Available online at www.sciencedirect.com
ScienceDirect
EPSL
Earth and Planetary Science Letters 268 (2008) 397–407
www.elsevier.com/locate/epsl

4±1.5 °C abrupt warming 11,270 yr ago identified from trapped air in Greenland ice

Takuro Kobashi [a,*], Jeffrey P. Severinghaus [a], Jean-Marc Barnola [b]

[a] *Scripps Institution of Oceanography, University of California, San Diego, La Jolla, CA 92093, USA*
[b] *Laboratoire de Glaciologie et Géophysique de l'Environnement, CNRS, Saint-Martin d'Héres, France*

Received 1 January 2008; received in revised form 20 January 2008; accepted 22 January 2008
Available online 8 February 2008
Editor: M.L. Delaney

Abstract

Nitrogen and argon isotopes in air trapped in a Greenland ice core (GISP2) show two prominent peaks in the interval 11,800–10,800 B.P., which indicate two large abrupt warming events. The first abrupt warming (10±4 °C) is the widely documented event at the end of the Younger Dryas. Here, we report on the second abrupt warming (4±1.5 °C), which occurred at the end of a short lived cooler interval known as the Preboreal Oscillation (11,270±30 B.P.). A rapid snow accumulation increase suggests that the climatic transition may have occurred within a few years. The character of the Preboreal Oscillation and the subsequent abrupt warming is similar to the Dansgaard–Oeschger (D/O) events in the last glacial period, suggestive of a common mechanism, but different from another large climate change at 8,200 B.P., in which cooling was abrupt but subsequent warming was gradual. The large abrupt warming at 11,270 B.P. may be considered to be the final D/O event prior to the arrival of the present stable and warm epoch.

Keywords: Preboreal Oscillation; abrupt climate change; Holocene; Greenland; ice core

1. Introduction

1.1. The Early Holocene and Preboreal Oscillation

After an abrupt warming at the end of the Younger Dryas interval (Severinghaus et al., 1998), Greenland temperature gradually rose by ~5 °C for ~2000 yr and accumulation rate increased by >40% (Cuffey and Clow, 1997). This temperature increase was interrupted by a brief cold event (the Preboreal Oscillation) at 11,400–11,270 B.P. (yr Before Present, where present means C.E. 1950; Fig. 1) (Bjorck et al., 1996, 1997). Abundant evidence of the Preboreal Oscillation has been found in the North Atlantic region from both low and high latitudes, and more evidence is now being found beyond the North Atlantic (Yu and Eicher, 1998; Hu et al., 2006). Bond et al. (1997) found that sea ice was advected toward the south in the North Atlantic during the Preboreal Oscillation (named as Event 8). Atmospheric methane concentration decreased during the Preboreal Oscillation by 8% or 60 ppb (Brook et al., 2000) and the tropical Atlantic had stronger trade winds and lower precipitation (Hughen et al., 1996), suggesting that a broad geographic area experienced this cooling (Brook et al., 2000). Many European pollen studies show that vegetation responded to this cool event (Bjorck et al., 1997). At the time of the Preboreal Oscillation, sea level was still about 50 m lower than present (Bard et al., 1996). A large meltwater pulse (MWP-1B) is inferred to have occurred around this time from an observed rapid sea level rise, potentially causing this brief cool event (Fairbanks, 1989; Bard et al., 1996).

* Corresponding author. Institute for Global Environmental Strategies (IGES), 2108 Kamiyamaguchi, Hayama, Kanagawa, 240-0115 Japan. Tel.: +81 46 855 3871; fax: +81 46 855 3809.
E-mail address: kobashi@iges.or.jp (T. Kobashi).

0012-821X/$ - see front matter © 2008 Elsevier B.V. All rights reserved.
doi:10.1016/j.epsl.2008.01.032

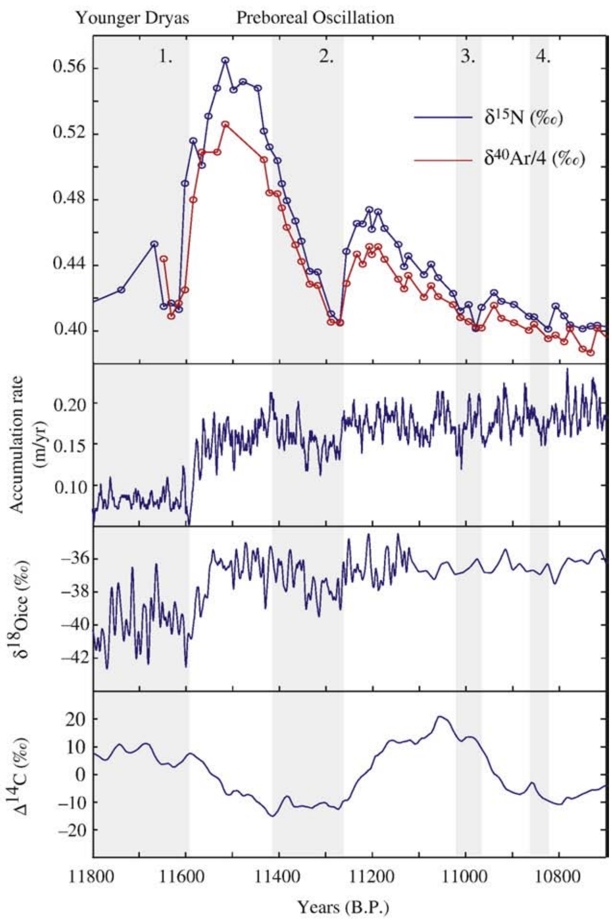

Fig. 1. Measured δ^{15}N and δ^{40}Ar in air trapped in ice (GISP2), accumulation rate (Alley et al., 1997b; Cuffey and Clow, 1997), $\delta^{18}O_{ice}$ (Stuiver et al., 1995), and residual Δ^{14}C from tree rings (Reimer et al., 2004) over the interval 11,800–10,700 B.P. The $\delta^{18}O_{ice}$ and accumulation data were smoothed with a 5-year running mean after a 1-year resolution time series was generated by linear interpolation. δ^{15}N and δ^{40}Ar for the period 11,800–11,447 B.P are from Severinghaus et al. (Severinghaus et al., 1998) and for the later part are from Kobashi (2007). The shaded areas numbered as 1 to 4 are cooler periods characterized by δ^{15}N, as follows: (1) the Younger Dryas, (2) the Preboreal Oscillation, (3) the 11.0 ka event, (4) the 10.8 ka event. Note that the resolution of the $\delta^{18}O_{ice}$ record changes after ~11,300 B.P. in the original data (Stuiver et al., 1995).

1.2. Nitrogen and argon isotopes and temperature reconstruction

Over the last decade, isotopic compositions of inert gases in ice cores such as nitrogen and argon have been extensively used to reconstruct past abrupt temperature changes (Severinghaus et al., 1998; Lang et al., 1999; Leuenberger et al., 1999; Severinghaus and Brook, 1999; Landais et al., 2004a,b; Grachev and Severinghaus, 2005; Huber et al., 2006b; Kobashi et al., 2007). Recent advances in measurement techniques have further sharpened the precision of estimates of temperature change (Kobashi, 2007), and a recently completed high resolution Holocene record (10–20 year interval) from GISP2 provides opportunities to understand past climate change during the period when human society went through major changes (Kobashi, 2007; Kobashi et al., 2007).

The basic principles of the method rest on the fact that the isotopic ratios of these gases in the atmosphere are constant for $>10^5$ yr (Mariotti, 1983; Allegre et al., 1987). Therefore, any deviations of isotopic composition in ice cores from atmospheric values are the result of isotopic fractionation in the firn layer (unconsolidated snow on top of glacial ice). The firn layer can be subdivided into three sections in terms of isotopic fractionation (Sowers et al., 1992). The upper few meters of firn is called the “convective zone”, in which air freely exchanges with the atmosphere by wind pumping (Colbeck, 1989; Kawamura et al., 2006). Therefore, air in this zone has the same gas composition as the atmosphere (Colbeck, 1989). Below this layer, there is a stagnant layer called the “diffusive air column”, where gas is nearly in diffusive equilibrium and transport is dominantly by molecular diffusion (Sowers et al., 1992). Isotopic fractionation occurs in this layer by gravitational settling and thermal diffusion (Severinghaus et al., 1998; Severinghaus and Brook, 1999; Severinghaus et al., 2001). The last layer is called the “non-diffusive zone” or “lock-in-zone”, where vertical air movement ceases owing to the existence of impermeable layers of higher-density firn (Sowers et al., 1992). The thickness of the convective zone and non-diffusive zone are likely constant during the Holocene in Greenland, as a model study reproduces gas isotope signals well with an assumption of constant thickness of these two layers (Goujon et al., 2003). Therefore, henceforth we call the thickness of the diffusive air column the “firn thickness”.

Gases in the diffusive air column fractionate by at least two mechanisms. First, a change in firn thickness induces a change in gravitational fractionation (Craig et al., 1988; Sowers et al., 1992). Second, a temperature gradient (ΔT) between the top and bottom of the firn induces thermal fractionation (Severinghaus et al., 1998). Measurements of both nitrogen and argon isotopes allow a deconvolution of these two effects, and can be used to infer past firn thickness and ΔT (Severinghaus and Brook, 1999; Landais et al., 2004a,b). The method is most effective for investigating decade-scale abrupt climate changes, which create a large ΔT and thus large isotopic signals (Severinghaus and Brook, 1999; Landais et al., 2004a,b). The method is not very effective on multi-centennial scales, because the firn thermally equilibrates on these timescales (Allegre et al., 1987).

The surface temperature reconstruction from ΔT is not straightforward for higher (>0.05 yr^{-1}) or lower (<0.005 yr^{-1}) frequencies because there is no single unique solution, due to smoothing of the record by gas diffusion and bubble close-off, and due to thermal equilibration. As various surface temperature histories can satisfy the observed gas-isotopic signals, oxygen isotope records of ice ($\delta^{18}O_{ice}$) have been used by some studies to provide constraints on the “shape” and rate of surface temperature change, thus reducing the dimensionality of the problem (Landais et al., 2004a; Huber et al., 2006b; Kobashi et al., 2007). However, $\delta^{18}O_{ice}$ is not only a proxy of temperature. It may vary without temperature change, or it may be biased by other climatic variables such as the evaporative origins of the moisture, or changes in the seasonality of precipitation (Charles et al., 1994; Jouzel et al., 1997; White et al., 1997). For these reasons, direct methods of temperature