

Deep-Water Flow over the Lomonosov Ridge in the Arctic Ocean

M.-L. TIMMERMANS, P. WINSOR, AND J. A. WHITEHEAD

Woods Hole Oceanographic Institution, Woods Hole, Massachusetts

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ABSTRACT

The Arctic Ocean likely impacts global climate through its effect on the rate of deep-water formation and the subsequent influence on global thermohaline circulation. Here, the renewal of the deep waters in the isolated Canadian Basin is quantified. Using hydraulic theory and hydrographic observations, the authors calculate the magnitude of this renewal where circumstances have thus far prevented direct measurements. A volume flow rate of $Q = 0.25 \pm 0.15$ Sv ($\text{Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$) from the Eurasian Basin to the Canadian Basin via a deep gap in the dividing Lomonosov Ridge is estimated. Deep-water renewal time estimates based on this flow are consistent with ^{14}C isolation ages. The flow is sufficiently large that it has a greater impact on the Canadian Basin deep water than either the geothermal heat flux or diffusive fluxes at the deep-water boundaries.

1. Introduction

The two main basins of the Arctic Ocean (Fig. 1), the Eurasian and Canadian Basins, are separated by the 1700-km-long, 20–70-km-wide Lomonosov Ridge with a mean depth around 2000 m. The latest International Bathymetrical Chart of the Arctic Ocean (IBCAO) (Jakobsson et al. 2000) indicates a section of the Lomonosov Ridge where deep-water exchange between the Canadian and Eurasian Basins is possible. This S-shaped gap is located near 88.7°N, 156.0°E with a sill depth of 2400 m where it is approximately 20 km wide. However, the exact bathymetry is still poorly known, and the actual sill depth and width must be assigned error bars of ± 100 m and ± 2.5 km, respectively (Jakobsson et al. 2000), with a further discussion of errors given in Jakobsson et al. (2002).

The Canadian Basin deep water is warmer and saltier than that of the Eurasian Basin (Fig. 2); this was attributed first by Worthington (1953) to the presence of a submarine ridge, now known as the Lomonosov Ridge. The relative freshness of the Eurasian Basin deep waters is likely linked to the exchange of these waters with

the Norwegian and Greenland Seas (Aagaard et al. 1985).

In a trans-Arctic section of ^{14}C Schlosser et al. (1997) demonstrated a large $\Delta^{14}\text{C}$ gradient between the Eurasian and Canadian Basins, clearly showing how the Lomonosov Ridge acts as an effective barrier to direct deep-water exchange between the basins. They calculated the mean isolation age of the Eurasian Basin bottom water (≥ 2500 m) to be about 250 years. This isolation age is the time that has elapsed after a water parcel leaves the surface having acquired its initial concentration via exchange with the atmosphere; it is distinct from the average time a water parcel spends in a particular deep basin. The deep waters of the Canadian Basin (≥ 2500 m) are older than those of the Eurasian Basin, with a ^{14}C isolation age estimate of 450 yr (Schlosser et al. 1997). Further, Schlosser et al. (1997) found no significant horizontal or vertical gradients in $\Delta^{14}\text{C}$ in the Canadian Basin (Makarov and Canada Basins) below 2250 m.

While the deepest water of the Canadian Basin is relatively isolated (Timmermans et al. 2003), it has been postulated that there may be inflows from the adjacent Eurasian Basin (Aagaard et al. 1985; Jones et al. 1995; Rudels et al. 2000). The flow of dense water through the S-shaped gap in the ridge, which acts as a sill, is likely to be hydraulically controlled and influ-

Corresponding author address: Mary-Louise Timmermans, MS#21, WHOI, Woods Hole, MA 02453.
E-mail: mtimmermans@whoi.edu

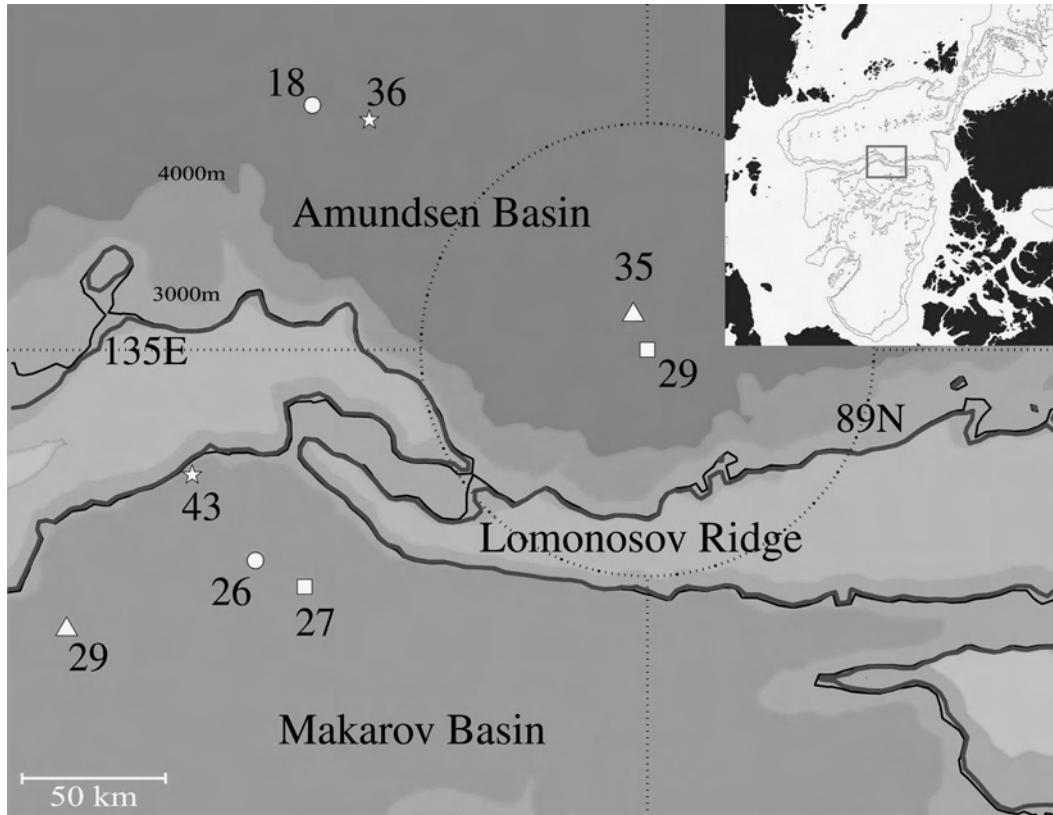


FIG. 1. Map of the Lomonosov Ridge and the Arctic Ocean (inset). The Canadian Basin consists of the Makarov and Canada Basins. The Eurasian Basin consists of the Amundsen and Nansen Basins. The detailed bathymetry (Jakobsson et al. 2000) of the Lomonosov Ridge and locations of CTD profiles taken on 1991 and 2001 *Oden* expeditions and the *Louis S. St-Laurent* AOS94 transect used in this study are shown. Matching symbols indicate cast pairs used in the volume flow-rate calculations (see Fig. 4). The thick black line is the 2400-m contour, and the thin line is 2300 m.

enced by the earth's rotation (Pratt and Lundberg, 1991; Whitehead 1998). A *weir* formula (Whitehead et al. 1974) relates volume flow rate Q of dense ($\rho + \Delta\rho$) water that lies in a deep upstream basin to a surface h_u above the sill (Fig. 3). That is,

$$Q = \frac{g'h_u^2}{2f}, \quad (1)$$

where $g' = g\Delta\rho/\rho$ and f is the Coriolis parameter ($f = 1.45 \times 10^{-4} \text{ s}^{-1}$). The side view of the channel connecting deep basins (Fig. 3) shows relatively flat density surfaces above a fixed depth with a distinct departure from horizontal in one of the basins below a certain depth. The difference from horizontal begins at the depth where the two density profiles from the upstream and downstream basins diverge, and we call this the bifurcation depth h_u . Equation (1) holds in the strong rotation limit for channels wider than the Rossby deformation radius $R = \sqrt{2g'h_u/f^2}$. Taking the upstream

fluid depth above the sill depth to be $h_u \approx 700 \text{ m}$ and $g' \approx 5 \times 10^{-4} \text{ m s}^{-2}$, we find $R \approx 6 \text{ km}$, which is less than the width $L \approx 20 \text{ km}$ of the opening at the bifurcation depth. Further, (1) is only valid where the thickness of the flow over the sill is small relative to its much larger depth in the upstream basin. However, the volume flow rate for smaller upstream depths lies within 22% of the result given by (1) (Whitehead 1989).

2. Results

To estimate the amount $\Delta\rho$ by which the flow over the sill is denser than the Makarov Basin water, we select density profile pairs from the Amundsen and Makarov Basins and compare them at the sill depth. The computed volume flow rate is shown in Fig. 4 as a function of sill depth h_s . Given the estimate and range of the depth of the gap in the ridge ($h_s = 2400 \pm 100 \text{ m}$), the volume flow rate from the Amundsen to the Makarov Basin is $Q = 0.25 \pm 0.15 \text{ Sv}$ ($\text{Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$).

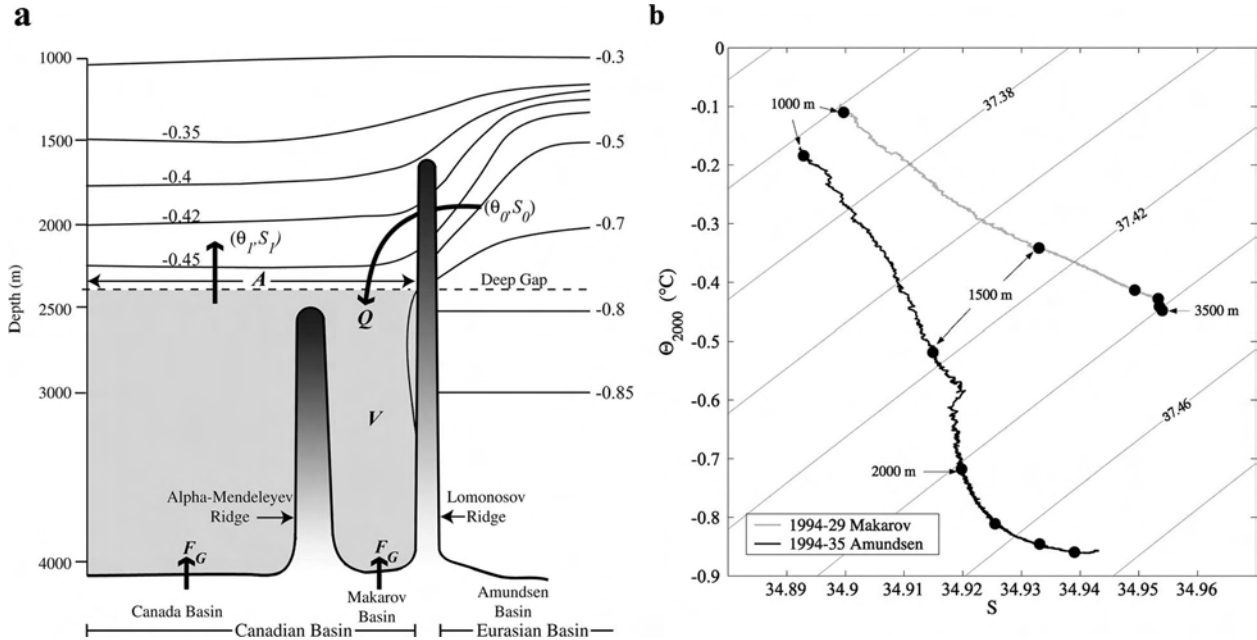


FIG. 2. (a) Arctic section of potential temperature. The depth of the deep gap in the Lomonosov Ridge is marked by the dashed line. Fluxes of heat and salt into and out of the Canadian Basin are labeled; θ_0 , θ_1 , S_0 , and S_1 are assumed to be constant. (b) Potential temperature (°C) vs salinity in the Makarov and Amundsen Basins. Lines of constant density are σ_{2000} . Circles indicate each 500-m depth change along the curves from 1000 to 3500 m.

This flow rate is comparable in magnitude to predictions derived from layered box models (0.13 Sv: Jones et al. 1995; 0.1 Sv: Anderson et al. 1999). We estimate the volume of the Canadian Basin below 2400 m (indicated by the shaded region in Fig. 2a) to be approxi-

mately $1.2 \times 10^6 \text{ km}^3$. Hence, assuming that the incoming water mixes completely with its surroundings, around 0.66% of the Canadian Basin is renewed each year. Therefore complete renewal occurs in about 150 years. This is in rough agreement with $\Delta^{14}\text{C}$ isolation

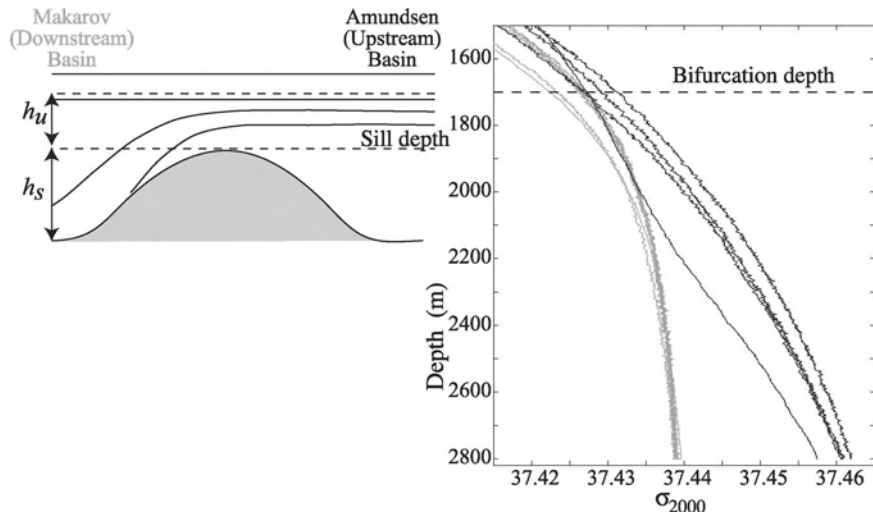


FIG. 3. (left) Schematic cross section of the Lomonosov Ridge and (right) potential density referenced to 2000 m (σ_{2000}) profiles from stations shown in Fig. 1. The σ_{2000} profiles indicate the bifurcation depth at which profiles from the up- and downstream basins diverge. The black curves are from the Amundsen Basin (upstream), and the gray curves are from the Makarov Basin (downstream).

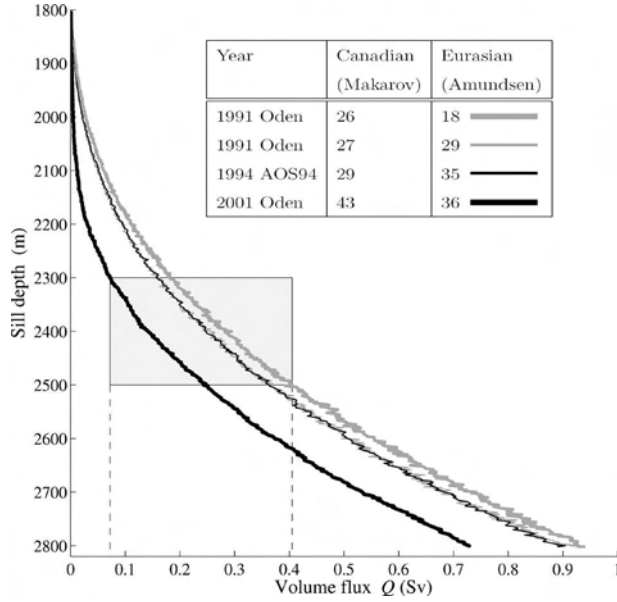


FIG. 4. Volume flow rate Q from the Eurasian (Amundsen) to the Canadian (Makarov) Basins as a function of sill depth h_s ; Q is calculated from (1) taking the bifurcation depth to be 1700 m. This depth is subtracted from the sill depth to determine h_u . Density profiles for σ_{2000} were used to calculate $\Delta\rho/\rho$; the shaded area shows the upper and lower limits of estimates for mean sill depth of the Lomonosov Ridge from detailed bathymetry (Jakobsson et al. 2000). The inset table lists cast number pairs for each year as shown in Fig. 1.

age estimates. For an isolation age of the deep Eurasian Basin of about 250 yr, the Canadian Basin water would have an isolation age of about 400 yr, comparable to the 450-yr estimate of Schlosser et al. (1997).

To assess how significant such a flow over the Lomonosov Ridge may be, we consider it in the context of heat and salt budgets for the Canadian Basin. The area of the Canadian Basin through a horizontal slice at the sill depth (about 2400 m) is $A \approx 1.6 \times 10^6 \text{ km}^2$ (see Fig. 2a). A volume flow rate Q enters the Canadian Basin with an average potential temperature $\theta_0 \approx -0.80^\circ\text{C}$ and salinity $S_0 \approx 34.925 \text{ psu}$, and leaves across the isopycnal $\theta_1 \approx -0.43^\circ\text{C}$, $S_1 \approx 34.953 \text{ psu}$ over a horizontal area A . Further, there is a geothermal heat input F_G ($\approx 50 \text{ mW m}^{-2}$) (Langseth et al. 1990) to the Canadian Basin. The potential temperature θ and salinity S of the volume V evolve according to

$$\begin{aligned} \frac{\partial\theta}{\partial t} &= \frac{Q}{V}(\theta_0 - \theta_1) + \frac{F_G A}{\rho c_p V} - \frac{\kappa_z A}{V} \frac{\partial\theta}{\partial z} \\ &= (-0.0024 + 0.0005 + 0.0001)^\circ\text{C yr}^{-1} \\ &\approx -0.002^\circ\text{C yr}^{-1} \end{aligned} \quad (2)$$

and

$$\begin{aligned} \frac{\partial S}{\partial t} &= \frac{Q}{V}(S_0 - S_1) - \frac{\kappa_z A}{V} \frac{\partial S}{\partial z} \\ &= (-0.00018 - 0.000017) \text{ yr}^{-1} \\ &\approx -0.0002 \text{ psu yr}^{-1}, \end{aligned} \quad (3)$$

where $\overline{\partial\theta/\partial z} \approx -2 \times 10^{-5}^\circ\text{C m}^{-1}$ and $\overline{\partial S/\partial z} \approx 4 \times 10^{-6} \text{ psu m}^{-1}$ are the potential temperature and salinity gradients at 2400 m in the Canadian Basin, $\rho \approx 1040 \text{ kg m}^{-3}$ is the density, and $c_p = 3900 \text{ J kg}^{-1}^\circ\text{C}$ is the specific heat capacity. We have chosen a representative turbulent diffusivity of $\kappa_z = 1 \times 10^{-4} \text{ m}^2 \text{ s}^{-1}$ (Munk and Wunsch 1998).

The largest terms in (2) and (3) arise from the inflow Q , while the geothermal heat flux into V and diffusion across A are relatively small contributions to the thermohaline evolution. We compute $\alpha\Delta\theta$ and $\beta\Delta S$, where $\alpha = -\rho^{-1}\partial\rho/\partial\theta$ and $\beta = \rho^{-1}\partial\rho/\partial S$ [$\alpha(\theta, p) = 1.2 \times 10^{-4}^\circ\text{C}^{-1}$ and $\beta(S, p) = 7.6 \times 10^{-4} \text{ psu}^{-1}$ at $\theta = -0.5^\circ\text{C}$, $S = 34.9 \text{ psu}$, and pressure $p = 3000 \text{ dbar}$], and find $|\alpha\Delta\theta| > |\beta\Delta S|$; hence, the water in the Canadian Basin is becoming denser as it cools and freshens. While the decadal salinity decrease ($\approx -0.002 \text{ psu}$) in the deep Makarov Basin would be at the accuracy of salinity determinations, this predicted deep-water cooling of the Makarov Basin over a decade (-0.02°C) should be detectable with well-calibrated CTD data between the 1991 and 2001 *Oden* profiles. In fact, we observed a warming of Makarov Basin water between 1991 and 2001, while over the same time we observed an even greater warming in the Amundsen Basin (0.04°C more than for the Makarov Basin). The apparent temperature increase could be attributed, for example, to calibration uncertainties between the Neil Brown CTD on the *Oden* 1991 cruise and the Sea-Bird 911 CTD on the *Oden* 2001 cruise. Any conclusions made based on these observations could be viewed as an over interpretation of trends in the data and, future observations, plus a comprehensive analysis of present observations, may confirm or question our hypothesis.

Our quantification of the renewal of the deep Canadian Basin, even where remoteness and extreme conditions have prevented measuring this flow directly, can now be used in the construction and testing of circulation models. It thus far appears that the deep water is being renewed continuously, largely independent of upper ocean variability. Future observational programs are necessary, including more precise bathymetry measurements of the gap in the Lomonosov Ridge, to guide such theoretical models of the processes that renew the deep water of the Canadian Basin and influence its thermohaline structure.

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