Glaciological investigations on Union Glacier, Ellsworth Mountains, West Antarctica

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ABSTRACT. Union Glacier, West Antarctica, was intensively mapped in December 2008, when an over-snow traverse was conducted by CECS and ALE, with the aim to determine the ice-dynamical characteristics of the glacier, through mapping the crevasse fields and by providing a glaciological baseline for future studies. A mean ice thickness of 1450 m was measured, confirming the presence of a deep subglacial topography (~900 m below sea level), much deeper than previously estimated. Ice velocities were also measured at 21 stakes drilled into the ice at the narrowest gate of the glacier between December 2007 and December 2008, yielding a mean value of $22.6 \, \mathrm{m \, a^{-1}}$. These velocities, combined with the measured ice thicknesses and a numerical model, yielded an ice flux of $0.10 \pm 0.03 \, \mathrm{km^3 \, a^{-1}}$ w.e. Considering the ice basin above this gate, a mean surface mass balance of $0.18 \pm 0.05 \, \mathrm{m \, a^{-1}}$ was estimated, a value consistent with a mean snow accumulation for nearby ice streams. These values indicate that the glacier is at present near equilibrium.

INTRODUCTION

In December 2008, a 320 km over-snow tractor traverse was conducted by the Centro de Estudios Científicos (CECS) with the support of Antarctic Logistics and Expeditions (ALE), through the southern part of the Ellsworth Mountains, West Antarctica, from Patriot Hills (80°18′ S, 81°22′ W) to Union Glacier (79°46′ S, 83°24′ W) (Fig. 1). Union Glacier flows into the Constellation Inlet of the Ronne Ice Shelf, which has been stable in recent decades in spite of recent atmospheric warming affecting the Antarctic Peninsula and West Antarctica (Steig and others, 2009). The Ronne Ice Shelf is fed by several ice streams, Union Glacier being one of the smaller and Rutford Ice Stream one of the main contributors (Rignot and Thomas, 2002).

Further north in the peninsula, several large ice shelves (e.g. Wordie, Larsen A and Larsen B) have collapsed in recent decades (Vaughan and Doake, 1996; Scambos and others, 2004). As a consequence, many glaciers have accelerated and thinned (Rignot and others, 2004, 2005). The main driving force behind these changes is the rapid recent atmospheric warming trend observed in the Antarctic Peninsula (Turner and others, 2005). This atmospheric warming has been confirmed further south by borehole measurements performed at Rutford Ice Stream (in a location 182 km north of Union Glacier). However, a similar measurement at Subglacial Lake Ellsworth (173 km west of Union Glacier) did not exhibit any significant trend (Barrett and others, 2009).

Rutford Ice Stream, which is much larger than Union Glacier, is characterized by tidally modulated ice flow (Gudmundsson, 2006) with surface velocities between 87 and 375 m a⁻¹ at distances between 240 and 40 km from the grounding line (King, 2009). Ice-flow velocities are stable over recent decades (Gudmundsson and Jenkins, 2009), and the glacier has a near-equilibrium net balance (Rignot and

Thomas, 2002). Recent studies indicate that its basin is thickening slightly (Zwally and others, 2005; Wingham and others, 2006; Vaughan and others, 2008). The Ronne Ice Shelf, in which Rutford Ice Stream flows, is not vulnerable to a disintegration process such as affects ice shelves in the Antarctic Peninsula (Vaughan and Spouge, 2002).

Considering the above characteristics, what kinds of changes are expected at Union Glacier?

Union Glacier has an estimated total area of 2561 km², according to the RADARSAT-1 Antarctic Mapping Project (RAMP) digital elevation model (DEM), with a total length of 86 km from the ice divide with Institute Ice Stream down to the grounding line at Constellation Inlet. It is composed of several glacier tongues. The main trunks (Union and Schanz; Fig. 1) feed through narrow valleys (9 km wide) from the interior plateau before joining together at the 'gate'. This gate, which is the narrowest section of the glacier (7 km wide), is characterized by a bare ice surface or blue-ice area (BIA) and a central moraine line (Fig. 2), the latter composed of clasts and small debris. The Union Glacier BIA is located where katabatic winds are enhanced. A 21-stake array for determining ice velocities was installed at this location in 2007, in order to determine surface mass loss.

Thus the aim of this research is to study the ice dynamics and basic glaciological characteristics of Union Glacier, providing a baseline for future modelling studies of glacier behaviour. Since the 1962/63 expedition conducted by the US Geological Survey (USGS), and an airborne survey by the British Antarctic Survey in 1974/75 (Lythe and others, 2001), very little work has been done on Union Glacier. In 2008, a survey was organized using dual-frequency GPS receivers, a radar for deep ice measurements and a ground-penetrating radar (GPR) for crevasse detection. These sensors were mounted on a convoy pulled by a Camoplast tractor. The traverse, from the Patriot Hills summer camp used by
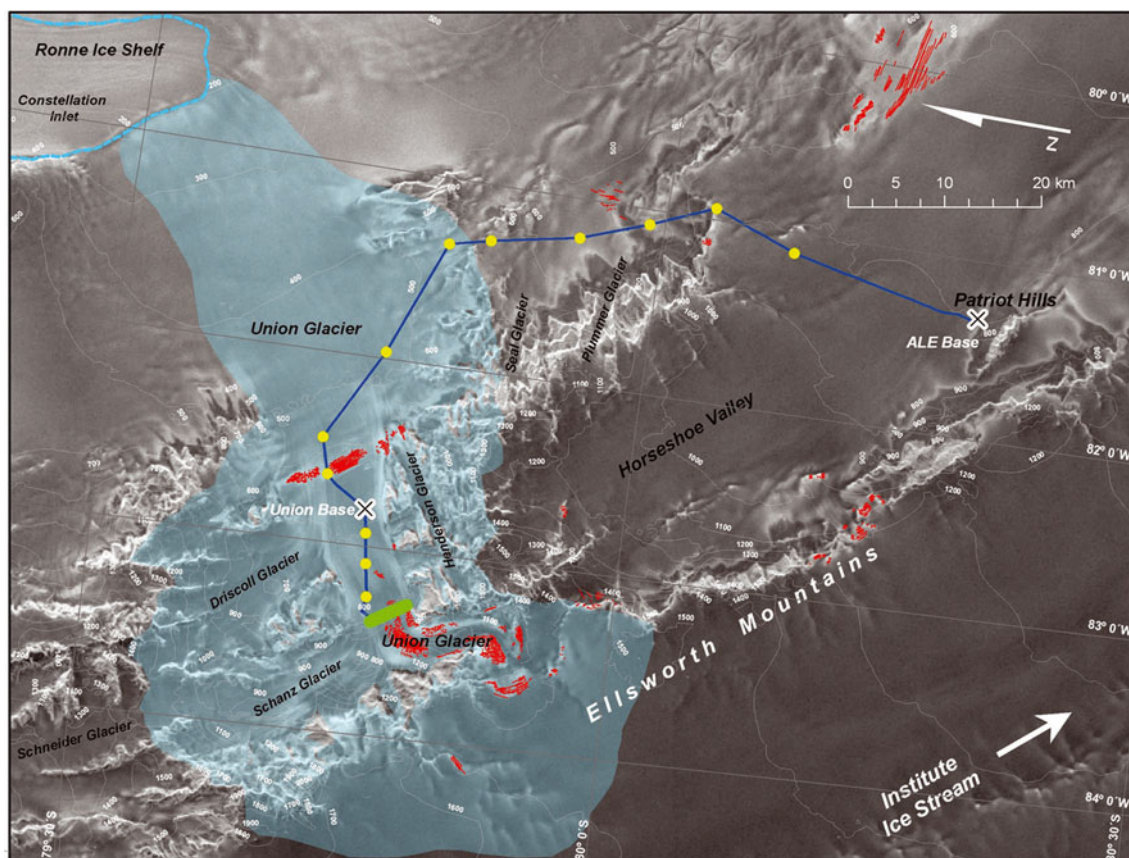
Fig. 1. Location map based on a RADARSAT mosaic from West Antarctica with contour lines from RAMP (light grey). The traverse route, where radar and GPS were measured, is shown in blue. Crevasse fields are shown in red. Green line shows stakes installed in 2007 and resurveyed in 2008, yellow circles show other stakes installed in 2008, and light blue dotted line shows the grounding line position. Union Glacier basin is shown in light blue shade. Base camps are indicated by 'X'.

ALE to Union Glacier, lasted 2 weeks. Several stakes were installed along the traverse to measure ice velocities, and continuous GPS measurements were performed all along the track to survey the surface topography.

METHODS
Remote sensing

Analysis of satellite imagery has been shown to be especially useful for mapping remote areas in Antarctica (Bindschadler, 1998). Applications include monitoring fluctuations in the extent of BIAs (Casassa and others, 2004), ice velocity determination (Stearns and Hamilton, 2005) and detecting the position and variations of the grounding line (Rignot, 1998). In this work we combine Advanced Spaceborne Thermal Emission Reflection Radiometer (ASTER) satellite images with ground-based GPS and radar data in order to map crevasse fields and plan safe over-snow traverse routes.

Crevasse fields can be detected on the ASTER visible/near-infrared (VNIR) colour composite images and in the RAMP mosaic. They appear as discontinuities on the snow surface, sometimes as open linear features but more often covered by snow and only evident by the shadows cast by snow built up at the edges of the crack. The fact that the crevasses are marked by linear discontinuities on the image allows them to be detected (Bindschadler and Vornberger, 2005), both visually and by applying filters (e.g. high-pass).

Differential GPS

A 21-bamboo-stake network was installed on Union Glacier in December 2007, using Topcon GR3 dual-frequency GPS receivers (Fig. 1, green colour). Due to logistic/weather constraints, differential GPS measurements lasted <30 s per stake, yielding position errors of the order of 1 m. An independent survey was done in December 2008, with dual-frequency Javad GPS model Lexon GD receivers, used to survey the ice surface as well as the stakes installed in 2007. To improve the accuracy of the second-year survey, a new GPS base metal pin was drilled and glued to a rock outcrop located near the base camp at Union Glacier (Figs 1 and 2). This point was used as a GPS base station, and was surveyed simultaneously with the other measurements (i.e. kinematic for radar and stop-and-go for the stakes). Each stake was measured for approximately 30 min, allowing a precise positioning of all of them (better than 10 cm accuracy). Comparing the position of these 21 stakes yields a detailed record of ice velocities over the 12 month period, with an estimated combined error of 1 m. In 2008, 13 stakes were also installed along the convoy track (Fig. 1, yellow colour), which will allow a better understanding of the ice dynamics in the future.

Ice thickness

A pulse-compression radar depth sounder (ULUR-1.5) was designed and built by CECS for thickness measurements in deep cold ice. The radar operates at a central frequency of
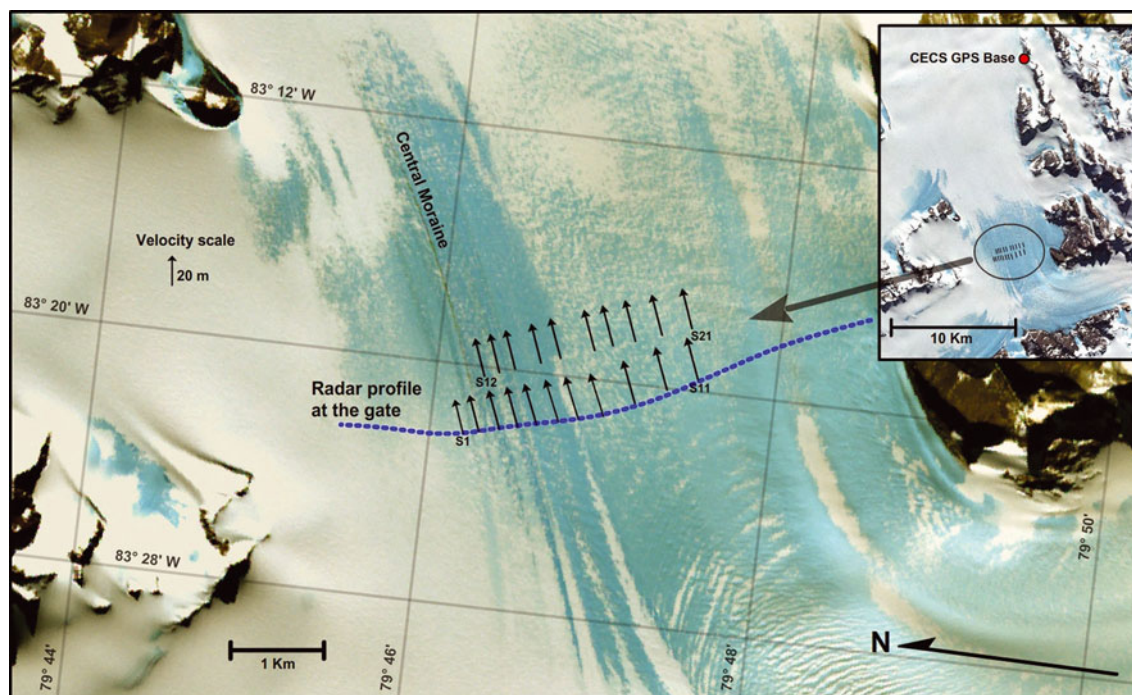
Fig. 2. Ice velocities at the Union Glacier gate measured between December 2007 and 2008. The inset shows the BIA and the GPS rock station installed in 2008 (CECS GPS point as a red dot), which is close to the USGS metal pin installed in 1962/63 (USGS, 1966) and resurveyed in 2008. Dashed blue line shows the surface radar track of the profile shown in Figure 3.

155 MHz, with a bandwidth of 20 MHz, a sample rate of 100 MHz and peak power of 200 W. Eight-element Yagi antennae were used for both the transmitter and receiver. Post-processing included an adjustment of the gain function to the raw data, mainly to enhance the bedrock echo.

RESULTS

Subglacial topography

During the radar survey, data were collected over 320 km (Fig. 1). The first part of the surveyed profile, crossing the Horseshoe Valley, yielded a maximum thickness of 2120 m, in an area where previous radar campaigns were unable to penetrate the total ice thickness (Casassa and others, 2004).

The smaller glaciers situated at the northern flank of the Horseshoe Valley (Plummer and Seal Glaciers) have maximum ice thicknesses at their centre flowline of 386 and 428 m, respectively. Several profiles were surveyed on Union Glacier, yielding a mean ice thickness of 1450 m along the gate (Fig. 3), where the stake network was installed in 2007. The subglacial topography of the valley has U-shaped flanks and very small roughness.

Mass balance

The stakes installed in 2007 at the glacier gate (Fig. 2) were resurveyed after 1 year, when a negative mass balance was detected, with a lowering of 16 ± 1 cm of the ice surface (error of 4 cm in each stake height difference). This value is consistent with the prevailing BIA condition in this sector of the glacier.

Ice-flux computation

The main input to compute the ice flux is the horizontal displacement of the 21 stakes. The mean displacement of the stakes is $22.6 \, m \, a^{-1}$, with maximum velocities of $24.5 \pm 1 \, m \, a^{-1}$ and minimum velocities of $19 \pm 1 \, m \, a^{-1}$ at the central moraine of the glacier. These velocities are one order of magnitude lower than those of Rutford Ice Stream, and are typical for ice frozen to the bed (Vaughan and others, 2008).

Ice flux was modelled through the gate where ice surface velocities were measured between 2007 and 2008 (Fig. 2). The glacier width at this gate is 7 km, and the model domain covers a subsection of 4.59 km from the central moraine (where the main flow of Union Glacier joins Schanz Glacier) to the lateral side (Fig. 3). Surface topography was measured with GPS, allowing positions to be determined approximately every 9 m along track, covering most of the modelled section. The 150 m separating the end of the surface topography profile and the glacier margin (where the terrain was inaccessible for the surveying system) were extrapolated with a sinusoidal curve, resulting in a 24 m height drop, approximately matching field observations.

Ice thickness at this gate, measured with the radar system, was approximately 1456 m. Over the last part of the profile before the glacier margin (approximately 550 m horizontally), where the radar profile was obscured by lateral reflection coming from the rock wall, the bed elevation (–697 m) was interpolated up to the ice/rock margin contact by a parabolic function (X distance 6500–7000 in Fig. 3).

Ice velocities were computed in two steps: (1) by defining the surface velocities across the gate by combining measured with modelled values and (2) by estimating the vertical change in horizontal velocities.

1. Surface velocities

The surface velocities between the measured stakes along the profile (Fig. 3) were linearly interpolated. For zones lacking stakes (the last 2 km), ice velocities were calculated
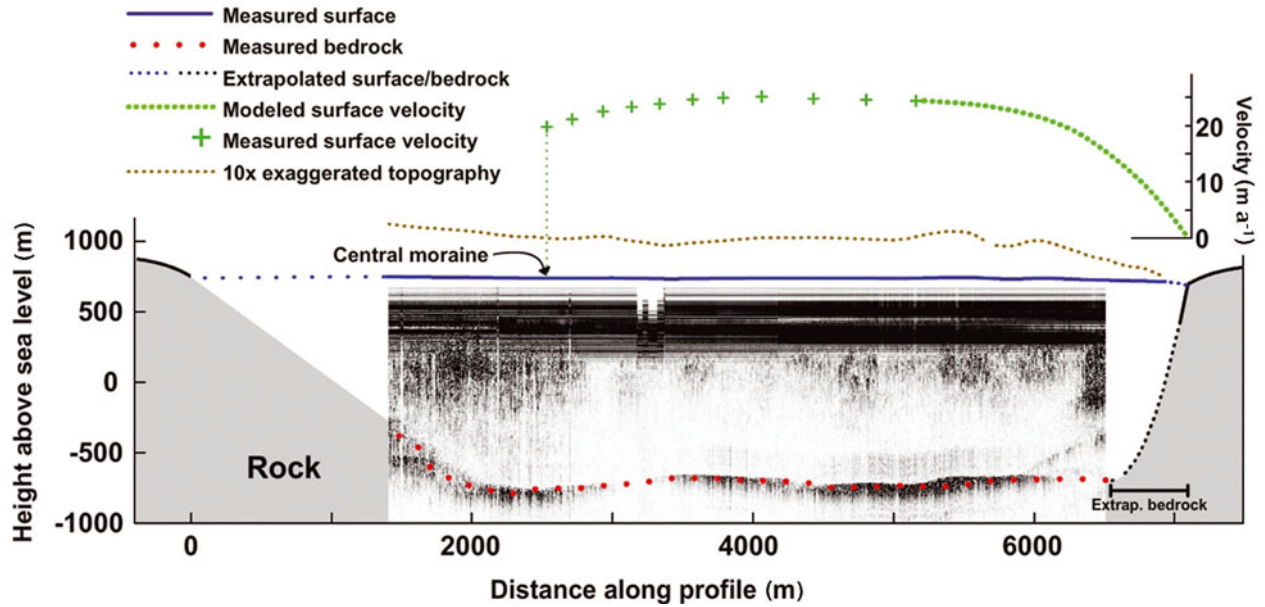
Fig. 3. Cross-section (gate) where ice flux was calculated. Surface topography measured with GPS is shown in blue. The blue dotted line is the surface topography extrapolated to the rock margins, and the brown dotted line shows the same topography with vertical exaggeration. Black lines represent rock at the glacier margin. Black dotted line shows the interpolated subglacial topography data as described in the text. Green crosses show measured surface velocities, and green dotted line shows modelled values. The radargram of the glacier cross-section is also displayed, indicating the interpreted subglacial topography as red dots. The banding that obscures the uppermost 500 m in the radargram is due to system saturation.

using the model for valley glacier velocities described by Nye (1965), where the horizontal and vertical variations are governed by

$$\frac{\partial \tau_{xy}}{\partial y} + \frac{\partial \tau_{xz}}{\partial z} = -\rho g \sin \alpha \qquad (1)$$

$$\frac{\partial u}{\partial y} = 2A\tau^2 \tau_{xy}, \quad \frac{\partial u}{\partial z} = 2A\tau^2 \tau_{xz}, \qquad (2)$$

where τ_{xy} and τ_{xz} are shear stresses $\tau = \sqrt{\tau_{xy}^2 + \tau_{xz}^2}$, α is the surface slope along the flowline, g is the gravitational acceleration constant and A is Glen's flow-law parameter. z is the vertical coordinate measured upwards, x is measured in the flow direction and y is the horizontal coordinate measured from the centre of the glacier towards the side perpendicular to the flux; u is the ice velocity along the x axis.

We used $\rho = 900\,\mathrm{kg\,m^{-3}}$, $g = 9.81\,\mathrm{m\,s^{-2}}$ and $A = 1.7 \times 10^{-16}\,\mathrm{kPa^{-3}\,s^{-1}}$, which is recommended for ice at $-20°\mathrm{C}$ (Paterson, 1994). The slope α in the profile section was estimated for the section immediately upstream of the gate, where slopes of $0.5–10°$ were found, with an average value of $1°$.

Since $\tau_{xz} = 0$ at the surface (Paterson, 1994), Equations (1) and (2) can be rewritten as

$$\frac{\partial u}{\partial y} = 2A\tau_{xy}^3$$
$$\qquad (3)$$
$$\frac{\partial \tau_{xy}}{\partial y} = -\rho g \sin \alpha.$$

Both equations were numerically integrated, using the measured surface velocities as initial condition, in order to complete the surface velocities to the gate margin. Integration constants were numerically adjusted during each iteration to match the boundary condition of $u = 0$ at the rock/ice contact (assuming no basal sliding).

2. Integrating velocities through the vertical column

Using the above surface velocities, subsurface velocities were calculated analytically, using Glen-type rheology under steady laminar flow for $n = 3$. Under these conditions, depth-averaged velocities are 80% of surface velocities (Paterson, 1994). This results in a total flux of $0.10\,\mathrm{km^3\,a^{-1}}$ w.e. (using $\rho = 900\,\mathrm{kg\,m^{-3}}$). Figure 3 shows the geometric setting of the gate with the surface measured velocities. Considering that the basin upstream of the modelled gate has an area of $559\,\mathrm{km^2}$, a mean net surface mass balance of $0.18\,\mathrm{m\,a^{-1}}$ w.e. is needed to keep the glacier in a steady state. This average net mass includes high spatial variability, with areas of net accumulation at the plateau at the top of the basin and areas with strong ablation/sublimation evidenced by the BIAs near the measured gate.

Multiple error sources remain in the total modelled ice flux, including differences between modelled and measured surface velocities. As a result, it is very difficult to assess the magnitude of this difference. However, it is very easy to establish a reasonable maximum value by considering a constant surface and subsurface ice velocity of $22.6\,\mathrm{m\,a^{-1}}$ (the mean value of the 21 measured stakes), leading to a flux of $0.13\,\mathrm{km^3\,a^{-1}}$ w.e., 30% more than the modelled flux. Assuming a 30% relative error, the total flux would be $0.10 \pm 0.03\,\mathrm{km^3\,a^{-1}}$ w.e., yielding an equilibrium net mass balance of $0.18 \pm 0.05\,\mathrm{m\,a^{-1}}$ w.e.

DISCUSSION AND CONCLUSIONS

Surface and subglacial topographies

The surface and subglacial topography measured during the 2008 campaign provide more precise and detailed results than those available until now (Fig. 4).

Surface topography, resulting from the GPS survey in 2008 (decimetre precision), is similar to the available Ice,
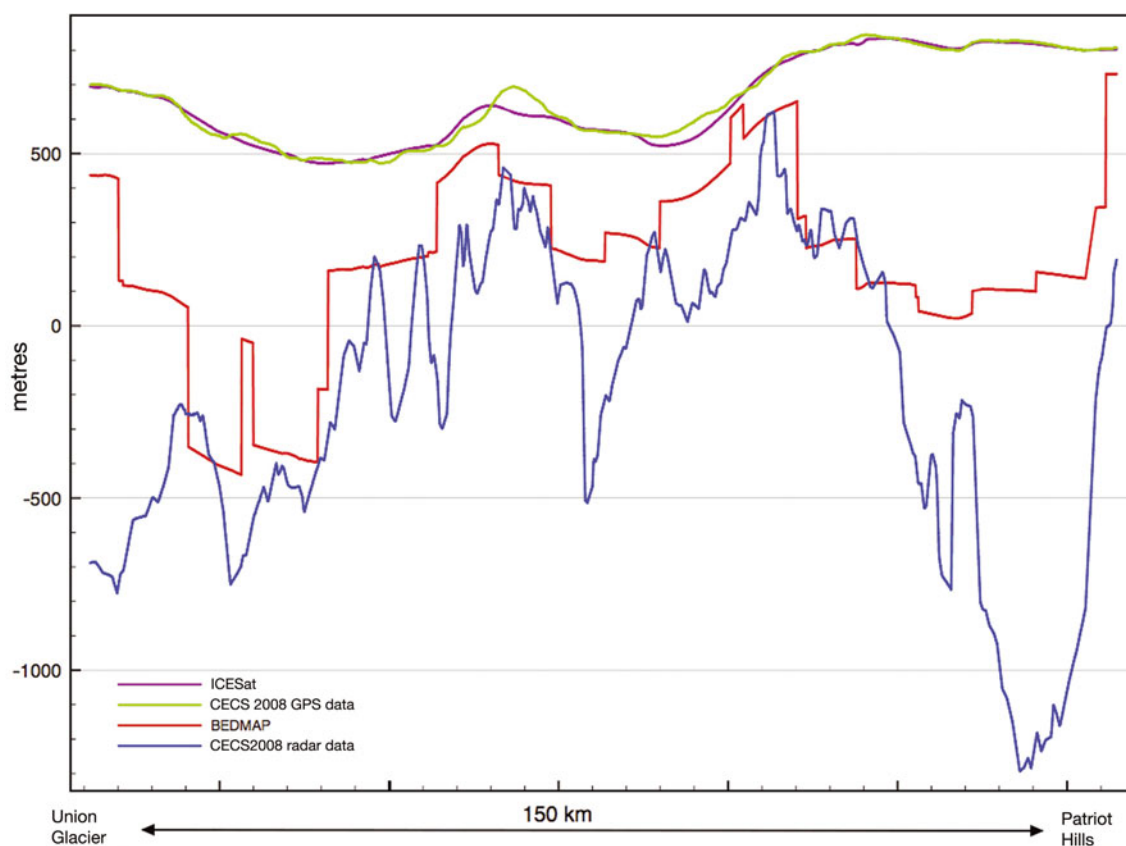
Fig. 4. Surface and bed profiles collected along the CECS 2008 traverse from Union Glacier to Patriot Hills, and compared to ICESat and BEDMAP data.

Cloud and land Elevation Satellite (ICESat) data general topography (500 m pixel size), with some anomalies in areas characterized by a rougher topography, but generally consistent with a mean difference of <5 m.

Subglacial topography obtained in 2008 is, however, very different to the BEDMAP data (Lythe and others, 2001), which is to be expected in view of the sparsity of available measurements in this area prior to the 2008 campaign. Comparison between BEDMAP and the new dataset shows differences of >50% along the same track where subglacial topography is below sea level (Fig. 4). For example, the deep valleys along this track are not represented correctly in BEDMAP. Thus it seems that Union Glacier is located in a much deeper subglacial valley than previously estimated.

Comparing 2008 ice-thickness measurements with BED-MAP shows a reasonable relation in general, but in areas with steep basal slopes and deep subglacial beds, differences are as high as 66% of the ice thickness. The difference between BEDMAP and the new ice-thickness data is most pronounced in the vicinity of Patriot Hills, where the comparison exceeds 1300 m (Fig. 4).

Glacier mass balance

It is not yet possible to establish the net mass balance of Union Glacier, but most evidence suggests a near-equilibrium state. Accumulation for this area inferred from regional atmospheric models indicates values of 0.09–0.34 m a^{-1} w.e. (Van den Broeke and others, 2006), consistent with the range of the estimated equilibrium net mass balance inferred from our ice-flux model (equivalent to an accumulation of 0.13–0.23 m a^{-1} w.e. neglecting ablation/sublimation). These values are similar to the results obtained at the Horseshoe Valley, where GPS surveys have been done in addition to stake mass-balance measurements since the mid-1990s, resulting in a balance accumulation of 0.17 m a^{-1} w.e. (Wendt and others, 2009).

In order to determine more precisely the present mass balance of Union Glacier and its possible future trends, it will be necessary to further constrain the bedrock topography (to anticipate dynamic glacier responses), as well as to acquire more representative accumulation/sublimation measurements. For this purpose, new radar profiles from the Union Glacier gate down to the local grounding line, following the central flowline, as well as along several transversal cross-sections, will be necessary. These profiles seem feasible in areas where crevasse fields are lacking. In addition, radar profiles and GPS surveys for both velocity determination and surface accumulation should complete the analysis.

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