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Dynamic Features of Thinning and Retreating Glaciar Upsala, a Lacustrine Calving Glacier in Southern Patagonia

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Abstract

Glaciar Upsala, which calves into a lake on the eastern side of the Hielo Patagónico Sur, South America, has significantly retreated by 5 km during the last 20 yr. The glacier near the terminus has thinned by about 33 m from 1990 to 1993. Bathymetric surveys made in 1998 at the proglacial lake revealed the existence of bedrock rises spreading out from exposed islands in the western part of the lake. Between 1978 and 1990, the western half of the glacier terminus was located at the bedrock rises, which suggests that the front fluctuations were strongly controlled by the bed topography. During 1990–93, the glacier terminus was located upstream from the bedrock rises, and it is considered to have been floating after estimation of the buoyancy. Significantly large extending strain rate of 0.22 a^{-1} was deduced from the continuity consideration in 1990–93. These results are discussed with those obtained at Columbia Glacier, Alaska.

Introduction

According to the estimates of global glacier mass balance in the IPCC Report (Warrick et al., 1995), the total amount of iceberg calving accounts for about 70% of the total mass loss of all glaciers including two polar ice sheets, and for about 7% of the total mass loss of Arctic ice caps and mountain glaciers. In spite of their importance, processes and mechanisms of calving are less well understood than dynamics of noncalving glaciers.

Research on calving glaciers has been mostly made at fiords in Alaska and the Arctic. It was known that, while some Alaskan tidewater glaciers were retreating catastrophically, some others were oscillating or advancing slowly (Mercer, 1961). Change in the terminus position of a calving glacier can be expressed as,

$$dL/dt = Ui - Uc, \tag{1}$$

where L is the width averaged terminus position, Ui is the width- and depth-averaged ice velocity at the terminus, and Uc is the calving rate defined as the volume rate of iceberg discharge divided by the cross-sectional area of the terminus. The equation implies that two physical processes, flow (Ui) and calving (Uc), control the position of the glacier terminus. However, because measurements of Uc are usually not easy, Uc is estimated from the other two parameters in equation (1).

Brown et al. (1982, 1983) estimated calving rates of 12 glaciers in Alaska, and showed that the annual mean calving rate is best fitted by a simple proportionality to average water depth at the terminus, as

$$Uc = k\,w, \tag{2}$$

where w is the water depth averaged over the width and over a year, and k a constant of proportionality. A similar form of linear relation has been widely applied to many other tidewater or freshwater calving glaciers (e.g., Funk and Röthlisberger, 1989; Pelto and Warren, 1991; Warren et al., 1995b). These calving equations seem to exhibit well the behaviors of tidewater glaciers: when the terminus of a glacier retreats into deeper water from the moraine shoal or the bed rise, the rate of calving increases as shown in equation (2), which may lead to the further recession of the glacier by equation (1).

It is sometimes pointed out that backstress (or backpressure) arising from the shoal or islands is crucial to the dynamics or stability of a glacier. Meier and Post (1987) explained rapid disintegration of grounded tidewater glaciers by the relation (equation 2) and also a feedback process including backpressure; namely "retreat decreases backpressure on the glacier, increasing its stretching and velocity, causing thinning, which decreases the effective pressure on the bed, causing further increase in stretching and further increase in calving." Hughes (1992) attempted to develop a calving theory for ice walls grounded in water, and derived a calving law in which calving rates are controlled by bending creep behind the ice wall. However, one cannot quantitatively estimate a calving rate of a glacier based on his theory, because it includes some parameters which are not easily measured.

According to Meier and Post (1987), no temperate tidewater glaciers have floating termini, although the floating condition has been reached locally and temporarily at Columbia Glacier, Alaska. This may be because the floating temperate ice can easily be broken off by tidal movements or currents due mainly to weakening of ice by the effect of water percolating through crevasses, cracks and veins. Consequently, if the terminus of a glacier becomes sufficiently thin to nearly float, the terminus portion calves to maintain a thickness somewhat in excess of the flotation thickness. This is the "height-above buoyancy model" proposed by Van der Veen (1996). According to this model, the position of the terminus (L in equation 1) is controlled by such geometric factors as ice thickness and water depth, so that calving rate Uc is not a cause but an effect given by equation (1). By analyzing data from Columbia Glacier, it was shown that, during the retreat, the thickness at the terminus appears to be linearly correlated with the water depth (Van der Veen, 1996). The thickness in excess of flotation at the terminus ranges from about 30 to 100 m, with a mean at about 50 m.

On the basis of force balance consideration and data analyses for Columbia Glacier, Van der Veen (1997) and Whillans and Venteris (1997) concluded that backstress emanating from

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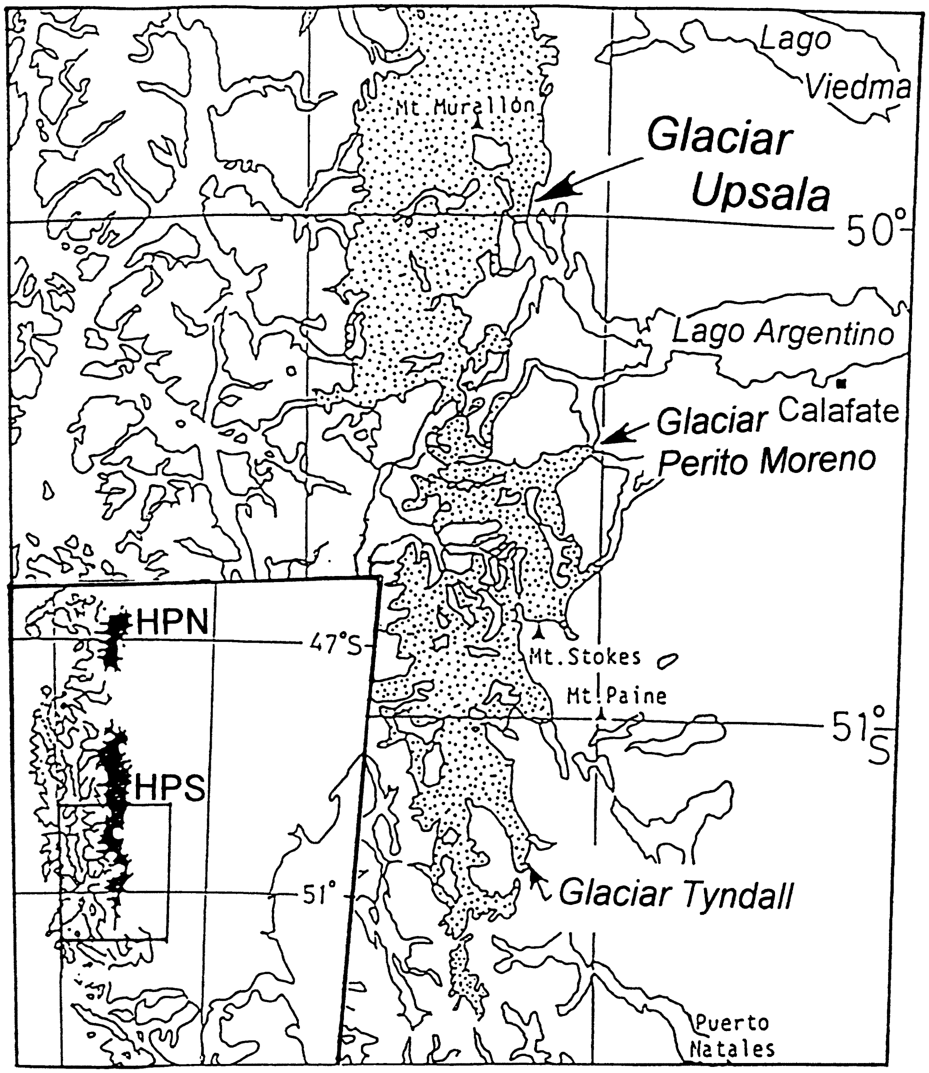

FIGURE 1. Map showing the location of Glaciar Upsala in the southern part of the Hielo Patagónico Sur (HPS: Southern Patagonia Icefield).

the glacier bed is unimportant for the dynamics and stability of grounded tidewater glaciers, and decrease in backstress was not the main cause for the retreat of Columbia Glacier. Venteris et al. (1997) emphasized the importance of the rate of longitudinal stretching for the dynamics of the glacier. Meier (1997) stated that the height-above-buoyancy model seems useful in explaining aspects of calving and suggests an increase in calving with ice thinning, even though it cannot apply when the effective pressure at the bed (ice pressure minus water pressure) approaches zero.

At Glaciar Upsala, a lacustrine calving glacier in southern Patagonia, discussion on the glacier dynamics has been insufficient so far due to lack of ice thickness data. Bathymetric surveys were therefore extensively carried out in 1997–1998 near the terminus of the glacier. Since the glacier has retreated by about 5 km during the last 20 yr, we can roughly reconstruct the geometry of the frontal 5-km portion of the former glacier using the water depth data (Skvarca, unpublished). In the present paper, we try to estimate and discuss such dynamic features as the effect of the bed topography, the height above buoyancy, and the longitudinal stretching of a freshwater calving glacier, comparing with a tidewater glacier (Columbia Glacier), both thinning and retreating significantly during these two decades.

Glaciar Upsala and Its Recent Behavior

In Patagonia, a number of glaciers calve into fiords on the western side and into large proglacial lakes on the eastern side of two icefields (Warren, 1994; Warren et al., 1995a). Among these glaciers, Glaciar Upsala (Fig. 1), 60 km in length and 902 km^2 in area (Aniya et al., 1996), is one of the largest freshwater calving glaciers. The accumulation area (611 km^2) is located to the east of the north-south ice divide on the Hielo Patagónico Sur (HPS: Southern Patagonia Icefield). The ablation area, in the form of a valley glacier about 3 km wide near the terminus, flows southward finally calving into Brazo Upsala (Fig. 2) at about 180 m a.s.l., a western arm of Lago (Lake) Argentino.

Recent fluctuations of Glaciar Upsala are worthy of note. Front positions of the glacier during the last half century were examined on the basis of topographic maps, aerial photographs, and satellite data (Aniya et al., 1992; Aniya and Skvarca, 1992; Skvarca et al., 1995a, 1995b). From 1945 to 1978 the glacier had been almost stable, and the retreat started in 1978. In Figure 2, front margins of the glacier in 1981, 1990, 1993, 1996, and 1998 are illustrated. We can clearly see in Figure 3 a remarkable change between 1993 and 1999 in the feature of the frontal part of Glaciar Upsala. Cumulative retreat distances averaged over the glacier width from 1968 to 1998 are shown in Figure 4. A considerably large recession of about 800 m a^{-1} occurred in mid-1994 (Naruse et al., 1997), then retreat slowed until 1997 when two Radarsat images revealed that a drastic calving occurred between January and May 1997, filling Brazo Upsala with icebergs (Aniya et al., 2000a). According to the information provided by a local park ranger, this major calving event took place in mid-March.

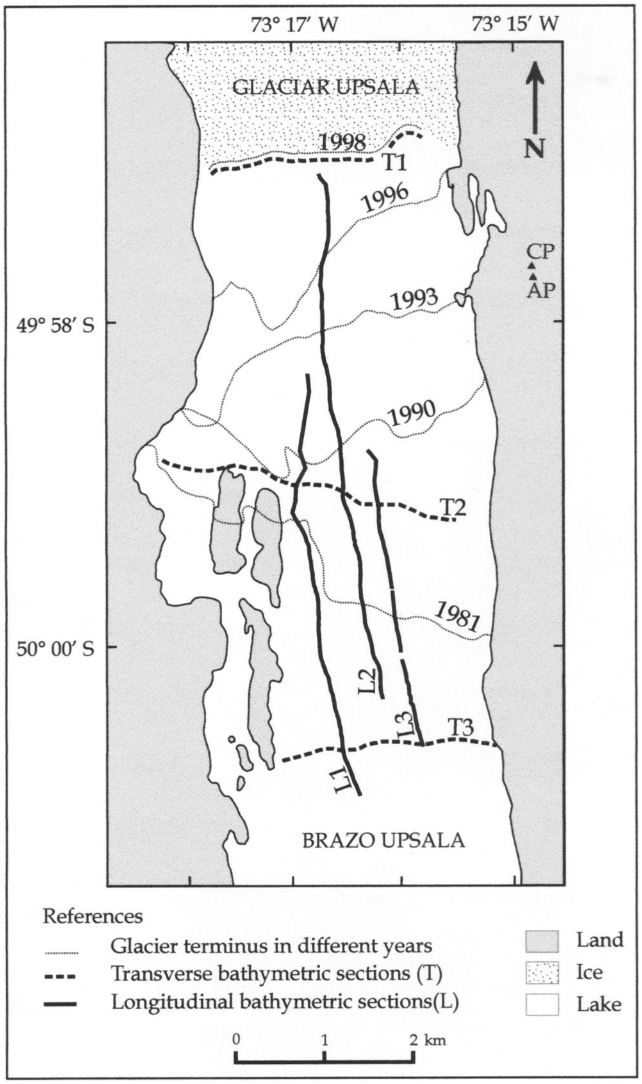

FIGURE 2. Plane view of the frontal part of Glaciar Upsala, which flows from the top (north) to the bottom (south) and finally calves into Brazo Upsala of Lago Argentino. The 1981 (February) front margin was determined by vertical aerial photographs (Aniya and Skvarca, 1992), the 1990 (November) margin is based on observations in situ, and the 1993 (November), the 1996 (December), and the 1998 (December) margins were measured by conventional angle and GPS surveys (Skvarca et al., 1995a, 1995b; Naruse et al., 1997; present study). The glacier width becomes narrower with retreating and thinning, therefore the lateral margins are drawn approximately as those in the mid 1990s. CP and AP on the eastern (lefthand side) bank indicate the control point and the azimuth point for surveys of the surface profile and the glacier flow. Thick solid lines show longitudinal sections of bathymetry L1, L2, and L3, and thick broken lines show transverse sections T1, T2, and T3, measured in December 1998.

Repeated surveys of the glacier surface were made in 1990 and 1993 along a transverse line fixed in space from the control point CP (Fig. 2) to the center line of the glacier at about 2.5 km and 1 km from the terminus in 1990 and 1993, respectively. Remarkably large thinning rates of ice were obtained ranging from 14 m a^{-1} near the eastern margin to 9.5 m a^{-1} in the central part, with a mean value of 11.1 m a^{-1} (Naruse et al., 1995a; Skvarca et al., 1995b). These thinning rates are much larger than values measured at other glaciers in Patagonia; namely, 5.2 m a^{-1} at Glaciar Soler (Aniya and Naruse, 1987), 4.0 m a^{-1} and 3.1 m a^{-1} at Glaciar Tyndall, and almost zero at Glaciar Perito Moreno (Naruse et al., 1995b; Skvarca and Naruse, 1997).

Very fast flow velocities were obtained at the same transverse line: 1300 m a^{-1} in November 1990 (Naruse et al., 1992) and 1600 m a^{-1} in November 1993 (Skvarca et al., 1995b) near the center line of Glaciar Upsala. Ablation rates were also measured in November 1993 with a mean of 58 mm d^{-1} (in ice thickness) near the glacier terminus (Skvarca et al., 1995b).

Naruse et al. (1997) discussed some possible causes of the thinning behavior of Glaciar Upsala. Using ablation data at Glaciar Perito Moreno and temperature data at the meteorological station Calafate, annual ablation rates near the Glaciar Upsala terminus were estimated to fluctuate from about 14 m a^{-1} to 18 m a^{-1} in ice thickness during the last 30 yr. The range (4 m a^{-1}) of year-to-year variations in annual ablation is much smaller than the measured ice thinning (11 m a^{-1}). Thus, it was concluded that the change (rise) in air temperature alone could not elucidate the thinning rate. The thickness change at the ablation area is a result of difference between the emergence velocity and the net ablation rate. Then, a scenario was suggested similar to a feedback proposed for Columbia Glacier by Meier and Post (1987). At Glaciar Upsala, the retreat may have resulted in reduction of longitudinal compressive stress exerted from bedrock rises and islands, causing a considerable decrease in the emergence flow. Thus, the ice may have thinned significantly in recent years because of reduction in emergence velocity.

Bathymetric Surveys at the Proglacial Lake

Measurements of water depth at Brazo Upsala were first made in February 1994 and March 1997 by H. Svetaz (pers. comm., 1997) with an echo-sounder (Furuno Electric Co.) at about 50 points near the glacier terminus. Secondly, extensive bathymetric surveys were carried out in December 1998 by one of the present authors (Skvarca, unpublished). An echo-sounder used in the surveys is the ECHOTRAC (DF 3200 MKII, Odom Hydrographic Systems Inc.) which can measure water depth from 7.5 m to 6000 m with an accuracy of 0.042% of the depth. The ECHOTRAC sounder was connected to a GPS instrument, and measurements were sequentially made every 5 s from a slowly cruising rubber raft. An accuracy of the absolute position of each point is estimated as ± 100 m. The surveyed area corresponds to the region in which the glacier terminus has receded during the last 20 yr (Fig. 2).

Longitudinal profiles of the lake bottom along three sections L1, L2, and L3 (Fig. 2) are shown in Figure 5. Hereafter, D(n) denotes the longitudinal distance n (km) in this coordinate. Along L1 stretching in parallel with two islands, very shallow depths (<200 m) are observed between D(1) and D(4). The lake bottom deepens upstream from D(3) in L1, from D(5) in L2, and from D(0.5) in L3.

Next, transverse profiles along T1, T2, and T3 are shown in Figure 6. The profile T1 represents a typical U- or V-shape valley with a maximum depth of 600 m slightly westward from the center of the channel. On the other hand, the profiles T2 and T3 exhibit the bedrock rises which may be connected to the three islands exposed in the western part of the channel. Particularly, a western one third of T2 indicates a shoal. The largest depth in T2 is found as 600 m at a slightly eastward site from the center. It became clear that the main trough of the bed turns slightly toward the east in the region from D(7) to D(1).

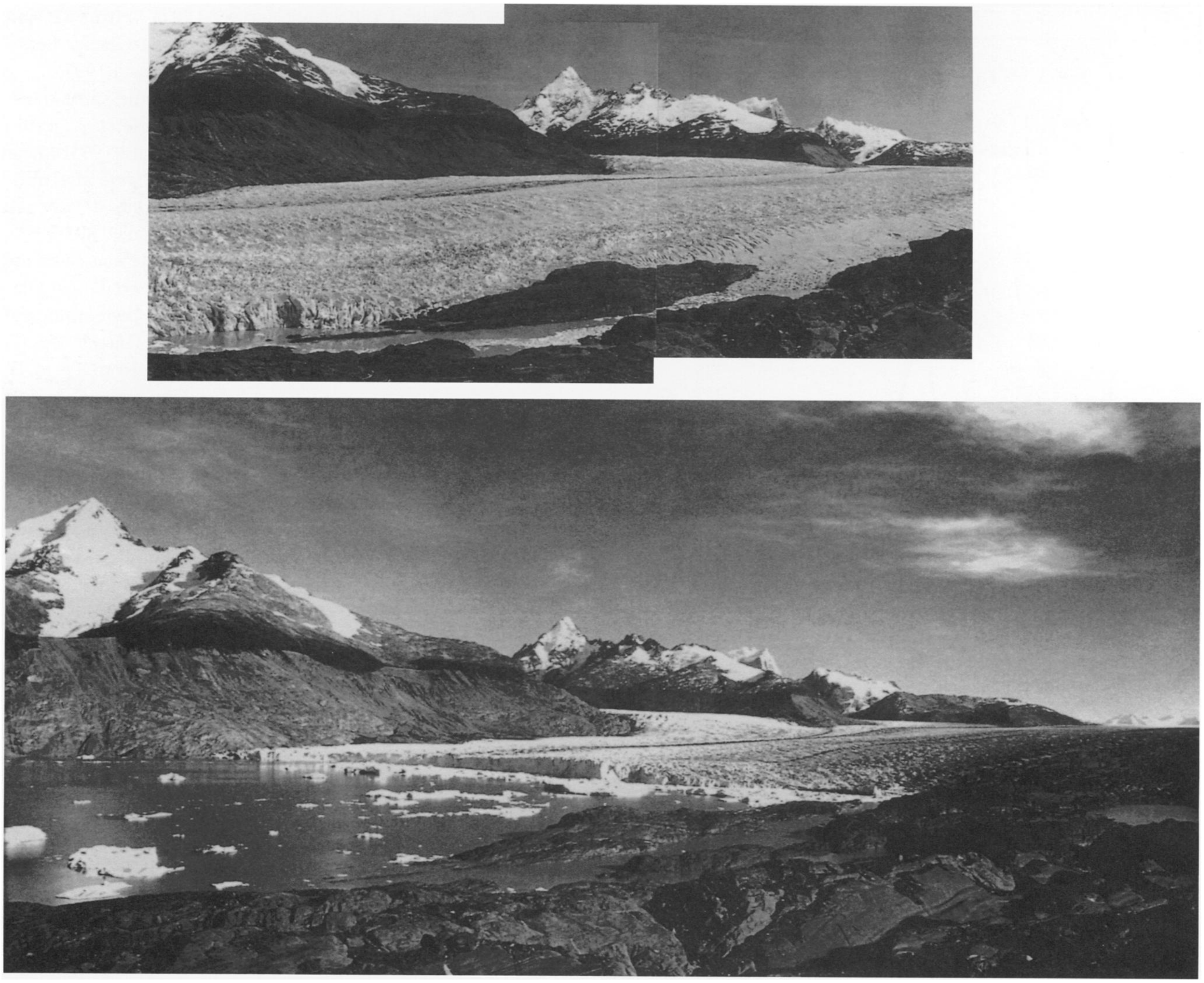

FIGURE 3. The frontal part of Glaciar Upsala in November 1993 (Upper photo) and March 1999 (Lower photo), viewed from the control station (marked by CP in Fig. 2) on the eastern (lefthand side) bank. A tributary glacier flowing from the eastern valley is Glaciar Bertacchi, and a snow-covered, pyramidal peak is Cerro Cono (2440 m). (Both photos by P. Skvarca.)

Discussion

EFFECT OF BED TOPOGRAPHY ON THE GLACIER FLUCTUATION

As mentioned above, large changes in front positions of Glaciar Upsala did not occur before 1978. Figure 4 shows that the width-averaged front retreated remarkably in 1981–1984, 1990–1994, and 1996–1998. However, it is noticed in Figure 2 that between 1981 and 1990 the large retreats occurred predominantly in the eastern half of the glacier. During this period, the western half of the glacier terminus was located at the bed bump as seen in L1 (Fig. 5) and T2 (Fig. 6). Hence, the frontal fluctuations before 1990 are considered to be strongly controlled by the bed topography in the western part of Brazo Upsala.

Whereas in the period from 1990 to 1998, though calving retreats occurred alternately in the western half and the eastern half, the cumulative retreat distance was almost uniform over the glacier width (Fig. 2). Bedrock topography in the region to the north of D(4) indicates a relatively gentle slope (L2 of Fig. 5), and a smooth concave cross profile (T1 of Fig.6).

Naruse et al. (1997) suggested that after 1990 the very small longitudinal compressive stress near the terminus might have resulted from a considerable reduction in backstress from bedrock rises and islands near the front. Although this remark may still be relevant, we cannot evaluate how significant the backstress was before 1990. However, there is no doubt that the islands and the surrounding bedrock rises should have affected the glacier dynamics, at least in the western half, when the glacier terminus was located there.

FLOTATION OF THE GLACIER FRONT

Van der Veen (1996), using aerial photogrammetry data from Columbia Glacier, showed that height above buoyancy at the terminus decreased gradually from about 100 m in 1976 to 30 m in 1991, synchronizing somewhat with the glacier retreat. Height above buoyancy, or ice thickness in excess of flotation F, is defined as

$$F = (h + w) - w\, \rho_w/\rho_i, \qquad (3)$$

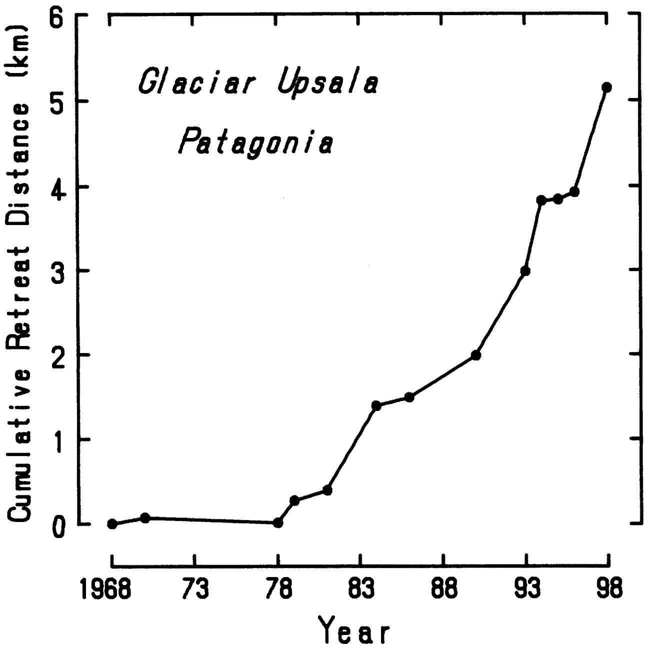

FIGURE 4. Width-averaged retreat distance of Glaciar Upsala from 1968 to 1998.

where, h is the surface height above the water level, w is the water depth, and ρ_w and ρ_I are densities of water and ice, respectively.

Heights of glacier surface above water near the terminus were measured by conventional angle surveys in November 1993, namely 45 m at D(5.3) for a front cliff (J. Leiva and R. Sendra, pers. comm., 1995) and 55 m at D(6.3) about 1 km upglacier from the front (Skvarca et al., 1995), both near the glacier center line. From the bed profile L2 (Fig. 5), we can see the water depths of about 400 m at D(5.3) and about 550 m at D(6.3). Hence the water depths divided by ice density 900 kg m^{-3} yield 444 m and 611 m (Table 1), which are almost equal to the estimated ice thicknesses (445 m and 605 m, respectively). In other words, the effective pressure at the bed is nearly zero. This estimate indicates that the frontal, central part of the glacier was just in flotation in summer 1993. Whereas, at 2.5 km upglacier from the terminus near the center in 1990, F was 23 m (Table 1).

From this consideration, it is found that a temperate glacier with a floating terminus in freshwater really exists. Also, Glaciar Nef in northern Patagonia exhibited the bending southward floating tongue on the JERS-1 image in 1993 (Aniya et al., 2000b). Furthermore, we can propose that the "height-above buoyancy model" with a mean excess height of 50 m may not be applied to freshwater calving glaciers.

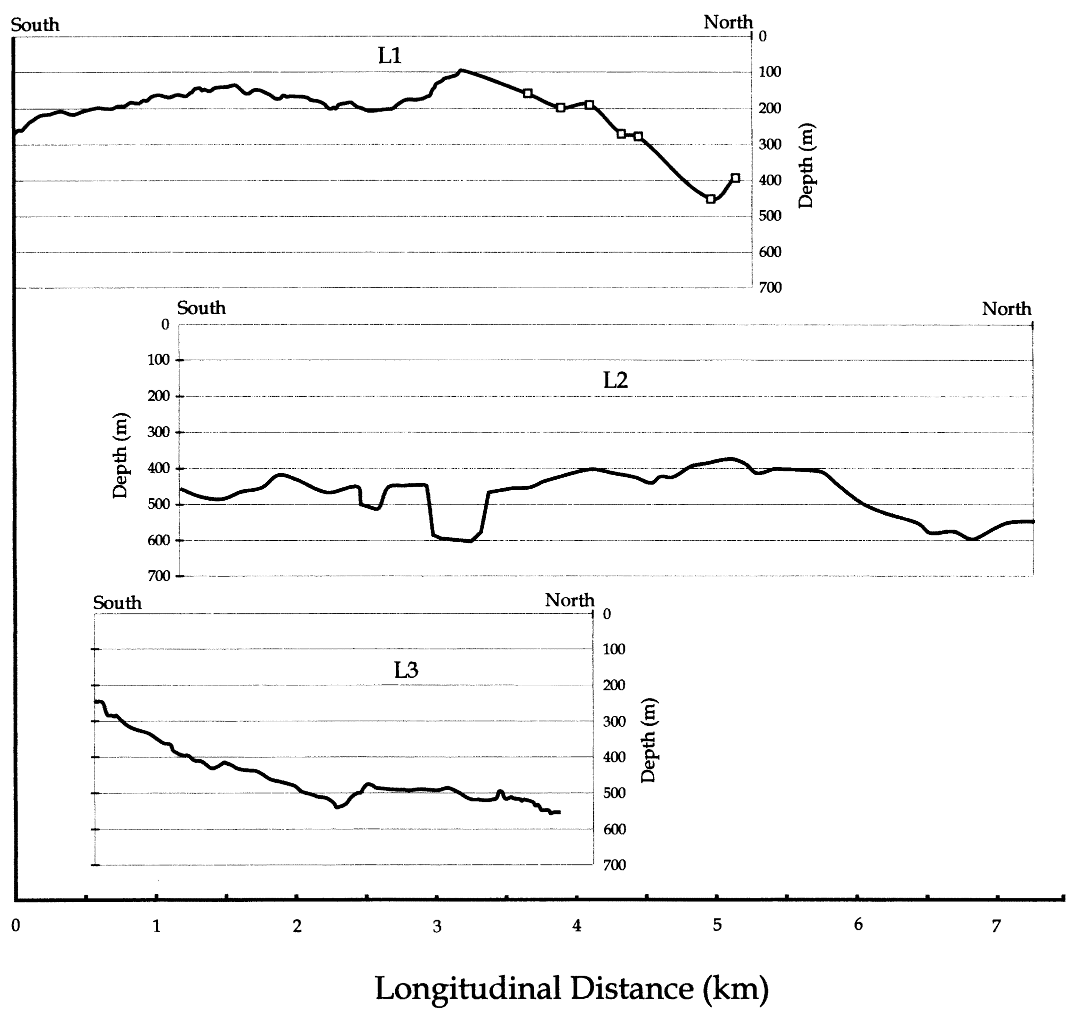

FIGURE 5. Longitudinal profiles of water depth along L1, L2, and L3 in Brazo Upsala, western arm of Lago Argentino. Seven open squares seen in L1 are data measured in 1994 and 1997 by H. Svetaz (pers. comm., 1997). Longitudinal distance originates from the southern end of line L1 (Fig. 2). A point at distance n (km) is expressed by D(n) in the text.

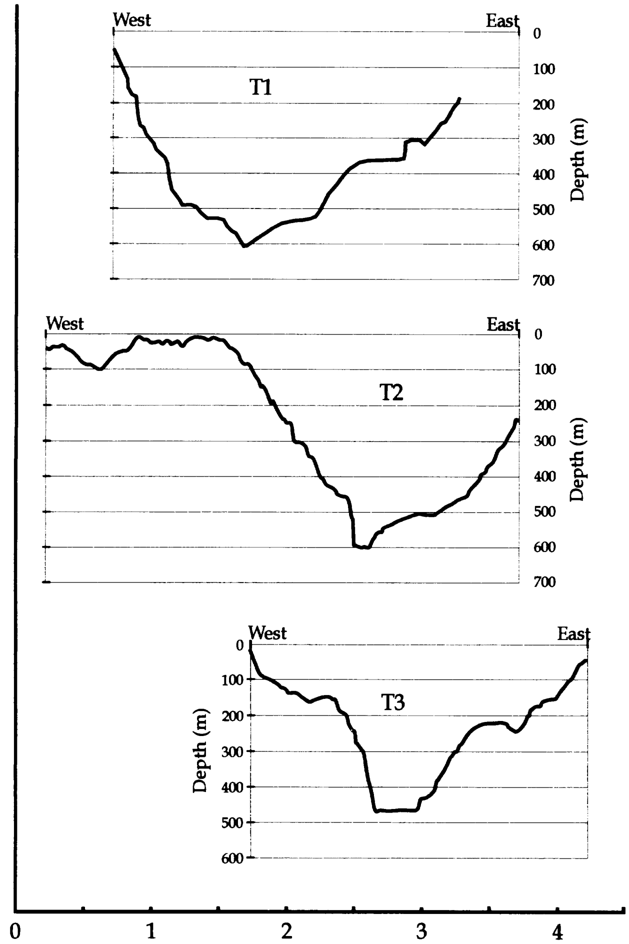

FIGURE 6. Transverse profiles of water depth along T1, T2, and T3 in Brazo Upsala. Transverse distance originates from the western shore near the line T2 (Fig. 2).

THE CONTINUITY CONDITION NEAR THE GLACIER TERMINUS

We now examine a mechanism of the glacier recession from a point of view of ice dynamics. The continuity condition (mass conservation) at any point of a parallel-sided glacier can be expressed as,

$$\partial H/\partial t = b - u(\partial h/\partial x + \partial w/\partial x) - H\partial u/\partial x, \tag{4}$$

where, H ($= h + w$) is ice thickness, b is surface mass balance (negative for ablation), u is depth-averaged flow velocity, x is longitudinal distance taken positive down-glacier, and t is time. Melting at the bottom and sides of glacier is neglected and ice is assumed as incompressible. We now evaluate the contributions of each term to the ice thinning, particularly the longitudinal strain rates $\partial u/\partial x$. This equation is applied to the frontal part of Glaciar Upsala in 1990–1993 when some field measurements were made.

Parameters in equation (4) near the glacier center line in 1990–1993 were estimated as follows: $\partial H/\partial t = -10$ m a^{-1}, measured at around D(6.3) from 1990 to 1993. $b = -16 \pm 2$ m a^{-1}, estimated from annual measurement of ablation at Glaciar Perito Moreno (Naruse et al., 1997). Flow velocities (1300 m a^{-1} and 1600 m a^{-1}) were obtained at D(6.3) in November. At Glaciar Perito Moreno, velocities in November are slightly, about 10 %, larger than the annual mean value. Depth-averaged velocity is

TABLE 1

Calculation on floatation near the terminus of Glaciar Upsala

Time	Position	Height h (m)	Depth w (m)	Excess floatation F (m)[a]
Nov. 1993	1 km from 1993 front glacier center line	56 m	550 m	−0
Nov. 1993	Cliff at 1993 front glacier center line	44 m	400 m	−0
Nov. 1993	1 km from 1993 front 0.5 km from eastern margin	26 m	(300 m)[b]	(−0)
Nov. 1990	2.5 km from 1990 front glacier center line	84 m	550 m	23 m
Nov. 1990	2 km from 1990 front 0.5 km from eastern margin	64 m	(300 m)	(30 m)

[a] When F is calculated as in a negative value, F is shown as −0, because (h + w) does not indicate ice thickness in such condition.

[b] Values in parentheses are less accurate, due to extrapolation of w.

assumed as 90% of the surface value. Hence, u = 1200 m a^{-1}, assumed for an annual mean value. From surface heights h (Table 1) at D(5.3) and D(6.3), $\partial h/\partial x = -0.012$. From the bedrock profile L2 (Fig. 5), a mean bed slope between D(5.3) and D(7.3) is derived as $\partial w/\partial x = -0.108$. As to a mean thickness at D(6.3) between 1990 and 1993, H = 620 m.

Thus, longitudinal (extension) strain rate, $\partial u/\partial x$, was calculated from equation (4) as 0.22 a^{-1}, which is significantly larger than typical values in temperate glaciers (Paterson, 1994, p.253). This result implies that the glacier near the terminus was stretching longitudinally with a significant degree though it was in the ablation area. The similar behavior was found at Columbia Glacier, where along-flow extension rate near the terminus increased rapidly from about 0.05 a^{-1} in 1981 to about 0.6 a^{-1} in 1985 in harmony with the front retreat (Venteris et al., 1997). The large extension was considered to be controlled by a longitudinal increase in basal sliding toward the terminus due to the basal water system (Van der Veen, 1996; Venteris et al., 1997).

Conclusions

Before 1990, when the terminus of Glaciar Upsala was located at or farther southward from the islands and the surrounding bedrock rises in the western part of Brazo Upsala, the dynamic behavior should be affected by the bed topography in terms of normal and shear stresses. During 1990–93, the glacier terminus is considered to have been floating, which may indicate a difference between freshwater and tidewater calving glaciers. Significantly large longitudinal extension strain rate of 0.22 a^{-1} was deduced for the same period, which should be caused by the enhanced basal sliding near the terminus. Stretching results in ice thinning, which may cause extensive calving, so that the glacier retreats farther. However, a question what phenomenon is a fundamental cause for variations of freshwater calving glaciers remains unsolved until future studies.

Acknowledgments

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