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The Younger Dryas in palynological records from the northern Northwest Atlantic: Does the terrestrial record lag the marine and air records?

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A B S T R A C T

Major changes in both pollen and dinocyst assemblages are recorded during the Younger Dryas in Bay of Islands and are associated with large drops in air and sea surface temperatures, and sea surface salinity. The changes in vegetation are similar with those observed at other sites in Newfoundland. Dinoflagellate cyst assemblages also show similar species succession as nearby sites from the Gulf of St. Lawrence and Laurentian Channel. The amplitude of changes in air temperature (a 7 °C drop) is similar to those recorded in northeastern North America. In the sea-surface, temperature dropped by 7 °C, salinity by 4 psu and the sea ice cover duration increased by 7 months. Sea-surface and air temperatures started cooling 250 and 110 years before the start of the YD, hence there is a 140 years delay between ocean and atmosphere. A warming trend is seen in the marine record 310 years before the end of the YD, while vegetation and air temperatures started recovering after the end of the event, some 370 years later. The major lithological and vegetation changes observed during the YD period in eastern North America might actually represent the full impacts of the cooling associated with the YD, since the cooling trend actually started before the start of the YD period, as defined by radiocarbon dates and pollen zones.

1. Introduction

The Younger Dryas (YD) was an abrupt cooling event recorded in the ice core record from 12,900 to 11,700 b2k that interrupted the warming trend of the last deglaciation (Alley, 2000; Rasmussen et al., 2006). The cooling was reflected by a decrease of several degrees in air and sea-surface temperatures (SST) (Shakun and Carlson, 2010). It is assumed to be followed by an abrupt warming (Steffensen et al., 2008). Many conclude that the cooling was caused by a reduction in northward heat transport during a slowdown in the Atlantic Meridional Overturning Circulation (AMOC) as a result of the catastrophic meltwater drainage of large glacial lakes during the last deglaciation (Broecker, 1987; Björck et al., 1996). Observations of large decreases in sea-surface salinity (SSS) and SST in the Laurentian Channel indeed indicate enhanced drainage via the St. Lawrence River valley at the time of the YD (Levac et al., 2015).

Despite strong evidence for meltwater and reduced AMOC as the main trigger for the YD, the fact that the first signs of the YD are associated with changes in wind patterns (Brauer et al., 2008) or changes in the source area for precipitation (Steffensen et al., 2008) lead to the conclusion that atmospheric circulation was responsible for increased melting, which then lead to reduced AMOC and large scale cooling (Bakke et al., 2009). Occurrence of YD induced climatic cooling around the globe (Carlson, 2013) is also seen as evidence that atmospheric circulation is involved as a mechanism to transfer the cooling to regions beyond the North Atlantic. Climate simulations matching observed paleoclimatic proxy data strongly suggest that the YD resulted from a combination of a weakened AMOC, an altered atmospheric circulation and moderate negative radiative forcing caused by elevated atmospheric dust load and lower greenhouse gases levels (Renssen et al., 2015). Support for these hypotheses thus requires comparison of atmospheric, terrestrial and marine records.

Inferences regarding lead and lag times of the reaction of the ocean versus that of the atmosphere, even with the best age models, are plagued by challenges in correlation of different records, each with its own uncertainties. For instance, deep-sea coral records suggest that the YD could have started in as little as 5 years but the average uncertainty of $^{230}Th/^{234}U$ chronometric method used to date the corals is 87 years (Smith et al., 1997). Marine records radiocarbon dated with carbonate fossils face problems of unresolved reservoir ages (Austin et al., 2011) and the YD period is concurrent with a plateau of almost unchanging radiocarbon dates (Reimer et al., 2009) hence making the dating of samples of YD age uncertain and further complicating correlations between records. A few ocean sediment records from eastern Canada were

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analyzed for both terrestrial and marine proxies, eliminating the problems associated with correlation of dating models (Anderson et al., 2007; de Vernal et al., 1993, 1996; Levac, 2003; Levac et al., 2011, 2015), but none presents reconstructions for both air and sea-surface temperatures. Our high resolution palynological (pollen and dinoflagellate cysts) record from Newfoundland's Bay of Islands both avoids the uncertainties associated with correlations of different records and provides a paired record of air temperature and sea-surface conditions.

In Eastern Canada, the YD is recognized on the basis of vegetation reversals and major lithological changes in lake deposits that are consistent from site to site (Mott et al., 1986; Mayle et al., 1993; Levesque et al., 1997). Radiocarbon dates for the onset and the end of the event (Mott et al., 1986; Mayle et al., 1993; Mayle and Cwynar, 1995), are consistent with the traditional chronozone boundaries of Northwest Europe (ca. 11,000 to 10,000 ^{14}C yr BP; Mangerud et al., 1974). Following calibration, the average dates for the onset and end of the YD in Eastern Canada are 11,000 and 12,900 cal yr BP, and in this paper we will refer to this specific time interval as the *YD period* (more details in the methods) to allow direct comparisons between different proxy records. The use of the term *YD period* follows similar practice by various authors (Rasmussen et al., 2006; Brauer et al., 2008; Renssen et al., 2015) even though this does not respect formal definitions of a period. Because the YD was linked to drainage of meltwater from glacial lakes via the St. Lawrence River (Levac et al., 2015) it should be detected in the Bay of Islands salinity record. Another advantage of the record from the Bay of Islands is its location near an ecotone so the vegetation should be sensitive to climate changes associated with the YD (Peteet, 2000).

Another difficulty in the comparison of event timing between proxy records is the lower temporal resolution of most Atlantic and Newfoundland records. The average thickness of the YD sediments in most cores from Atlantic Canada and Newfoundland is around 50 cm while it is 250 cm in Bay of Islands core MD99-2225. Even with multiple ^{14}C dates, when the error margins for these dates are considered, dates could “overlap” as noted by Mayle and Cwynar (1995). Here we present results of palynological analysis performed at a high resolution for the downcore depth interval 20 to 25 m, i.e. slightly before and after the YD.

Paleoclimatic studies suggest that changes in atmospheric circulation are needed to transfer the climatic signal from the YD inland and affect vegetation in locations beyond coastal regions (Shuman et al., 2002; Rach et al., 2014; Renssen et al., 2015) implying there will be a lag between ocean and land, i.e. between the start of the meltwater forcing and the shifts in vegetation. In our study we assess whether there is a lag between the onset of changes in the marine environment and changes in vegetation at a coastal location, where we should expect air temperatures to be affected more rapidly by the cooling in sea-surface than air temperatures at inland locations. At present, since lake records from Eastern Canada have not been correlated with nearby paleoceanographic records, it is possible that proxies within the easily recognizable YD period in lakes might only represent the full expression of the event's impacts, i.e. the conditions after the cooling is fully established, lake productivity has dropped and local vegetation has changed, perhaps with reduced cover, making watersheds subject to greater erosion. Our record will enable the assessment of the duration of the lag (if there is one) between the drainage from Lake Agassiz and changes in vegetation.

2. Study area

The Bay of Islands is located on the west coast of Newfoundland (Fig. 1). Lowlands around the Bay are surrounded by the Long Range Mountains, which protect them from cold winds and the area experiences a warmer and drier climate than the rest of Newfoundland (Banfield, 1983). The average air temperature at nearby Corner Brook is −7.2 °C in February and 16.9 °C in August (Environment Canada, 1989).

Local vegetation is dominated by *Abies balsamea*, accompanied by *Pinus strobus*, *Picea mariana* and *Picea glauca*, as well as small stands of shrub *Betula* spp. and *Populus tremuloides* (Damman, 1983). Three tree species reach their northern limit here: *Pinus strobus*, *Acer rubrum*, and *Populus tremuloides*. This is also the only ecoregion on Newfoundland where *Fraxinus nigra* is present, and where *Acer spicatum* is resilient enough to compete with other species. *Alnus rugosa* thickets are common, and *Salix* thickets are sporadic (Damman, 1983).

The area of Bay of Islands is ~200 km^2, and its water depths reach 200 m. Shallow sills (40 to 100 m deep) are located at the entrance of the bay (Fig. 2). SST and SSS are influenced by water exchanges between the bay and the Gulf of St. Lawrence (Levac, 2003). The average SST is −0.4 °C in February and 16.7 °C in August, while SSS is 32 psu in February and 30.65 psu in August (NODC, 1994). Sea ice is present in the Bay for 2 months a^{-1} on average, typically between the end of January and the end of March (Drinkwater et al., 1999) and rivers are frozen from mid-December to mid-April (Banfield, 1983).

Core MD99-2225 was extracted in 1999 by the research vessel *Marion Dufresne* from Humber Arm (48°59.88′N; 58°05.08W), a fjord extending from the east side of the bay, at a water depth of 104 m, about 14 km from the Humber River (Fig. 2). This location was chosen because of its close proximity to the mouth of the Humber River to provide a high sedimentation rates which enable higher temporal resolution compared to many other oceanic records. Bay of Islands was also selected because it was deglaciated relatively early with a record starting before the Younger Dryas.

Deglaciation of the area around the Bay of Islands started ~14 ka when relative sea level was higher due to isostatic depression, and deglaciation was accompanied by a marine incursion. The start of the marine incursion, i.e. the highest level, is not dated. However, marine deposits located 50 m above present sea level at Goose Arm (northern part of Bay of Islands) date to 15,419 cal yr BP (Batterson et al., 1993) giving a minimum age to the marine incursion into Bay of Islands and deglaciation of the area.

3. Methods

3.1. Core description and dating

The section of core MD99-2225 studied here (20 to 25 m) is composed of bioturbated dark grey to black clayey silt (Hillaire-Marcel et al., 1999). Radiocarbon ages are available, two based on hinged mollusc shells at 1440 cm and 2166 cm (Table 1) and two based on pollen grains at 1240 and 1260 cm (Neulieb et al., 2013; Table 1). These dates were calibrated with CALIB 7.1 MARINE13.14c calibration datasets (shells) and INTCAL13 (pollen) (Stuiver et al., 2014; Reimer et al., 2013). All ages are reported in calibrated years before radiocarbon present (1950 CE) or cal yr BP.

A Delta R value of 150 ± 42 was used for both marine dates (McNeely et al., 2006). A sea ice correction (see Bard et al., 1994) of 145 ± 36 years was used for the date at 2166 cm. The error for this sea ice correction comes from the difference between this and a value obtained for an added 1.1 month, the standard error of the sea ice dinoflagellate transfer function (Levac et al., 2015).

Additional dates obtained via correlation with nearby onshore lake cores were used to better constrain ages in core MD99-2225, based on similarities in pollen diagrams or pollen zone boundaries (see Results section and Table 1) and were also calibrated with the CALIB 7.0.2 program (Reimer et al., 2009) using the INTCAL13 calibration dataset (Reimer et al., 2013). Based on these available dates, we defined the Younger Dryas at the Bay of Islands as the depth interval between 2350 cm and 2100 cm. The start of the YD period is defined by correlations with the pollen records (see Section 4.1) from Woody Hill Brook Pond and Southwest Brook Lake where it is dated at 11,100 ± 120 (Anderson and Macpherson, 1994) or 12,940 ± 220 cal yr BP. The end

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