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Ice-sheet radar layering and the development of preferred crystal orientation fabrics between Lake Vostok and Ridge B, central East Antarctica

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Abstract

Airborne radar data at 60 MHz are analysed to examine the flow of ice between the Ridge B ice divide and the Vostok Station subglacial lake in central East Antarctica. Interferometric SAR ice surface velocities show how three radar transects are aligned along the general direction of ice flow. Internal layering within the ice sheet is used to determine changes in the vertical component of ice flow along the transects. These data indicate ice sheet flow is related strongly to subglacial topography. In one transect, ice flows in a conventional base-parallel manner down a relatively uniform subglacial slope in the direction of ice flow. In the other two transects, internal layers show the flow of ice over a large subglacial hill upstream of Lake Vostok. Across the stoss face of this subglacial hill, analysis of the radar reflections from internal layers show that layers of ice with a preferred crystal orientation are likely to develop. This process results in layers of ice with a preferred crystal fabric across Lake Vostok at ice depths greater than ∼2.8 km. Evidence to support our analysis comes from the Vostok ice core where layers of ice with noticeable crystal alignment are abundant at ice depths greater than 2.7 km.

Keywords: Antarctica; ice; crystals; Vostok Station; ice cores

1. Introduction

The existence of a 200 km long subglacial lake beneath Vostok Station, central East Antarctica (Lake Vostok) has been known for a number of years [1–4]. At Vostok Station, a deep ice core which holds palaeo-environmental data spanning over 400 000 years [5] terminates 150 m from the surface of the sub-ice lake. Ice flows onto Lake Vostok from the Ridge B ice divide located ∼200 km west of the lake. There has been no recent attempt to identify the glacial processes that operate upstream of Lake Vostok and the Vostok ice core [6]. However, advances in our understanding of how ice-sheets flow, and new methods by which airborne radar and satellite data are processed and interpreted, make it timely to reinvestigate these processes. Here we present an assessment of several geophysical datasets between Ridge B and Lake Vostok from which we are able

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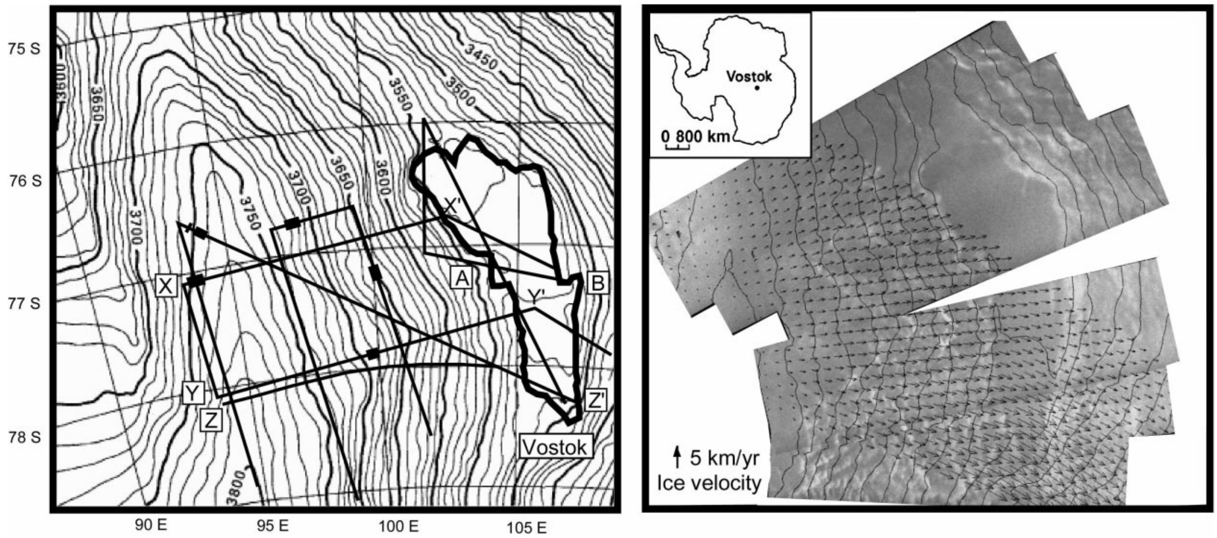

Fig. 1. (Left panel) ERS-1 altimetry of the ice sheet surface between Ridge B and Lake Vostok. Positions of radar flightlines, and subglacial lakes (denoted as black boxes) are indicated, as is the approximate outline of Lake Vostok [12]. (Right panel) InSAR ice-sheet surface velocities between Ridge and Lake Vostok [13].

to identify information on ice-sheet dynamics that occur upstream of the lake and ice core site.

Our geophysical datasets include ERS-1 altimetry of the ice-sheet surface, interferometric SAR (InSAR) data that reveal the ice sheet surface velocity and airborne radar sounding information in both single-pulse and pseudo cross-section formats. ERS-1 and InSAR data allow the determination of the ice divide position and the velocity of the ice sheet to be established. Airborne radar can be used to identify ice thickness, the presence of subglacial water, internal ice-sheet structure and a judgement of where boundaries between different crystal orientation fabrics are likely to develop in the ice sheet due to enhanced englacial stresses.

Spatial coverage of surface altimetry and InSAR data is continuous across the study area (Fig. 1). However, airborne radar data of use for our purpose are limited to three flow-parallel transects separated by between 10 and 100 km, and a small number of radar lines aligned non-parallel to ice flow (Fig. 1). The InSAR data allow us to measure ice surface velocity along the airborne radar lines and, by comparing results, assess how the ice sheet flows across subglacial topography upstream of Lake Vostok.

2. Interpretation of airborne radar Z-scopes

Airborne radar data, at 60 MHz with 250 ns pulse length, can be recorded in two formats. The most often used is the Z-scope mode, where single radar pulses are stacked next to each other in a time-dependent manner and, similar to seismic sections, represent pseudo-cross sections of the ice sheet. The second is the A-scope format, which records ice-sheet reflections from a single pulse of radio-waves as a graph of two-wave travel time (or distance) against reflected power. These radar data can be used to measure ice thickness, subglacial topography and internal ice-sheet layering. Internal layers are caused by changes to the dielectric properties of ice due to three distinct processes. The first is by changes in ice density at depths less than 0.7–0.9 km. The second is by acidic layers of ice, formed when an aerosol product from a large volcanic event is deposited on the ice surface [7]. The third is when changes in crystal orientation fabrics develop between adjacent layers. They are formed due to small differences in the initial strain rate, and enhanced englacial stresses [8]. Ice layers with different ages will contain different impurities and, so, may possess different mechanical properties which may result in

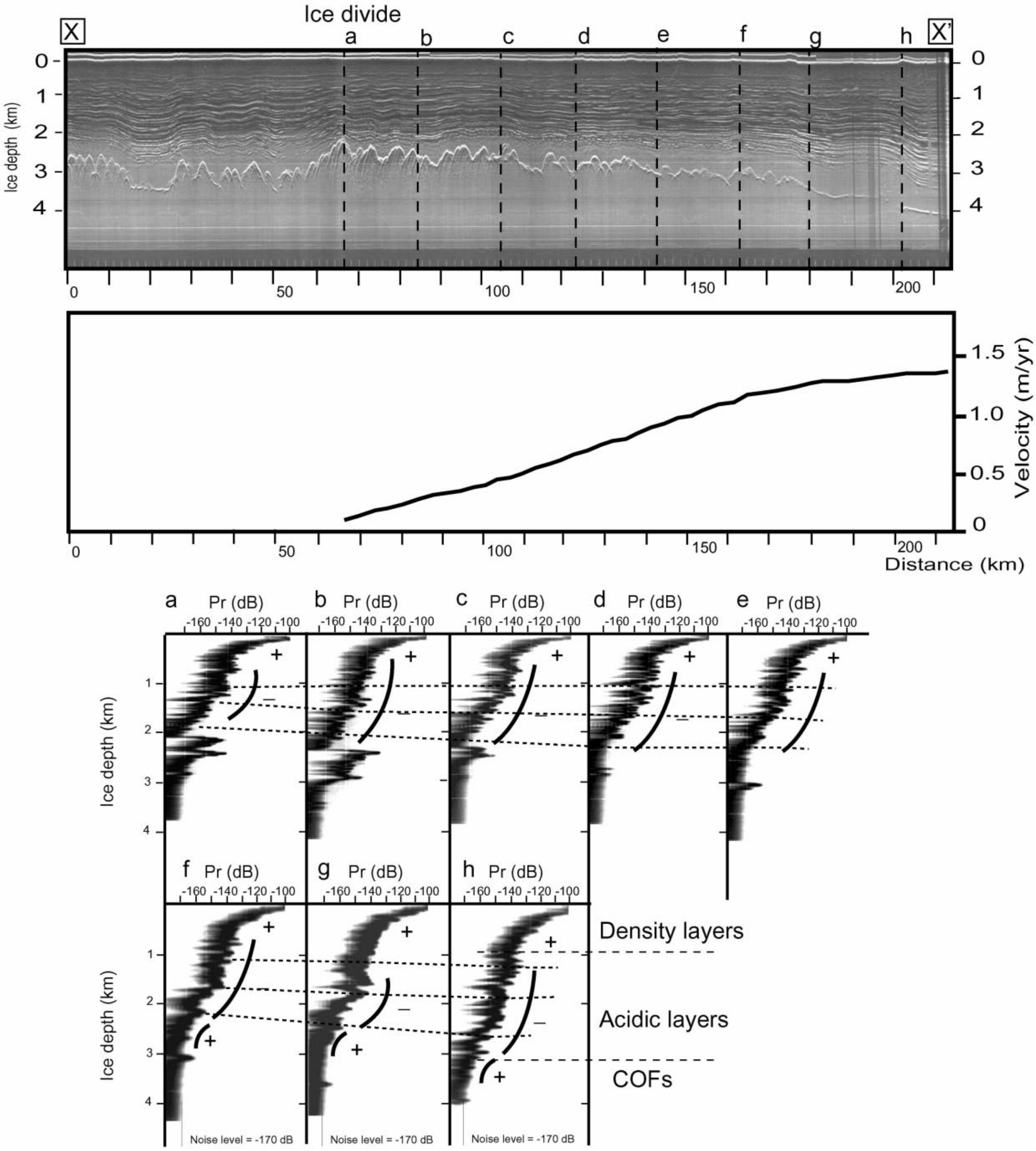

Fig. 2. Radar line XX′. (Upper panel) Z-scope record. (Middle panel) Ice surface velocity along XX′ from the InSAR data (Fig. 1). (Lower panel) A-scopes at eight sites along the transect. Change in A-scope power in the last 2 A-scopes.

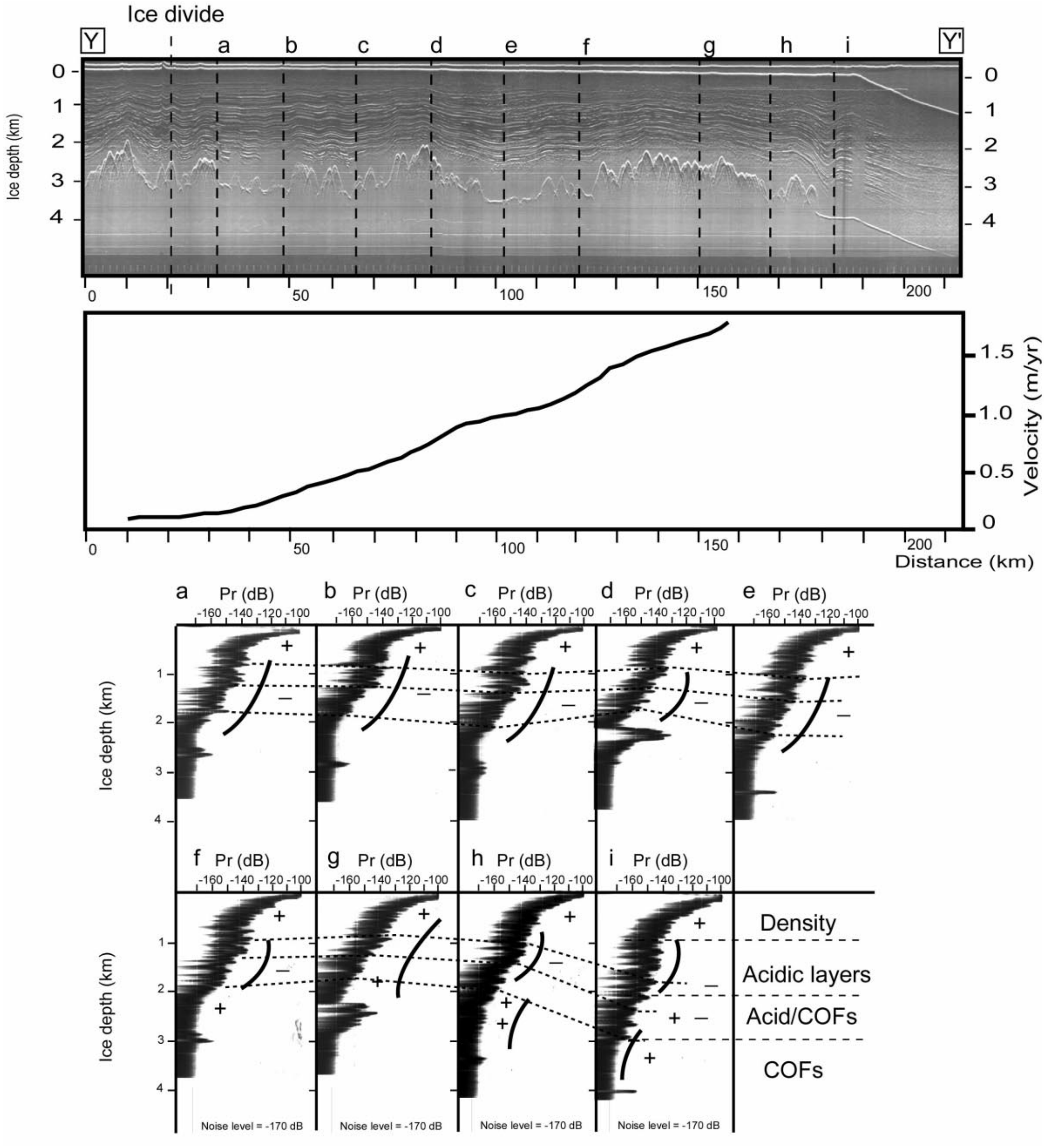

Fig. 3. Radar line YY′. (Upper panel) Z-scope between 720 and 830 CBD (1 CBD unit = 15 s) (line 2). (Middle panel) Ice surface velocity along YY′ from the InSAR data (Fig. 1) (Lower panel) A-scopes at eight sites along the transect.

the non-steady distribution of stress between these layers. Internal radar layers are assumed to be isochronous and, deep beneath the ice surface, parallel to the direction of ice flow [7,8].

Inspection of the InSAR ice velocity field reveals that three radar flightlines are aligned approximately along the line of ice flow between Ridge B and Lake Vostok (Fig. 1). Two of these

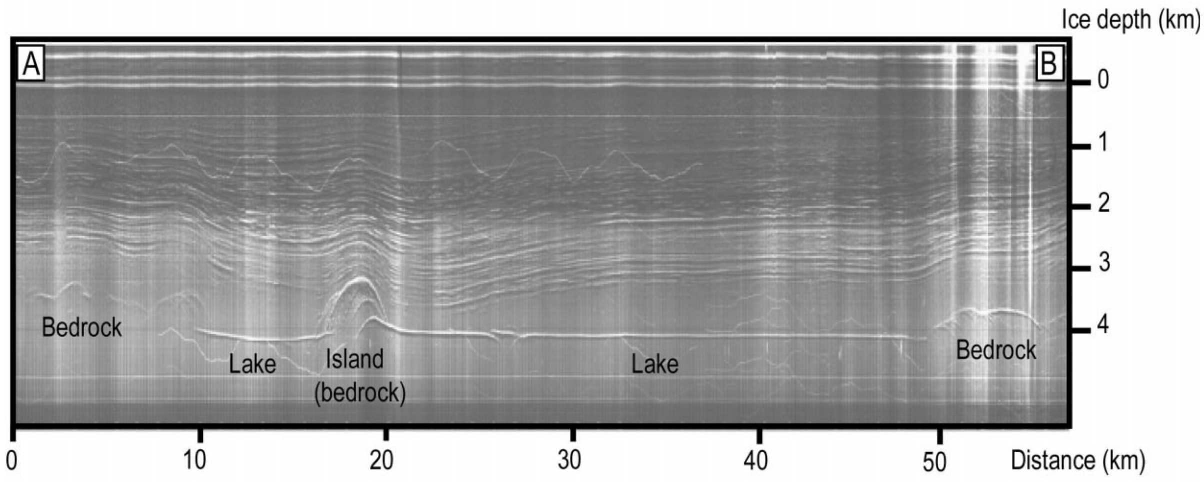

Fig. 4. 60 MHz radar across Lake Vostok [3]. The locations of this transect is provided in Fig. 1.

lines, XX′ and YY′ have radar data in both Z-scope and A-scope mode. In the Z-scope transects of these two lines ice flows from left to right (Figs. 2 and 3). There is a marked difference between the subglacial topographies, and the ice surface velocity, along these two transects. In XX′ (Fig. 2), the crest of Ridge B lies above a subglacial peak 2.2 km beneath the ice surface. Here isochrons are at their highest elevations in the transect. Between the ice divide and the subglacial lake, the subglacial topography is characterised by a uniform gradient of 0.01 ($\sim 0.6°$), with smaller-scale (< 10 km wide, < 200 m amplitude) undulations superimposed. Internal layers generally ride over these minor bedrock perturbations, and dip towards the lake at an angle which is similar to the general bedrock slope in the deeper layers, and similar to the ice surface slope in the upper layers. Thus, there is a vertical divergence of internal layers as ice flows from Ridge B to Lake Vostok across XX′. This divergence is due partly to an increase in the rate of ice accumulation between Ridge B and Lake Vostok, and the change in ice-sheet dynamics from the longitudinal spreading of ice at the divide, to the base parallel shearing that operates a few tens of kilometres from the divide. Ice surface velocity along XX′ increases in a very steady manner from 0 m yr^{-1} at Ridge B to 1.7 m yr^{-1} at the western edge of Lake Vostok (Fig. 2).

In contrast, the subglacial bedrock morphology along YY′ (Fig. 3) is characterised by two large subglacial hills with elevations of around 1 km. Bedrock perturbations, larger than those across XX′, are superimposed over these hills, resulting in a morphology that is more diverse than the relatively smooth profile in XX′. The ice divide is situated above a small subglacial trough (Fig. 3). Ice flows over the stoss face of a number of subglacial hills across YY′. Internal layers follow the general shape of the bedrock profile, converging in vertical section over subglacial peaks and diverging across subglacial valleys. The surface elevation of the ice sheet is also influenced by the subglacial topography. Downstream of the large hill in YY′ (Fig. 3), the surface slope of the ice sheet is noticeably steeper than over the valley (Fig. 1). The ice surface velocity along transect YY′ shows a slight yet noticeable variation in the otherwise steady increase in ice velocity in response to ice flow over subglacial mountains and valleys (Fig. 3). Generally, across the valleys as englacial divergence occurs and the ice surface is flatter, the rate of increase in ice surface velocity reduces (e.g. site ‘e’, Fig. 3). However, as englacial convergence occurs across the stoss face of hills, and the surface slope gets steeper, the velocity of ice increases (e.g. downstream of site ‘f’, Fig. 3).

Since internal layering deep beneath the ice-sheet surface can be regarded as being similar to particle flow paths, there is an obvious difference in the topographic influence on ice flow in our two transects.

In both radar lines (XX′ and YY′), internal layers flow over the smaller-scale bedrock perturbations. This leaves the subglacial valleys free of internal layering. Analysis of the continuous layering above these echo-free zones (EFZs) demonstrates that ice above 'rides' across them. EFZs are likely regions where ice flow is more complex than the simple base-parallel shearing of ice above, which may lead to the break-up of internal layers [8]. In this situation, ice within EFZs is effectively stagnant, because it is confined within subglacial topography. Actual ice dynamics within EFZs are not known, and require numerical modelling (and ice coring) to work out fully. However, it is likely that the processes will involve the re-crystallisation and re-circulation of ice.

Unfortunately, in the radar data used here we are unable to ascertain whether the region where no internal layers are observed is due to the EFZ or because the reflections are weaker than the −170 dB detection limit. However, the base of the internal layering is also observed across Lake Vostok along several radar lines. Importantly, one of these lines shows significant variation in the thickness between the lowest internal layer (continuously traceable across the lake) and the ice-sheet base (Fig. 4). This indicates that (1) internal layers can be detected to ice depths of 4 km, and (2) the zone below the detection level of the radar equipment (BDLZ) is not simply a function of ice depth, but that it is related to the flow of ice. These suggest the extent of the BDLZ in our radar cross sections is similar to the true EFZ.

Recent numerical modelling of ice flow over sinusoidal topography indicates that 'stagnant' ice (where radar layers would be absent) is likely to occur when the bedrock amplitude/wavelength ratio is greater than 0.28 [9]. The value of this ratio for much of the topography across our transects is well above 0.28.

Upstream of Lake Vostok, the locations where radar echoes are absent appear (1) primarily within topographic hollows and (2) often at ice depths shallower than detectable layers from other nearby locations. This provides additional evidence to suggest that the BDLZ areas in our transects correspond closely with the actual EFZs. We therefore suggest that, within the relatively small topographic hollows around Ridge B, there is an abundance of ice that is effectively stagnant beneath the 'mobile' ice sheet. These regions act to 'smooth' the effective mobile base of the ice sheet from the complex bedrock beneath.

3. Radar A-scopes and the development of crystal orientation fabrics within internal layers

Recent crystallographic examination of the Vostok ice core [5,10], located at the downstream end of ZZ′ (Fig. 1), reveals two main types of crystal fabric: a 'griddle-type' fabric associated with coarse-grained ice, and fine grained ice with a vertical orientated *c*-axis [10]. This latter ice is formed under simple shear conditions, whilst the former requires uniaxial tension. The difference between these two ice textures increases with ice depth such that below 2700 m simple shear induced crystal alignments are very clear. These are the only data concerning ice crystal fabrics in our study area.

Under regions of enhanced stress (e.g. as ice flows across the stoss face of subglacial hills) the power reflected off internal layers alters due to the development of layers with a preferred crystal orientation fabric (COFs) [8]. Japanese glaciologists have been able to demonstrate this phenomenon by utilising a two-frequency radar system, because acidic layers of ice yield radar reflection amplitudes independent of frequency, whilst reflections from the boundary between COFs and normal glacier ice are frequency dependent [8]. This is because contrasts in COFs, between layers of ice with a preferred crystal orientation and anisotropic glacier ice, cause a change in the electrical permittivity, whilst acidic layers affect the ice conductivity. Potential layers of preferred COFs can be observed in single 60 MHz radar data because the depth-related shape of the A-scope is similar to an exponential-type decay, whilst acid layers of ice cause a noticeable 'S' shape to the A-scope. In other words, the sign of the second derivative of

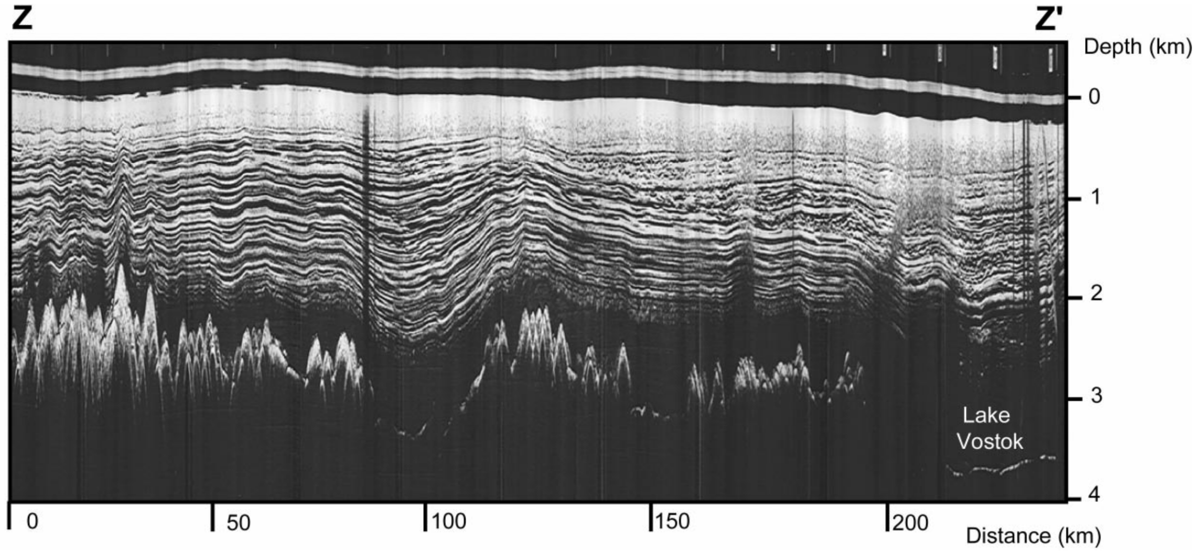

Fig. 5. Radar line ZZ′. Z-scope record along the line of ice flow between Ridge B and the Vostok ice core site.

the signal envelope with ice depth, in a decibel scale, indicates whether internal layers are from conductivity changes (i.e. acidic ice yielding a negative second derivative) or permeability changes (i.e. density variations and crystal orientation fabrics yielding a positive second derivative).

Changes in COFs, from anisotropic glacier ice to ice with a preferred crystal orientation, are likely to form in regions of enhanced stress, usually across the stoss face of subglacial obstacles [8]. Because internal layers can be traced continuously across such hills, these boundaries of COFs must form within existing 'acidic' layers of ice [8]. It is expected that the rate of COF formation will increase with stress. Therefore, changes in COFs are more likely to form in the deepest layers of the mobile ice sheet (above the EFZ) than at relatively shallower depths.

The most obvious place to examine this idea within our radar data, is between sites 'f' and 'g' in YY′, where ice flows over a 1 km high subglacial hill (Fig. 3) because stresses in the ice are likely to be enhanced. The radar A-scopes at and upstream of site 'f' demonstrate the familiar 'S' shape. That is, at ice depths less than 1000 m, a positive second derivative indicates permittivity reflections due to ice density changes. At ice depths greater than 1000 m, the negative sign of the second derivative indicates internal layers are caused by acidic layers of ice. However, at site 'g', the positive second derivative sign is altered, suggesting that ice permittivity reflections dominate the ice column. Downstream from this site (site 'h'), the same internal layers beneath 1 km ice depth at 'g' are located much deeper beneath the ice surface. However, the exponential decay pattern, indicative of reflections from permittivity variations, is retained in ice depths greater than about 1.8 km at site 'h'. We assume that permittivity variations at ice depths greater than 1000 m can only be caused by contrasts in COFs [8]. We therefore conclude that ice with a preferred COF develops below ice depths of 1 km between site 'f' and 'g', and are maintained at greater depths downstream from 'g' as ice flows down the subglacial hill. The internal layers where contrasts in COFs could occur at site 'h' are located at ice depths greater than 2.8 km at site 'i', over Lake Vostok (Fig. 3). However, at site 'i' some evidence for COF boundaries can be seen at depths as shallow as 2 km (Fig. 3).

In contrast, XX′ exhibits less change in the pattern of A-scope returns, and these are limited to ice depths greater than 3 km, implying that crystal orientation fabrics are developed only near the ice base (Fig. 2). Therefore, acidic layering dominates the majority of the ice column below 1 km. However, between site 'f' and Lake Vostok, there is evidence for COF boundaries below 3.25 km (Fig. 2).

A further Z-scope transect is located directly along the line of ice flow from Ridge B to Vostok Station (ZZ′, Figs. 1 and 5). These data show a very similar subglacial, englacial and supraglacial morphology to that in YY′ (Fig. 5). In both YY′ and ZZ′, ice flows over a 1 km high subglacial hill about 100 km west of the lake. We contend that, in YY′, ice flow over this hill causes layers of ice with a preferred COF to form within existing internal layers. We expect the same process to occur across the subglacial hill in ZZ′, and for preferred ice crystal orientations to exist in internal layers below about 2800 m (and possibly as shallow as 2000 m). Unfortunately, A-scope data are not available to examine this radar line in more detail. However, our analysis of where ice with preferred COF is likely to form across ZZ′ (Fig. 5) ties in well with ice crystallographic information from the Vostok ice core, mentioned at the beginning of this section.

Ice layers with a preferred COF will have a rheology different to normal glacier ice. The amount of measured strain within mono-crystalline ice is 10 times that of glacier ice under simple shear conditions, and less than 1/10 for other stress configurations [8,11]. This suggests that the zone identified in YY′ where boundaries of COFs are observed, will have rheological consequences downstream of Lake Vostok.

4. Conclusions

Three flow-parallel airborne radar datasets were analysed in conjunction with the InSAR-derived ice-sheet velocity field to indicate ice-sheet behaviour between the Ridge B ice-sheet divide and Lake Vostok. The first, XX′ (Fig. 2), showed a simple relation between the flow of ice and subglacial conditions. Here, the ice divide is located beneath a topographic hill, from which ice flows down a smooth bedrock slope to the subglacial lake 140 km away.

In contrast, YY′ (Fig. 3) shows far more complex ice flow features, associated with the underlying bedrock physiography. Ice flows across two major (>1 km high) subglacial hills, causing convergence of internal layers in vertical section over the crests, and divergence across the valleys. Analysis of radar A-scopes shows that ice with a preferred COF develops in the ice sheet across the stoss face of the largest of these hills. Over Lake Vostok, the crystal fabrics are likely to be present beneath 2.0 km, and dominate below ice depths of 2.8 km.

A third radar transect along the line of ice flow from Ridge B to Vostok Station (ZZ′) shows a very similar relationship between internal layering and topography to YY′. We therefore expect layers with ice crystal alignment to develop across the large subglacial hill (Fig. 5) and to be present in ice below $\sim$2.8 km at the Vostok ice core site.

Independent support for our analysis comes from crystal fabrics (ice with a vertical *c*-axis) measured in the Vostok ice core. Ice which has a shear-stress induced crystal alignment may deform under further shear-stress 10 times as much as normal glacier ice [11]. This means that the ice flow models assuming homogeneous isotropic ice and smoothed bedrock topography may over simplify the nature of ice flow at the centre of Antarctica.

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