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Irregular glacial interstadials recorded in a new Greenland ice core

S. J. Johnsen*†, H. B. Clausen*, W. Dansgaard*, K. Fuhrer†, N. Gundestrup*, C. U. Hammer*, P. Iversen*, J. Jouzel§, B. Stauffer‡ & J. P. Steffensen*

* Geophysical Institute, University of Copenhagen, Haraldsgade 6, 2200 Copenhagen N, Denmark
† Science Institute, University of Reykjavik, Dunhaga 3, Reykjavik 107, Iceland
‡ Physikalisches Institut, Universität Bern, Sidlerstrasse 5, Bern 3012, Switzerland
§ Laboratoire de Modélisation du Climat et de l'Environment, CEA/DSM, CE Saclay 91191; and Laboratoire de Glaciologie et Géophysique de l'Environment, BP 96, 38402 St Martin d'Hères Cedex, France

THE Greenland ice sheet offers the most favourable conditions in the Northern Hemisphere for obtaining high-resolution continuous time series of climate-related parameters. Profiles of $^{18}O/^{16}O$ ratio along three previous deep Greenland ice cores[1–3] seemed to reveal irregular but well-defined episodes of relatively mild climate conditions (interstadials) during the mid and late parts of the last glaciation, but there has been some doubt as to whether the shifts in oxygen isotope ratio were genuine representations of changes in climate, rather than artefacts due to disturbed stratification. Here we present results from a new deep ice core drilled at the summit of the Greenland ice sheet, where the depositional environment and the flow pattern of the ice are close to ideal for core recovery and analysis. The results reproduce the previous findings to such a degree that the existence of the interstadial episodes can no longer be in doubt. According to a preliminary timescale based on stratigraphic studies, the interstadials lasted from 500 to 2,000 years, and their irregular occurrence suggests complexity in the behaviour of the North Atlantic ocean circulation.

Three deep Greenland ice cores, drilled from surface to bedrock at Camp Century, Dye 3 and Renland (Fig. 1) in 1966, 1981 and 1987, respectively, reach the ice deposited during the last glaciation[1–3]. This conclusion is based on the occurrence of very low $\delta^{18}O$ values ($\delta^{18}O$ is the relative deviation of the ^{18}O concentration from that in the standard mean ocean water) in the deeper parts of the cores, and on evidence that $\delta^{18}O$ in polar snow and ice depends mainly on the temperature of formation of the precipitation[4,5]. In the mid- and late glacial parts of the cores, $\delta^{18}O$ is alternately very low and intermediate (Fig. 2), corresponding to abrupt shifts between two apparently quasi-stationary climate stages[6–8]. Similar features appear in both marine[9,10] and other terrestrial records from the North Atlantic region[11,12], but only weakly in Antarctic ice cores[1,5], if at all, which points to circulation changes in the North Atlantic Ocean as the driving force[6].

Nevertheless, there has still been doubt about the climatic interpretation of the δ shifts. Intermittent occurrence of isotopically light surface meltwater in the North Atlantic Ocean was suggested as an alternative explanation of the low $\delta^{18}O$ values

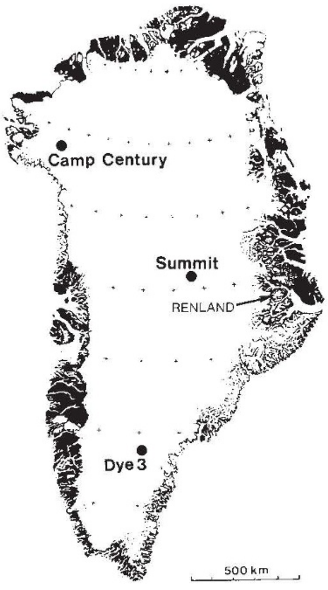

FIG. 1 Greenland with the four drill sites mentioned in the text.

during the last cold period (the Younger Dryas)[13], but deuterium excess data ($d = \delta D - 8\delta^{18}O$) excluded this possibility[8]. Most recently, the meltwater peaks have been shown to coincide with the high-δ rather than the low-δ parts of the Dye 3 record (see Fig. 3*c* and *d* in ref. 14). Another suggested cause of the δ shifts at greater depths was disturbed stratification of the deeper part of the ice, as this ice has travelled a long way over a hilly bedrock[15], or, in the case of Renland, is situated only a few metres above the bedrock[3].

The Summit location on the top of the ice sheet (Fig. 1) is an almost ideal depositional environment in which to recover an ice core: the flow pattern is simple, because there is no horizontal ice movement at present, and little in the past, when the ice divide might have been slightly displaced from its present position. Furthermore, the Summit surface temperature seldom rises above the freezing point, in contrast to Dye 3 where summer melting often causes post-depositional changes in the firn, for example absorption of additional soluble gases from the atmosphere. Summit was therefore chosen as the target of a new deep core drilling under the multinational European Greenland Ice-core Project (GRIP) 1990–1992. By the end of the second field season in 1991, the drill reached a depth of 2,321 m, where the ice is ~40,000 years old.

The timescale shown in Table 1 was established by stratigraphic methods. Back to 8,600 yr BP (before present) the timescale rests on identification of reference horizons in the form of acid volcanic fall-out dated in the Dye 3 core by counting annual $\delta^{18}O$ variations downwards from the surface[16]. Before 8,600 yr BP, the dating was accomplished by a multi-parameter method that identifies the seasonal variations of the concentrations of Ca^{2+}, microparticles, NH_4^+ and nitrate. This is extremely laborious, of course, and the resulting timescale (Table 1) is preliminary. Details will be reported elsewhere.

Figure 2 shows continuous $\delta^{18}O$ profiles along the four Greenland ice cores, plotted on linear depth scales that span the depth intervals listed in Table 2. The Summit depth scale is shown at the outer left along with the stratigraphically derived Summit timescale. The heavy and thin vertical lines indicate δ levels characteristic of low and intermediate δ, respectively.

The close correlation between the Summit and Dye 3 records rules out disturbed stratigraphy as a potential cause for the two δ levels, which must therefore be of climatic significance. Periods of relatively high δ reflect mild interstadials (IS), detectable in all four records (except for IS 3 at Camp Century). The absence of IS 3 at Camp Century, as well as the 'extra' interstadial just below IS 1e, may be ascribed to the less advanced ice-core processing techniques applied to the cores 26 years ago. IS 1 is identical with the Bølling and Allerød mild periods in Europe, but it is not clear how to separate them by an 'Older Dryas' cold spell.

Smoothed versions of the records have minima between IS 2 and IS 5, around 25,000 yr BP. The temperature minimum thus occurred before the often quoted time of maximum glacier extent (18,000 yr BP[16], corresponding to ~21,500 calendar years BP[18]). This suggests a time lag of more than 3,000 yr for the build-up of great ice sheets, in essential agreement with a climate model that simulates the transient response of the cryosphere to astronomical forcing of climate[19].

The right-hand side of Table 2 lists the mean δ values, δ_p, δ_m and δ_c (for present, mild-glacial and cold-glacial stages, respectively), and the differences $\Delta_{mc} = \delta_m - \delta_c$, $\Delta_{pm} = \delta_p - \delta_m$ and $\Delta_{pc} = \delta_p - \delta_c$. As regards the three deep cores, the Δ_{mc} values are similar, but the Camp Century values of both Δ_{pm} and Δ_{pc} are significantly higher than those found at Summit and Dye 3. This may be explained by the changing altitude of the deposition area at Camp Century during the glacial to post-glacial transition[20], and the termination of the shadow effect of the shrinking Laurentide ice-sheet. The conditions at the small isolated Renland ice cap are more complicated[3], but all of the interstadials are easily recognized.

The present geographical relationship between $\delta^{18}O$ (in ‰) and temperature (T in °C)[8]

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

seems valid also for substantial temporal changes[4,21], and

TABLE 1 Depth/time for Summit ice core

Depth (m)	Time (yr)			Event
58.85	1,816	AD	V	Tambora
67.60	1,783	AD	V	Laki
110.95	1,601	AD	V	Unknown
139.40	1,477 ± 2	AD	V	Unknown
187.25	1,259 ± 2	AD	V	Unknown
203.80	1,179	AD	V	Katla
256.15	934 ± 3	AD	V	Eldgá
455.35	2,039 ± 5	BP	V	Unknown
736.45	3,636 ± 7	BP	V	Thera
802.75	4,040 ± 10	BP	V	Unknown
1,334.50	8,210 ± 30	BP	C	$\delta^{18}O$ minimum
1,380.50	8,600 ± 50	BP	V	Unknown
1,623.60	11,550 ± 70	BP	C	Transition to Pre-boreal
1,661.55	12,700 ± 100	BP	C	Onset of Younger Dryas
1,753.40	14,450 ± 200	BP	C	Onset of the Bølling interstadial
1,766.00	15,000 ± 250	BP		
1,884.00	20,000 ± 800	BP		
2,006.00	25,000 ± 1,000	BP		
2,113.00	30,000 ± 1,300	BP		
2,225.00	35,000 ± 1,600	BP		
2,321.40	40,000 ± 2,000	BP		End of 1991 core

Depth against time-of-formation relationship for strata in the Summit ice core, of which those marked by V contain acid fall-out from discrete volcanic eruptions back to 8,600 yr BP (before AD 1990). The strata marked by C indicate notable climatic events from 8,210 to 14,450 yr BP detectable in all of the ice cores listed in Table 2. The ± values are estimated error limits. No error has been put on the time of historically well dated volcanic events.

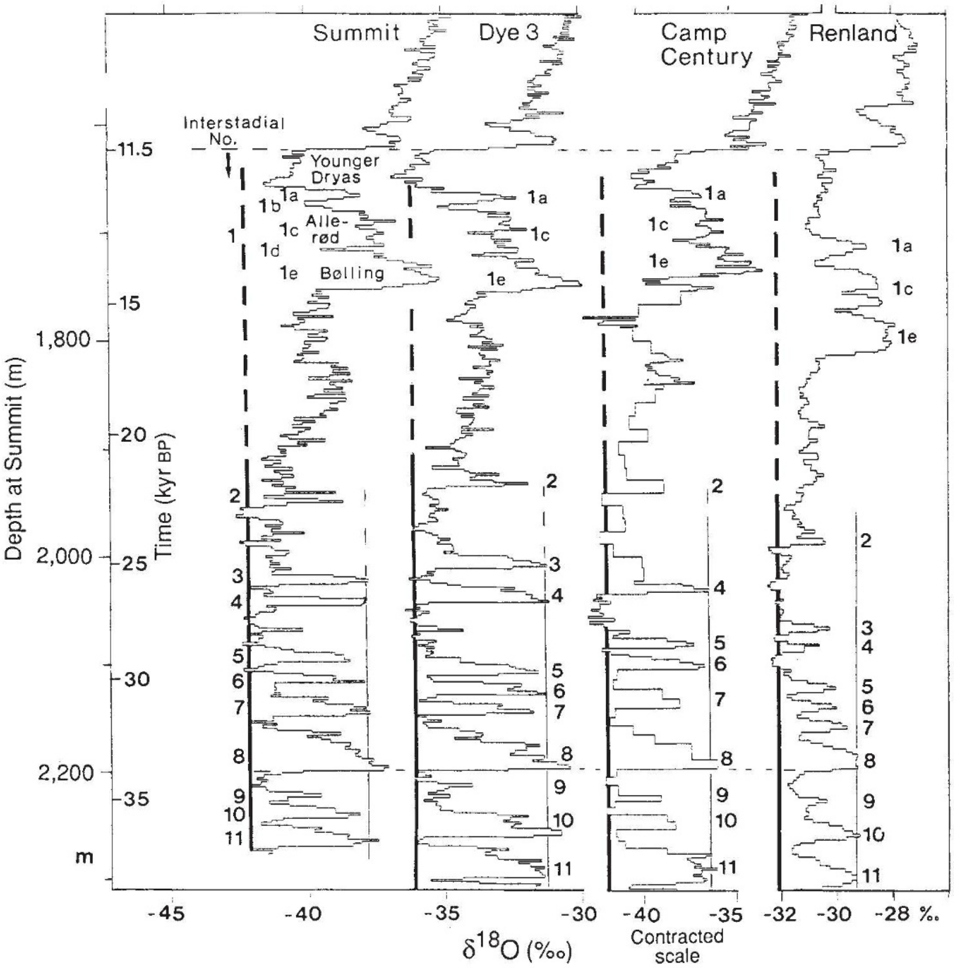

FIG. 2 Continuous $\delta^{18}O$ profiles along sections of four Greenland ice cores from Summit (Central), Dye 3 (Southeast), Camp Century (Northwest) and Renland (East Greenland), spanning nearly the same time interval. The four records are all plotted on linear depth scales, of which only the Summit depth scale is shown to the left along with a Summit timescale. The heavy and thin vertical lines indicate estimated δ levels characteristic of late-glacial cold and mild stages respectively. The figures close to these lines define a suggested numbering of significant mid- and late-glacial interstadials.

TABLE 2 Ice core data

Drill site	Location	Altitude above sea level (m)	Mean air temperature (°C)	Accumulation rate (cm ice per yr)	Ice-sheet thickness (m)	Ice-core length (m)	Increment shown in Fig. 2 (m)	δ_p (‰)	δ_m (‰)	δ_c (‰)	Δ_{mc} (‰)	Δ_{pm} (‰)	Δ_{pc} (‰)
Summit	72° 34′ N 37° 37′ W	3,230	−32	23	3,030	2,321	1,500–2,281	−35.3	−37.7	−41.9	4.2	2.4	6.6
Dye 3	56° 11′ N 43° 50′ W	2,490	−20	56	2,037	2,037	1,762–1,919	−27.9	−31.2	−36.1	4.9	3.3	8.2
Camp Century	77° 11′ N 61° 07′ W	1,890	−24	38	1,390	1,390	1,149–1,248	−29.0	−36.4	−41.9	5.5	7.4	12.9
Renland	71° 18′ N 26° 44′ W	2,350	−18	50	325	325	305.4–313.0	−27.1	−29.3	−32.1	2.8	2.2	5.0

indicates that the cold glacial stages (Fig. 2) were ~7 °C colder than the mild ones, and 12–13 °C colder than at present (when corrected for the higher glacial δ value of the oceans[22]), in agreement with combined ice and heat-flow modelling[21]. Temperature fluctuations of this size in Greenland must have been associated with significant environmental changes in the entire North Atlantic region, particularly in Europe, as seen for example in the botanical evidence of 'swift climatic changes' during the last glaciation[12].

Figure 2 further shows that the duration of the interstadials ranges from about 500 to 2,000 yr. They begin abruptly, perhaps within a few decades[14,23] (the estimated 50-year duration of the Younger Dryas to Pre-boreal transition in ref. 23 has been confirmed by new annual layer counting on the Summit core), but they terminate gradually or in a stepwise fashion. Furthermore, the interstadials apparently occur at irregular intervals. The driving force is probably changing intensity and/or direction of the North Atlantic Current, associated with changing sea ice cover[8,22] and deep water formation[10,14], but the interstadials have not been satisfactorily modelled. Even the mechanism behind the most thoroughly studied interstadial (IS 1) is still debated[24,25].

If the interstadials turn out to reflect a randomness in the general pattern of atmosphere–ocean circulation, comprising two or more possible flow schemes for a given set of primary parameters, climate modellers will have to reconsider the feasibility of predicting natural climate change. The glacial interstadials were indeed linked to changing circulation in an ocean partly covered by sea ice, but it should be borne in mind that within the last millennium, the well documented medieval warmth in Europe was gradually replaced by the 'little ice age', when Iceland was frequently surrounded by sea ice, and the succeeding warming culminated in an abrupt temperature increase in the 1920s, too abrupt to be explained by the increasing greenhouse effect. This oscillation had smaller amplitudes than its glacial predecessors, but a similar sequence of events (gradual cooling followed by abrupt warming). □

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Reconciling particulate organic carbon flux and sediment community oxygen consumption in the deep North Pacific

K. L. Smith Jr, R. J. Baldwin & P. M. Williams

Marine Biology Research Division, 0202,
Scripps Institution of Oceanography, La Jolla, California 92093, USA
Marine Research Division, 0218, Scripps Institution of Oceanography,
La Jolla, California 90293, USA

THE mineralization of organic carbon in the deep ocean has been considered an enigma in the North Pacific. Comparisons of the supply of particulate organic carbon (POC) with estimates of its mineralization to CO_2 by the sediment community have indicated that the supply of POC sinking into the benthic boundary layer may be as much as 97% short of meeting the organic carbon demand of the sediment community[1,2]. These previous findings were based on short-time-series measurements (<14 days) conducted *in situ* with sediment traps and benthic respirometers. We report here a comparison of long-time-series measurements (2.3 years) of POC flux into the benthic boundary layer and concurrent, seasonal measurements of sediment community oxygen consumption. We chose a single abyssal station (4,100 m depth) in a region of the eastern North Pacific where there is a strong seasonal fluctuation in primary production, and where previous studies[1,2] showed the supply of POC to be as much as an order of magnitude lower than the demand by the sediment community. Our measurements, using the same methods, show agreement to within 15% between organic carbon supply and demand. This reconciliation is attributable to the inclusion of previously undetected episodic inputs of POC into the benthic boundary layer. These episodic inputs are critical to sustaining the sediment community at this station, and are probably important in other deep-sea environments where seasonal fluctuations in primary production are prominent in surface waters.

Short-time-series measurements of POC flux into the benthic boundary layer (BBL) and its utilization by the sediment community, as estimated from sediment community oxygen consumption (SCOC), have been made in the North Atlantic and North Pacific oceans (Table 1). POC fluxes at the Atlantic stations are sufficient to meet the demands of SCOC. But similar methods used in the Pacific have shown that there is a shortfall in the supply of POC necessary to meet the demands of the sediment community at numerous stations stretching from the continental margin off California to abyssal stations in the central gyre of the North Pacific. Several hypotheses have been suggested to account for the discrepancy in the North Pacific,