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POST-1850s CHANGES IN GLACIER BENITO, NORTH PATAGONIAN ICEFIELD, CHILE

VANESSA WINCHESTER[1], MARTIN SESSIONS[2], JAMMIE VALDIVIA CERDA[3], OLAF WÜNDRICH[4], SAMUEL CLEMMENS[5], MEGAN NASH[6], NEIL F GLASSER[5]

1. Oxford University Centre for the Environment, University of Oxford, OX1 3QY, UK.
vanessa.winchester@geog.ox.ac.uk
2. 9 Eliza St, Amaroo, ACT 2914, Australia.
3 Universidad Tecnologica Metropolitana (UTEM) Santiago, Chile
4. Colibri Ventura, Casilla 113, Coyhaique, XI Region de Aysén, Chile
5. Institute of Geography & Earth Sciences, Aberystwyth University, Llandinam Building, Penglais Campus, Aberystwyth, SY23 3DB, UK.
6. Australian National University, ACT

Abstract

In southern South America field studies validating glacier recession are scant and of brief duration. This studyusing field data from 1972/73, 2007 and 2001, presents the longest glacier study yet undertaken in the region. Rates of thinning of Glacier Benito, a temperate outlet glacier on the west side of the North Patagonian Icefield (NPI), were derived using data collected by the British Joint Services Expedition in 1972/73 and subsequent data collected in 2007 and 2011.. Rates of recession are based on dendrochronological dating for the terminal moraines; these dates indicate that the last cold period reached its maximum in the 1850s: the earliest date yet estimated for the beginning of “Little Ice Age” glacier retreat around the NPI. Estimated ice front recession from the LIA (1858) to 2002 is almost 2 km, with rates increasing dramatically from 17.7 m yr^{-1} between 1975-1998, to almost 170 m yr^{-1}, 1998-2002. Over the 34-year period between xxxx and yyyy from the first survey in 1972/73, the lower glacier thinned by nearly 150 m.

Introduction

There is now compelling evidence that climate warming is impacting glaciers worldwide (EPICA Community Members 2006), with glacier fluctuations widely acknowledged as relatively reliable indicators of climate change (Rosenblüth *et al.* 1997). However, warming varies spatially and temporally between regions and the magnitude of changes can be difficult to determine especially in climatically critical regions in southern South America such as the North and South Patagonian Icefields (NPI and SPI) where, due to challenging weather conditions, instrument records are few, widely separated, often incomplete and of relatively short duration; thus proxy records are required to fill the gap. Proxy indicators of climatic change in this region have included lichenometry and dendrochronology (Winchester and Harrison 1994, 1996; Villalba et al. 2003, Koch 2009), pollen analysis and sediment cores (Bennett et al. 2000; Markgraf, et al. 2007), radiocarbon and cosmogenic dating of moraines (Glasser et al. 2002; Harrison et al. 2008) and ice cores (Matsuoka and Naruse 1999; http://www.glaciologia.cl/spi.html#index 2002; Vimeux et al. 2008). Together with geomorphological mapping, these indicators present a picture of icefields highly sensitive to climate change (Glasser at al. 2004; 2011).

The climate on the west coast of southern South America is mediated by three ocean-atmosphere systems: the cold Humboldt Current flowing northwards from the circumpolar Southern Ocean, the westerly winds that, rising over the Andes, create a steep west-east climatic gradient, and a teleconnection with the equatorial El Niño/Southern Oscillation **(**ENSO) (Winchester et al, 1999; Daniels and Veblen 2000); east of the Andes inputs from continental sources from the E–NE are sometimes mixed with circumpolar air masses (Vimeux et al. 2008).

Together the NPI and SPI cover some 16,950 km^2 (Barcaza et al. 2009), and of this the total ice area of the NPI in 2001 was 3953 km^2 (Rivera et al, 2007). Separated today by a 100 km gap the two Icefields form the largest temperate ice mass in the Southern Hemisphere stretching from 46^0.28'S to 51^0.35'S, but they are now shrinking due to changes in temperature and precipitation affected by latitudinal migration of the westerlies (Cai 2006; Sallée et al. 2010; Shevenell et al. 2011) and altered storm tracks (Lamy et al. 2010).

The shrinking Icefields are globally important owing to their contribution to sea-level rise (Rignot et al. 2003; Glasser et al. 2011). Locally, the famous calving San Rafael Glacier descending the western flank of the NPI is a focus for tourism and, of national importance on the east side of the Andes, new hydroelectric dams are projected for Rio Baker (Vince 2010): this river, the main drainage channel for the NPI's eastern glaciers, empties into the Pacific between the Icefields. Hence, measurements of rates of change on the Icefield and its outlet glaciers are of interest not only with regard to quantifying sea-level rise, but the prospects for tourism and, more importantly, the water supply for Rio Baker lying in the rain shadow east of the Andes (Dussaillant *et al*. 2011).

Glaciological mass balance is controlled by difference between the rate of accumulation in the source area and ablation, with ablation dependent on temperature, debris cover and calving at the terminus, especially where there is tidal activity that may obscure the link between glacier retreat and climate (Warren et al.1995; Glasser et al.2002; Aniya 2007). Besides climate, individual glacier dynamics can be attributed to a range of variables including basal lubrication and topographic features, ice thickness, slope, moraine shoals (Powell 1991) and tectonic events, in particular with regard to the NPI, those associated with the Liquiñe-Ofqui fault that here defines the western margin of the Andes.

To date, rates of glacier retreat around the NPI have been estimated for twenty-one of the twenty-eight larger outlet glaciers. Although the behavior of individual glaciers has varied, the overall trend has been of retreat since the mid-nineteenth century (Harrison et al. 2007). Extensive glacier studies around the NPI have also been carried out based on aerial surveys and satellite imagery derived from Landsat images (Aniya op. cit.), ASTER-derived digital elevation models and Shuttle Radar Topography Mission information (Rignot et al. 2003; Barcaza et al. 2009). A further study of snow cover on the NPI (2000-2006) using Moderate Resolution Imaging Spectro-radiometer (MODIS) images found that snow cover fluctuated not only inter-seasonally but also intra-seasonally especially on the western side of the Icefield in winter where it has been concluded that temperature determines the extent of snow cover whereas on the east side both temperature and precipitation are implicated (Lopez et al. 2008). Studies have produced proxy estimates for ice thinning, ice-front retreat rates, area loss and snow cover, but there is as yet scant field evidence (Ohata et al. 1985) to substantiate estimates. Thus, longer-term field studies are required.

This study focuses on Glacier Benito, a temperate outlet glacier on the west side of the NPI within the Laguna San Rafael National Park. Glacier Benito was first surveyed by the British Joint Services Expedition in the austral summer 1972-1973 (Sessions unpublished) and this study has extended this record with additional surveyscarried out in March 2007 and 2011, providing the longest repeat field study of any glacier around the NPI. with. Our aim is to supply comparative data for estimation of the lower glacier's changes in ice thickness over the 34-year period; supply ice surface movement and ablation rates for comparison with 1972-1973 data, and provide glacier recession rates since the start of retreat at the end of the 'Little Ice Age' (LIA) a period conventionally regarded as culminating in maximum positions sometime between AD 1550 and 1890. Additionally in 2007, a bathymetric map of the proglacial lake was constructed and dendrochronology and aerial photography were used to provide minimum dates for the moraines and rates of ice front recession.

Site description

Glacier Benito (47°02'S, 73°53'W) lying 10.5 km inland from Abra (inlet) Kelly on the Pacific foreshore descends to approximately 14 m a.s.l. from a maximum elevation of 2500 m (Figure 1). Its surface is largely free from debris with a surface area of 169 km^2 and an estimated ratio of accumulation to ablation areas of 1:0.7 and an ELA averaging 908 m a.s.l. (Rivera *et al*. 2007) compared with the start of partially-compacted névé at 850 m a.s.l ± 10 m (observed on 10/03/73, Sessions unpublished). The glacier terminates in a proglacial lake that discharges into Rio Benito flowing southwest down a glaciofluvial channel to meet the ocean at an inlet on the Golfo de Peñas.

Previously there was another outlet flowing northwest down a wide valley to join Rio Andre discharging into Abra Kelly. This former outlet is now closed 0.54 km from the current lakeshore on the northern corner of a prominent terminal moraine that describes a 1.5 km arc across the valley floor. The moraine has a maximum ridge height of 10 meters and, at the northern end three ridge crests diminish in height towards the lake. A stream, running along the proximal edge of the terminal moraine has incised a channel in the outwash plain. Other dry or intermittent channels also cross the plain and at its northern end there are a number of kettle holes and boggy areas. A series of moraines fringe the lakeshore. Large southern beech trees (*Nothofagus nitida*) up to 20 m tall grow on the main terminal moraine and smaller trees and bushes (*Pernettya mucronata*) grow on or near the crest of the moraines within the complex. The moraine fronting the southwestern shore of the lake re-emerges as an island in the middle of Rio Benito’s outlet channel. On the island there are two trees at approximately 5 m and 9 m above lake level, one on the proximal side of the moraine and the other close to the crest on the distal side.

A forest trimline, visible as a sudden change in vegetation characteristics, cuts across the northern valley wall. The trimline occupies a broad, glacier-cut bench sloping longitudinally across the side of the mountain and ending in a steeper slope above the terminal moraine (Figure 2). On the glacier-cut bench, rounded cobbles line the flat bed of a former lateral channel. Up-slope from the channel a small, deeply incised stream flows between the mountain wall and small moraines fragments. The trimline on the southern valley wall, due to the more broken nature of the terrain, is only distinguishable in its lower reaches.

Methods

Starting in December 1972, a network of three fixed survey stations was established on the valley side (Figure 3, Camp, Boulder and Cairn); from these four stakes (Figure 3, st1 tp st4) were positioned in the lower ablation zone using a Wild T2 theodolite. The baseline was determined from a chained distance on the outwash plain whilst the datum altitude was estimated from pressure corrected altimeter readings over four months. The positions of the four stakes were re-measured 71 days later. Ablation readings were taken over a 112 day period involving up to 16 visits to some stakes. A fifth stake was added for ablation measurement later in the expedition (Figure 3, point 2). The theodolite survey produced results that could be recalculated in 2007 and 2011, with the baseline re-surveyed using new, in this case Differential Global Position System (DGPS) technology.

In 2007, two new base stations (Figure 3, Stn1 and Stn2) were measured using a Trimble 5700 Series DGPS base station with a rover station. Satellite observations over several days were used to establish the positions of Stn1 and Stn2. Using the rover with both real time and post survey processing, two of the 1972/1973 fixed stations (Figure 3, Boulder and Camp) were re-occupied so that the 1972/1973 survey observations could be corrected both for position and altitude to fit the 2007 network. The rover was then used in February 2007, on three occasions, to measure the positions of six stakes inserted into the ice (Figure 3, Sb, S1 to S5). Contours were measured on the glacier and the terminal moraines and other features were mapped.

As a crosscheck of the 2007 survey work, Sessions, Wündrich and Dowling revisited the glacier in 2011. With a Hiperlite + and a Sokia GRX-1, they re-measured Stn1 and Stn2 using three times as many satellites as in 2007. They also revisited three of the 2007 ice measurement positions. Post processing using Automatic Precise Positioning System (APPS) from JPL/NASA enabled survey errors to be further reduced.

Ice front retreat rates and distances were calculated based on the survey network measured during the 2007 expedition and two overlapping 1998 vertical aerial photographs (N°12529 and N°12530, scale approximately 1:70,000) purchased from Servicio Aerofotogrametrico, Fuerza Aerea de Chile **(**www.saf.cl**).** Orthorectification of the images was carried out using PCI Geomatic software and based on ground-control points taken from data provided by Google Earth. The outline of the glacier in 2002 was also taken from Google Earth imagery and likewise superimposed over the 1998 base map. An unrectified vertical 1974 aerial photograph (8V 58WRS USAF 9-28-74 10DEC74 - AF75-1 R19B No 1235/6), was used to approximate the 1974 outline of the glacier tongue. The location of the glacier tongue in 1944 was estimated from the trimetrogon oblique aerial photographs (US Airforce, Sortie 91-PC-5M-4028).

Depths of the proglacial lake were taken using a portable eco sounder (a Garmin Fishfinder 90) enabling construction of a bathymetric contour map with positions fixed using the GPS (Figure 4). Panoramas were photographed to include sites photographed in 1972-1973.

We collected 63 tree cores and stem cross sections from *Nothofagus nitida* (Phil.) Krasser*,* (Southern Beech) and noted their GPS coordinates. The cores were mounted on wood supports and polished to a shine to reveal their annual rings; these were then counted both in the field and later under a microscope. The sampled trees were growing on moraines, the lake foreshore, the island and on the forest trimline on the northern flank of the glacier.

The samples were later located by their GPS coordinates on a map constructed from an orthorectified, oblique 1944 aerial photograph of the moraines, lake shore and glacier ice front, superimposed over satellite imagery (also orthorectified) from 1974 and 2002. Moraine dates and rates of glacier recession were derived from tree age and distances to the glacier ice-front positions shown on the map are dated as described below.

Dating parameters

Dates for surface exposure of recently de-glaciated terrain are derived from the sum of a tree core’s annual ring count to pith + estimate of number of years growth below the core + an estimate of the delay before germination and establishment (ecesis).

Annual-ring counts only give the age of a tree above coring height; age below the core is unknown: the missing age was obtained by sectioning seven small trees just above ground level and dividing the annual ring counts by tree heights to establish the average annual growth rate for young *N. nitida* on the Benito moraines (Table 1). Thus, moraine-dating estimates are based on core height divided by the average annual growth of seedling trees plus core-ring count (Winchester and Harrison 2000) plus estimate of delay before ecesis (representing the time taken for the freshly exposed ground to stabilize become fertile and seed to germinate after ice retreat). This value was deduced from the age of three trees: two on the island (shown on figure x?) and one on the continuation of the island moraine on the southern lakeshore moraine, with their ages then compared with the date the island emerged from the ice as shown in two aerial photographs: one taken in 1973 showing the ice front touching the shoreline at the northern end of the lake and the other in 1974 showing the glacier ice-front retreated a little from the shore with the future island just visible as a dark area emerging from the ice some 80 m offshore (Figure 5).

Results

Surveys

Table 2 shows the corrected altitude of stakes measured in 1972/1973 to the 2007 datum. Each stake observation in 1972/1973 involved at least three sightings to determine altitude. The maximum difference in altitude measurement at a stake was 0.67 m with a mean difference of 0.38 m. The altitudes of the stakes measured in February 1973 were cross-correlated with the

December 1972 measurements, taking into account ablation and movement down slope, as a further check to confirm that the altitudes are within a ±0.5 m error margin.

In 2007, the observed error on each occasion was 0.1 m with respect to the datum. As the baseline of the 1972/1973 network could not be established until the last day of the 2007 expedition, the stakes inserted in 2007 do not correspond exactly to the 1972/1973 stake positions. Table 3a shows the corrected altitudes for the 2007 network of 1972/1973 stakes. In summary, the glacier thinned by an average of 148.5 m ±5 m in 34 years amounting to 4.37 m yr^{-1} ±0.18 m yr^{-1} (Figure 6). The results of the 2011 revisit are shown in Table 3b. Over the interval, the glacier surface lowered on average 24.4 m ±3 m as measured near three of the 2007 stake positions. The thinning rate of 6.1 m yr^{-1} ±0.5 m yr^{-1} is a significant increase on the rate for the previous 34 years.

In 1972/1973 the average down-glacier surface movement was 0.48 m d^{-1} ±0.01 m d^{-1} (175 m yr^{-1}). In 2007 the average measured movement was 0.44 m d^{-1} ±0.01 m d^{-1} (161 m yr^{-1}). Given that the rate of movement of a temperate glacier can change significantly both inter and intra-annually, the difference is probably not significant. Average surface ablation rate for the measured period in 1972/1973 was 0.051 m d^{-1} and in 2007 it was 0.059 m d^{-1}. Given that the 1972/1973 observed ablation rates range from 0.033 to 0.068 m d^{-1} for the lower four stakes and that local meteorological data were not available for either survey, no conclusions can be drawn from this comparison.

Dendrochronology

Ecesis estimates are derived from tree age on the island (as from the austral 2006 growing season). On the island's distal side tree age was 16 years and 13 years on the proximal side below the ridge top. A further small tree on the proximal side of the equivalent moraine on the southern shore of the lake was also 13 years. These ages supply dates of 1993 and 1990, which subtracted from the 1974 aerial exposure date indicates maximum possible ecesis delays of 19 and 16 years on the proximal and distal sides of the island respectively.

Dendrochronological dating (Table 4) for the southern end of the arcuate terminal moraine suggests an 1858 exposure date for the beginning of glacier retreat from its LIA maximum and an 1859 date for the northern valley-wall trimline; the age of trees growing in the adjacent run-off channel imply that the ice surface had downwasted leaving the channel dry by 1881. The dating of three moraines on the outwash plain shows that the downwasting glacier retreated in phases at the northern end of the terminal arc over a distance of about 50 m from 1886 to 1901.

The 1944 oblique aerial photograph shows the northwest channel still flowing out towards Inlet Kelly: tree dates in a channel running along the proximal foot of the terminal moraine indicate channel activity prior to 1950. Trees on the outermost of the recent suite of lakeshore moraines supplied minimum tree-ring dates of 1955/1956. Stillstands produced two further moraines marking the present northern lakeshore dating to 1966 and 1970, and finally the 1974 moraine was deposited appearing as an island half way across Rio Benito's outlet and reappearing on the lake's southern shore. The course of this latter moraine is further defined by the bathymetric survey showing the lake shallowing as it crosses the neck of the Rio's outlet (Figure 4). The continued existence of a moraine here, rather than being washed away by the outgoing flood, is likely due to debris deposited by stranded melting icebergs. The following estimates for average recession rates and distances covered between dated glacier positions are measured along a notional centre line of the glacier passing through the island (Figure 7). Potential measurement errors are discussed below.

- 1858-1886 glacier thinning with minimal retreat.
- 1886-1944 recession over a distance of 519 m at a rate of ~9 m yr^{-1}.
- 1945 -1974, recession over 426 m, averaging 14.7 m yr^{-1}.
- 1975–1998 recession over 407 m, averaging 17.7 m yr^{-1}.

- 1999-2002 recession increased over a distance of 509 m averaging 169.7 m yr^{-1}.
- Total retreat 1858-2002 was 1935 m.

Tree-ring dating during the first 28 years suggest that there was major ice-loss in the vertical dimension as the glacier thinned, with retreat on the horizontal axis at the southern end of the terminal moraine starting around 1858 and at the northern end in 1886. Initial slow downwasting (1859-1881) is evidenced by 148 and 126-year old trees growing above and in, respectively, a meltwater channel on the northern mountainside bench (Figure 7). The extent of thinning to the present is shown by the change in terminal-surface gradient from ~13.5^0 in the 1850s, as shown by the angle of the trimline on the valley-side (Figure 2), to between 3^0 and 4^0 by 1973 (Figure 6); the slope of the glacier terminus in 2007 was under 1^0, although three km up glacier the slope was still 3^0 between stakes S1, S5 and S3 (Figure 4) similar to the slope in that area in 1973.

Discussion

Errors and dating estimates

The exact dates, distances and rates of retreat are approximations owing to a number of potential error sources both in approaches to measurement and to dating. Since ice-front configurations are highly variable, retreat values depend on where precisely on the glacier terminus measurements are taken. The arbitrary choice of an approximate centre line on Glacier Benito passing through the island thus only provides a relative measure of ice loss between periods. The dating of surface exposure from tree ages can also be problematic since older un-cored trees may exist and simple tree-ring counting without any cross dating to identify missing or extra (false) rings could affect dating accuracy (Fritts 1976; Koch 2009). Additionally, growth below core height may vary for individual trees (Winchester and Harrison 2000; Winchester et al.2001)

Several factors point to a more remote date, nearer 1850 than 1860, for the beginning of glacier retreat here: our estimated delay before ecesis is based on differences in tree ages on the proximal and distal sides of the island in 2007 (with differences implying that trees are sensitive to microclimatic variations) whereas ecesis is likely to have taken longer during the harsher mid-nineteenth century; further, a sensitive response to local climatic and topographic conditions is indicated by the 19-year ecesis delay at Glacier Benito as compared with 6-years at San Rafael (Winchester and Harrison 1996). A difference that could be explained by Benito's terminus being 10-km from the sea and enclosed by a moraine arc, with icy down-glacier katabatic winds forming a frost pocket in the sheltering bowl of forest-clad moraines; compared with the termini of the much bigger San Rafael and San Quintin Glaciers (Figure 1) where prevailing winds have a much greater influence due to less shelter from surrounding ridges. The ecesis error is likely to be small where values are closely controlled by aerial photography as at Benito, with secure 1970s dating for the lakeside-moraines (photographs show them bare of vegetation in 1973) and aerial photographic evidence showing the island just emerging in 1974.

Other intrinsic error sources lie in orthorectification of aerial images, with the accuracy of the process depending on the quality of the Digital Elevation Model and the correction formula. Potere (2008) describes Google Earth horizontal positional accuracy as effectively 50 m. This would invalidate our retreat-rate estimates if it were not for the dendrochronological and photographic evidence that provides secure dating for the recent moraines including an absolute date for the island. Additionally, the resolution of the Google Earth images of Glacier Benito for 2011 has increased significantly since 2007. The 2011 expedition was able to measure nine ground control points, visible in Google Earth, and determine that the difference in position of the features in the Google Earth imagery was 17 m ± 3 m on a bearing of 200° and 7 m ± 8 m higher compared to the measured points. The glacier position in 1944 is the most approximate of the orthorectified images since it is derived from a Trimetrogon oblique photograph of poor quality and no 1944 trees were found (Figure 5). Thus the estimated 14m yr^{-1} rate of retreat 1945-1974 is an approximation.

Climate, ice movement and ablation and ice-surface thinning

Assessing climate change around the Icefields is hampered by the scarcity of continuous longer-term meteorological records. Existing records on the western seaboard (Rosenblüth et al. 1995; Villalba et al. 2009) together with tree-ring reconstructions (Villalba et al. 2003), filling in the areal instrument gap on the eastern side of the Andes, show that temperature trends differ north of 46^0S compared with south of that latitude; precipitation also differs being higher to the north during the austral winter months whereas at Isla San Pedro due west of the NPI precipitation is highest during the summer (Winchester and Harrison 1996). We propose that the position of the northern limit of the NPI at $46^0$30'S is sensitively dependent on climate.

Incomplete records from stations at Cabo Raper and Isla San Pedro due west of the NPI show a tendency for rainfall decrease during the 20th century (Rivera and Casassa 1999) with temperatures increasing southwards from 1976 by 0.4^0C at latitude 46^0S to 1.4^0C at 53^0S. By contrast, north of 46^0S (between 41^0S-43^0S) there was no trend over the period (Rosenblüth et al.1995). Villalba et al. (2003: 177) observe, based on tree-rings, that south of latitude 46^0S "the rate of temperature increase from 1850 to 1920 was the highest over the past 360 years" including "a notable increase in the warming trend after 1976, with summer warming responsible for much of the increase". These findings support our estimate of an 1850s retreat from Benito's LIA moraine; they also highlight the importance of seasonality in melting and discharge rates.

Concerning seasonal variability (Vimeux et al. 2008), it should be noted that Benito's surface movement rates will be higher in summer during periods of peak meltwater discharge and ablation than in winter and thus average annual movement will be rather less than the rates given here for 1973 (0.48 m d^{-1}) and 2007 (0.44 m d^{-1}): a potential scale for the error is suggested by differences in ablation at Glacier Soler on the warmer, eastern side of the NPI where mean mid-summer rates are as high as 0.131 m d^{-1} (Kobayashi and Saito 1985b) while spring rates are only 0.03 m d^{-1} (Fukami and Naruse 1987). Ablation rates in 1972 and 2007 at Benito, 0.051 and 0.059 m d^{-1}

respectively (Table 2) are close to those near the terminus of San Rafael where there was a measured loss of 0.068 m d^{-1} water equivalent in 1983/84 (Ohata et al.1985c).

Surface thinning, averaging 4.37 m yr^{-1} (1973-2007) on Benito Glacier, 3.76 km from the island (Figure 3), exceeds that of all other NPI glaciers measured by Rivera et al. (2007): a contributing factor may be the additional six years (2001-2007) of possibly dramatic ice-surface wasting.

Between stakes st1 and st2 (Figure 3) our data indicates that the surface gradient was $3.5^{0,}$ in 1973 (close to Aniya's 1988 estimate of 3.8^{0}). The average gradient between similar stakes in 2007 was 3^{0} with the glacier surface flattening considerably below stake Sb (Figure 3).

Calving and frontal retreat

Calving credited as a major control on glacier dynamics (Warren and Aniya 1999) is not a feature of Benito Glacier despite Aniya's (2007: 67) assertion, based on aerial surveys, that Benito was a calving glacier between 1986 and 1991, but for calving to take place an ice front needs to be free floating: the shallow lakebed profile (Figure 4) suggests that the ice was grounded over the period. A problem in identification of calving from aerial photography is that it may be difficult to distinguish calved icebergs from floating, melting ice initially fractured by impact against a shelving lake floor. Changes in the size of 'icebergs' (commented on by Aniya op. cit.) could be the result of an increase in the depth of the lakebed profile as the glacier recedes.

The dendrochronological dating that supplies the earliest date for retreat on the southwestern corner of the terminal moraine and a later date of 1886 at its northern extremity is consistent with the asymmetrical valley profile revealed by the bathymetric survey (Figure 4). The deepened trough along the northeastern valley side signals maximal erosion at this point due to glacier dynamics influenced by the bend in the ice stream (see ice contours Figure 3). Hence, although surface thinning began in the 1850s on the southern side of the glacier front, on the northern mountainside the dating evidence shows that the lateral channel fed by the melting glacier did not cease to flow until 1881 implying that downwasting was initially very slow on the northern valley side with the ice still level with the trimline bench until then.

Benito's retreat rate, 1999-2002, of 169.7 m yr^{-1} can be compared with the larger San Rafael Glacier that between 1990-2002 retreated at an average rate of 84 m yr^{-1} while the northern part of San Quintin's tongue retreated, 1997-2003, approximately 338 m yr^{-1} (as measured from the Google Earth image, 2002) or 33 km^{2} of ice (Rivera et al. 2007). Since these three glaciers are very different in many respects including source area, size, and debris cover (Benito's debris cover is among the lowest of all the NPI glaciers; Rivera et al. op. cit.) and since the evidence for changes in precipitation is equivocal, we propose that increasing retreat rates from 1974 and especially since the 1990s (Table 2) are a response to climate warming (Glasser et al. 2011).

The bathymetric data (Figure 4) were collected to provide information on the current depth of the lake, the non-calving status of the glacier and the maximum thickness of its terminus. The lakebed's slope, northeastwards from the island moraine, is also of interest with respect to the asymmetrical response of the ice front which, taken with the 1944 oblique view of the glacier front, strongly infers that the glacier front has been grounded over most of its width for the whole 150-year period.

Conclusions

The extent of thinning since the LIA maximum to the present is revealed by differences between the terminal-surface gradient on the northern valley side from ~13.5^{0} in the 1850s, as shown by the angle of the LIA trimline (Figure 2), to an average of 3^{0} by 1974 reducing to less than 1^{0} by 2007 (Figure 6). Although our data show that neither the down-glacier movement rates nor surface ablation in 1972 and 2007 differed significantly over the period, total ice surface thinning of 148.5 m in the lower ablation zone was substantial, proceeding at an average rate of nearly 4.4 m yr^{-1}, with results from the visit in 2011 showing an increasing rate. This pattern of downwasting has resulted in the typical concave terminal profile of a fast retreating glacier, with terminal recession rates increasing from ~9 m yr^{-1}, for the period 1886-1944, to almost 170 m yr^{-1} between 1999-2002. Ice-front retreat at the terminus has been non-uniform over the period due to an asymmetrical valley profile.

The early onset of retreat from the southern end of the terminal moraine in the 1850s suggests that Benito Glacier is highly sensitive to warming and among the most reactive of other NPI glaciers previously studied, with this dating placing the LIA maximum on the NPI two decades earlier than previously recorded (Glasser et al 2011) coinciding with the start of glacier retreat in the western USA (Bradley and Jones 1993; Hall and Fagre 2003) and the majority of glaciers elsewhere (Glasser op. cit.). Our results highlight accelerating climate warming in southern South America south of the current northern margin of the NPI. We propose that this margin describes a climatic boundary responsive to changes in the westerly winds, their related storm tracks and ocean/atmosphere warming.

More long-term field studies on the NPI are still required and reliable forecasts of Icefield dynamics in a climate-warming scenario must wait until studies of mass wasting, glacier retreat and ice-surface topography can be combined with data showing the ice-bed interface.

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Vanessa Winchester, School of Geography and the Environment, OUCE, University of Oxford, Oxford, OX1 3QY, UK. E-mail: vanessa.winchester@geog.ox.ac.uk
Martin Sessions, 9 Eliza St, Amaroo, ACT 2914, Australia. sessions@cyberone.com.au).
Jammie Valdivia, Universidad Tecnologica Metropolitana (UTEM) Santiago, Chile.
jammievaldivia@yahoo.es
Olaf Wüendrich, Colibri Ventura, Casilla 113, Coyhaique, XI Region de Aysén, Chile
olafwuendrich@yahoo.de
Sam Clemmens and Neil Glasser, Institute of Geography & Earth Sciences, Aberystwyth University, Llandinam Building, UK. Samuel.Clemmens@Wales.GSI.Gov.UK
Megan Nash, Australian National University, ACT, Australia. u4194419@anu.edu.au

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Table 1. Average growth rate of *Nothofagus* up to 250 cm tall. The positions of the measure trees are shown in….Fig…

Tree height cm	Ring count	Growth cm yr^{-1}	Location
250	14	17.8	NE flank up glacier
178	12	16.2	N. lake front M.
135	15	9	N. lake front M.
150	5	30	N. end lakeshore
50	3	16.6	N end lakeshore
100	5	20.0	N end lakeshore
19	1	19.0	Rio foreshore
Average		18.4	n=7

Table 2. Stake movements, ablation rates and altitudes 1972/73 and 2007. Position of stakes shown in Figure 3.

Year	Stake No	Altitude m a.s.l.	Error m	Movement m day^{-1}	Error m	Ablation m day^{-1}	Notes
	Datum	14.7					Surface level of lake
1972/3[2]							
	st1	201.0	±0.5[3]	0.48	±0.01[4]	0.059	Movement 14/12/72 to 23/2/73, Ablation 28/2/73 to 13/3/73
	st2	231.0	±0.5[3]	0.49	±0.01[4]	0.063	Movement 14/12/72 to 22/2/73, Ablation 28/2/73 to 13/3/73
	st3	236.0	±0.5[3]	0.45	±0.01[4]	0.045	Movement 14/12/72 to 22/2/73, Ablation 1/3/73 to 13/3/73
	st4	230.0	±0.5[3]	0.49	±0.01[4]	0.038	Movement 14/12/72 to 22/2/73, Ablation 22/2/73 to 13/3/73
	Av.			**0.48**	±0.01	**0.051**	
2007							
	Sb	45.2	±0.1[5]				2007 - Test stake
	S1	62.3	±0.1[5]	0.44	±0.01	0.063	28/2/07 to 13/3/07
	S2	86.8	±0.1[5]	0.44	±0.01	0.062	28/2/07 to 13/3/07
	S5	93.0	±0.1[5]	0.44	±0.01	0.035	5/3/07 to 13/3/07
	S4	104.3	±0.1[5]	0.38	±0.01	0.045	28/2/07 to 13/3/07
	S3	126.7	±0.1[5]	0.47	±0.01	0.066	28/2/07 to 13/3/07
	Av.			**0.43**	±0.01	0.0**59**	

Notes:
1. Varying measurement periods.
2. 1972/3 data converted to 2007 datum.
3. Altitude measured on last day of period. Difference in altitude observations using theodolite range from 0.24 m to 0.67 m with a mean of 0.38 m (standard deviation 0.54 m).
4. Differences in closing range from 0.13 m to 0.64 m with a mean of 0.35 m (standard deviation 0.166 m). Distance travelled by stakes in period was between 31 and 36 m.
5. Altitude measured on first day of period. Measured using Trimble DGPS with Base Station and Rover. Post processed. Observational accuracy with respect to base station is 10 cm.

Table 3a Comparative data for glacier surface elevation changes of the stake network 1973-2007. 3b: surface elevation changes of the stake network 2007-2011 (data reduced to 2007 datum). Stake positions shown in Figure 3.

a

Stake ID 1973	Stake Altitude[1] m a s l	Error m	Closest Stake 2007	Stake Altitude[1] m a s l	Error m	Altitude from contour map[2] m a s l	Error[3] m	Change 1973 to 2007 m	Error m
st1	201	±0.5	S1	62.3	±0.1	50	±4.5	151	±5.0
			Sb	45.2	±0.1				
st4	230	±0.5	S2	86.8	±0.1	84	±1.5	146	±2.0
st2	231	±0.5	S5	83.5	±1.2	79	±1.5	151	±2.0
st3	236	±0.5	S4	104.3	±0.1	90	±4.5	146	±5.0
Average								148.5	±5.0
Annual	change							4.37	±0.18

b

Stake ID 2007	Stake Altitude 2007[1] m a s l	Error m	Position Altitude 2011 m a s l	Error m	Distance from 2007 stake m	Topographic Error[2, 4] m	Change 2007 to 2011 m	Error m
S2	86.8	±0.1	61.1	±0.1	22	±3.0	25.6	
S3	126.7	±0.1	100.8	±0.1	23	±3.0	25.8	
S4	104.3	±0.1	82.4	±0.1	35	±2.0	21.9	
Average							24.4	±3.0
Annual	change						6.1	±0.5

Notes 3a, 3b

1. Data from Table 2.
2. Derived from 2007 Expedition Map, created from DGPS “Rover” readings.
3 Error when correcting the 1973 ice surface elevation to the 2007 position.
4. Error when correcting the 2011 ice surface elevation to the 2007 position.

Table 4. Dendrochronological data.

Latitude S	Longitude W	Location	Core ID	Ring count	Yrs to pith	Yrs to core ht	Yrs to ecesis	Est. date
47001'51.4"	73^{0}54'18.9"	Island crest W. flank	8	15	0	1	16	1974
47001'51.4"	73^{0}54'18.9"	Island E flank	9	9	3	1	19	1974
47^{0}01'51.4"	73^{0}54'22.3"	S.Rio lakeside M	10	9	3	1	19	1974
47^{0}01'51.4"	73^{0}54'22.4"	S.Rio lakeside M	11	15	0	1	16	1974
47^{0}01'46.6"	73^{0}54'21.2"	Camp M S.	4	31	2	2	16	1955
47^{0}01'44.4"	73^{0}54'22.1"	Camp M S.	6	24	3	4	19	1956
47^{0}01'40.1"	73^{0}54'21.2"	Camp M channel	15	28	2	2	16	1958
47^{0}01'34.5"	73^{0}50'18.6"	Lake-end outer M base	17	30	3	2	16	1955
47^{0}01'34.5"	73^{0}54'17.9"	N lake-end stream	18	26	3	2	16	1959
47^{0}01'34.5"	73^{0}54'13.7"	N lake end middle M	22	21	0	2	19	1964
47^{0}01'34.5"	73^{0}54'16.1"	N lakefront M	19	13	3	1	19	1970
47^{0}01'34.5"	73^{0}54'12.2"	N lakefront M	24	17	1	2	16	1970
47^{0}01'34.5"	73^{0}54'12.1"	Lakefront M NE corner	27	15	3	1	19	1970
47^{0}01'28.4"	73^{0}54'13.2"	Gt M channel edge	35	30	5	2	19	1950
47^{0}01'41.4"	73^{0}54'28.0"	Trimline forest	38	110	10	7	19	1860
47^{0}01'41.4"	73^{0}54'18.4"	Trimline ravine edge	37	107	11	6	19	1863
47^{0}01'41.1"	73^{0}54'28.0"	Trimline channel	41	97	1	6	19	1883
47^{0}01'26.0"	73^{0}54'39.2"	N end Gt M outer crest	48	82	12	7	19	1886
47^{0}01'26.9"	73^{0}54'42.8"	N end Gt M mid. crest	50	85	4	6	19	1892
47^{0}01'28.7"	73^{0}54'37.2"	N end Gt M inner crest	51	63	19	4	19	1902
47^{0}02'05.0"	73^{0}55'25.9"	S end Gt M crest	2	115	6	8	19	1858

Ring count dated from last ring formed in 2006.
Gt M = Great Moraine. Camp M. S. = Camp moraine south end
Years to pith: number of rings added to count where core did not reach centre. The number derived by fitting concentric circles scribed on acetate to the curves of oldest rings visible in cores.
Years to core height: estimated number of rings above ground and below core height based on average growth rate of 18.4 cm yr (Table 1).
Years to ecesis: estimated number of years taken for seeds to germinate after ice retreat based on delay before germination on proximal side of island = 19 years, and distal side = 16 years.
Estimated date = Ring count+years to pith+years to core height+ecesis (19 years added to all 19th century dates).