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Constraining Ice Dynamics at Dome C, Antarctica, Using Remotely Sensed Measurements

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Abstract. A first time description is given of the ice flow at Dome C, Antarctica, around the EPICA drilling site. We used satellite radar altimetry to obtain the precise ice surface topography, airborne radio echo sounding to obtain the ice thickness and satellite SAR interferometry to derive one component of the surface velocity field. The balance flux around the Dome C area is then accurately mapped and comparisons made between driving stress, surface and balance velocity to help us describe the ice flow in the region. As a by-product of the study, we also recover anomalies in the ice flow conditions in sub-glacial lake locations. These effects result from locally invalid shallow-ice approximation. The results of this study form the basis for future investigations of the ice flow conditions at Dome C in relation to ice core interpretation.

Introduction

The EPICA (European Project on Ice Core drilling in Antarctica) chose Dome C as the site to perform the deepest drilling in Antarctica. The choice of this site was guided by the fact that around the summit of Dome C, the horizontal movement of the ice surface is negligible and thus the total ice column is formed by snowfall accumulation. Hence analysis of the ice cores gives a direct measure of the climate evolution at this place. The aim of this paper is to demonstrate how the use of remotely sensed data can help understand the ice movement around the EPICA drilling site. We map the balance, deformation and surface velocities using surface topography, thickness and interferometric synthetic aperture radar data. Finally, we compare all of these velocity fields and draw some conclusions about the ice movement around Dome C.

Datasets. The ice surface topography has been computed using radar altimeter data from the ERS1 geodetic cycle during 1994-1995 [Remy *et al.*, 1999].

The chosen grid spacing is a compromise between the radar footprint (2-3 km diameter) and the along and cross track resolution. The mapped spatial resolution is thus 1/30 degree (i.e., 3.7 km in a north-south direction and 1 km in east-west direction). All calculations presented in this paper refer to this grid unless otherwise indicated. The height precision of the mapped topography has been evaluated by comparison with a kinematic GPS survey [Ceffalo *et al.*, 1996] and agreement is at an rms level of 20 cm. The ice thickness data are the result of an airborne radio echo sounding survey [Tabacco *et al.*, 1998] collected over a 80*120 km rectangular grid with a 10 km across track grid spacing, refined to 5 km in the central region. These data have been interpolated [Remy and Tabacco, 2000] to the same 1/30 degree grid as the ice surface topography and the difference gives the bedrock topography. Finally, two ERS SAR acquisitions have been used to perform the interferometry. The SAR scene selection was difficult as there was no data available from ascending tracks over Dome C and only a few useful scenes on the descending tracks. At Dome C, the ice flow is slow (0-20 cm/a) and thus it is necessary to use a long temporal baseline (typically of the order of 70 days or more) to resolve a significant signal. It is also important to keep any radar coherence between the two acquisitions to form the interferogram. The coherence between scenes decreases with time separation (the surface state changes) and with orbital separation between the 2 acquisitions (if the ground target is not seen with the same incidence angle, the target response will differ). It is therefore important to have short orthogonal baselines. Many attempts at forming interferograms with long baselines lead to incoherence and no result. We only found one image pair (descending pass), centered at Dome C, with a 45 m orthogonal baseline. These data were from orbits ERS1-24781 in April 1996 and ERS2-4106 (frame 5193) in February 1996, with the time separation being 69 days.

Velocity field construction.

Balance velocity. The balance velocity results from the equilibrium between the snow accumulated on the ice cap (vertical flux) and the ice flow (horizontal ice

flux). The ice cap is assumed stationary and the total ice column is assumed to flow in the same direction (i.e., in the direction of greatest slope), in other words the shallow ice approximation is used. The balance velocity can thus be represented as the mean value of the ice column horizontal velocity assuming a plane flow:

$$U_B(x) = \frac{1}{H} \int_{top}^x acc(x') dx' \quad (1)$$

where H is the ice thickness, acc the accumulation rate and x the abscissa along the flow line from the top of the dome. The integration is performed along a flowline following the greatest slope. We computed the balance velocity on a 5 km grid using the method of *Budd and Warner*, [1996]. We experimented with many horizontal spatial scales to compute the slope, ranging from 10–30 km (3–10 times the ice thickness), but did not find any major differences, perhaps because the slopes were small. The value of 10 km (which corresponds to the resolution of our basic datasets) gave a more detailed resolution for the flow direction. Balance fluxes ($U_B * H$ in m/a) are computed over the whole domain surrounding Dome C since they do not need the knowledge of the ice thickness. The accumulation rate (0.04 m/a) is taken from *Petit et al.*, [1982], based on a nuclear analysis of snow pits. Since the correlation length scale of various accumulation rate compilations is of the order of 100 km, the assumption of constant accumulation over our area of interest is reasonable. The balance velocity (Figure 1) is then obtained by dividing by the ice thickness and projected to a geographical grid. To first order, the ice velocity field is a divergent flow, starting from the top of the summit, with iso-values form-

ing concentric ellipses. More rapid flow corresponds to bedrock troughs, where the ice flow seems to converge. For example, part of the west-east flux coming from Dome C towards the mountain near 125E is redirected northward and southward. The horizontal balance velocity values range from 0 near the drilling site to 25–30 cm/a at 25 km from the summit or 3200 m elevation.

Driving stress. The driving stress (τ_d) is the main force acting on the ice column and is responsible for the ice deformation. We estimated τ_d components using the precise ice topography (h), slope and thickness (H) as

$$\begin{cases} \tau_{dx} = \rho g H \frac{\partial h}{\partial x} \\ \tau_{dy} = \rho g H \frac{\partial h}{\partial y} \end{cases} \quad (2)$$

where ρ is the density of ice, g the gravity constant, x and y are horizontal coordinates. Here, longitudinal stress is not accounted for since it is difficult to estimate a-priori. But the high spatial resolution of the dataset would help to decide whether this stress is significant or not. Since there has not been significant ice deformation at Dome C to give it a preferential orientation, horizontal isotropy is assumed. The deformation velocity is defined as a function of the driving stress :

$$U_D = -HA\tau_d|\tau_d|^{(n-1)} \quad (3)$$

where A and n are ice constitutive law parameters (see [Glen, 1955; Paterson, 1994] for detailed description). This deformation velocity only refers to local quantities (thickness and slope) and is not a priori dependent on the history of the flow from Dome C, in the same way as the balance velocity is. Thus this calculation provides an independent estimation of the way that the ice flows.

SAR Interferometry. We computed phase differences between the two SAR acquisitions using both NASA/JPL and CNES (DIAPASON) software. The coherence level between the images is of the order of 0.35. The fringe pattern is based on [Rignot et al., 1995]:

$$\delta\phi = \frac{4\pi}{\lambda} (B \cos(\alpha + \theta) + \delta t V \cdot r) + \phi_{orb} \quad (4)$$

where λ is the radar wavelength (5.6 cm), B the orbital separation between the two data acquisitions, α the baseline angle with respect to the horizontal, r the unit range vector, θ the topography dependent angle of r , δt the time separation between the two acquisitions, V the surface velocity vector, ϕ_{orb} is the residual phase signal linked to any data uncertainties in the area. The height of the phase ambiguity (topographic change necessary to produce one fringe) is of the order of 150 m and is computed at each point with the topographic correction applied using the surface elevation data. The topographic correction is however less than one fringe given the flatness of the area. The main signal present in the computed interferogram is composed of a constant slope, which is also the case for the balance and deformation velocities projected in the SAR line of sight. In such cases, it becomes difficult to sep-

Figure 1. Map of the ice balance flux (arrows) in the Dome C area, the gray scale corresponds to the high resolution bedrock topography, the isoline to the high precision surface topography, the star to the EPICA drilling site and circles to sliding areas detected by comparing InSAR surface velocities to driving stresses.

arate between velocity-induced fringes and residual orbital fringes, which usually appear as linear ramps. We computed the interferogram using both the ESA precise orbits and Delft orbits data [Scharoo and Visser, 1998]. The first set of orbits introduced many fringes and we separated the orbital fringes by least squares fitting of the interferometric fringe pattern to a simulated fringe pattern based on the balance velocities. The Delft orbits lead to the same result, but without the need for any fitting. We therefore assumed that the Delft orbits were precise enough and did not contain any residual orbit signal. The final interferogram consisted of 2 1/2 fringes. The phase was further unwrapped using the Goldstein *et al.*, [1994] method. The error in the phase estimation is expected to be very small, since the troposphere is particularly dry at Dome C. Since the fringe pattern is simple, we estimated an error of 6 mm/a in the velocity based on the above phase noise. However, an absolute error from orbital uncertainties can still remain in the data at the same error level. We further assume a total error of 1 cm/a for the data in the following section. The resulting SAR velocities represent ice surface velocities, whilst the balance and deformation velocities refer to column-averaged velocities.

Comparison of the different velocity fields.

Having only one direction of SAR acquisition, we cannot extend the SAR measurement into two directions. We thus project all datasets in the horizontal plane and in the SAR line of sight direction. Figure 2 shows the maps of the projected driving stress, balance velocity and SAR velocity. We computed correlation coefficients to compare the different datasets. Comparing balance and SAR velocity (Figure 3a) leads to a 0.82 correlation, the scatter plot indicating a clear linear correspondence between both velocity estimations. (Standard theory for isothermal flow predicts a $(n+1)/(n+2)$ factor between both, but the noise level prevents any conclusion between $n=1, 2$ or 3). A discrepancy arises, however, for high velocities (>0.12 m/a), and we can see from Figure 2 that this mostly comes from the fact that the balance velocity is sensitive to the bedrock holes - see for example, near location (74.75S, 122.5E) or in the large valley near (124.5E). Adopting a vertical profile of the horizontal velocity, with most of the variation in deeper layers and a constant over the higher layers, the ratio between surface and average velocity varies inversely proportional to the ice thickness. This could also be the result of the vertical profile to longitudinal stresses present in the ice around bedrock depressions. The comparison between τ_{ud} and V_{SAR} is interesting - see Figure 3b. The correlation coefficient is -0.83. This relation between V_{SAR} (the surface velocity) and τ_d is just an expression of the constitutive law of the ice flow. One can observe a power relation between both variables but it is difficult to compute a unique curve fit using least squares estimation as the distribution of data is too variable. A value for the Glen exponent

Figure 2. Maps of the computed flow parameters, projected in the SAR look direction : (a) the driving stress in Pa, (b) the balance velocity in m/a and (c) the IN-SAR surface velocity in m/a.

[Glen, 1955] of 2 or 3 provides a good fit to the data, most of the misfit coming from our imprecise knowledge of the absolute velocity field and the error level. The values close to zero in Figure 3b correspond to (i) the dome area and, (ii) areas where the driving stress and velocity are mostly orthogonal to the line of sight. The region of Figure 3b having the same sign for stress and velocity (lower left, see box) is anomalous as the constitutive law of ice flow requires opposite signs and so these values are physically implausible. Regions where

Figure 3. Scatter diagrams of the three computed flow parameters : (a) the balance velocity against the surface velocity, (b) the surface velocity against the driving stress. Correlation coefficients are indicated within the diagram.

these anomalous values occur correspond to sliding areas. When sliding occurs on small size lakes, the lower part of the ice column is accelerated to ensure continuity, while the upper part of the ice column maintains a uniform velocity. A small anomaly (depending on the size of the lake) appears in the surface topography and hence in the driving stress. This mechanism can lead to very low stresses and surface velocities remaining of the same order of magnitude. Furthermore, sliding may locally change the direction of the ice flow. Hence, the projections of driving stress and surface velocity can have the same sign. In other terms, the shallow-ice approximation no longer holds. In any case, the areas corresponding to this kind of condition correspond to lakes detected by radio echo sounding, for example at (74.95S, 123.7E), the lake closest to Dome C and also at (74.9S 124.5E), a lake referenced by [Siegert *et al.*, 1996]. The areas showing these properties are displayed as circles on Figure 1. Finally the correlation between U_B and τ_d is 0.91.

Summary. In this paper, we investigated the flow of ice at Dome C in the area surrounding the deep drilling site. We demonstrated that remotely sensed data may help describe the ice flow by mapping a number of physical parameters of concern for ice flow studies. Ice surface velocities are well recovered from a 69 day temporal baseline ERS interferogram, the balance velocity is accurately mapped using surface and bedrock topographies and the driving stress can also be well described. We found that the shallow-ice approximation is not valid over small sub-glacial lakes. It appears that a method to recover sub-glacial lakes can be achieved using precise topography and velocity. However, our analyses lack an ascending ERS track for a full INSAR description of the surface velocity and of an absolute velocity reference. This situation can be overcome easily with future acquisitions of ERS2 or ENVISAT data. In

situ measurements of ice surface velocity will be beneficial to extend the analysis using in situ Doris and GPS surveys (C. Vincent and A. Cappra, personal comm.). The very slow movement (less than 1 cm/a) around the drilling site indicates that the ice present in the core may not have moved more than 1 km from that point, even for very old (deep) ice particles. Van der Veen and Whillans, [1992] showed that there can be discrepancy between the summit of the topography and the flow centre, but addressing this point will require additional and absolute measurements. Additionally, it would be of importance to have an improved measurement of this surface velocity and use modeling to constrain the flow divergence with depth, i.e., the longitudinal strain rate vertical profile [Bolzan, 1985].

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