Quaternary Science Reviews 27 (2008) 882–887

Why there was not a Younger Dryas-like event during the Penultimate Deglaciation

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Received 16 August 2007; received in revised form 14 February 2008; accepted 15 February 2008

Abstract

The Younger Dryas cold event is a relatively unique feature of the last deglaciation when compared to previous deglaciations, suggesting a unique trigger rather than the commonly held forcing mechanism of North American freshwater routing to the North Atlantic. Here, I compare the last (T-I) and penultimate (T-II) deglaciations and provide new support for the argument that the lack of a Younger Dryas-like event during T-II is due to the rapidity of Northern Hemisphere ice sheet retreat under greater boreal summer insolation forcing. Faster ice retreat suppressed Atlantic meridional overturning circulation (AMOC) until near the end of T-II, while during T-I AMOC increased relatively early. During T-I, the eastward routing of freshwater that caused the Younger Dryas happened after AMOC resumption, whereas during T-II this routing occurred prior to the resumption of AMOC. Thus the increased flux of freshwater to the North Atlantic during T-II had little effect on AMOC, explaining the lack of a Younger Dryas-like climate oscillation during this deglaciation.

1. Introduction

The Younger Dryas cold event (~12.9–11.5 kyr BP) interrupted the last deglaciation (~19–6 kyr BP) with an abrupt return to cold, glacial conditions (Fig. 1a). This event is recorded in numerous records from around the North Atlantic region (Clark et al., 2002), in subtropical records influenced by monsoon intensity (Yuan et al., 2004; Kelly et al., 2006; Weldeab et al., 2007), and in records of atmospheric methane concentration mainly reflecting precipitation and temperature impacts on boreal and tropical wetlands (Brook et al., 2000; Monnin et al., 2001; Schaefer et al., 2006; Sowers, 2006) (Fig. 1b and c). The Younger Dryas was originally hypothesized to be caused by the retreat of the southern Laurentide Ice Sheet (LIS) out of Lake Superior with attendant routing of western Canadian Plains freshwater from the Mississippi River to the St. Lawrence River ~12.9 kyr BP (Johnson and McClure, 1976; Rooth, 1982; Broecker et al., 1989). This increased freshwater discharge freshened the northern North Atlantic, reducing Atlantic meridional overturning circulation (AMOC) and northward heat transport (McManus et al., 2004) (Fig. 1a). Subsequent paleoceanographic studies questioned this mechanism because of the lack of evidence for St. Lawrence Estuary freshening during the Younger Dryas (Keigwin and Jones, 1995; deVernal et al., 1996), which in part led modeling studies to suggest alternative freshwater sources as the cause of this event (Tarasov and Peltier, 2005). Furthermore, minimum limiting radiocarbon dates from the western Lake Superior region have been interpreted as indicating that ice retreated out of Lake Superior after the onset of the Younger Dryas (Lowell et al., 2005; Teller et al., 2005), but these dates are only minimum limiting and thus do not refute the original routing hypothesis (Teller and Boyd, 2006). Recently, geochemical records from the St. Lawrence Estuary contested these studies and likely confirmed the original routing hypothesis as the cause of the Younger Dryas (Carlson et al., 2007). However, evidence for a Younger Dryas-like event is equivocal during earlier deglaciations (Sarnthein and Tiedmann, 1990; Adkins et al., 1997; Oppo et al., 1997, 2001; Chapman and Shackleton, 1998; Petit et al., 1999; Spahni et al., 2005; Kelly et al., 2006;

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0277-3791/$ - see front matter
doi:10.1016/j.quascirev.2008.02.004

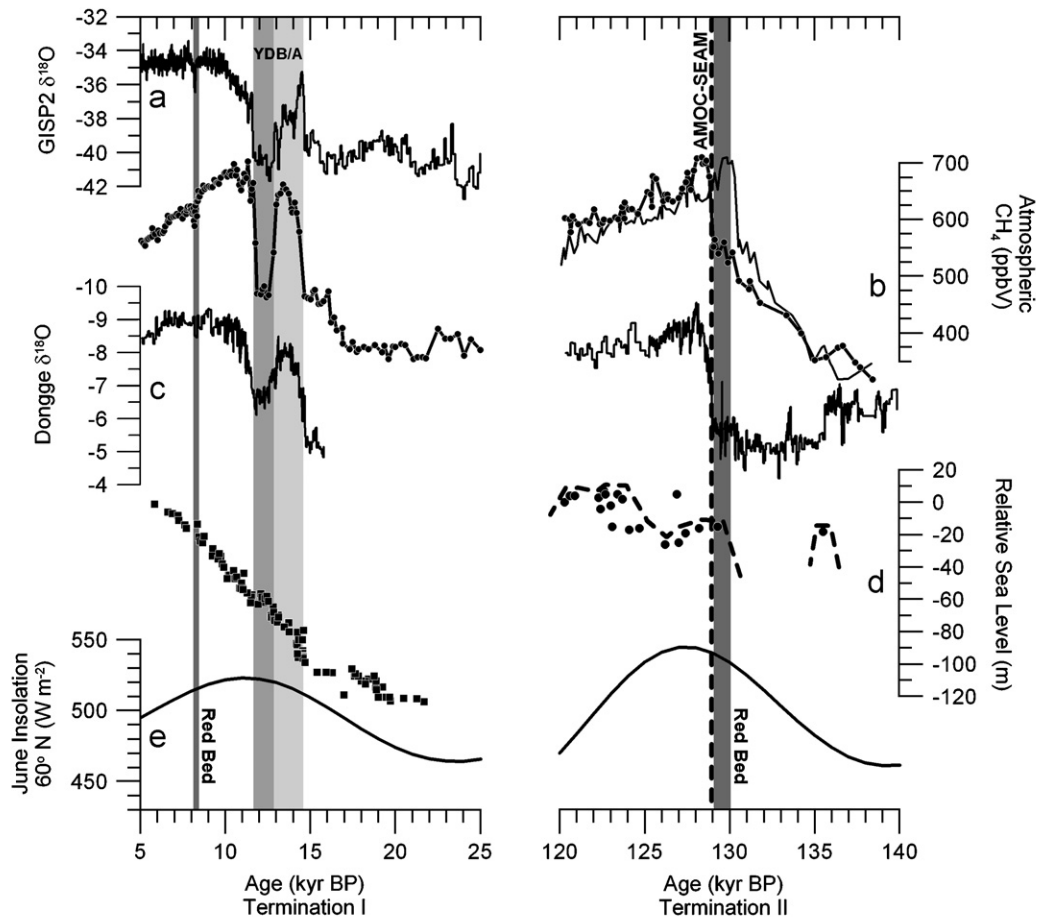

Fig. 1. Comparing North Atlantic influenced proxies during T-I (left) and T-II (right). (a) GISP2 $\delta^{18}O$, a proxy of Summit Greenland temperature (Grootes et al., 1993). (b) T-I Dome C (Monnin et al., 2001) and T-II Vostok (Petit et al., 1999) atmospheric methane concentration (black line with round symbols). Also shown during T-II is the alternate age model for methane of Ruddiman and Raymo (2003) (thin black line). The Shackleton (2000) age model falls between these two. (c) Dongge Cave speleothem $\delta^{18}O$ (Yuan et al., 2004; Kelly et al., 2006), a proxy for SEAM intensity. Note that these records are similar to proxy records of West African monsoon intensity (Weldeab et al., 2007). (d) Relative sea level; T-I is from the Clark and Mix (2002) compilation, and T-II is the revised coral dating of Thompson and Goldstein (2005) with their interpreted sea level curve (dashed black lines). Note that the magnitude of the T-II record is in agreement with Red Sea records (Siddall et al., 2006; Rohling et al., 2008). (e) June insolation at 60° N (Berger and Loutre, 1991). Dark gray bars denote the timing of red bed deposition in the Labrador Sea. Medium gray bar during T-I is the timing of the Younger Dryas (YD) cold period. Light gray bar is the timing of the Bølling/Allerød (B/A) warm period. Dashed vertical line during T-II denotes the timing of increased SEAM intensity and the inferred turn-on of AMOC, similar to the onset of the Bølling during T-I.

Desprat et al., 2007; Weldeab et al., 2007), implying a unique trigger for the event rather than the routing mechanism (Firestone et al., 2007), which would presumably operate during every deglaciation (Broecker, 2006). Here, I synthesize key deglacial records along with new evidence that support the hypothesis that the lack of a Younger Dryas-like event during the penultimate deglaciation (Termination II; T-II) is due to the magnitude of deglacial forcing and the resulting climate setting relative to the last deglaciation (Termination I; T-I), not a unique trigger.

2. Deglacial events during T-I and T-II

The major warming of the North Atlantic region following the last glacial maximum (LGM; ~21 kyr BP) occurred ~14.7 kyr BP due to an increase in AMOC (Fig. 1a) (McManus et al., 2004) and is called the Bølling/Allerød warm period (B/A). This abrupt warming is also expressed as an intensification of the Southeast Asian monsoon (SEAM) (Yuan et al., 2004) (Fig. 1c) and increased atmospheric methane concentration (Fig. 1b) reflecting greater production in tropical and boreal wetlands (Brook et al., 2000). The subsequent routing of freshwater to the St. Lawrence River reduced AMOC ~12.9 kyr BP with attendant cooling of the North Atlantic region, diminished monsoon activity and decreased wetland productivity (Fig. 1) (Brook et al., 2000; Clark et al., 2002; Yuan et al., 2004; Carlson et al., 2007). By the beginning of the Younger Dryas, the LIS had lost ~1/3 of its overall LGM extent (Dyke, 2004), contributing ~20 m to the ~60 m of deglacial sea level rise (Fig. 1d). Note that the Younger Dryas and ~1/2 of the overall deglacial sea level rise occurred near the peak in boreal summer insolation (Fig. 1d and e).

During T-II, paleoceanographic data suggest a short-lived deglacial cooling event of a smaller magnitude than the Younger Dryas that is restricted to the eastern North Atlantic, possibly in response to variations in shallow North Atlantic overturning (Oppo et al., 2001; Desprat et al., 2007). This local climate oscillation is not reflected in deeper AMOC records (Sarnthein and Tiedmann, 1990; Adkins et al., 1997; Oppo et al., 1997, 2001; Chapman and Shackleton, 1998), or records of atmospheric methane concentration and SEAM intensity (Petit et al., 1999; Kelly et al., 2006), suggesting that a large increase and subsequent decrease in AMOC did not occur during T-II (Fig. 1). Rather, atmospheric methane and SEAM abruptly increased ~129 kyr BP[1] similar to the B/A onset during T-I (Fig. 1). Given the correlation between AMOC, SEAM, and atmospheric methane (Brook et al., 2000; McManus et al., 2004; Yuan et al., 2004; Broecker, 2006), the SEAM-methane increases at ~129 kyr PB imply a coeval abrupt increase in AMOC. Similarly, Oppo et al. (1997) demonstrated that North Atlantic benthic δ^{13}C records suggest a resumption of AMOC at this abrupt transition. This abrupt, B/A-like event occurred ~2 kyr before peak boreal summer insolation when sea level was ~15 m below modern, indicating a greatly reduced LIS relative to the LIS volume/area at the B/A onset during T-I (Fig. 1 and Table 1).

3. Comparing T-I and T-II

Northward retreat of the southern LIS out of the Great Lakes with attendant eastward routing of western Canadian Plains freshwater should occur at approximately the same LIS and global ice volumes during T-II as during T-I, given that maximum ice sheet volumes and extents have not varied significantly over the past 450 kyr (Fig. 2b). During T-I, this routing occurred after the resumption of AMOC at ~14.7 kyr BP, which allowed the increased freshwater discharge to reduce AMOC and cause the Younger Dryas. During T-II, determining the timing of when the LIS was ~1/3 of its volume/area, or when sea level had risen ~60 m, is difficult due to a lack of absolute dated sea level records from this time interval. However, it must have occurred some time before ~130 kyr BP (Fig. 1d), indicating that the eastward routing of freshwater to the North Atlantic took place prior to the increase in AMOC strength (Oppo et al., 1997) inferred at ~129 kyr BP (Fig. 1c) (Kelly et al., 2006). As originally suggested by Ruddiman et al. (1980) and subsequently by Oppo et al. (1997), without vigorous AMOC, the increased freshwater discharge to the North Atlantic would only reinforce an already suppressed AMOC and thus not significantly influence regional climate.

The contrasting relative timings of deglacial AMOC resumption during T-I and T-II may be in response to the greater boreal summer insolation increase that forced T-II relative to T-I (Fig. 1 and Table 1) (Ruddiman et al., 1980; Oppo et al., 1997). During T-II, global ice sheet retreat lasted ~14 kyr from the initial increase in boreal summer insolation to the achievement of full deglaciation[2] (0 m of relative sea level), relative to ~19 kyr during T-I (Fig. 1). The effects of greater insolation forcing are also evident in the time the LIS took to retreat to a greatly diminished size. One indicator of a greatly diminished LIS is the deposition of red clay sediment from northern Hudson Bay in the Labrador Sea, which indicates the collapse of ice over Hudson Bay and attendant eastward drainage of northwestern Hudson Bay water (Barber et al., 1999). During T-I, the deposition of this red bed in the Labrador Sea occurred ~8.4 kyr BP (Barber et al., 1999) when sea level was ~−15 m (Fig. 1). During T-II, a similar carbonate red bed was deposited in the Labrador Sea 129–30 kyr BP (Fig. 1) based on orbital tuning of foraminifera δ^{18}O associated with the layer (J. Stoner, personal communication). Supporting this timing is relative sea level, which was ~−15 m to 129 kyr BP (Fig. 1, Table 1). Thus during T-II, the LIS was greatly diminished prior to peak boreal summer insolation, whereas during T-I, the LIS did not reach this diminished size until ~2 kyr after peak insolation (Fig. 1, Table 1). Essentially, the LIS retreated faster during T-II than during T-I.

Given several assumptions of ice sheet volume outlined in the caption of Table 1, I estimate the meltwater discharge to the North Atlantic and Arctic Oceans during T-I and T-II from retreat of the surrounding LIS, Scandinavian, Barents-Kara and British Ice Sheets from 5000 years after the insolation minimum to the turn-on of AMOC (i.e. increased SEAM intensity and atmospheric methane concentration) (Table 1). During T-I, meltwater discharge to the North Atlantic and Arctic Oceans is 0.05–0.08 Sverdrups (1 Sverdrup (Sv) = $10^6\,m^3\,s^{-1}$) and during T-II ~0.19 Sv. According to general circulation models, 0.19 Sv is sufficient to reduce AMOC, whereas 0.05–0.08 Sv would have less of an effect on AMOC (Stouffer et al., 2006) and could allow the resumption of AMOC at ~14.7 kyr BP. The retreat of North Atlantic ice sheets despite reduced AMOC and a cold North Atlantic during T-I and T-II may be in response to direct insolation forcing (Oppo et al., 1997) and/or tropical warming during periods of reduced AMOC, which may have caused negative mass balances for Northern Hemisphere Ice Sheets (Clark et al., 2007). Thus greater meltwater forcing during T-II can explain why AMOC remained suppressed until near the end of deglaciation (Oppo et al., 1997). The eastward routing of freshwater that caused the Younger Dryas (Carlson et al., 2007) happened during T-II while

[1]Note in Fig. 1b, I have plotted along with the Petit et al. (1999) Vostok gas-age chronology, the chronology of Ruddiman and Raymo (2003). While this chronology differs in the timing of the abrupt increase in methane, the overall conclusions there from remain the same (Table 1). However, because of the agreement between the Petit et al. (1999) chronology and the absolute dated Dongge Cave SEAM record, I use the Petit et al. chronology in my discussion.

[2]Note that this is based on the Thompson and Goldstein (2005) revised U/Th dates. If the original ages are used of Stirling et al. (1998), the timing for T-II reduces to ~9 kyr.

Table 1
Comparison between T-I and T-II

Termination	$\Delta W\,m^{-2}$	RSL peak Insol (m)	RSL CH_4 rise (m)	RSL red bed (m)	Δ kyr red bed	NA runoff (Sv)
T-I	58.8	−55	−80 to −95	−15 to −20	~15.6	0.05–0.08
T-II-GT4	83.46	−15	−15	−15	9–10	0.19
T-II-RR03	83.46	−15	−20 to −30	−15	9–10	0.20–0.23

(From left to right) Change in June insolation at 60°N that forced deglaciation (Berger and Loutre, 1991).
Relative sea level (RSL) at peak insolation, the rapid rise in atmospheric methane, and at the deposition of the red bed in the Labrador Sea.
Amount the sea level rise is from 5 kyr after initial June insolation increase and the timing of the abrupt increase in atmospheric methane concentration and SEAM intensity, assuming linear ice sheet melting (Fig. 1). Note the two different age models for Vostok methane used: GT-4 (Petit et al., 1999) and RR03 (Ruddiman and Raymo, 2003). The Shackleton (2000) age model falls in between these two age models (Fig. 1b).
Number of kyr from the timing of initial June insolation increase to the timing of red bed deposition. Estimated North Atlantic and Arctic Ocean discharge from the melting of North Atlantic ice sheets in Sverdrups (Sv, $10^6\,m^3\,s^{-1}$) assuming an LGM and Penultimate Glacial maximum relative sea level lowering of 120 m (Clark and Mix, 2002), excluding an Antarctic contribution of 20 m and Cordilleran/other smaller ice sheet contributions of ~6 m (Clark and Mix, 2002).

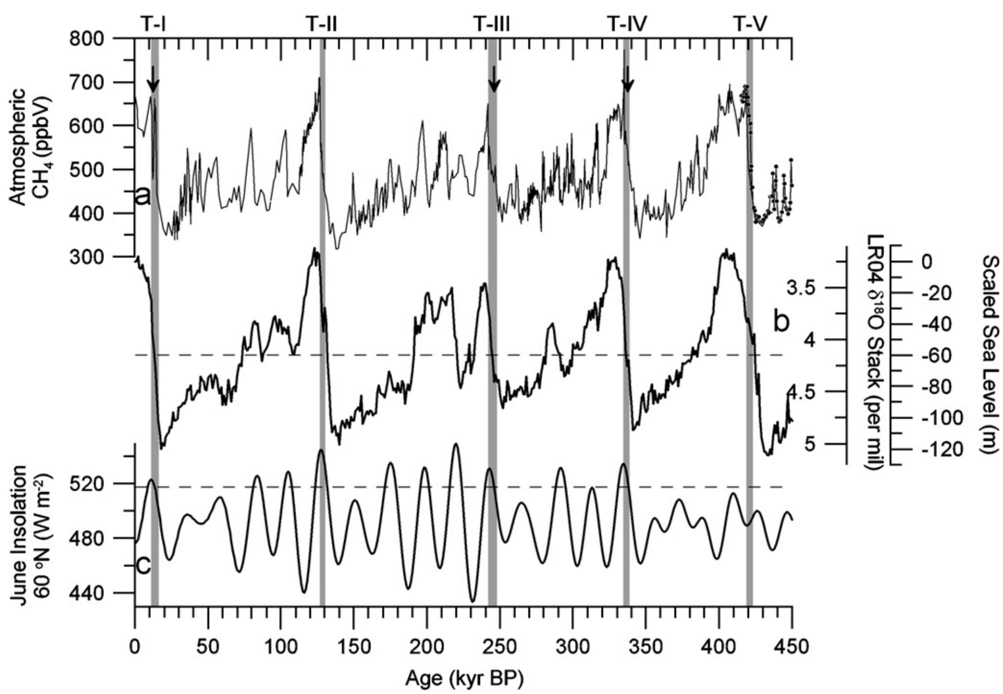

Fig. 2. 450,000 years of atmospheric methane concentration, approximate global ice volume and June insolation at 60°N. (a) Vostok (black line) and Dome C (black symbols) atmospheric methane concentration on the newer EDC2 time scale to match the Vostok record to the higher resolution Dome C record at ~420 kyr BP (Petit et al., 1999; Spahni et al., 2005). (b) Benthic $\delta^{18}O$ stack LR04, a proxy of sea level and global ice volume (Lisiecki and Raymo, 2005), scaled to LGM sea level lowering to produce the equivalent change in sea level. (c) June insolation at 60°N (Berger and Loutre, 1991). Arrows denote the timing of the Younger Dryas and possible similar events in the atmospheric methane record during earlier deglaciations. Vertical gray bars denote the timing of rapid atmospheric methane concentration increases during deglaciations. Horizontal dashed lines denote the benthic $\delta^{18}O$ (b) and June insolation (c) levels during the Younger Dryas.

AMOC was still suppressed, and therefore had less of an effect on AMOC and climate.

North American freshwater routing to the northern North Atlantic should occur during even earlier deglaciations with attendant effects on AMOC and climate presumably when rates of deglaciation and sea level rise were less than during T-II. Boreal summer insolation increases during Terminations III, IV and V were less than the insolation change during T-II with attendant slower rates of sea level rise (Fig. 2). While large oscillations in atmospheric methane are not clearly apparent across these terminations, a small oscillation in atmospheric methane did occur during T-III and possibly during T-IV (Fig. 2a), with the smaller magnitude than the Younger Dryas oscillation possibly reflecting the lower sample resolution (Petit et al., 1999). These methane oscillations imply potential increases and subsequent decreases in AMOC, similar to T-I and in agreement with paleoceanographic evidence for climate oscillations in the North Atlantic that were larger than the local event during T-II (e.g. Desprat et al., 2007).

In contrast, atmospheric methane appears to have abruptly increased during T-V without an oscillation (Spahni et al., 2005). However, the orbital forcing behind this deglaciation was smaller than T-I and drawn out over ~20 kyr with a similarly long period of ice sheet melting, and thus may not be comparable to the later deglaciations (Fig. 2).

4. Conclusions

During the last deglaciation, North American freshwater routing to the North Atlantic reduced AMOC and caused the Younger Dryas, because AMOC had already resumed and could be influenced by increased freshwater discharge. During the penultimate deglaciation, AMOC did not resume until near the end of deglaciation, likely reflecting greater boreal summer insolation forcing with faster ice sheet retreat and greater meltwater discharge to the North Atlantic. Thus eastward routing of western Canadian Plains freshwater could not decrease an already suppressed AMOC, explaining the lack of a Younger Dryas-like climate event. Therefore, a unique trigger for the Younger Dryas is not required to explain the lack of a similar event during the penultimate deglaciation. Resolving earlier Younger Dryas-like deglacial events will require the development of higher resolution atmospheric methane data and well-dated speleothem records.

Acknowledgments

The author would like to thank Delia Oppo, William Thompson, Jeremy Shakun, Luke Skinner, Peter Clark, Summer Praetorius, Stephanie Desprat, and Joseph Stoner for discussion and comments on this topic. Comments and suggestions from two anonymous reviewers greatly improved this manuscript.

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