.** Elevation profile of the Lake Sumner glaciolacustrine basin, over-deepenings are marked by O.

certainly drop and ice velocity will decline due to dewatering of the substrate under the ice (Shulmeister et al., 2019).

Some LGM lakes in New Zealand were the size of small ephemeral ponds (e.g. Lake Cleardale, Lake Doubtful; Supplementary Information) and others were tens of kilometres long, several kilometres wide, and lasted many thousands of years (e.g. Lake Wakatipu, Lake Wanaka, Lake Hawea, Lake Pukaki, Lake Tekapo). Nearly all of the larger glacial lakes have stood at higher levels than present, as evidenced by flights of shorelines (Figs. 1, 4 and 10C) and associated glaciolacustrine sediments perched above the modern water levels (Gage, 1975; Suggate et al., 1978; Wellman, 1979). These higher-level lakes, although probably short-lived, were of considerable lateral extent (Table 3). For example, a much-enlarged Lake Wakatipu was previously contiguous with Lake Hayes (Park, 1909; Brodie and Irwin, 1970; Bell, 1982), and represents the greatest increase in volume between the modern day and LGM of 16.5%. Similarly, Lakes Wanaka, Hawea, Ohau, Pukaki and Tekapo have previously been much larger (Table 3), and probably existed in similar positions during earlier interglacials (Wellman, 1979; Fitzharris et al., 2007).

The bathymetry datasets that we have compiled are also helpful for noting that many of the LGM lakes have prominent retrograde bedrock slopes (Figs. 3, 6, 9, 10, 12 and 13). In the glacial systems presented here, the LGM glaciers managed to extend beyond the margins of their over-deepened troughs (Barrell et al., 2011). The initial ice terminus retreat would have been terrestrial before the ice dropped back into its trough down a retrograde slope into progressively deeper water, and consequently with increased potential for flotation of the ice margin. Flotation of the ice margin is important because the onset and development of calving can partly disengage ice-margin dynamics from climate forcing (Carrivick and Tweed, 2013). These LGM lakes would have caused a shift from land-terminating to lacustrine-calving glacier termini that would have accelerated ice terminus recession in many valleys (Thackray et al., 2009; Shulmeister et al., 2010a,b; Rother et al., 2015; Shulmeister et al., 2018b), and consequently ice mass loss would likely have accelerated too.

5.2. Post-glacial adjustment

Throughout this paper we have relied on indicators of palaeo-water levels such as shorelines and delta-top edges that are truly horizontal upon deposition. There has been a suggestion of tilted shorelines in the literature by Wellman (1979), and Putnam et al. (2010a) due to either a glacio-isostatic response, or tectonic processes and differential rates of uplift away from the crest of the Southern Alps. However, no convincing criteria for these tilted shorelines or maps of them is given in either case. This is problematic because even in our high-resolution 'seamless' topography data the shorelines are represented disparately due to spatial variability in formation and preservation. Furthermore, we have realised in the progress of this study that shorelines may be easily confused with kame-moraine sequences. Not all shorelines are depositional. They can be incised into pre-existing sediments by erosional wave action, such as those around Lake Alexandrina, and as the mapping of this study has demonstrated, they are often preserved discontinuously and often in disparate flights such as at Jacks Point and Bible Terrace at Lake Wakatipu where the number of shorelines in each flight varies in number and elevation (Fig. 4).

In this study, we looked for pronounced steps in the profiles of continuous, uninterrupted palaeo-shorelines both in the field and with the high-resolution topography data at Lake Wakatipu and Lake Tekapo. We did not find any evidence of pronounced steps, nor for that matter any evidence of (gradual) tilting. Notwithstanding that an absence of evidence is not evidence of the lack of an event, we note that Cook et al. (2013) also reported that there is no differential, along-lake post-glacial rebound signal from the last ~ 17 000 yrs around Lake Wakatipu. An explanation for the lack of tilting may be that uplift or rebound has not occurred in the vicinity of the LGM lakes because the volume of ice this far down the glacial system is relatively small compared to the area covered, or, the ice margins oscillated rapidly and so was only present for a short duration. We also refute tectonic tilting/faulting because although the intensity of uplift along the Alpine Fault is up to 10 mm/yr^{-1} and amongst the highest rates in the world (Coates and Cox, 2002), the uplift pattern of the Southern Alps is also known to be highly spatially variable (Adams, 1980). The rates of uplift reduce rapidly with distance from the Alpine Fault. Both Lake Wakatipu and Lake Tekapo are a considerable distance away from the Alpine Fault, therefore uplift may not have occurred in the vicinity of these LGM lakes. For example, the Mt John NETT station, which is 68 km from the Alpine Fault, records no significant vertical rise (Beavan et al., 2007).

5.3. Sediment-landform evidence

Palaeo-shorelines are the most abundant glaciolacustrine feature identified in our mapping but as mentioned in the previous section there is widespread variation in shoreline formation and preservation. At some lakes, e.g. Lake Ohau, palaeo-shorelines are completely absent. Palaeo-lake shorelines might be absent at Lake Ohau either because (i) steep bedrock valley sides exist at the northern end of the lake; (ii) the lake is currently at its highest level (due to damming since 1979) and therefore any lacustrine evidence remains underwater; or (iii) there were no major still-stands in lake level long enough to preserve evidence.

Gilbert-type deltas have previously been recognised in sediments at Lake Wanaka (Evans, 2015), and Lake Hawea (Wyshnytzky, 2013). These are from tributary inputs along the sides of the valley and lake rather than at proximal or distal ends of the glacier. Other deltas mapped by us in this study (e.g. the Greenstone River Fan, and the Glenorchy delta at Lake Wakatipu) are consistent in morphology with a Gilbert-type delta, as described by Armstrong (2010), and Hyatt et al. (2012) for similar deposits in the Hawea moraine and in the Bayfield Cliff exposure within the Rakaia valley (Supplementary Information).

Our review of previous sedimentological studies has highlighted unequivocal evidence that these lakes were in contact with active ice during recession from the LGM limits. Dropstones, including some with striae, are diagnostic evidence for these ice-contact environments and this angular, supraglacial material, attributed to iceberg calving at the glacier margin, has been found in sediments at Lake Wanaka (Evans, 2015), Lake Hawea (Armstrong, 2010), Lake Tekapo, and in the Rakaia Valley (Hyatt, 2010; Supplementary Information Fig. 3). Additionally, evidence of the lakes being ice-contact comes from (i) structural evidence for iceberg scouring that has been reported from the Rakaia Valley (Hyatt et al., 2012; Supplementary Information), and (ii) evidence of buried glacial ice at the margins of Lake Pukaki (Evans et al., 2013).

5.4. Process evolution

We cannot deduce absolute timings or durations of lake level stands from the literature or from our mapping. However, there is a consensus in the literature that the major outlet glaciers receded or collapsed over tens of kilometres within decades (e.g. Shulmeister et al., 2005; Putnam et al., 2013a,b). Some evidence of glacier readvances over pre-existing landforms and sediments is represented by glaciotectonic disturbance e.g. at Lake Wanaka (Evans, 2015), Lake Hawea (Armstrong, 2010), and Lake Tekapo (Fig. 11C, F). During the early stages of ice-margin thinning and retreat from LGM terminus positions, meltwater became impounded behind terminal moraines and outwash fan-heads (*sensu* Kirkbride, 2000) and marked the inception of the lakes (e.g. Shulmeister, 2017), for example as at Lake Hawea and lakes in the Rakaia valley (Supplementary Information). During the first phase of lake development the formation of outwash fan-heads, which are primarily a product of high sediment fluxes, was further promoted by the glacier termini being at the lip of their over-deepened basins.

With the onset of glacier margin recession down retrograde slopes, outlet streams cut progressively deeper channels through the unconsolidated deposits of the fan-heads. The lake outlet incision and resultant lowering of lake water levels apparently proceeded in marked flights; i.e. with episodes of lake level stability as evidenced by prominent palaeo-shorelines at all the major LGM lakes, except Lake Ohau as discussed above. In the cases of Lakes Wakatipu and Hawea, this lowering process has been greatly retarded by the outlet streams exposing their beds of comparatively hard bedrock which has formed resistant sills at the lake outlets. The cause of the lowering proceeding in steps is unknown, but using the present-day Lake Tasman as an analogue could indicate a second stage of lake evolution due to ice stagnation and thinning, development of supraglacial ponds which coalesce, and consequently progressively lower ice surfaces to a level of their meltwater rivers (Kirkbride, 1993; Kirkbride and Warren, 1999; Warren and Kirkbride, 2003). As ice surfaces lowered and lake levels lowered, sediment delivery to a proglacial meltwater river would also reduce and even cease as all debris became captured by the (enlarging) lake. The lake length is also important for suspended sediment load. Thus the proglacial river regime would change from aggradation to incision of its channel (Evans et al., 2010; Carrivick and Tweed, 2013) so long as a notch in the moraine walls were eroded. These channels became erosional very quickly. If notch incision could not keep pace with lake level lowering; i.e. ice-margin recession and thinning, then it would become abandoned and this is a possible mechanism for the switch in lake outlet drainage observed at Lake Wakatipu. A third stage of lake development could occur if calving were instigated by the glacier margins both thinning and retreating into water that was sufficiently deep enough to permit flotation. Flotation of an ice margin depends on water depth and lake bed topography and in the bathymetry data we note the existence of underwater pinning points, for example the submerged major hummocks in Lakes Wakatipu, Wanaka, Hawea (Figs. 3 and 6), which would have interrupted the rate of recession and perhaps style as well.

Stagnation of ice and a predominance of thinning over ice-margin recession would have been promoted by extensive debris cover across the glacier termini. Outcrops through moraines on the west coast clearly show that glaciers on the west coast had extensive debris cover during the LGM (Evans et al., 2010). The terrestrial margins of the Pukaki Glacier are diagnostic of active temperate recession by a debris-charged glacier (Evans et al., 2013) as were those in the Ohau valley due to the evidence of coarse, angular boulders in the moraine surrounding the lake (Webb, 2009). Where debris cover was extensive in LGM glaciers it may have preserved thinning tongues, and retreat rates may have lagged behind climate, but only up until the calving threshold was crossed.

The transition from land- to lake-terminating glaciers would have fundamentally altered the primary mechanisms of mass loss and glacier dynamics and could have partly decoupled the ice-margin retreat rate from climatic influences. The retreat rates would have increased, but may still have fluctuated with climate through varying ice velocity to the terminus (Warren and Kirkbride, 2003). The glaciers would have transitioned into a new glacio-climatic regime of enhanced ablation which persisted throughout the period of ice-margin retreat through the lake basin. In this lake-glacier interaction regime, annual net balances are consistently more negative than prior to the non-calving/calving transition. These proglacial lakes most likely established heat advection and wave impact as significant additional mechanisms of ice loss, probably leading to locally enhanced rates of ice recession. This mode of post-LGM retreat was a common feature of glacier systems across the Southern Alps, as evidenced by widespread glaciolacustrine deposits, particularly in the eastern Alps that have been documented in this review and by numerous others (e.g. Speight, 1926; Gage, 1958; Clayton, 1968; Suggate, 1990; Hart, 1996; Mager and Fitzsimons, 2007). Despite the importance of recognizing the influence of the LGM lakes on glacier recession and mass loss, numerical reconstructions of deglaciation from the LGM in New Zealand have not taken calving into account: e.g. Golledge et al. (2012).

5.5. Chronology

Geochronology of deglaciation from the LGM in New Zealand has often been the inspiration for geomorphological mapping and landform-sediment analysis to date, rather than establishing glacier dynamics or further understanding of landscape evolution processes. Cosmogenic dating is now widespread across New Zealand (e.g. Shulmeister et al., 2005, 2010a,b; Winkler, 2009; Putnam et al., 2010a,b; 2013a,b) and has superseded ^{14}C dating. However, few glacial lake deposits in New Zealand have been dated directly; geochronological campaigns have targeted moraines rather than glaciolacustrine landforms and sediments. Most of the targeted moraines represent maximum ice advances, but others represent ice re-advances or still-stands during recession. Furthermore, an unequivocal chronology of Late Quaternary glacial lake deposits has been difficult to attain largely because of sparse material for ^{14}C dating (Burrows, 1975; Suggate et al., 1978; Soons, 1982). Suggate (1965) suggested that the main glacial lakes in the South Island were all last occupied around the same time. However, the timing of greatest glacial extent in the last glacial cycle is not simultaneous across New Zealand (Shulmeister et al., 2019). This timing of greatest glacial extent may or may not be important for the lakes which mostly lie some distance (hundreds of metres to kilometres) up-valley of the hindmost 'LGM' moraine. Where absolute dates are available we have compiled them in Table 4 to assess whether the timings of lake formation and evolution are comparable between sites. We are also able consider if initial glacier terminus retreat rates were synchronous at each lake.

Table 4
Absolute dates available for the formation of each lake investigated in this study. Proglacial lake formation was in progress shortly after c. 18 ka in most lakes. All C^{14} dates have been converted to calendar years using the IntCal09 calibration curve (Riemer et al., 2009). Note, no absolute dates exist for the Tekapo or Hurunui (Lake Sumner) valleys. TCN: Terrestrial Cosmogenic Nuclide. OSL: Optically Stimulated Luminescence. IRSL: Infrared Stimulated Luminescence. *Note: Evans (2015) interpreted this age as retreat from local NZ LGM.

Lake	Age of Formation (ka)	Method	Reference
Wakatipu	17.1 ± 2.6	OSL (feldspar)	Cook et al. (2013)
Wanaka	$29 \pm 5^*$	OSL (quartz)	Evans (2015)
Hawea	17.8	^{14}C	McKellar (1960)
	17.5	TCN (^{10}Be)	Graham et al. (1998)
	<18	IRSL (feldspar)	Wyshnytzky (2013)
Ohau	17.69 ± 0.35	TCN (10Be)	Putnam et al. (2013a)
Pukaki	17.4 ± 1.0	TCN (^{10}Be)	Schaefer et al. (2006)
	18.35 ± 0.39	TCN (^{10}Be)	Putnam et al. (2010b)
	13.492 ± 0.29	^{14}C	Read (1976)
Tekapo	Unknown		
Heron	25	TCN (^{10}Be)	Mabin (1984)
	18.6 ± 1.2	TCN (^{10}Be)	Pugh (2008)
	18.6 ± 0.3	TCN (^{10}Be)	Rother et al. (2014)
Coleridge	16.5 ± 0.9	TCN (^{10}Be)	Shulmeister et al. (2010a)
	17.02 ± 0.070	TCN (^{10}Be)	Putnam et al. (2013b)
Sumner	Unknown		

It is likely that all glacial lakes in their present form post-date the maximum ice advances associated with the LGM (Suggate and Moar, 1970; Moar and Suggate, 1979; Bell, 1982; Nelson et al., 1985). For example, the moraines that encircle the former lakes in Westland (Supplementary Information; Fig. 1A) are inferred by Suggate and Almond (2005) to have formed ca. 19 000–17 000 cal yr BP. Thomas (2018) noted that some lakes (such as the Whataroa and Arahura palaeo-lakes on the west coast) undoubtedly existed at earlier stages too.

Chronological studies from Lake Pukaki (Schaefer et al., 2015), Lake Oahu (Putnam et al., 2013a), and in the Rakaia valley (Putnam et al., 2013b; Supplementary Information) all report recession or collapse of major ice tongues over tens of kilometres during a short period. This rapid recession has been attributed by Shulmeister et al. (2010a) and Evans et al. (2013) to the formation of large and deep proglacial lakes during deglaciation. For example, at Lake Pukaki deglaciation is thought to have commenced c. 18 350 yrs ago (Putnam et al., 2010a,b). By c. 16 400 yrs ago, the glacier had retreated to a position that was more than half the distance from the LGM extent to present day glacier limits (Moar, 1980; Putnam et al., 2010b). Moraine dating in the Rakaia and Ohau valleys (Putnam et al., 2013a,b) indicates that rapid and sustained glacier recession began c. 18 ka and each glacier lost at least 40% of its length within no more than c. 1000 yrs. In further support of these rates of recession being very rapid, ^{10}Be surface-exposure ages for moraine boulders in the Upper Rangitata valley were interpreted by Shulmeister et al. (2018b) to indicate a gradual reduction in glacier surface height between 21 and 17 ka due to a lack of a proglacial lake in the valley. The alternative is that this synchronous retreat was due to accelerated warming at this time (e.g. Putnam et al., 2013a,b). In contrast, Mabin (1987) and Barrell et al. (1996) note the evidence for the existence of a long-lived proglacial lake in the Upper Rangitata valley. Barrell et al. (2019) argues that the geomorphology is consistent with sustained glacier recession and the newer, revised glacial chronology implies that substantial ice-surface lowering of the LGM Rangitata glacier was in progress c. 17.7 ± 0.3 ka, compatible with dating from other eastern valleys of the Southern Alps.

5.6. Modern analogues

The distribution, geometry and evolution of the NZ LGM lakes reviewed and mapped in this study are most obviously similar to the modern proglacial lake-glacier systems in the Mount Cook region. In this modern environment, glacial diamicts are rare and proglacial fluvial reworking of primary glacial deposits leads to a low preservation potential for tills and related subglacial facies (Hambrey and Erhmann, 2004). The literature synthesis in this review has identified that the majority of glaciogenic landforms and sediments produced during the LGM were dominated by water (Shulmeister, 2017; Shulmeister et al., 2019). This is unsurprising given the size of the ablation zones of the NZ LGM glaciers and their altitude. The best modern analogues for such active, temperate, glacial land systems are the humid mountain ranges of the Alaskan west coast and those in southern Iceland where coupled landsystem associations exist. Such glacier systems transport large volumes of englacial and supraglacial debris, derived from extraglacial sources in the mountainous catchments, on to coastal lowlands (Price, 1969; Boulton, 1986; Evans, 2006; Evans and Twigg, 2002; Evans and Hiemstra, 2005; Evans et al., 2010).

In scale, situation and dam type the modern glaciers in South Iceland (e.g. Schomacker et al., 2010; Evans and Orton, 2015) are analogous to the NZ LGM lakes. For example, the Heinabergsjökull/Skalafellsjökull foreland in southeast Iceland constitutes a modern analogue for active temperate piedmont lobes associated with the construction of large outwash fan-heads fed by high glaciofluvial sediment yields (Evans and Orton, 2015). The southern Iceland glaciers have the characteristics of valley glaciers due to topographic confinement by mountain spurs. Since the 1950s the snout of Heinabergsjökull has been in contact with a proglacial lake which has developed on the ice-proximal side of an ice-contact fan/outwash head (Benn et al., 2003; Evans and Orton, 2015). However, it is worth noting that these Icelandic glacier-lake systems are more likely to have stronger winter freeze than the New Zealand systems, even at the LGM.

Comparisons of the NZ LGM lakes can also be made with the massive, debris-covered glaciers in the Everest region, central Himalaya, where very large lakes are forming as supraglacial ponds coalesce. Glacial lake development across the central Himalaya is similar in process to that identified in New Zealand, where ice surface lowering and increasing glacier stagnation has been highlighted to promote supraglacial pond formation and their potential to coalesce into larger lakes where a low glacier surface gradient exists (Watanabe et al., 1994; Richardson and Reynolds, 2000; Quincey et al., 2007; Röhl, 2008; Thompson et al., 2012; King et al., 2017, 2018). The evolutionary trajectory of these large proglacial lakes can be attributed to the development of supraglacial ponds (Kirkbride, 1993; Watanabe et al., 1994; Röhl, 2008; Sakai, 2012; Carrivick and Tweed, 2013; Watson et al., 2018).

The NZ LGM lakes differ in their landform and sediments from those in other Alpine regions of the world. Most obviously, moraine-dammed proglacial lakes are common in mountain valley systems where recession of high altitude, semi-arid mountain glaciers, such as those of the central and northern Andes and the Himalaya. In contrast, New Zealand's mountain valley system favoured barriers created by ice-contact fans and outwash heads rather than terminal moraines. For example, the southern margin of Lake Pukaki is at least partially defined by an outwash fan-head (Hart, 1996; Mager and Fitzsimons, 2007; Armstrong, 2010; Evans et al., 2013). This prevalence of outwash fan-head barriers in New Zealand is due to the efficiency of sediment transfer from the glacier to the fluvial zones which both limits moraine development and encourages outwash fan-head aggradation (Carrivick and Russell, 2007). Similar outwash fan-head dams also occur on the eastern side valleys of Patagonia such as in the Leones, Soler, Nef, and Colonia catchments (Glasser and Jansson, 2005; Harrison et al., 2006).

6. Conclusions

In summary, we have conducted a holistic review of the landform and sedimentary record of ice-contact proglacial lakes during the LGM in South Island, New Zealand. We have combined the synthesis knowledge and specific site details from this review with new high-resolution mapping validated by extensive field investigation, to objectively identify and quantify the distribution and geometry, and evolution of these proglacial lakes.

There are localised constraints to proglacial lake development across New Zealand related to topography, glacier size and meltwater and sediment fluxes. LGM glaciolacustrine sedimentary deposits are widespread in the eastern Southern Alps and whilst present in the west, they are far-less well discerned. In the nine valleys east of the Main Divide examined in this study, it is primarily the geometry of over-deepened glacial troughs that has promoted the formation of a proglacial lake upon ice retreat. Besides providing the over-deepened basins within which the LGM lakes formed, valley geometry, especially the juncture between confined valley sides and piedmont aprons, also promoted the deposition of voluminous outwash fan-head gravel. This gravel evidences the very large, tectonically-exacerbated sediment supply. Additionally, Hyatt et al. (2012) and Shulmeister et al. (2019) have suggested that there was a year-round availability of meltwater during the LGM. These two elements; a virtually unlimited sediment supply and a year-round meltwater supply, have dictated the nature of the glaciolacustrine landforms and sediments. Most noticeably, lake damming was initiated by fan-heads blocking the outlet, such as Lake Hawea, Lake Pukaki, Lake Tekapo, and in the Rakaia Valley (Supplementary Information). The thickness of the fan barriers are typically tens to hundreds of metres thick. The barriers formed during long-lived, pervasive obstructions to water and sediment flow down-valley. There is unequivocal evidence that these lakes were ice-contact, receiving icebergs from a calving, and thus probably floating ice-margin. LGM lakes that have drained and are no longer present (e.g. Soons and Gullentops, 1973) were also unequivocally ice-marginal (Supplementary Information).

The synthesised proglacial lake evidence of this paper may be used to produce detailed reconstructions of ice margin recession, evolving ice dynamics and the development of palaeo-lake systems. For example, determining where the ice margin was and how thick it was in relation to the lake levels would be instructive for comparing the rate of retreat and thinning of glacier lobes with the rate of lake level changes. These data are also useful for improving the parameterisation of numerical models aiming to refine reconstructions of regional ice dynamics over long timescales.

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Appendix A. Supplementary data

Supplementary data to this article can be found online at https://doi.org/10.1016/j.quascirev.2019.03.035.

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