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Stable isotopes in the basal silty ice preserved in the Greenland Ice Sheet at Summit; environmental implications

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Abstract. Modelling ice sheet behaviour in the context of climatic changes depends on initial and boundary conditions which can be better defined by studying the composition of basal ice. This study deals with basal ice reached by deep drilling at Summit in Central Greenland (GRIP core). The isotopic composition of this ice indicates that ice formed at the ground surface in the absence of the ice sheet largely contributed to its formation. The basal silty ice is a remnant of a growing stage of the ice sheet, possibly the original build up.

The basal ice of the GRIP core

The GRIP ice core was drilled into silty ice close to bedrock at Summit (72° 34'N ; 37° 37'W) on the main ice divide of the Greenland Ice Sheet (G.I.S.) in summer 1992 after having penetrated 3022.54 m of ice. About 6 m of basal ice were recovered. The temperature at the bottom turned out to be -9°C, that is well below the pressure melting point. A striking discontinuity exists at the top surface of the silty ice. The clean glacier ice above it shows evidence of disturbed layer structures (Dansgaard et al., 1993). The GRIP core does not reach bedrock. In the nearby GISP 2 core, about 13 m of similar silty ice without any coarse material at the ice-rock contact were recovered before bedrock penetration occurred (A. Gow, personal communication). The remaining thickness of silty ice in the GRIP core cannot however be estimated from this data since fluctuations in thickness of basal ice are known to occur over short distances.

The silty ice preserved in the ice core can be subdivided into three parts on the basis of its characteristics in thin sec-

tion and of its stable isotopic composition. An upper part, about 1 m thick, displays interbedding of silty ice layers and clear ice layers. The clear ice layers have properties close to those of glacier ice just above the discontinuity: very few visible particles, large crystals (1.2 to 10.5 cm²) and similar δ -values more negative than the adjoining silty layers. The middle part, about 3 m thick, shows more homogeneous ice with only diffuse banding. The dispersed silty and sandy particles are evenly distributed at the boundary of small ice crystals (0.09 to 0.4 cm²). An occasional pebble up to about 1 cm in length is also present. This middle part exhibits δ -variations similar to the one between the Last Glacial Maximum and Holocene values. The lower part, about 2 m thick, is similar to the middle part but the δ -values are less variable and close to the maximum observed. The dirt content of the silty ice is in fact relatively low: from 0.01 to 0.30 in weight percent. These values are in the lower range of those observed in the Camp Century and in the Dye 3 cores.

Stable isotope composition

The stable isotope composition of the silty ice, studied by determining both the δD and the $\delta^{18}O$ by 3.45 cm long continuous samples, shows striking features (Fig. 1). First, the maximum $\delta^{18}O$ value in the lower part of this ice reaches -25‰, a value which is never observed anywhere else in the whole GRIP core. Holocene and Eemian $\delta^{18}O$ values are respectively -35‰ and -32‰ (Dansgaard et al., 1993). Johnsen et al. (1992) have shown that the $\delta^{18}O$ of the ice is closely related to the local mean annual surface temperature T (in °C) by the relationship:

$$\delta^{18}O = 0.67T - 13.7\text{‰}$$

This relation holds for Greenland for temperatures <-14°C. If this equation is applied to typical $\delta^{18}O$ values of the GRIP core (Table 1), then the value of the deduced surface temperature for the basal silty ice is much too high for an extensive ice sheet to exist. In fact, one can consider the limiting case where this part of the silty ice originated on the ground surface in the absence of an ice sheet. The ground surface would

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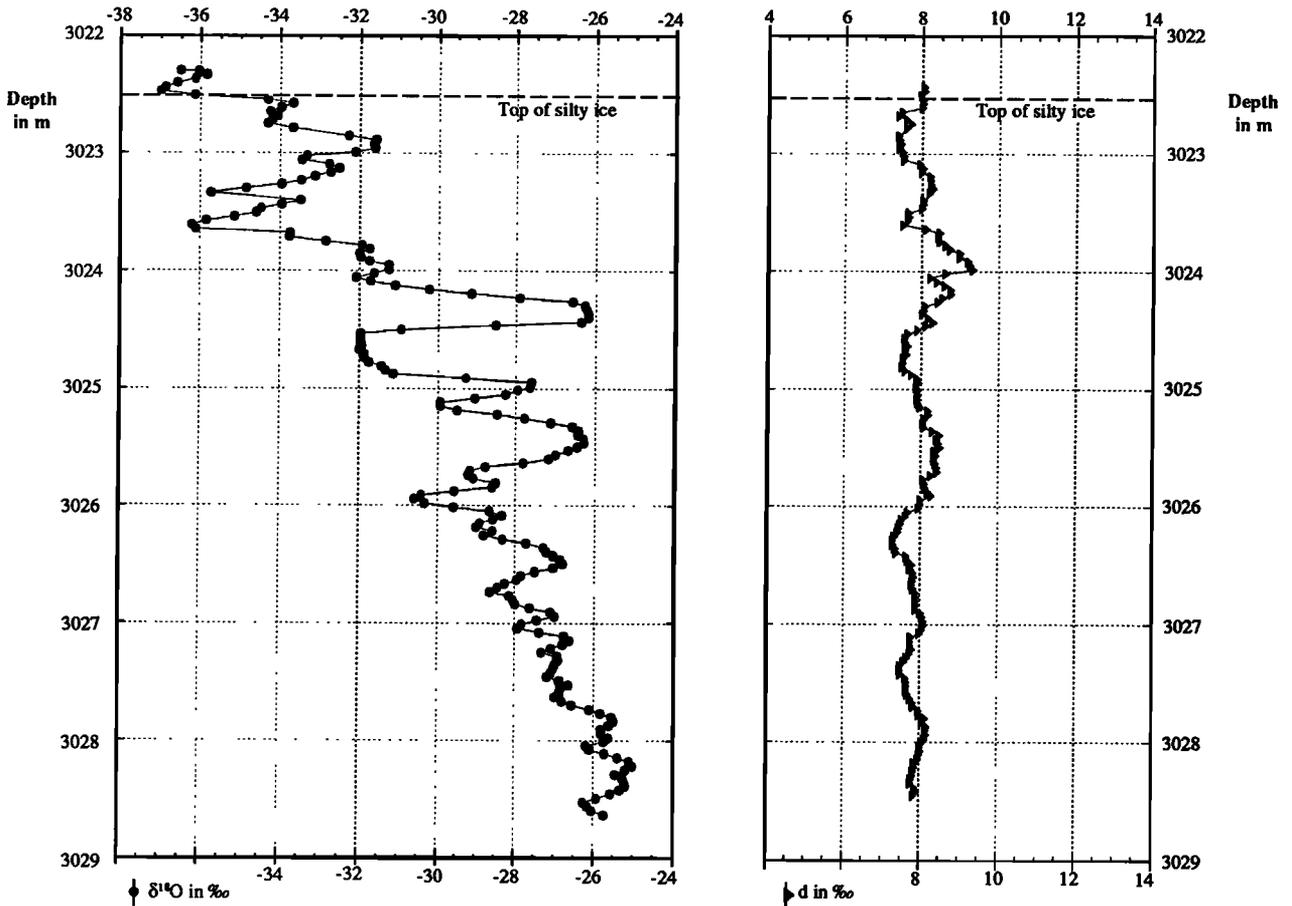


Figure 1. The $\delta^{18}\text{O}$ and deuterium excess profiles in the silty ice of the G.R.I.P. core. The deuterium excess profile is smoothed by a 10 points running mean in order to remove single point peaks. The δD 's were measured at the Centre d'Etudes Nucléaires de Saclay in France with an accuracy of 0.5‰, the $\delta^{18}\text{O}$'s at the Science Institute of the University of Reykjavik in Iceland with an accuracy of 0.05‰.

in that case be at a higher elevation because of the absence of an ice load. If a 3000 m thick ice sheet entirely disappears, the ground surface will be about 1000 m higher in elevation when the isostatic rebound is completed. A difference of 2000 m in elevation is thus involved and, with the current $\delta^{18}\text{O}$ change of 6‰ per 1000 m elevation (Johnsen et al., 1989), it can be computed from Table 1 that the ice would have a $\delta^{18}\text{O}$ value of -23‰, assuming temperature conditions similar to those prevailing during the Holocene. For conditions similar to those prevailing during the formation of the glacier ice just above the silty ice, a $\delta^{18}\text{O}$ value of -25‰ would be obtained. Such calculated values are in close agreement with the δ -values in the lower part of the silty ice.

The above reasoning assumes that the isotope / temperature relationship observed today applies to the silty ice, thus that the isotopic signal is not explained by phase changes occurring at or near the ice-bedrock interface. In our mind this is beyond question for the following reasons :

1. In the GRIP core, the maximum shift between the $\delta^{18}\text{O}$ value of the ice just above the silty ice and the $\delta^{18}\text{O}$ value of the silty ice itself reaches 12‰. This considerable shift, never before observed in basal ice, cannot be due to phase changes since fractionation by freezing leads to an enrichment in heavy oxygen which does not exceed 3‰ in $\delta^{18}\text{O}$. Primarily on the basis of gas analyses, Herron and Langway (1979) interpreted the silty ice in the Camp Century core

(C.C.) as resulting from an adfreezing process. Koerner (1989) viewed this silty ice, together with that of the Dye 3 core (D 3), as superimposed ice formed during the Eemian when the GIS was thought to have had reduced size. No positive shift between glacier ice just above the silty ice and the silty ice itself can be detected in the C.C. core and a maximum shift of 5‰ is observed in the D 3 core. So, the situation in

Table 1. Deduced surface temperature from the $\delta^{18}\text{O}$ of several ice types from the G.R.I.P. core

Ice types	$\delta^{18}\text{O}$ (in ‰)	T (in °C)
Holocene	- 35	- 31.8
Younger Dryas	- 41	- 40.7
Last Glacial Maximum	- 43	- 43.7
Eemian	- 32	- 27.3
Glacier ice just below the Eemian	- 41	- 40.7
Glacier ice just above the silty ice	- 37	- 34.8
Basal ice	- 25	- 16.9
	(maximum value)	

the GRIP core is fundamentally different from that observed in the silty ice from these other Greenland cores and no phase changes could explain its isotopic signal.

2. Other features of the isotopic composition of the silty ice of the GRIP core can be put into evidence if both δD and $\delta^{18}O$ are considered as shown in Figs. 1 and 2. δD and $\delta^{18}O$ are well related by the equation:

$$\delta D = 8.006 \delta^{18}O + 8.145 \quad (r = 0.999 ; 184 \text{ samples})$$

which can be considered as a local Meteoric Water Line (MWL). The mean value of 8‰ for the deuterium excess d ($d = 8 \delta^{18}O - \delta D$) is identical to that observed at continental, high elevation sites for Arctic precipitation (Johnsen et al., 1989). The alignment of the samples on a MWL together with the absence of any trend in the deuterium excess profile are arguments for isotopically unmodified solid precipitation. Indeed, if surface meltwater percolates into the snow forming superimposed ice and inducing some run-off, then the deuterium excess values will display a clear trend (Arnason, 1981). This is not the case here so that, although some melting could have occurred at grain boundaries, no significant meltwater was lost during or after the formation of the silty ice. The superimposed ice hypothesis can thus be ruled out.

3. Other arguments can be developed which indicate that an origin at the base of the ice sheet by phase changes at the interface is quite unlikely. Various mechanisms for phase changes have been proposed including pressure melting and regelation or freezing on (Weertman, 1961 ; Gow et al., 1979; Herron and Langway, 1979). The co-isotopic composition, both in δD and $\delta^{18}O$, of the basal ice formed as a consequence of these interfacial mechanisms and the isotopic changes produced by melting and freezing at the base of a glacier have been extensively studied (Jouzel and Souchez, 1982 ; Souchez and Jouzel, 1984 ; Souchez et al., 1988). Samples of ice due to freezing on at the base are usually aligned, on a δD - $\delta^{18}O$ diagram, on a freezing slope which has

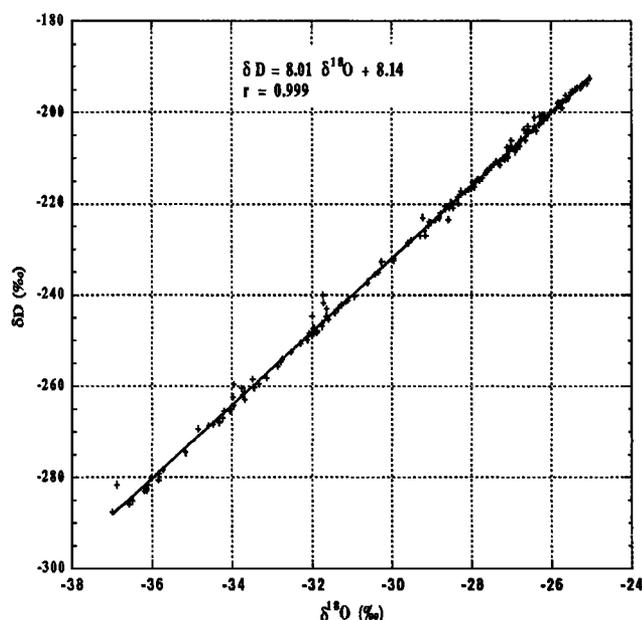


Figure 2. The δD - $\delta^{18}O$ relationship in the basal silty ice from the GRIP core.

a lower value than the MWL (Jouzel and Souchez, 1982 ; Souchez and Jouzel, 1984). If ion-exchange between ice and OH-bearing minerals occurs, the ice samples will also deviate significantly from the MWL (Souchez et al., 1990). Samples of basal ice produced by pressure melting and regelation have, in most cases, identical δ -values as glacier ice samples; only in a few cases have more negative δ -values been observed because of more negative subglacial meltwater. Such an origin for the basal ice cannot be considered in the case of the silty ice studied here because the isotopic shift would be much lower than 12‰ in $\delta^{18}O$.

Both the considerable increase in δ -values between glacier ice above the silty ice and the lower part of the silty ice and the major shifts in δ -values within the silty ice are characteristics which are not present in previous studies of basal ice. Such properties can be explained if local ice formed on the bedrock is mixed and / or interbedded with ice from the ice sheet during a growing phase. In our opinion this means that the isotopic trend in the silty ice (increasing δ -values towards the base) and the sharp variations within the silty ice are the result of deformational processes and do not represent depositional features. The increasing δ -values towards the base is the result of an increasing contribution of the local ice component in the mixing process. Sublimation of this local component of the basal ice prior to overriding is unlikely to have changed its isotopic composition since this phase change occurs without or with a very weak fractionation (Moser and Stichler, 1980). The local ice formed at the ground surface cannot be ice from permafrost soil. Indeed, in order to produce the low dirt content observed in the silty ice, glacier ice would have to be the major constituent involved in the mixing process ($\approx 99\%$) but this would have produced a quite different isotopic signature. Wind-drift ice or / and surface ice formed in a marshy environment are most probable candidates for the local component. Let us note here that mixing of two ice types located on a MWL will give rise to ice located on the same line which can then be considered as a mixing line on a δD - $\delta^{18}O$ diagram.

Gas composition analyses of the silty ice which are developed in another paper (in preparation) give further arguments. Extremely high levels of carbon dioxide and methane reaching respectively 150.000 ppmv and 6.000 ppmv have been detected. Such high values are completely out of range from the usual concentration in glacier ice. They have never been measured from basal ice at other drilling sites. They require that soil processes, possibly occurring in a marshy environment, be taken into account and imply incorporation of ice formed at the ground surface.

Conclusion

The stress conditions under the central part of an ice dome render very unlikely the formation, at its base, of silty ice with debris originating from the ice-bedrock interface. A different ice sheet geometry is required to allow such an occurrence, which is quite possible since the dome could have migrated in the past. Here, we show an extreme case where part of the silty ice originated in the absence of the ice sheet but probably near its edge. The northern margin of the Sydkap ice cap in Ellesmere Island (Arctic Canada) where marsh and snow apron containing dirt and an occasional small pebble coexist at close proximity of an ice cliff with temperatures below the melting point is probably a good analogue.

This implies that the G.I.S. was not present at Summit when the basal ice was formed and that a growing stage of the ice sheet was responsible for incorporation at its base of ice of local origin formed at the ground surface. As indicated by the marine sediment record (Shackleton et al., 1984 ; Funder, 1989), the G.I.S. probably originated about 2.4 million years ago. As suggested by a recent comparison of the isotopic profiles of the GISP 2 and of the GRIP cores (Grootes et al., 1993), the ice sheet was developed in the Summit region during the Eemian and beyond although no precise dating can yet be obtained on this older ice. However, the uncertainty on the age of the oldest glacier ice just above the silty ice does not affect the interpretation of the basal ice from the GRIP core as given above. It can be assumed that the G.I.S. was also developed during the interglacials preceding the Eemian since the climate during these periods was less warm than that of the Eemian (Kellog, 1980). The basal ice thus possibly represents the original build-up of the G.I.S.

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