Global and Planetary Change 147 (2016) 1–11

Contents lists available at ScienceDirect

Global and Planetary Change

journal homepage: www.elsevier.com/locate/gloplacha

Bathymetric evolution of Tasman Glacier terminal lake, New Zealand, as determined by remote surveying techniques

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ARTICLE INFO

Article history:
Received 10 December 2015
Received in revised form 17 October 2016
Accepted 18 October 2016
Available online 19 October 2016

ABSTRACT

Processes that drive iceberg calving at the margins of freshwater terminating glaciers are still poorly understood. This knowledge-gap is in part due to the challenge of obtaining good in situ data in a highly dynamic and dangerous environment. We are using emerging remote technologies, in the form of a remote controlled jet boat to survey bathymetry, and Structure from Motion (SfM) to characterize terminus morphology, to better understand relationships between lake growth and terminus evolution. Comparison of results between the jet boat mounted dual-frequency Garmin fish-finder with an Odom Echotrac DF3200 MKII with 200/38 kHz dual-frequency transducer, showed that after a sound velocity adjustment, the remote survey obtained depth data within ±1 m of the higher grade survey equipment. Water depths of up to 240 m were recorded only 100 m away from the terminus, and subaerial cliff height ranged from around 6 to 33 m, with the central region of the terminus more likely to experience buoyancy. Subaqueous ice ramps are ephemeral features, and in 2015 multiple ice ramps extended out into the lake from the terminus by 100–200 m. The consistent location of some of the subaqueous ramps between surveys may indicate that other processes, for example, subglacial hydrology, also influence evolving terminus morphology.

1. Introduction

Global glacier recession is increasing the number of glaciers that terminate in proglacial lakes (Thompson et al., 2012; Carrivick and Tweed, 2013), resulting in a shift in the importance of ice-loss processes from melting to iceberg calving (Kirkbride and Warren, 1999). Research at calving glacier margins has been, and continues to be, predominantly focussed on marine terminating glaciers (Benn et al., 2007; Sakakibara et al., 2013; Åström et al., 2014), and although many processes are similar, lake-calving rates are generally an order of magnitude lower than those in marine environments (e.g. Warren et al., 2001; Warren and Kirkbride, 2003; Boyce et al., 2007; Truffer and Motyka, 2016). While proglacial lakes are a key water resource, they can also be a hazard, with floods from moraine dam failure causing loss of life and infrastructure (Henderson et al., 2011; Bolch et al., 2012; Thompson et al., 2012). However, instabilities at the terminal face of calving glaciers means that data acquisition can be hazardous and therefore data are sparse.

New Zealand has over 3100 glaciers and only a small proportion (<1%) of these have developed proglacial lakes (Chinn, 1999), but over one third of New Zealand's perennial ice is contained in these lake-calving glaciers (Chinn, 2001). There is still no 'law' that can adequately explain all the variability recorded at calving margins (Benn et al., 2007; Amundson and Truffer, 2010; Nick et al., 2010). Previous research has shown that calving rates increase with water depth (Warren et al., 1995; Benn et al., 2007) and can be inversely proportional to height above buoyancy, but calving rates tend to have weak relationships with velocity gradients (van der Veen, 2002; Benn et al., 2007). However, seasonal advance and retreat recorded at marine outlet glaciers has been simulated by consideration of surface and basal crevasse penetration (Nick et al., 2010), and ice thickness at the terminus has been identified as a first order control on calving rate (Amundson and Truffer, 2010). Although water depth only explains part of the calving problem, knowledge of bathymetry in the terminus region and how it evolves over time is still of value, as water depth strongly influences terminus buoyancy, and consequently calving behaviour (Warren et al., 2001; Benn et al., 2007). Prior to a glacier tongue reaching flotation, iceberg calving is dominated by melting at the waterline, which drives high-frequency, low magnitude subaerial calving events (Röhl, 2006). Subaerial calving can result in the development of a subaqueous ice ramp, which will eventually buoyantly calve (van der Veen, 2002; Robertson et al., 2012). Ongoing surface thinning (downwasting) results in the terminus becoming buoyant (Warren et al., 2001; Boyce et al., 2007; Trüssel et al., 2015), resulting in low frequency, high magnitude calving events (Dykes et al., 2011).

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http://dx.doi.org/10.1016/j.gloplacha.2016.10.010

Whether the terminus of a calving glacier is buoyant depends on the flotation thickness (H_F) at the calving margin (Cuffey and Paterson, 2010, p 121):

$$H_F = \frac{\rho_w}{\rho_i} D_W \quad (1)$$

where ρ_w is water density (≈ 1000 kg m^{-3}, varying with temperature, suspended sediment concentration and conductivity), ρ_i ice density (917 kg m^{-3}) and D_W water depth. In freshwater environments $H_F \approx 1.1\ D_W$. The height required by the terminal ice cliff in order for the terminus to remain above buoyancy (H_O) is (Benn et al., 2007, p 148):

$$H_O = H_T - \frac{\rho_w}{\rho_i} D_W \quad (2)$$

where H_T is the ice thickness at the terminus.

Although D_W is technically straightforward to measure, it is hindered by safety issues. However, a remote-controlled survey boat with appropriate instrumentation, can improve both safety and spatial resolution of survey data (Neal et al., 2012), mapping areas deemed too unsafe for peopled boats. Likewise, the technique of constructing high-resolution digital elevation models (DEMs) from digital photography, known as Structure from Motion (SfM), can, if well defined in real-world coordinates, be used to derive topographic measurement (Westoby et al., 2012). For example, Ryan et al. (2015) recently employed a combination of an unmanned aerial vehicle (UAV) and SfM to quantify ice flux and terminus evolution at Store Glacier, Greenland.

Here we present results derived from these new technologies to: 1) Update the bathymetry of the rapidly enlarging Tasman Lake, 2) Determine the size and extent of the ice ramp, 3) Combine bathymetric results and 3-D modelling of the subaerial cliff to determine H_F and consider spatial variability in terminus buoyancy.

2. Study site and previous surveys

Tasman Glacier, New Zealand's largest store of glacial ice (Chinn, 2001), is rapidly retreating (Dykes et al., 2011; Purdie, 2013). High net annual accumulation of around 4 m water equivalent (Purdie et al., 2011a,b), and subdued melting under debris cover (Purdie and Fitzharris, 1999), cannot offset large losses from iceberg calving (Dykes, 2013). Since 2007, Tasman Glacier appears to have exhibited a switch to buoyancy driven calving (Dykes, 2013), meaning that calving events are less frequent, but typically of larger magnitude. To fully understand the changes occurring at the terminus of the Tasman Glacier, we need to more accurately characterize the physical evolution at the ice-lake interface.

As a large debris-covered glacier located on the eastern side of the Southern Alps, the Tasman Glacier has been slow to adjust to climate warming since the termination of the Little Ice Age (LIA) (Schaefer et al., 2009). Prior to the development of a proglacial lake in 1990, the Tasman Glacier maintained its LIA position, losing volume, by significant surface lowering (down-wasting) (Hochstein et al., 1995; Kirkbride and Warren, 1997; Chinn et al., 2012). In contrast over this same time period, the fast reacting Fox and Franz Josef Glaciers immediately west of the Southern Alps, have undergone a number of advance and retreat phases, with the most recent advance (~300 m) ceasing in 2009; both glaciers have since rapidly retreated (~700 m) (Purdie et al., 2014). Today Tasman Glacier is around 22 km long and covers an area of ~95 km^2 (excluding tributaries) (Fig. 1).

In the early 1960s, aerial photography revealed that melt ponds were already forming on the debris-covered tongue of the glacier (Kirkbride, 1993). Landsat imagery from November 1989 showed that these melt ponds had become quite extensive, although they had yet to coalesce into a single body of water. By December 1990, the newly formed Tasman Lake covered an area of around 1.65 km^2 (Fig. 1C).

The first bathymetric survey conducted in 1993, showed that Tasman Lake was 1.95 km^2 and up to 125 m deep (Hochstein et al., 1995). Since then surveys have been completed in 1995 (Watson, 1995), 2002 (Röhl, 2005), and 2008 (Dykes et al., 2011). During that period Tasman Lake increased to 5.96 km^2 (Fig. 1) and attained a maximum depth of 240 m. The 2008 survey also confirmed that there was a subaqueous ice ramp, around 130 m long, extending into the lake (Dykes et al., 2011; Robertson et al., 2012).

Accurate estimates of H_T at Tasman Glacier have proved challenging, due to large ice thickness, the presence of surface-debris, uncertainty about till thickness at the bed, and more recently, potential flotation (Broadbent, 1974; Watson, 1995; Hart, 2014). However, a previous estimate by Broadbent (1974) that ice was 220 m thick in the terminus region is in agreement with recent bathymetric surveys (Röhl, 2005; Dykes et al., 2011).

3. Methods

3.1. Bathymetric survey

In April 2013, May 2014 and November 2015 bathymetric surveys were undertaken on Tasman Lake. The 2013 and 2014 surveys each collected in excess of 40 km of data over the entire lake area, whereas the 2015 survey focused solely on the terminus region. For the 2013 and 2014 surveys an Odom Echotrac MKII echo sounder with a 200/38 kHz dual-frequency transducer was rigidly attached on an over-the-side mount to the Glacier Explorers Macboat, providing a very stable platform (Fig. 2). A Richard Branker Research (RBR) XR620 conductivity-temperature-density (CTD) probe was used to measure temperature, salinity and pressure through the water column at the start of the survey, which were then used in the UNESCO standard Chen and Millero formula (Chen and Millero, 1977) to calculate sound velocity. A bar check – where a horizontal bar is lowered below the transducer and held at known depths, thus correcting for the draught of the sounder and confirming sound velocity, was completed to calibrate the Odom sounder.

A Trimble R8 real time kinematic (RTK) global navigation survey system (GNSS) using 10 Hz frequency sampling was used for positioning and heave compensation. The base station was set-up <5 km from the boat for short baselines. In 2013, the base location was established using a network adjustment in Trimble Business Centre and static survey data from the intended base point, four additional established Land Information New Zealand (LINZ) survey marks nearby, and the Mount John Position NZ station. All lake depths reported have been adjusted to the 2013 level to enable comparison.

The horizontal uncertainty for the Odom was ± 0.18 m (Table 1). Trimble HYDRO PRO software was used to combine the soundings with position and heave, providing real-time information on water depth, speed and run-lines. Critically, accurate water levels were measured with RTK GNSS around the lake edge each year, allowing true comparisons to be made between bathymetric lake-floor models regardless of the lake water level, an important aspect of the project and of vital use to future studies. These measurements also enable land measurements to be combined seamlessly with depths. A final estimate for the vertical uncertainty of data from the Odom echo sounder is ± 1.1 and ± 5.9 m at 50 and 250 m depth respectively (Table 2).

Although survey data density was good in 2013 (Fig. 3), the threat of rock-fall and ice collapse meant that data collection was sparse within 100 m of the calving face. Therefore in 2014 the Odom data were complimented by a remote controlled boat survey proximal to the terminal face.

A dual-frequency Garmin Fishfinder 400C sonar, with an estimated accuracy of ± 5–10 m, was mounted on the bottom of a JetTec 1.3 m aluminium hulled jet boat. A Trimble R8 GNSS antenna was mounted