# The Holocene–Younger Dryas Transition Recorded at Summit, Greenland

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Analysis of ice from Dye-3, Greenland, has demonstrated that the transition between the Younger Dryas and Holocene climate periods occurred over a 40-year period. A near annually resolved, multiparameter record of the transition recorded in the GISP2 core from Summit, Greenland, shows that most of the transition occurred in a series of steps with durations of about 5 years. Some climate proxies associated with mid-latitude sources appear to have changed about 15 years before climate proxies associated with more northern regions. Changes in atmospheric water vapor are likely to have played a large role in the climate transition.

The Greenland Ice Sheet Project II (GISP2) and Greenland Ice Core Project (GRIP) ice cores were collected at the summit of the Greenland ice sheet at locations selected to obtain high–temporal resolution climate records of the last 100,000 years or so. The large-scale features of the transition between the Younger Dryas and the Holocene are illustrated by the GISP2 oxygen isotope ($\delta^{18}O$) ice record (1) (Fig. 1), which shows that, although most of the transition occurred in a few decades, it took ~1500 years [from ~11,700 to ~10,200 years before the present (B.P.), the present being defined as 1950 A.D.] for the $\delta^{18}O$ of the ice to reach typical Holocene values. The rapid nature of the transition had been previously noted at Dye-3, Greenland (2). Comparison of the GISP2 and GRIP (3) cores, which were collected 30 km apart, demonstrates that in this interval, features that span more than several centimeters of ice, which represents less than 2 years, are well replicated in both cores and can be

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considered to be climatologically significant (1, 4).

Here we focus on the ~200-year period during which most of the climate transition occurred. The multiparameter GISP2 record (Fig. 2) shows the sequence of events associated with the transition from the cold, dry, windy Younger Dryas period to the warmer, wetter, calmer Holocene. Annual layer counting reduces the uncertainty in the absolute ages to ±200 years (5, 6). It is possible to determine the duration of intervals between events with an uncertainty of ±3% or 2 years, whichever is greater (5, 6). Except for the methane data, we interpret the onset of an environmental change to occur when there is a noticeable change in the slope of a proxy indicator; this point is not necessarily when the proxy indicator exceeds the range of variability characteristic of the Younger Dryas. In 11,660 B.P., the concentration of non–sea-salt sulfate (7) increased above background levels by a factor of 4 for <5 years. This increase is unlikely to be associated with emissions of dimethylsulfide from the sea surface because there is no corre-

sponding feature in the methanesulfonate record (8). The elevated direct-current electrical conductivity (9) and ionic balance associated with this non–sea-salt sulfate indicates that the anomalous sulfate was a strong acid, most likely sulfuric acid (10). This elevated electrical conductivity signal is also seen in the GRIP (11) and Dye-3 (12) cores, collected 30 and 600 km away, respectively. The sulfate feature (13) coincides with a <10-year decrease in the deuterium (expressed as $\delta D$) in the snow (14), suggesting that the climate at Summit cooled. The feature was also associated with an increase in non–sea-salt calcium (15). This excursion in sulfate levels and $\delta D$ most likely resulted from a volcanic eruption. The source of this eruption is unknown and could be from a nearby volcano (Icelandic, Alaskan, or Kamchatkan) or a large, explosive eruption elsewhere. There were six similar periods of elevated sulfate that are interpreted to be associated with volcanic activity during the 2000 years before the Younger Dryas–Holocene climate transition (13). We do not believe the ~10-year cooling associated with this eruption would have been sufficient to tip a precariously poised climate system to a warmer state, but the coincident occurrence of the eruption and climate transition is intriguing.

The deuterium excess ($\delta D - 8\delta^{18}O$), which is more strongly influenced by the isotopic composition of the source water and evaporation along the transport path than $\delta D$ (16, 17), began to decrease in 11,660 B.P. and continued decreasing for 15 years. It has been suggested (2) that this dip was caused by the northward movement of the polar front, which moved the location of the moisture source area, thought to be in waters with sea surface temperatures around 25°C (currently 30°N) (16), to a more northern, colder location. The $\delta D$ levels remained low during the first part of this period, suggesting that temperatures at Summit remained cold. This interpretation
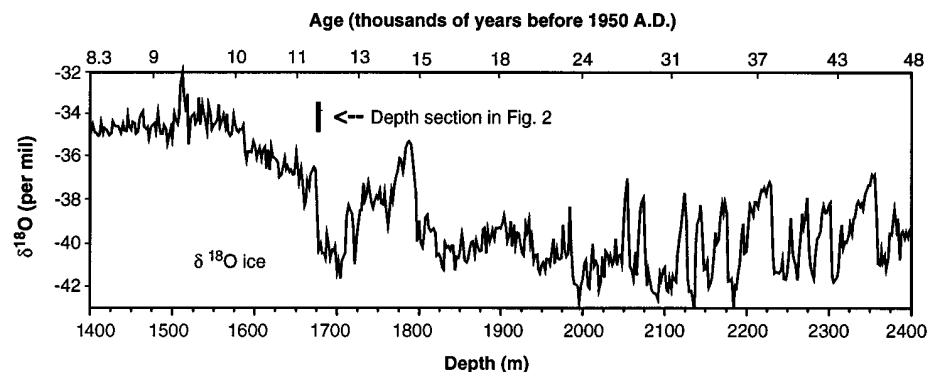
## Age (thousands of years before 1950 A.D.)



**Fig. 1.** The GISP2 $\delta^{18}O$ ice record (1) for a 40,000-year period at the end of the Wisconsin and start of the Holocene, plotted on a linear depth scale. The corresponding ages are on the top scale. The time period discussed in this report, and presented in Fig. 2, is indicated by the bar.

of δD is complicated by the change in moisture source area indicated by the deuterium excess; however, the magnitude of the observed change in deuterium excess is sufficiently small that δD can be interpreted as indicating continued cold conditions at Summit. The amount of non–sea-salt calcium and the concentration of insoluble particles decreased during this time, suggesting that an increase in moisture in non-Arctic source areas (i) reduced the size of terrestrial dust source areas and increased the washout rate of airborne particles or (ii) decreased wind speeds. During this 15-year period, lack of change in the concentration of sea-salt sodium, the sodium/chloride ratio (15), and mean particle size (18) suggests that there was no notable change in the extent of sea ice in the oceans adjacent to Greenland or in Arctic atmospheric transport processes.

At about 11,645 B.P., many indicators moved significantly out of their band of Younger Dryas variability in a second abrupt step. The δD and snow accumulation rate (5) increased abruptly, suggesting that the mean temperature at Summit increased rapidly. The δD increased several years before the accumulation rate increased and the sodium/chloride ratio decreased, suggesting a lag of a few years between warmer temperatures and the reduction in the distance to open ocean. The increase in snow accumulation rate is in excess of what is expected in association with the warming indicated by the stable isotopes (19). The mean particle size and particle concentration decreased, suggesting a reduction in wind speeds. Although there is likely to be a reduction of sea ice associated with the warming indicated by the δD, the concentration of sea-salt sodium decreased. This pattern indicates that (i) atmospheric transport was reduced by lower wind speeds or an increased washout rate (18), (ii) the source area contained less ocean area because of decreased meridional zonal transport (15, 20), or (iii) an increased percentage of precipitation occurred during summer, when the concentration of sea-salt sodium in precipitation was lower (21). The doubling of the snow accumulation rate may have been the result of the winters becoming sufficiently warm that greater amounts of winter snowfall oc-

curred (22). The increase in winter snow would cause annual mean δD to give an underestimate of the true warming at this time.

The possibility of different moisture source areas and changes in the relative seasonal precipitation rates make it difficult to quantify the magnitude of the temperature change indicated by the δD record. However, estimates from the temperature distribution in the GISP2 borehole (23) and the isotopic compositions of trapped gases (24) suggest that temperatures increased by ~15°C over the ~1500-year transition period. The δD record suggests that about half of this change occurred over <15 years. Between 11,645 and 11,612 B.P., there are excursions in δD, deuterium excess, mean particle size, non–sea-salt calcium, and accumulation that last <5 years. In 11,612 B.P., δD increased and remained high, indicating that temperatures increased at Summit. Sea-salt sodium concentrations and mean particle size decreased, indicating a move to less windy conditions. The accumulation rate did not change significantly, indicating that the seasonal timing and amount of snowfall remained constant.

Atmospheric gas becomes trapped in ice ~800 years after the ice surrounding it fell as snow, precluding a depth-based comparison between the gas and ice records. Atmospheric methane, from low- to mid-latitude wetlands and possible higher latitude source areas, increased dramatically at the transition (Fig. 3) (24, 25). Comparison to $^{15}N/^{14}N$, which is sensitive to near-surface temperature gradients when air is trapped in the ice, suggests that methane began to rise 0 to 30 years after an abrupt warming event (24). This warming is assumed to be the warming in 11,645 B.P., and the ages relative to this point are calculated (25). Within the dating uncertainty, and allowing for a lag between environmental changes and changes in methane production, the warming at Summit was approximately synchronous with the increase in temperatures and moisture in methane source areas. The widespread distribution of methane source areas indicates that the climate transition
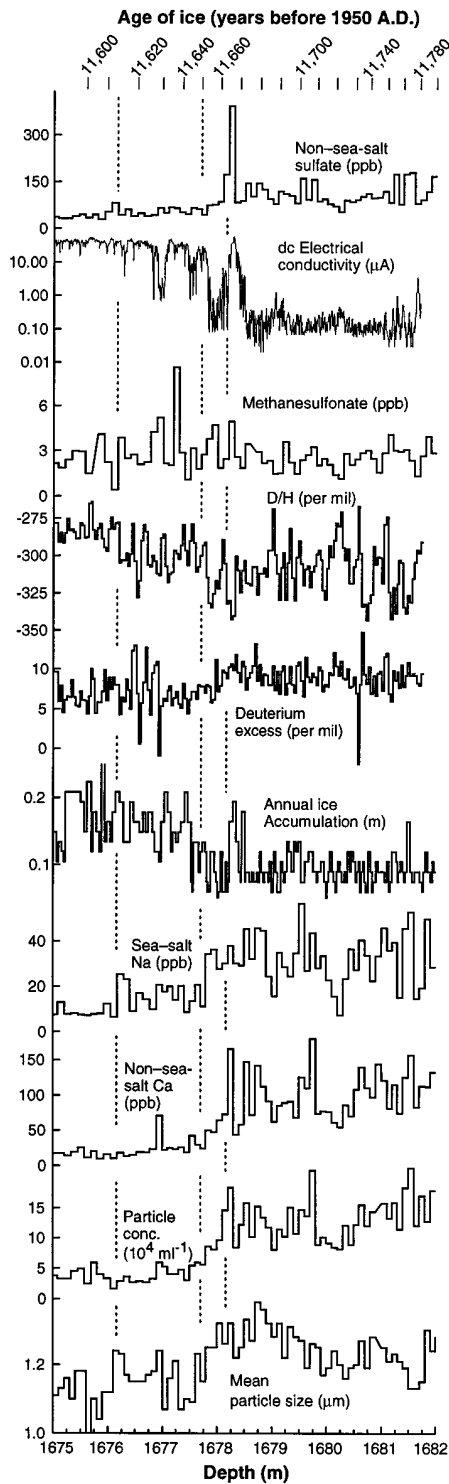


**Fig. 2.** The Younger Dryas–Holocene transition as recorded in the GISP2 core. All data are plotted with depth as a linear scale on the bottom of the plots. The corresponding ages are on the top scale. The dashed vertical lines are time horizons to facilitate comparison between plots.

**Fig. 3.** The GISP2 methane record (25), obtained from measurements of discontinuous samples and plotted as a function of depth. The solid vertical time lines correspond to the time lines in Fig. 2. The uncertainty in the age of the time lines relative to the age of the time lines in Fig. 2 is shown by the horizontal bars.
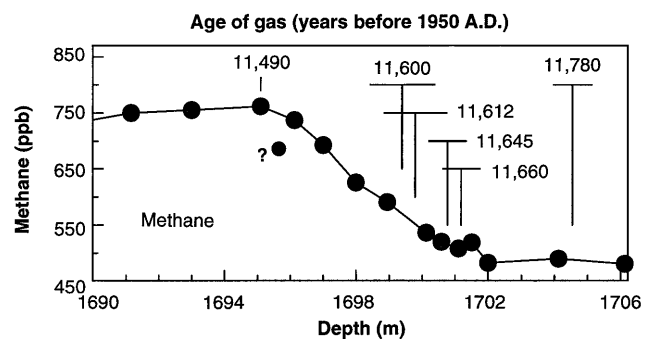
was rapid not just around Greenland but was also nearly synchronous over a large part of the terrestrial biosphere (*24, 25*).

A hemisphere-wide increase in water vapor at the onset of the transition is suggested by the increase in methane source areas, a possible increase in particle washout rate and decreased size of dust source areas, and increased snow accumulation at Summit. Although it is unclear if the change in environmental conditions resulting in the increase in methane led temperature change in Greenland, other climate proxies associated with non-Arctic areas (deuterium excess, non–sea-salt calcium, and particle concentration) appear to change before climate proxies associated with Arctic areas and Arctic atmospheric transport (δD, sea-salt sodium, and mean particle size). This difference suggests that some climate conditions changed outside of the Arctic in 11,660 B.P. and were followed by changes at higher latitudes in 11,645 and 11,612 B.P. Because the proxy records that are available are not sensitive to all of the climate changes occurring in the Arctic, it is not clear if the climate transition started at lower latitudes or if earlier changes that we are unable to detect occurred in more northern regions.

We speculate that the transition was characterized by a change in ocean circulation that significantly increased atmospheric water vapor. This added vapor could have increased retention of long-wave solar radiation and created a feedback that stabilized the new climate state. Regardless of the mechanism, the transition was a 50-year period during which wind speeds, precipitation, temperatures, and sea ice were changing throughout the Northern Hemisphere on subdecadal time scales.

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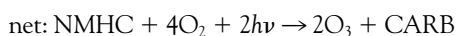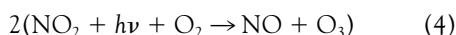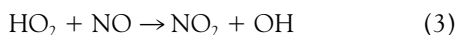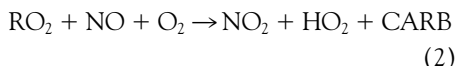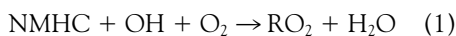
# The Impact of Aerosols on Solar Ultraviolet Radiation and Photochemical Smog

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Photochemical smog, or ground-level ozone, has been the most recalcitrant of air pollution problems, but reductions in emissions of sulfur and hydrocarbons may yield unanticipated benefits in air quality. While sulfate and some organic aerosol particles scatter solar radiation back into space and can cool Earth's surface, they also change the actinic flux of ultraviolet (UV) radiation. Observations and numerical models show that UV-scattering particles in the boundary layer accelerate photochemical reactions and smog production, but UV-absorbing aerosols such as mineral dust and soot inhibit smog production. Results could have major implications for the control of air pollution.

$M$ore than 100 counties in the United States regularly violate the Environmental Protection Agency's (EPA) Ambient Air Quality Standard for ozone ($O_3$) of 120 ppbv (parts per $10^9$ by volume averaged over 1 hour) (*1*). This ozone results from the interaction of pollutant oxides of nitrogen and nonmethane hydrocarbons (NMHCs) with solar radiation, for example, via reactions (1) to (4), and is thus sometimes called photochemical or Los Angeles–type smog.

$$NMHC + OH + O_2 \rightarrow RO_2 + H_2O \quad (1)$$

$$RO_2 + NO + O_2 \rightarrow NO_2 + HO_2 + CARB \quad (2)$$

$$HO_2 + NO \rightarrow NO_2 + OH \quad (3)$$

$$2(NO_2 + h\nu + O_2 \rightarrow NO + O_3) \quad (4)$$

net: $NMHC + 4O_2 + 2h\nu \rightarrow 2O_3 + CARB$

(where CARB is carbonyl compounds, which can further break down to produce additional $O_3$, and $h\nu$ represents a quantum of light). The rate of production of smog depends on the concentrations of these pol-

lutants and [for reactions such as (4)] on the intensity of solar near-UV ($300 < \lambda < 400$ nm) radiation (*2, 3*).

On hot, smoggy summer days in many North American and European cities the cloud-free sky shows a milky white color—the result of particulate air pollution scattering solar radiation. The impact of these aerosols on Earth's radiative balance and on climate (aerosol radiative forcing) has been studied extensively (*4*). The amount of radiation available to drive these reactions depends on the solar zenith angle (θ), absorption and scattering by gases and particles, and the surface albedo (the fraction of light reflected from Earth's surface). In the atmosphere, ozone is usually the only important absorbing gas in the near-UV spectrum. Scattering of radiation by gases (Rayleigh scattering) redistributes much of the UV radiation; Bruehl and Crutzen (*5*) showed that Rayleigh scattering can increase the effective path length and relative importance of tropospheric ozone in shielding us from harmful UV radiation. Optically thick clouds reduce the radiation below them, but calculations by Madronich (*6*) showed that scattering by cloud droplets can actually increase the actinic flux and the rate of photochemical reactions in the upper parts of clouds.

Theoretical calculations and observa-

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