Twentieth century warming in deep waters of the Gulf of St. Lawrence: A unique feature of the last millennium

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[1] The impact of human activities on Earth's climate is still subject to debate and the pattern of a sharp recent global temperature increase contrasting with much lesser variable temperatures during preceding centuries has often been challenged, partly due to the lack of unquestionable evidence. In this paper, oxygen isotope compositions of benthic foraminifer shells recovered from sediments of the Lower St. Lawrence Estuary and the Gulf are used to reconstruct temperature changes in a water mass originating from ~400 m deep North Atlantic waters. The data demonstrate that the 1.7 ± 0.3°C warming measured during the last century corresponds to a δ^{18}O shift of 0.4 ± 0.05‰, encompassing the temperature effect and related change in the isotopic composition of the corresponding water mass. In contrast, δ^{18}O values remained nearly constant over the last millennium, except for a small positive shift which we attribute to the Little Ice Age. We conclude that the 20th century warming of the incoming intermediate North Atlantic water has had no equivalent during the last thousand years.
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1. Introduction

[2] Dissolved oxygen concentrations [O_2] in bottom waters of the Lower St. Lawrence Estuary (LSLE) have decreased from ~125 to less than 65 μmol L^{-1} over the past 75 years, leading to severe and persistent hypoxia [*Gilbert et al.*, 2005]. An increase in bottom water temperatures, concomitant with the [O_2] decline [*Gilbert et al.*, 2005] may have promoted the remineralisation rate of organic matter in the water column and sediment [*Gillooly et al.*, 2001] and, hence, the magnitude of the oxygen sink. The instrumental records of [O_2] and temperature are discontinuous and postdate 1932 AD. Hence, these fragmentary records make it difficult to quantitatively demonstrate the relationship between temperature and [O_2] and impossible to know if significant variations in bottom water oxygenation occurred prior to this time. In this context, sedimentological records of δ^{18}O from benthic foraminifera may be useful to document changes in temperature and/or water mass origin beyond the existing instrumental records [*Frew et al.*, 2000].

[3] In this paper, we present oxygen isotope measurements ($\delta^{18}O_c$) in assemblages of the benthic foraminifera *Globobulimina auriculata* extracted from two cores recovered in the LSLE. One spans the last century, the other, the last millennium (Figure 1). Instrumental temperature and salinity records in the ambient water mass are used to assess the temperature and salinity impact on the $\delta^{18}O_c$ record of the last century. Finally, we infer temperature conditions during the last millennium from $\delta^{18}O_c$ data in the longer sedimentary record.

2. Regional Context

[4] The dominant topographic feature of the Estuary and Gulf of St. Lawrence is the Laurentian Channel, a submarine valley 250–500 m deep that extends over 1240 km landward from the continental shelf edge of the Eastern Canadian coast to Tadoussac (Figure 1). The circulation is estuarine and is characterized by three water layers: 1) a thin surface layer (down to 50 m) of low salinity water that flows seaward and originates from mixing of seawater with freshwater runoff from the Great Lakes, St. Lawrence River and Northern Quebec river drainage system, 2) a cold intermediate layer (50–150m), which originates from winter cooling of denser surface waters as tributary flow decreases and ice forms [*Gilbert and Pettigrew*, 1997; *Galbraith*, 2006], and 3) a warmer and saltier deep layer that flows landward and originates from the mixing of the Labrador Current Water (LCW) and the North Atlantic Central Water (NACW) at the shelf edge [*Dickie and Trites*, 1983]. The LCW and NACW have very different temperatures, salinities, and [O_2] but similar densities (Table S1 of the auxiliary material) and, thus, mix to form the St. Lawrence Bottom Water (SLBW) [*Gilbert et al.*, 2005].[1] Hence, even a small change in the proportion of the two parental water masses may result in variations in the properties of the SLBW. The SLBW is isolated from the atmosphere by a permanent pycnocline situated between 100 and 150 m water depth. Under these conditions, it gradually loses oxygen through respiration and remineralization of organic matter that settles through the water column, as it flows landward from the mouth to the head of the Laurentian Channel, i.e., from Cabot Strait to Tadoussac (Figure 1). At depths greater than 150 m, the oxygen consumed through respiration cannot even be replenished by winter convection [*Petrie et al.*, 1996]. Therefore, the oxygen levels are very sensitive to variations in the properties of the water that

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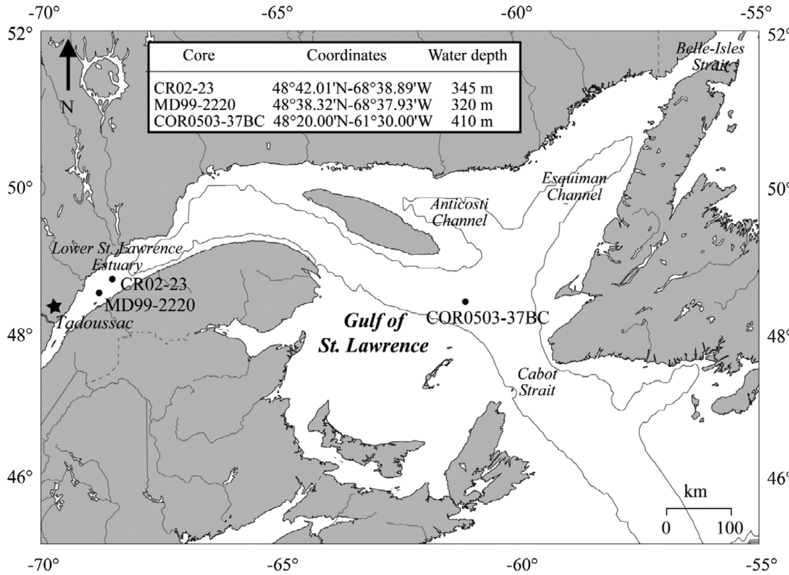

Figure 1. Map of the St. Lawrence marine system and the location of coring sites. The contour is the 400 m isobath.

enters the Laurentian Channel and the rate of organic matter remineralisation.

3. Methods

3.1. Sedimentation Rates

[5] The chronostratigraphy of the sediment cores was established based on ^{14}C, $^{210}Pb_{ex}$, ^{137}Cs, and isothermal remanent magnetizations measurements [*St-Onge et al.*, 2003; *Thibodeau et al.*, 2006]. Interpretation of these data yields an average sedimentation rate of 0.42 ± 0.04 cm/year for box core CR02-23 which, thus, spans about the last 100 years [*Thibodeau et al.*, 2006]. With a mean sedimentation rate of 0.2 cm/year [*St-Onge et al.*, 2003], the upper 200 cm of piston core MD99-2220 represents the last thousand years. The top 14 cm of sediment in core MD99-2220 was disturbed by the coring process [*St-Onge et al.*, 2003] and was of little use (see auxiliary material for detail about sediment recovery). Core COR0503-37BC sedimentation rate was estimated at ~0.15 cm/year [*Genovesi et al.*, 2008].

3.2. Isotopic Analyses

[6] In order to evaluate the temperature from the isotopic composition of foraminifers ($\delta^{18}O_c$), we used the equation of *Shackleton* [1974]:

$$t = 16.9 - 4.38^*\left(\delta^{18}O_c - \delta^{18}O_w + 0.27\right) + 0.10^*\left(\delta^{18}O_c - \delta^{18}O_w + 0.27\right)^2 \quad (1)$$

where t represents the temperature (in °C) of the water from which calcite was precipitated, $\delta^{18}O_c$ is the isotopic composition of the calcite with respect to V-PDB, and $\delta^{18}O_w$ is the isotopic composition of the ambient water versus V-SMOW. The uncertainty on temperature estimates, propagated from the analytical precision, is ± 0.2°C.

[7] We also calculated $\delta^{18}O_w$ from instrumental salinity measurements, since they are strongly correlated in the LSLE and Gulf bottom waters (>250 m; $\delta^{18}O_w = 0.67^*S - 23.1$, $r^2 = 0.86$; see Figure S2 of the auxiliary material). The standard deviation of $\delta^{18}O_w$ measurements, based on the analysis of duplicate samples, is ± 0.05‰.

4. Results and Discussion

[8] $\delta^{18}O_c$ values in core CR02-23 decrease in a stepwise manner from the base (~3.0 ‰ at 45 cm) to the top of core, and reach a minimum value of ~2.7 ‰ in the upper 3 cm (Figures 2 and S1 of the auxiliary material). In core MD99-2220, $\delta^{18}O_c$ values below 28 cm depth average 3.09 ± 0.08 ‰ (Figure 3). These values are higher, and are constrained within a far narrower range than those recorded in the upper 28 cm (3.0 to 2.6 ‰) of cores MD99-2220 and CR02-23 (Figure 3). In addition, a set of heavier values averaging 3.15 ± 0.05 ‰. (n = 36) is recorded between 55 and 90 cm in core MD99-2220 (n = 36). This set of values is significantly heavier than those recorded within the 28 to 55 cm interval and below 90 cm (see Figure 3).

[9] Core CR02-23 encompasses the last century, allowing a comparison with historical, instrumental temperature data. $\delta^{18}O_c$ values were converted into temperatures using equation (1), assuming, as a first approximation, a constant isotopic composition of about −0.05 ± 0.04 ‰ for the bottom water mass. This value has been retained as a first estimate, based on seven $\delta^{18}O_w$ measurements carried out in 2007 and 2009, near the study site , which yielded values of −0.04 ± 0.03 (n = 6) and −0.11‰ (n = 1), respectively. As illustrated in Figure 2, the calculated and measured temperature curves show very similar patterns but the magnitude of the temporal temperature variation estimated from isotopic data is +1.3 ± 0.2°C (Figure S1) *vs* +1.7 ± 0.3°C, based on instrumental data [*Gilbert et al.*, 2005]. This suggests that part of the temperature-driven shift in $\delta^{18}O_c$–values may have been masked by a small positive shift in $\delta^{18}O_w$–values. In the observed temperature range, one would expect a nearly constant $d\delta_c/dt$ relationship of about −0.3‰/°C (from equation (1)). The actual 1.7°C

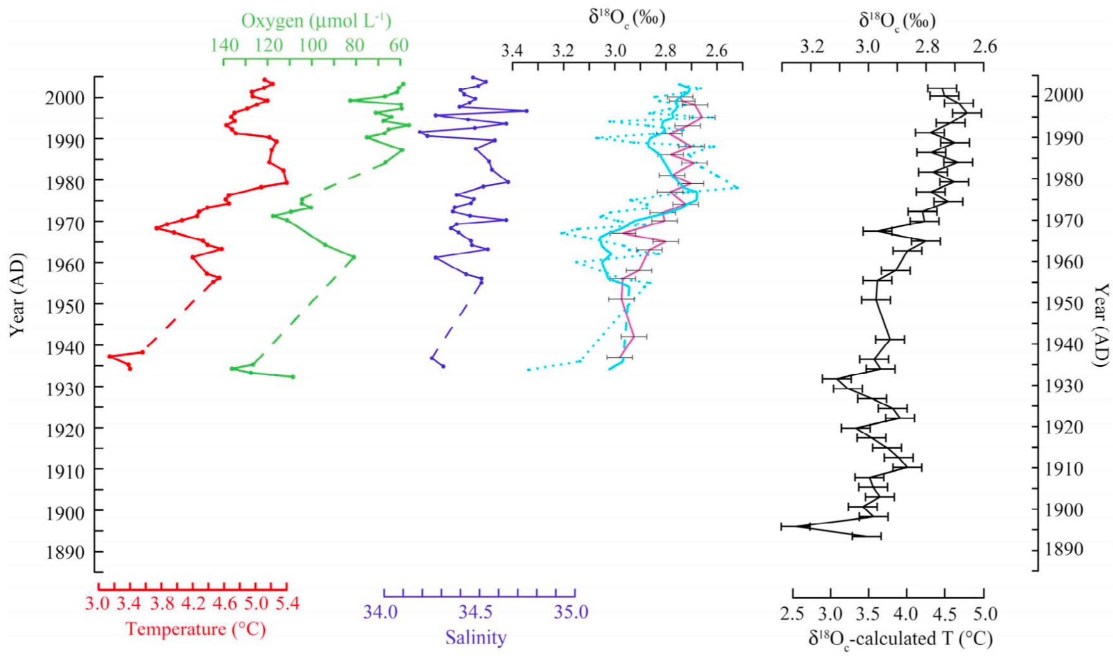

Figure 2. Available instrumental temperature (red), oxygen (green) and salinity (dark blue) records for the St. Lawrence bottom water at a depth >300m against age (from the CLIMATE database). Theoretical $\delta^{18}O_c$ (dotted light blue) and smoothed theoretical $\delta^{18}O_c$ (solid light blue) were calculated from the relationship between salinity, $\delta^{18}O_w$, and the instrumental temperature in the isotope paleotemperature equation (1) and is plotted beside $\delta^{18}O_c$ measured in core CR02-23 (purple). $\delta^{18}O_c$ measured in core CR02-23 was used to estimate paleotemperature with the equation (1) (black). Error bars report uncertainty for the estimated temperature calculation (±0.2°C), propagated from the analytical precision of the $\delta^{18}O_c$ measurements (±0.05 ‰), and for the analytical precision itself.

increase in bottom water temperature should thus have induced a shift of about −0.5‰ in the $\delta^{18}O_c$–values of *G. auriculata* shells. The measured shift (−0.3‰) thus suggests a +0.2‰ increase in $\delta^{18}O_w$ values to account for the difference. *Gilbert et al.* [2005] proposed that the recent increase in SLBW temperature results from the decrease of the proportion of the cold and well oxygenated LCW relative to the warm and less oxygenated NACW in the water

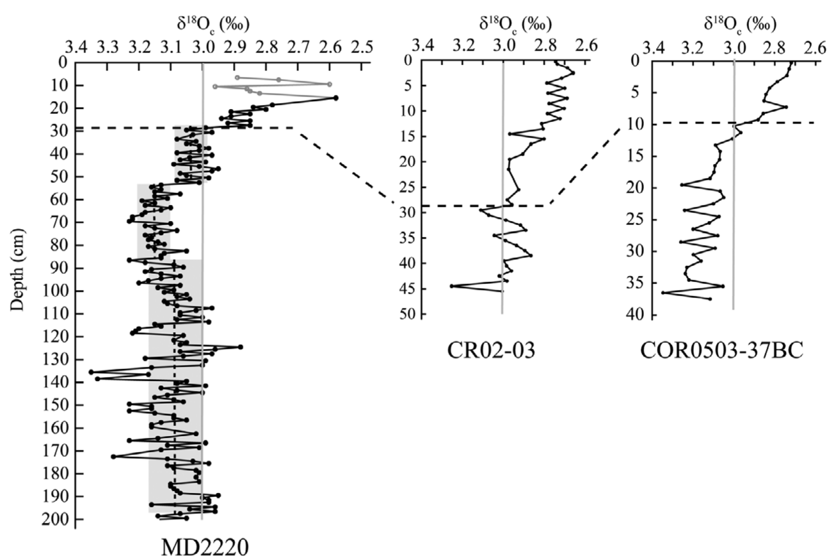

Figure 3. $\delta^{18}O_c$ vs. depth in cores MD99-2220, CR02-23 and COR0503-37BC. The dashed line is the +3 ‰ threshold in $\delta^{18}O_c$–values. In core MD2220, the vertical dashed lines represent the mean values for different sections of the core and the shaded zones correspond to the standard deviation (1 sigma). Below 28 cm, 3 sections are defined based on the $\delta^{18}O_c$ values (see auxiliary material, section 5). The top 14 cm of sediment in core MD99-2220 is shaded since it was disturbed by the coring process and was of little use [*St-Onge et al.*, 2003].

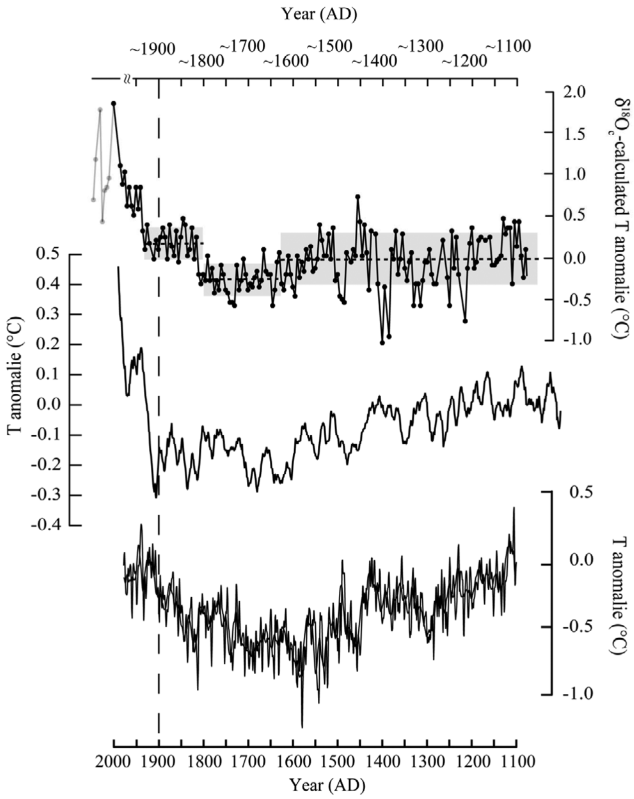

Figure 4. (top) $\delta^{18}O_c$ in core MD99-2220 vs. age in cal. years BP. Before ca. 1920 AD, 3 sections are defined based on statistically significant $\delta^{18}O_c$–calculated temperature anomalies from the mean pre-15th century value (see auxiliary material, section 5). (middle and bottom) Northern Hemisphere atmospheric temperatures anomalies based on tree ring, ice core, and instrumental records reconstructed by *Crowley* [2000] and *Moberg et al.* [2005].

mass that feeds the Laurentian Channel. Using temperature and salinity as tracers, *Gilbert et al.* [2005] estimated that the proportion of NACW in the water mass entering the Laurentian Channel trough Cabot Strait increased from 28 to 48% relative to the LCW between 1930 and 2003 AD. According to recent measurements, the two water masses have $\delta^{18}O_w$–values of about −0.5 (LCW) and +0.5 ‰ (NACW) [*Khatiwala et al.*, 1999]. The $\delta^{18}O_w$ values measured in the LSLE (~−0.05‰) is consistent with Gilbert et al.'s estimate of the relative contribution (~50:50) of the two water masses to the modern SLBW. The increasing proportion of NACW in this water mass, from ~28% to 48%, since the 1930s, should have resulted in a ~+0.2‰ shift of the SLBW $\delta^{18}O_w$ values over the same period. This shift in SLBW δ^{18}Ow value would account for the discrepancy between the temperatures inferred from the measured $\delta^{18}O_c$–record and the instrumental data since the 1930s. This conclusion rests on the assumption that the isotopic compositions of the NADW and LCW have been invariant since 1930. It should also be considered with caution given the small isotopic offsets involved. In order to evaluate the effect of a variation in $\delta^{18}O_w$ on the estimated temperatures, we also calculated a theoretical $\delta^{18}O_c$ from equation (1), using our estimates of $\delta^{18}O_w$ based on salinity measurements and instrumental temperature data (Figures 2 and S2 of the auxiliary material). The resulting profile is similar to the measured $\delta^{18}O_c$ in CR02-23, suggesting that the $\delta^{18}O_c$ record can be used as a rough paleothermometer in the SLBW. Therefore, irrespective of the offset value, a significant warming (>1°C) during the course of the last century is indisputable. This is confirmed by a similar trend, with a comparable amplitude, observed at another site of the Laurentian Channel in the Gulf of St. Lawrence (Figure 3) [*Genovesi et al.*, 2008]. The similarity of $\delta^{18}O_c$ records from the three sites in the Estuary and Gulf of St. Lawrence demonstrates that the warming trend of the last century is a regional feature. In contrast, prior to the last 100 years, the isotopic record from core MD99-2220 is nearly invariant, with a mean $\delta^{18}O_c$ value of 3.09 ± 0.08 ‰ below 28 cm (Figure 3), i.e., for the time interval spanning from about 1000 to 1900 AD. However, a possible feature of significance is the set of heavier values recorded in the 55 to 95 cm interval (i.e., between ~1630 and 1800), peaking at about 70 cm (~1740). It suggests a cooling of less than 0.5°C, possibly linked to the Little Ice Age (LIA) (Figure 4). Paradoxically, data from cores collected in the area of the Laurentian fan, at the outlet of the Laurentian Channel in the North Atlantic, indicate sea-surface warming during the LIA [*Keigwin and Pickart*, 1999]. This has been interpreted as a northward shift of the slope water current in response to a dominant negative North Atlantic oscillation mode. The hydrography in the area of the Laurentian Fan is complex, as it is located at a front marked by the mixing of three water masses (the Labrador Sea Current, the Labrador Sea Water, and the North Atlantic Drift). Any change in the strength and trajectory of any or several of these water masses may have resulted in changes of surface water characteristics. The hypothesis of *Keigwin and Pickart* [1999] may be correct for surface waters south of 44° but it does not necessarily apply to a water mass collected below 400 m in the NW North Atlantic and carried into the SLBW.

[10] Following the interval which we associate with the LIA, the warming of the SLBW appears to have occurred in two main steps. The first one is tenuous (~0.3°C) but, nonetheless significant. It started at the beginning of the 19th Century and is consistent with the Northern Hemisphere compilation of climate changes by *Moberg et al.* [2005] (Figure 4). This early warming could reflect the recovery from the LIA, although one may argue that it results from anthropogenic forcing. The second warming phase started at the turn of the 20th Century and is more pronounced, >1°C over the last century. Such a warming is seen in a large array of paleotemperature records [e.g., *Crowley*, 2000]. In the SLBW, the warming likely results from the increased temperature of North Atlantic waters entering the Gulf of St. Lawrence through Cabot Strait. Hence, it would reflect either the change in proportion of parent water masses or the distal effect of heat accumulation in constituent waters masses or both. Whether or not the post-1900 warming is due to anthropogenic forcing is a matter of debate, which we do not address here. Nonetheless, irrespective of the precise mechanisms responsible for the temperature variations reconstructed from core MD99-2220, it is unquestionable that the last century has

been marked there by a warming trend having no equivalent over the last millennium.

5. Conclusion

[11] The isotope records of benthic foraminifer shells in sediment cores recovered from the Laurentian Channel in the Lower Estuary and Gulf of St. Lawrence confirm that the last century was marked by a progressive increase in bottom water temperatures, which could have contributed to the establishment of hypoxia [cf. *Gilbert et a*l., 2005]. A warming trend was previously reported by *Keigwin and Pickart* [1999] for the last 3 or 4 centuries, based on sedimentary time series from the Laurentian Fan on the continental shelf and slope. It appears to be a consistent feature of many proxy climate records of the northern Hemisphere [e.g., *Moberg et al.*, 2005]. This warming might originate from anthropogenic activities but it could also be superimposed on a longer timescale natural oscillation that has not yet been captured in our records. To elucidate the origin of this warming, longer time series of $\delta^{18}O$ in benthic foraminifer shells should be compared, notably, with proxies of overlying water oxygenation, such as the vertical distribution of redox-sensitive metals [e.g., *Sundby et al.*, 2004] and benthic foraminifer assemblages [e.g., *Gooday*, 2003; *Gooday et al.*, 2009]. This would possibly better document the relationship between temperature, bottom water oxygenation, and their linkages with either natural, long-term variations in the North Atlantic thermohaline circulation and surface currents or with the current, anthropogenically-driven global warming.

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