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Radar internal layers from the Greenland summit

Robert W. Jacobel

Department of Physics, St. Olaf College, Northfield, MN

Steven M. Hodge

Ice and Climate Project, U.S. Geological Survey, University of Puget Sound, Tacoma, WA

Abstract. Ice penetrating radar measurements made over the summit region of Greenland show returns from internal layers which can be used to augment the interpretation of climate information from the two deep cores recently recovered from this area. These reflecting surfaces, believed to represent isochrones, give information about the stress regime near the summit, and may aid in a better calibration of the age depth scale between the two cores - particularly in the lowest 10% of ice thickness where there is currently disagreement. The approximate depth at which internal echoes become discontinuous corresponds with the observations of steep inclinations and overturned folds on the scale of centimeters in the core samples. However the deepest internal layers which can be distinguished in the profiles place constraints on the scale and location of high angle or overturned folds.

Introduction

In 1987 ice penetrating radar measurements were made over the summit region of Greenland as a part of the site selection process for the deep drilling projects which have since taken place there. Radar data were acquired using the 60 MHz Technical University of Denmark (TUD) radar along 26 flight lines comprising a grid of approximately 180 km on each side, and a map of surface and bed topography was produced from analog records, (Hodge et al., 1990). At the same time, high resolution digital data were recorded in a parallel data stream using a system described by Wright et al., 1989. Because of recent interest in the deep internal layers generated by conflicting interpretations of the lowest portions of the GRIP and GISP II cores, (Taylor et al., 1993; Grootes et al., 1993), we have analyzed portions of our digital records and present here results from 3 flight lines which cross the summit region nearest the core holes.

Figure 1 modified from Hodge et al., (1990), shows these three radar flight lines, our coordinate system and the location of the two cores. The flight line N375 passes 2.5 km north of GISP2 and 1.5 km south of GRIP. Line E125 passes 3.9 km to the east of the GRIP core, and line W125 passes 4.7 km to the east of the GISP2 core. Figures 2, 3, and 4 show the radar data from each of these flight lines. The surface and bed echoes are indicated on the plots as well as an approximate depth scale which assumes an average wave velocity of 167 m/microsecond. Vertical exaggeration is approximately 26:1. Vertical resolution of the data is 40 ns, corresponding to approximately 3.3 m in ice thickness, just slightly greater than the 2.7 m wavelength of the radar in ice. In the horizontal direction, wave forms were written to tape after stacking 512 returns which corresponds to approximately 17 m of surface travel at average air speeds. In figures 2a, 3a, and 4 the data have been further averaged by an additional factor of 8, corresponding to approximately 8 wave forms displayed per kilometer along the surface. Figures 2b and 3b display enhanced portions of these same flight lines in the area near the bed at an expanded horizontal resolution of 16 wave forms per kilometer. Vertical exaggeration is approximately 16:1. Further enhancements to the full horizontal resolution of the data do not significantly improve the image quality.

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Radar Internal Layers

In addition to the prominent surface and bed echoes, returns from many internal layers can be distinguished, and most of these can be traced across the full 180 km of the grid. Of particular interest are those in the deepest portion of the records. The causes of radar echoes arising from internal layers in ice have been discussed by Moore, (1988), and more recently by Fujita and Mae, (in press) who conclude that for frequencies in the range of several tens of MHz, changes in conductivity are responsible for the dielectric contrast which produces reflections. Acidic fallout from volcanic eruptions is the most likely source of these episodic surface deposits, and this hypothesis accounts for a

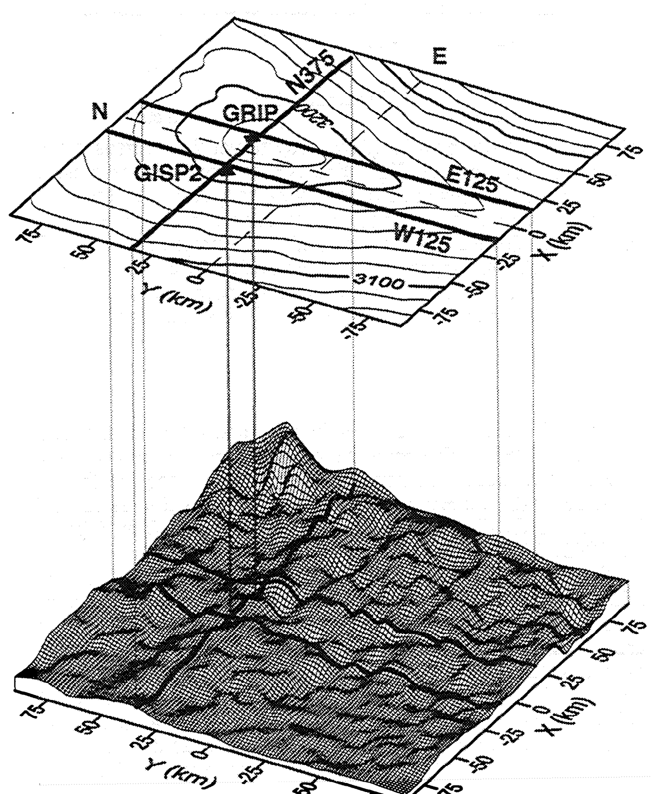

Figure 1. Map of the Greenland summit region showing surface contours and bed topography, the GRIP and GISP2 core locations, and the flight lines used in this study. Flight lines are designated with a letter indicating ordinal direction and a nominal grid coordinate giving the location of the line with respect to the grid origin. Thus, for example, line N375 is a west-to-east line, located 37.5 km north of the grid center, passing near both core sites.

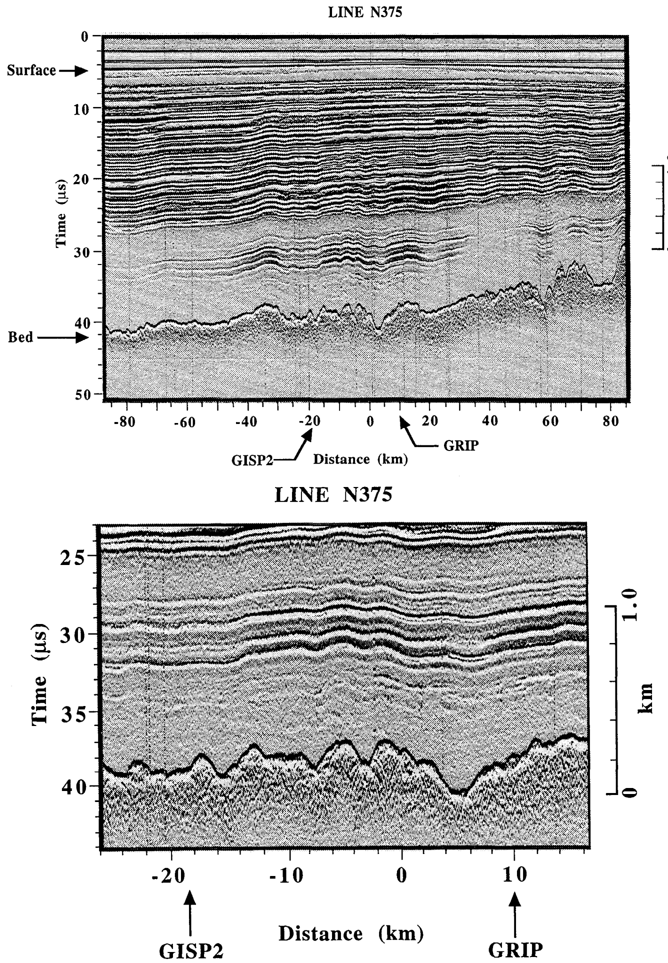

Figure 2a. Radar profile of two-way travel time versus distance along the west to east flight line N375 which passes closest to both coreholes. Distance coordinates correspond to the grid of Figure 1. The east-west position of each core is projected onto the profile. GISP2 lies 2.5 km south of its projection, and GRIP, 1.5 km north. An approximate depth scale as well as the position of the surface and bed echoes are also indicated on the figure. **(2b).** Enhancement of the bed region of the same flight line.

In a number of instances, bright internal echoes within the upper 1000 m of the record appear to die out and later reemerge a few kilometers away. Some of this may result from discontinuous deposition, though neither core shows evidence for this. More likely it is due to small changes (~ 10%) in layer separation which determine the condition of constructive interference. These could result, for example, from local differences in accumulation rate. The radar also has a finite footprint on any surface within the ice, and therefore changes in reflected power arise from contributions everywhere within the beam. Thus signal loss results from a convolution of the antenna beam pattern and these variations in layer separation occurring over the entire area illuminated by the beam.

The weak signal strength from the lower layers causes the number of them per unit of ice thickness to vary at different depths. Nonetheless, the continuity of individual reflecting horizons along the flight lines is quite good until the last 300 to 400 m above the bed. Below that depth the layers become increasingly broken and ill-defined, though traces of a layer just 160 m above the bed can be found nearly all the way across flight line E125. There are several reasons that the layers appear to become discontinuous as the bed is approached. One is simply that the overall signal strength from weaker internal layers is low at those depths, and the power returned is only marginally above the noise level. This can be inferred from the fact that small changes in the energy reflected from the surface appear to mute the returns from entire sections of the lower layers, as can be seen in portions of each of the figures (for example, km 40-60 in line N375). Fortunately, this does not appear to be the case in the number of the general features in this (and other) radar records (Hammer, 1980). The prominent break in the pattern of internal layers which occurs at approximately 55% of the ice thickness coincides with the Holocene - Wisconsin transition as identified in both the GRIP and GISP II cores at about 1600 m, (Taylor et al., 1992; Alley et al., 1993; Johnsen et al., 1992).

The apparent spacing of the internal layers between 10 and 20 microseconds (the portion of the record where they are most dense) corresponds to about 41 m. This is more than five times greater than the theoretical minimum layer separation resolvable by the system, and is also greater than typical separations of volcanic layers (Taylor et al., 1993). The reason for this difference is that detectable internal echoes in a radar system arise from constructive interference between acid layers with a favorable separation (Moore, 1988). Thus, every horizon with a strong dielectric contrast does not produce an echo, and in general there will be no simple correspondence between the radar internal echoes and electrical conductivity measurements, for example. On the other hand, the mechanism for producing echoes from internal layers is consistent from one profile to the next for a given radar system, and in most surveys the layers can be correlated from one line to the next, (Jacobel et al., 1994). For example, the three prominent dark-light bands at approximately 75% of the ice thickness in the Wisconsin ice can be identified in all three of the figures.

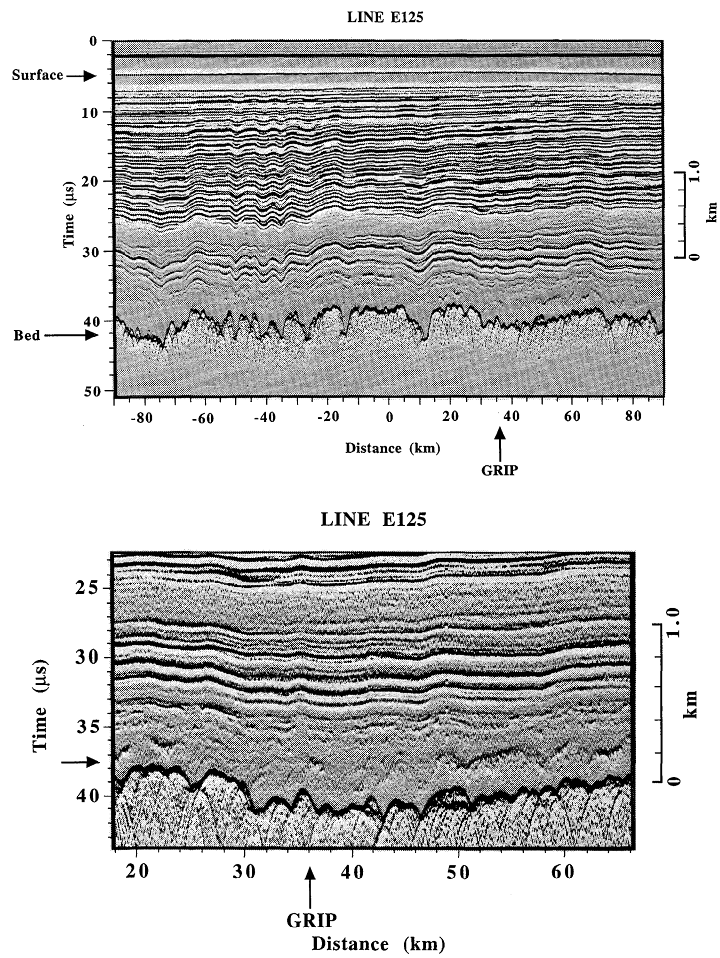

Figure 3a. Radar profile of two-way travel time versus distance along south to north flight line E125. The GRIP core location is 3.9 km west of this flight line at the position projected onto the figure. **(3b).** Enhancement of the bed region of the same flight line. Arrow indicates the discontinuous but persistently visible layer 160 m above the bed.

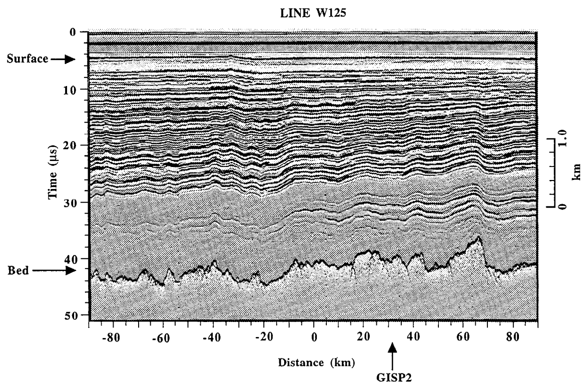

Figure 4. Radar profile along the south to north flight line W125 which passes 4.7 km east of the GISP2 core site at the location projected onto the figure.

vicinity of the two core sites where the deep internal echoes are as good or better than anywhere else in these flight lines.

However, from the above argument it is also clear that an additional and important reason for echoes to appear discontinuous must be that the internal layer separation is no longer spatially constant, and the condition for constructive interference is fulfilled only intermittently. This would naturally be the case as layers nearer the bottom conform more closely to the bedrock topography, and steeper inclines in both directions perpendicular to the beam make it increasingly difficult to produce constructive interference from two or more surfaces as the component of their separation in the direction of the radar beam changes rapidly. Evidence that this is occurring can be seen in the bed regions of figures 2-4 where the steepest bed slopes approach 30%, and the deepest internal layers become increasingly discontinuous. Where visible, they also appear to vary in their separation, especially over regions of greatest topographical relief (for example, near km 20 and km 60 in line E125, and km 60-65 in line W125), giving support to the arguments made above. Apparent layer separations do not appear to vary in the upper 90% of the ice thickness, but as the layers become more deformed by shear stresses toward the bed, they depart from conformity with each other as discussed by Robin and Millar, (1982), and thus become more difficult to detect.

Layer Deformation

Taylor et al., (1993) report observations of deformation on a scale of a few centimeters with inclinations above 20°, and overturned layers beginning at about 600 m above the bed at GISP2, and 200 m above the bed at GRIP (possibly higher). Together with possible boudin-type structures, these deformational patterns are hypothesized to account for the differences in physical and chemical properties between the two cores in their deepest 10%. An important question then is, do these features exist at larger spatial scales as well, where they would go undetected in a typical core section and thus invalidate layer chronology? Overturned layer structures at centimeter scales could have the same effect as changes in layer separation in destroying constructive interference and thereby giving rise to discontinuities in the radar echoes, provided the folding occurs over regions comparable to a wavelength or more.

The discontinuities we observe are thus consistent with centimeter scale overturned folding or high angle folding - both of which persist on extended spatial scales, or with boudinage. Unfortunately the radar can not resolve structures horizontally on scales less than about 20 m, and so we can not extend these observations of overturned and steeply dipping layers seen in core samples from centimeter to meter wavelengths. All we can say is that the discontinuities we do observe are the result of processes which cause the layer separation and or slope to vary in such a way that constructive interference is lost.

However it does appear that the radar records can place some constraints on overturned folding near the bottom. The general pattern in all profiles is that the layers simply conform more and more closely to the bed topography as the bottom is approached. Nowhere do we see folding which grows steeper than the bed slopes below. When layers do become indistinct, it is typically in the areas of greatest bed curvature (particularly troughs), and then they generally reemerge at the same height above the bed when the slope lessens. These observations constrain the existence of high angle or overturned folds on scales larger than the horizontal resolution (approximately 20 m) to those places where layers can *not* be distinguished. The question then becomes, what scales of folding might occur where there are no clear echoes.

As an example, the lowest easily-recognized layer in line E125 occurs approximately 160 m above the bed, and can be identified intermittently across the entire profile. It is depicted most clearly in the bed detail section of figure 3b where the longest gaps are typically 2 to 3 km long. Assuming that the absence of clear echoes results from high angle or overturned folding, the location and length of these gaps constrains the region of folding, and places an upper limit on the maximum fold horizontal scale. Such hypothetical high angle or overturned folds could then possibly occur at this depth on scales from centimeters up to a few kilometers at most.

The depiction of any layers at all in the lower 300 m of the ice column, however discontinuous, provides useful information about the stress regime, and indicates that ice flow in the summit region has not been highly irregular. These layers, together with model studies, may also enable a better calibration of the age-depth relation between the two core sites, assuming that they are indeed depositional isochrones, (Whillans, 1976). Presently there is disagreement about this age-depth scale below about 2750 m, (Taylor et al., 1993).

Ground-based radar studies have also been carried out in the vicinity of the two core sites by Hempel and Thyssen, (1992). The pattern of echoes from internal layers along their profile running directly between the two core holes is very similar to our results from the corresponding section of line N375 shown in figure 2b. The spatial pattern of bright and faint echoes (including gaps) is remarkably similar in the two data sets in both the horizontal and vertical dimensions. However, one feature discussed by them which is not evident in our profile, is a slight upwarping in the internal layers over the topographic basin 4.5 km west of the GRIP core.

In summary, the fact that observations of steep inclinations and overturned folds in core samples coincide with the approximate depth at which internal echoes in radar records become discontinuous is unlikely to be mere coincidence. The mechanism responsible, though not entirely evident from either data set, is probably the same or at least related. From these data, it would appear that radar records of internal layer deformation, at least in the present circumstance, can be a useful tool for augmenting records from deep ice in reconstructing climate history.

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R. Jacobel, Department of Physics, St. Olaf College, Northfield, MN 55057. (e-mail: jacobel@stolaf.edu)

S. Hodge, U.S. Geological Survey, University of Puget Sound, Tacoma, WA 98416. (e-mail: smhodge@orca_paktcm.wr.usgs.gov)

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