(2,670 m at GRIP), it is not possible to determine the duration of anomalous features thinner than 20 m. In both cores, some of the features near the marine isotope stage 5e/6 boundary[4,8] that are thinner than 5 m may contain ice that is out of chronological sequence. These findings raise concerns about the interpretation of previously reported climate variations during the Eemian period (marine isotope sub-stage 5e, 2,790–2,870 m at GRIP)[4,8]. At present we are unable to say which, if either, core contains a more complete climate record.

Detailed consideration of the records may be able to determine the extent of deformation. For example, the several-century age difference (a few tenths of a metre at these depths) between the ice and the gas it contains[35,36] will result in abruptly changing chemical and gas features being preserved in ice that has different flow properties. Deviations from the anticipated offset between the records of ice chemistry and rapidly responding gases such as methane, may permit identification of features in the records that are introduced by ice flow. Detailed analysis of the crystal fabric, dust layers and chemistry may also be able to identify folded sequences. The upper 90% of the records are unaffected by ice flow and preserve a climate record with annual resolution. By detailed analysis of both cores, we anticipate that it will be possible to unravel many of the complexities that ice flow has introduced in the marine isotope stage 5e portion of the records. Deep Antarctic cores, where the marine isotope stage 5e portion occurs further from the bedrock, will also help resolve these issues. □

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Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores

P. M. Grootes*, M. Stuiver*, J. W. C. White†, S. Johnsen‡§ & J. Jouzel‖

* Department of Geological Sciences and Quaternary Research Center, University of Washington, Seattle, Washington 98195, USA
† INSTAAR, University of Colorado, Boulder, Colorado 80304, USA
‡ The Niels Bohr Institute, Department of Geophysics, University of Copenhagen, Haraldsgade 6, DK-2200 Copenhagen, Denmark
§ Science Institute, Department of Geophysics, University of Iceland, Dunhaga 3, IS-107 Reykjavik, Iceland
‖ Laboratoire de Modelisation du Climat et de l'Environnement, CEA/DSM CE Saclay 91191, Gif-sur-Yvette Cedex, France

RECENT results[1,2] from the Greenland Ice-core Project (GRIP) Summit ice core suggest that the climate in Greenland has been remarkably stable during the Holocene, but was extremely unstable for the time period represented by the rest of the core, spanning the last two glaciations and the intervening Eemian interglacial. Here we present the complete oxygen isotope record for the Greenland Ice Sheet Project 2 (GISP2) core, drilled 28 km west of the GRIP core. We observe large, rapid climate fluctuations throughout the last glacial period, which closely match those reported for the GRIP core. However, in the bottom 10% of the cores, spanning the Eemian interglacial and the previous glaciation, there are significant differences between the two records. It is possible that ice flow may have altered the chronological sequences of the stratigraphy for the bottom part of one or both of the cores. Considerable further work will be necessary to evalu-

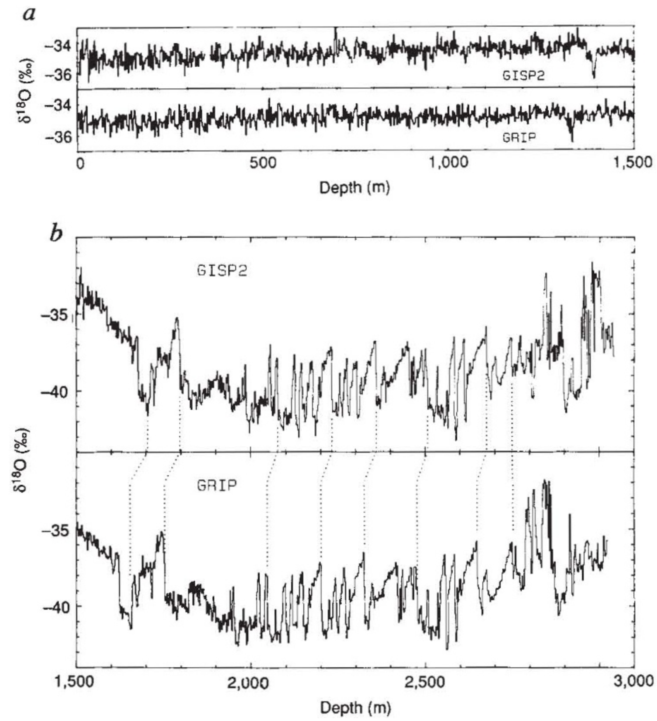

FIG. 1 GISP2 and GRIP $\delta^{18}O$ versus depth of the Greenland ice sheet at its summit area. Plot resolution is 2 m and 2.2 m for GISP2 and GRIP respectively, based on 1-m and 0.55-m data sets. The Holocene records (*a*) show relatively stable $\delta^{18}O$ values (mean value −34.7‰ at GISP2, −34.9‰ at GRIP). Corresponding (correlation coefficient r=0.945) major $\delta^{18}O$ changes of the earliest Holocene and Pleistocene (*b*) are marked by dotted lines down to ~2,750 m. For the record of the penultimate interglacial and beyond, below 2,750 m near bedrock, this correlation breaks down.

ate the likelihood of this, and the extent to which it will still be possible to extract meaningful climate information from the lowest sections of the cores.

On 1 July 1993 the US Greenland Ice-Sheet Project 2 completed a 5-yr ice-core drilling effort when bedrock was reached near the summit of the Greenland ice sheet (72.58° N, 38.48° W, 3,208 m above sea level, mean annual temperature −31 °C). The 3,053.44-m-long, 0.132-m-diameter core was drilled with the new US Polar Ice Coring Office (PICO) electro-mechanical drill[3] using *n*-butylacetate as a drilling fluid. The bottom 13.11 m of the core consists of distinctly banded brown silty ice, with silt and rock inclusions. A 1.55-m rock core was recovered from bedrock underneath the ice. About 40% of the core cross-section was sampled continuously in the field in a multi-investigator sampling program[4]. We report here (Fig. 1) the basic features of the GISP2 oxygen isotope profile, expressed as $\delta^{18}O$, the relative difference between the $^{18}O/^{16}O$ abundance ratios of the ice and standard mean ocean water (V-SMOW) expressed in per mil (‰). The $\delta^{18}O$ value of precipitation generally reflects its temperature of formation[5] with lower $\delta^{18}O$ values corresponding to lower temperatures.

We compare the GISP2 isotope record (Fig. 1*a* and *b*) with that of the GRIP core drilled 28 km to the east at the present ice divide[1]. The distance from the divide (28 km, about nine ice thicknesses) places GISP2 in a flank-flow regime. Surface and bedrock topography between the two drill sites is relatively flat[6]. As major climatically induced $\delta^{18}O$ perturbations have to occur at both sites simultaneously, a simple and direct comparison can be made between the $\delta^{18}O$-depth profiles of the cores to evaluate the climatic significance of the observed isotope fluctuations on a summit-wide scale. This comparison by depth avoids the present uncertainties in the time–depth relationship that are to be reduced by further joint evaluation of the GISP2 and GRIP records[7]. The records are presented at 2-m (GISP2) and 2.2-m (GRIP) resolution in two parts, 0–1,500 m (Holocene) and 1,500 m depth to bedrock (mostly Pleistocene) as was previously done for the GRIP core[1].

The Holocene is a period of relatively stable climate in both cores with mean $\delta^{18}O$ values of −34.7 and −34.9‰ for GISP2 and GRIP respectively. The small Holocene $\delta^{18}O$ fluctuations of 1–2‰ occur too frequently to allow an unambiguous correlation between the cores. As the timescales for the cores are currently still under debate, it is only possible at this stage to compare the changes in $\delta^{18}O$ with depth between the two cores. Doing so yields virtually no correlation between the cores for the Holocene; it is likely that local differences in deposition played a major role in the small-amplitude, high-frequency changes during this period. The major $\delta^{18}O$–temperature fluctuations in the Pleistocene are another matter. Down to 2,700 m the frequent fluctuations are quite similar in both cores, not just in general appearance, but down to the structure of the various warmer episodes. The excellent agreement of $\delta^{18}O$ down to 2,700 m between two cores drilled 28 km apart and analysed independently establishes the regional significance of the detailed isotope fluctuations for this part of the record.

The transitions between cold and warm conditions in the $\delta^{18}O$ record of the summit cores are rapid, of the order of decades (see also refs 8, 9). Corresponding changes in dust content, electrical conductivity measurement (ECM) and ice accumulation are even more abrupt[9,10]. This excludes changes in insolation, global ice volume or oceanic water transport at depth as the cause of the rapid transitions in the $\delta^{18}O$ record, as these have all longer response times. Mayewski *et al.*[11] relate the rapid climate changes at GISP2 to expansion and contraction of the polar atmospheric cell, based on a factor analysis of the covariance of groups of major cations and anions found in the ice. The sequence of stadial–interstadial episodes at GRIP has recently been correlated with new, highly detailed records of sea surface temperatures from North Atlantic sediments[12]. Evidently the Summit record of climate fluctuations during the glacial period down to 2,700 m reflects atmospheric conditions and forcing by the North Atlantic Ocean mixed layer, and thus is of hemispheric and probably global importance.

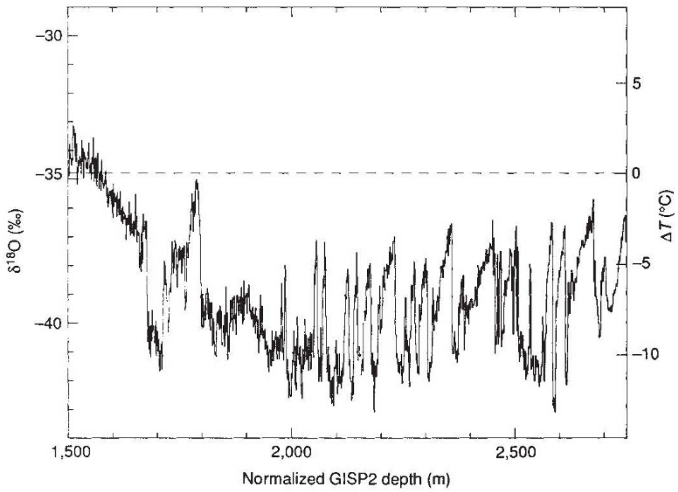

FIG. 2 Unified oxygen isotope climate record for the Summit area of the Greenland ice sheet based on the GISP2 and GRIP $\delta^{18}O$ records. Corresponding sections of the datasets were identified by using rapid $\delta^{18}O$ changes associated with Dansgaard–Oeschger interstadial events as chronostratigraphic markers. We adjusted the depth interval of the GRIP dataset between two markers to that of GISP2, and calculated simple averages over 1-m intervals. The record covers the earliest part of the Holocene, the glacial–interglacial transition and most of the last glacial. The right-hand scale shows deviations from mean Holocene temperatures (ΔT) assuming that the $\delta^{18}O$ deviation from the mean Holocene $\delta^{18}O$ of −34.8‰ is caused by local temperature changes with a $\delta^{18}O$-temperature coefficient of 0.63‰ °C^{-1} (ref. 13).

Although a detailed timescale is still under construction[7], a statistical comparison of the GISP2 and GRIP isotope records can be made for the 1,500–2,750 m depth interval by using the major rapid changes in $\delta^{18}O$ as chronostratigraphic markers to correct for the depth differences between the cores. An average offset between the records of 0.09‰ and a correlation coefficient $r = 0.945$ for the depth-normalized records, indicating 89% common variance, confirm the excellent visual agreement between 1,500 and 2,750 m depth. Averaging the GISP2 and GRIP $\delta^{18}O$ records with 1-m resolution over the chronostratigraphically defined intervals reduces the influence of the remaining variance, which probably reflects local variability, and produces a glacial climate history for the Summit area of the Greenland ice sheet based on both ice cores (Fig. 2).

The Summit records provide a wealth of environmental information in a detail hitherto unavailable for the last glacial. Figure 2 clearly shows an unstable glacial climate. Isotope values during the last glacial switch between periods of high and low $\delta^{18}O$ values lasting several hundred to a few thousand years[1,8,10], except during the last glacial maximum (marine isotope stage 2) from about 1,800 to 2,050 m at GISP2. A steep rise in $\delta^{18}O$ from a low (−41 to −43‰) range to a high of −37‰ is followed, in the longer interstadials, by a gradual decrease to ~−39‰ and then a precipitous drop. If we assume that the isotope values reflect central Greenland temperature and use a $\delta^{18}O$ temperature coefficient of 0.63‰ °C^{-1} (ref. 13), then the mean annual temperatures of the maxima occurring during the glacial for the Summit area were 3.5 °C colder than the Holocene average temperature (mean $\delta^{18}O$, −34.8‰). Mean temperatures dropped by as much as 3 °C during the longer interstadials. The $\delta^{18}O$ change from −37‰ to −39‰ during the longer interglacials corresponds to a temperature decrease of ~3 °C. Full glacial conditions were 10–13 °C colder than during the Holocene. Close agreement between fluctuations in sea surface temperature indicated by *N. pachyderma* (s) in two North Atlantic cores and the GRIP $\delta^{18}O$ record[12] shows that the frequent, large temperature fluctuations at Summit must have been part of a larger atmospheric pattern that extended over the North Atlantic, and probably also over northwest Europe. The large number of interstadials revealed by the ice cores may be the cause of some of the confusion about the number of interstadials and their timing in northwest European climate records[14-17] and offer an opportunity for their interpretation.

The bottom 10% of the isotope records differ significantly. Below interstadial 22 (ref. 1) (2,700 m depth at GISP2, 2,676 m at GRIP, that is 87 kyr BP in the GRIP preliminary timescale[1]) layer thicknesses differ substantially; below interstadial 23 (ref. 1) (~2,750 m in both cores) the correlation between the isotope values deteriorates drastically. Taylor *et al.* report a similar breakdown of correlation between the cores for electrical conductivity measurements[18]. Thus, although the bottom part of both records clearly contains ice from the penultimate interglacial and beyond, the loss of a simple relationship between depth and age prevents a detailed dual-core climate reconstruction deeper than the early Wisconsin glacial. The GISP2 record of the Eemian/Sangamon interglacial period is probably affected by flow deformation and thus cannot confirm the conclusions regarding extremely high climate variability during this period that were based on the GRIP record[1,2].

The excellent agreement observed above a depth of 2,700 m suggests that surface climate conditions at the GISP2 and GRIP localities were similar for both cores for the entire accumulation interval. This agrees with similar peak-to-trough $\delta^{18}O$ values in the bottom part of the cores. The differences between the records must therefore result from flow deformation at one or both of the record localities. Inclined layering, first observed in visual stratigraphy at a depth of 2,678 m in the GISP2 core and noticed at 2,847 m in the GRIP core (see Taylor *et al.*[18]), may indicate such flow deformation. The location of the GRIP core at the ice divide presently protects the deep ice at GRIP from folding. Under different conditions of sea level, ice accumulation rate and temperature during the glacial period, the ice divide may have moved 10–50 km, probably towards the east[19]. This would have placed both GISP2 and GRIP in a flank-flow regime, with the GRIP site downstream of bedrock slopes immediately to the east[6]. The greater depth at which inclined layering was reported for the GRIP core fits the predicted protection of the divide flow and suggests that the GRIP record can be interpreted climatically to a greater depth than the GISP2 record. Further work on the ice cores and the glaciology of the Summit area is needed to provide a detailed timescale for the well established glacial record down to 2,700 m, and to interpret the disturbed bottom part of the records. □

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Endogenous growth of persistently active volcanoes

Peter Francis*†, Clive Oppenheimer† & David Stevenson†

* Planetary Geosciences Division, University of Hawaii, 2525 Correa Road, Honolulu, Hawaii 96822, USA
† Department of Earth Sciences, The Open University, Milton Keynes MK7 6AA, UK

LAVA lakes and active strombolian vents have persisted at some volcanoes for periods exceeding the historic record. They liberate prodigious amounts of volatiles and thermal energy but erupt little lava, a paradox that raises questions about how volcanoes grow. Although long-lasting surface manifestations can be sustained by convective exchange of magma with deeper reservoirs, residence times of magmas beneath several basaltic volcanoes are ~10–100 years[1,2], indicating that where surface activity continues for more than 100–1,000 years, the reservoirs are replenished by new magma. Endogenous growth of Kilauea volcano (Hawaii) through dyke intrusion and cumulate formation is a well-understood consequence of the steady supply of mantle-derived magma[3,4]. As we show here, inferred heat losses from the Halemaumau lava lake indicate a period of dominantly endogenous growth of Kilauea volcano during the nineteenth century. Moreover, heat losses and degassing rates for several other volcanoes, including Stromboli, also indicate cryptic influxes of magma that far exceed visible effluxes of lavas. We propose that persistent activity at Stromboli, and at other volcanoes in different tectonic settings, is evidence of endogenous growth, involving processes similar to those at Kilauea.

When Europeans first visited Kilauea volcano in 1823, they found the Halemaumau crater filled by an active lava lake. In 1880, the lake had an estimated area of 6.51×10^3 m^2 (ref. 5). It persisted with various fluctuations until 1924 (ref. 3). Significant flank eruptions of lava took place only in 1840, 1868 and 1919–20 (refs 3, 6). The Kupaianaha lava pond which was active in the 1980s provides an analogue for the 19th century Halemaumau lava lake, although it formed part of an effusive lava system, whereas Halemaumau was located directly above a shallow reservoir at 1–2 km depth[4,7]. We use measurements made at Kupaianaha to estimate heat losses from the Halemaumau lake. During the period 1987–8, the surface of the Kupaianaha pond, though crusted, typically remained at a temperature of 200–300 °C, corresponding to an average radiant emittance of $\sim 5 \times 10^3$ W m^{-2} (ref. 8), and a convective loss (forced or natural) of $\sim 2 \times 10^3$ W m^{-2}. If these values are used for the pre-1924 Halemaumau, the inferred surface thermal losses *alone* imply a magma flux into the volcano of ~ 0.6–3.0×10^3 kg s^{-1} (depending on the enthalpy model chosen; see Table 1 and Fig. 1). When conductive losses through the walls of the reservoir and conduit[9], and hydrothermal losses (not easily modelled) are considered, a much larger magma influx is indicated, greatly exceeding the time-averaged surface eruption rate of $\sim 3 \times 10^2$ kg s^{-1} for the period[10], but consistent with the observed eruption rate of $\sim 10^4$ kg s^{-1} from Kilauea since 1924 (refs 4, 7). This in turn provides a lower bound to the rate of supply of magma to the volcano from the mantle plume powering the Hawaiian hotspot[4,11].

In order to sustain the rate of heat loss from Halemaumau, we suggest that fresh magma was arriving beneath Kilauea during the 19th century at about the same rate as it has during prolonged periods of visible lava effusion in the 20th century. It follows that Kilauea was steadily 'growing'. The difference