JOURNAL OF GEOPHYSICAL RESEARCH, VOL. 89, NO. C1, PAGES 637–640, JANUARY 20, 1984

Winter Mixed Layer Entrainment of Weddell Deep Water

A. L. GORDON

Lamont-Doherty Geological Observatory of Columbia University

C. T. A. CHEN

Oregon State University

W. G. METCALF

Lamont-Doherty Geological Observatory of Columbia University

Observations from the *Somov* during the US-USSR Weddell Polynya expedition show that the mixed layer below the sea ice just prior to the austral spring retreat in the 60°S Greenwich meridian region has an oxygen content of 7.4 ml/l. This is 86% of full oxygen saturation, representing an oxygen deficit, relative to full saturation, of 1.1 ml/l. The source of this deficit is believed to be a consequence of oxygen-poor (4.5 ml/l) Weddell Deep Water (WDW) entrainment by the winter mixed layer. Assuming effective cutoff of ocean-atmosphere oxygen exchange by the nearly complete snow and sea ice cover with no net impact of oxygen content as a result of biological factors, a mixing ratio of 1:3 for WDW to "beginning of winter" surface water is required to explain the end-of-winter mixed-layer oxygen content. Accompanying the WDW transfer into the mixed layer is heat transfer of approximately 7×10^3 cal/cm^2 (2.9×10^8 J/m^2) during the five winter months of sea ice coverage as well as salt transfer, which requires 34 cm of fresh water to produce typical mixed-layer salinity. During the seven ice-free months when entrainment is expected to be minor, diffusive heat and salt flux continues. A mean annual heat flux of 12 W/m^2 is suggested, with an annual demand for freshwater of 46 cm/yr. Consideration of the winter period salinity budget indicates net sea ice melting of 20 cm, which can be attributed to regional convergence of sea ice. The remaining freshwater is derived from excess precipitation and possibly iceberg melt. Oxygen undersaturation of the winter surface water suggests slightly less potential for abyssal water ventilation than might be expected from a fully saturated condition.

1. INTRODUCTION

The joint US-USSR Weddell Polynya expedition aboard the Soviet ship *Mikhail Somov* of the Arctic-Antarctic Research Institute of Leningrad [*Gordon and Sarukhanyan*, 1982; *Gordon*, 1982] permitted, for the first time, detailed observations below the Southern Ocean sea ice for investigation of the cumulative winter effects. The CTD (conductivity, temperature, depth) hydrographic stations [*Huber et al.*, 1983] were taken in nearly full ice cover conditions [*Ackley and Smith*, 1983] in the vicinity of 60°S, just east of the Greenwich meridian (see station map and hydrographic sections in *Gordon and Huber* [this issue]).

In this study we present evidence of significant winter period transfer of deep water to the surface water mixed layer and estimate the associated rate of upward heat and salt flux.

2. WINTER VERSUS SUMMER STRATIFICATION

The thermohaline stratification below the sea ice is a simple two-layer structure of a homogeneous cold, relatively fresh, mixed layer separated from the nearly homogeneous warmer, more saline, Weddell Deep Water (WDW) with an abrupt but weak pycnocline (*Gordon and Huber* [this issue]; Figure 1). The mixed layer varies in thickness from 50 to 135 m, averaging 108 m, with a mean temperature and salinity of -1.844°C and 34.287‰, respectively, which is 0.035°C above the freezing point at 1 atm. The thinner mixed layers are somewhat warmer and occur over anomalously warm WDW cells [*Gordon and Huber*, this issue]. The slight but meaningful temperature excess relative to the freezing point is believed to be a consequence of introduction of WDW heat into the mixed layer.

Summer stratification differs primarily by the presence of a seasonally warmed and slightly fresher surface layer overlying the remnant of the winter mixed layer marked by a temperature minimum [*Toole*, 1981; *Gordon and Molinelli*, 1982, plates of vertical sections and plates 196–199]. Comparison of *Somov* station 21 [*Huber et al.*, 1983] with the summer *Islas Orcadas* station 102 [*Huber et al.*, 1981]—the stations being separated by 64 km (*Somov* station 21 was obtained on November 3, 1981, at 62°20′S and 3°07′E; the *Islas Orcadas* station 102 was obtained February 2, 1977, at 62°02′S and 4°10′E)—shows the summer temperature minimum to be about 0.5°C warmer, with little salinity change relative to the winter mixed-layer base. Above the temperature minimum, the seasonally warmed surface water represents storage of 2.4 $\times 10^4$ cal/cm^2 (1.0×10^9 J/m^2) relative to the freezing point. The *Islas Orcadas* summer water salinity is lower than the *Somov* station 21 by 0.063 ‰, which requires storage of 20 cm of freshwater relative to the winter mixed layer. Two other *Somov-Islas Orcadas* station pairs, composed of stations within 100 km of each other (9-101; 2-99), indicate the same heat storage as the 21-192 pair, but somewhat different freshwater storage (16 and 37 cm, respectively). Seasonal freshwater storage in the range of 16 to 37 cm seems small in view of the approximately 1 m of sea ice and snow cover characteristic of the region as observed from *Somov* [*Ackley et al.*, 1982] prior to the spring melt. A possible solution is that some of the spring sea ice melt, rather than diluting resident surface water,

Paper number 3C1478.
0148-0227/84/003C-1478$05.00

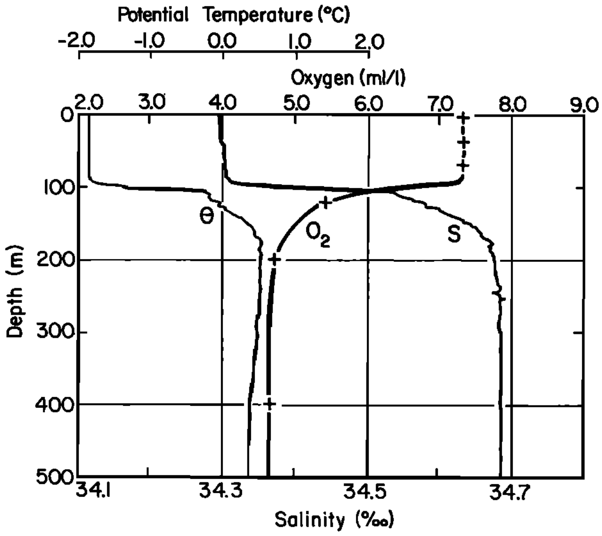

Fig. 1. Potential temperature, salinity, and oxygen stratification at *Somov* station 22 at 62°12'S, 1°05'E on November 5, 1981. The thermohaline data are derived from the CTD system. The oxygen data are from titration of water samples obtained from a rosette system. There was a 100% ice cover with few leads [*Ackley and Smith*, 1982]. The hydrographic data are reported by *Huber et al.* [1983].

mixes with the saline WDW that is diffused into the mixed layer during the spring melting period (as discussed below).

3. Mixed Layer Oxygen

The oxygen content of the entire mixed layer below the sea ice averages 7.40 (with a ±0.1 ml/l distribution about the mean) or 86% of saturation (Figure 2). This is about 1.1 ml/l below the full saturation value and is similar to the oxygen levels of the summer period temperature minimum (*Gordon and Molinelli* [1982], see temperature-oxygen relation for the vertical sections, plates 109 and 195). The WDW saturation levels are seasonally invariant near 57% to 60% of full oxygen saturation.

The origin of the mixed-layer oxygen undersaturation is likely due to entrainment of oxygen-poor WDW by the winter mixed layer. The oxygen content can be used to determine the amount of entrained WDW if we assume: (1) oxygen exchange between ocean and atmosphere is effectively blocked by the presence of a more or less complete (greater than 90%) snow and sea ice cover, (2) the net biological production and utilization of oxygen below the sea ice is zero, and (3) the mixed layer immediately prior to the ice cover is at the freezing point and at full oxygen saturation of 8.54 ml/l. Support for these assumptions is offered below.

1. The sea ice collected in the study region is dominated by frazil ice structure [*Ackley et al.*, 1983] with few brine channels. Thus the gas permeability is expected to be low [*Weiss et al.*, 1979; *Chen*, 1982]. The depletion in oxygen is probably not due to rapid oxidation of organic matter after the pack ice is formed because the surface water oxygen consumption rate is not large enough to generate a 14% oxygen undersaturation within a period of months [*Packard et al.*, 1971; *Skopintsev*, 1976; *Knauer et al.*, 1979].

2. The bacterial and biological activities are also low at the time of observation [*Marra et al.*, 1982], suggesting a low rate of oxygen production and consumption. The oxygen concentration in the remnant winter water at GEOSECS station 89 (60°01'S, 0°01'E, January 22, 1973; *Bainbridge* [1981]), which is less than 35 km from *Somov* station 33, is only 0.02 ml/l lower than our measured value, while the temperature and salinity are only 0.01°C and 0.07 ‰ higher, respectively. The excellent agreement in oxygen concentration collected in different seasons again suggests a low rate of oxygen production and consumption. Additionally, *Somov* oxygen, *p*H, calcium, alkalinity, total CO_2, nitrate, phosphate, and silica concentrations and the GEOSECS station 89 oxygen, alkalinity, total CO_2, nutrients, and ^{13}C values (*Chen* [1982]; also, see data listings of *Huber et al.* [1983]) all correlate linearly with salinity from surface down to the WDW core, which again indicates that the distribution of these chemical properties in the upper water column below the ice is predominantly controlled by conservative mixing processes with little production or utilization of chemicals [*Weiss et al.*, 1979; *Edmond et al.*, 1979; *Minas*, 1980].

3. The assumption regarding the mixed layer prior to the ice cover period is straightforward, since oxygen equilibrium is expected during the cooling period prior to the ice cover formation. Possibly some entrainment might occur before ice formation, and the resulting water does not equilibrate; however, this is reflected in the total entrainment determined below.

4. Rates of Deep to Surface Water Heat Flux

The potential temperature versus oxygen relationships for all *Somov* data points (Figure 3*a*) can be used to determine the ratio of WDW to the surface water characteristic of the period immediately prior to winter ice formation. A one-part WDW (0.5°C and 4.50 ml/l) to three parts surface water (−1.87°C and 8.54 ml/l) blend is required to produce the observed

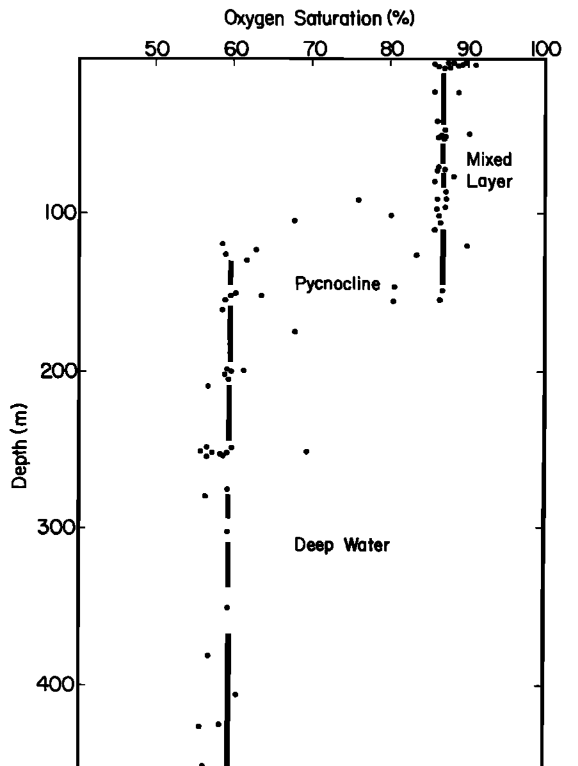

Fig. 2. Oxygen saturation from all rosette samples obtained by *Somov*. The saturation is determined by using the equation given in the *UNESCO International Oceanographic Tables* [1973].

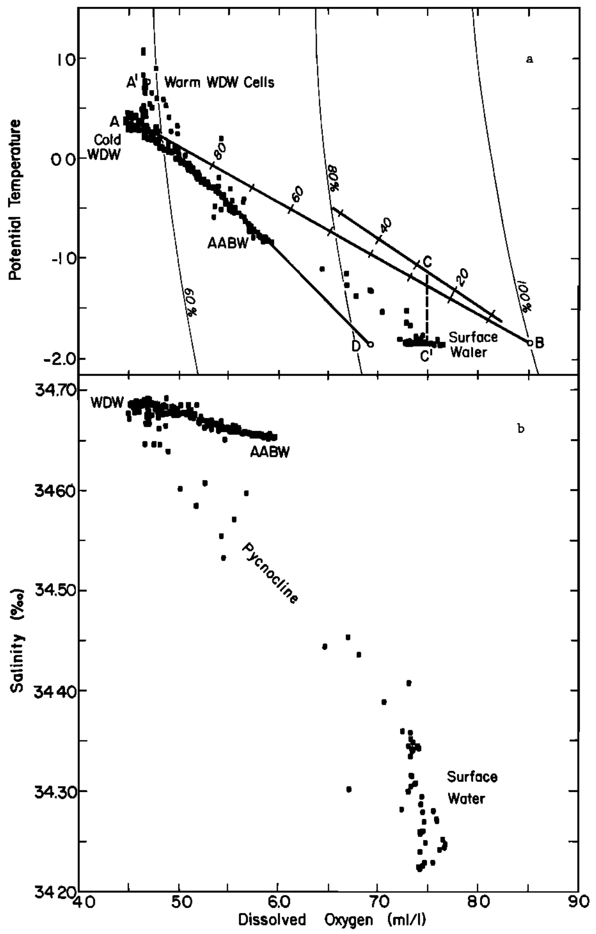

Fig. 3. Potential temperature versus oxygen content (*a*) and salinity versus oxygen content (*b*) from all rosette samples obtained by *Somov*. WDW is Weddell Deep Water and AABW is Antarctic Bottom Water. The two linear mixing lines AB and A'B are marked with percent of A water type (WDW) relative to B water type. Point D is the freezing temperature extension of the line fit to the temperature-oxygen trend of deep and bottom water colder than 0°C.

mixed-layer oxygen of 7.4 ml/l (line AB on Figure 3*a*). However, the temperature of this mixture would be 0.63°C above the freezing point. For the average mixed-layer thickness of 108 m the heat accompanying the upward transfer of WDW is 6.66×10^3 cal/cm^2 (2.78×10^8 J/m^2). If the warm WDW cells [*Gordon and Huber*, this issue] are responsible for the regionally depressed mixed-layer oxygen, the heat transfer is slightly increased (see line A'B on Figure 3*a*) to 7.93×10^3 cal/cm^2 (3.31×10^8 J/m^2). Thus the average heat flux during the 5-month ice-covered period [*Gordon and Huber*, this issue] is 20 to 25 W/m^2. If the vertical heat transfer across the pycnocline is taken as zero for the 7-month ice-free period, the average annual heat flux is 9 to 11 W/m^2.

While entrainment of WDW would cease with the onset of the seasonal pycnocline, some diffusive heat flux is expected during the ice-free part of the year. A vertical mixing coefficient K_z of 0.1 cm^2/s would accomplish a cross-pycnocline heat flux during the 7-month ice-free period of 1 W/m^2; a K_z of unity yields 10 W/m^2. Therefore, the total deep to surface water heat flux (ice cover plus ice free periods) may be in the range of 9–16 W/m^2. A value of 12 W/m^2 is suggested for a working value, reflecting a K_z of 0.5 cm^2/s.

How realistic is the winter period entrainment value and heat flux? For an annual steady state the rate of winter entrainment must balance the annual Ekman-induced upwelling of the pycnocline. The above calculation suggests an effective WDW entrainment of 27 m into the mixed layer during the winter-ice-covered period. This agrees quite well with the annual average upwelling rate of 1×10^{-4} cm/s (31 m/yr) determined from Ekman pumping calculations by using a climatically averaged wind field for the region of the *Somov* data [*Gordon et al.*, 1977].

The annual average heat flux from WDW to the mixed layer of 9–16 W/m^2 is somewhat below the annual ocean-to-atmosphere heat flux calculated from meteorological parameters. *Zillman* [1972] calculates a heat loss of 18 W/m^2 at 60°S south of Australia (Zillman's Figure 3); while *Gordon* [1981] finds 31 W/m^2 average for the 60°–70°S circumpolar belt. However, it is noted that the *Somov* results pertain to the 59°–62°S region and would be expected to be lower than the 60°–70°S average but somewhat above the Zillman value because the Weddell pycnocline is weaker than the circumpolar average [*Martinson et al.*, 1981]. In view of the uncertainty in the diffusive heat flux determined during the ice-free period and the difference in the type of calculation, agreement is reasonably good.

5. RATE OF DEEP TO SURFACE WATER SALT FLUX

The salt introduction to the surface water mixed layer by the effective entrainment of 27 m of WDW during the five winter ice cover months requires freshwater input of 34 cm to reduce the WDW salinity (34.680‰) to the annual mean surface water salinity (34.250‰, representing a 7 : 5 blend of the 34.224‰ average summer salinity from *Islas Orcadas* data with 34.287‰ winter salinity from the *Somov* data). During the ice-free period, salt diffusion across the pycnocline would place additional requirements for compensating freshwater input. Using K_z of 0.1 and 1.0 cm^2/s introduces 0.078 to 0.78 gm of salt per centimeter squared, respectively, into the surface water over 7 months, requiring a freshwater input of 2.3 to 23 cm to produce the annual mean surface water salinity.

Thus the annual freshwater input required to balance the total WDW salt flux into the surface layer is 36 to 57 cm/yr. A value of 46 cm/yr would correspond to the 12 W/m^2 best guess (K_z of 0.5 cm^2/s).

Estimates of annual freshwater flux into the 60°–70°S belt by excess precipitation over evaporation and continental runoff (glacial ice meltwater) are as high as 40 cm/yr [*Gordon*, 1981]. However, the mean snow cover on the sea ice was only 20 cm [*Ackley et al.*, 1982], which translates to only 2–6 cm of water, thus a value of 40 cm/yr may be an overestimate. Additional freshwater input could be derived if there is a net advection of sea ice into the region.

Support for this suggestion is derived by considering the approximate salinity balance for the five ice-covered months. A 1 : 3 mix of WDW to the surface water just prior to the ice cover would result in an end-of-winter salinity of 34.338‰. To reduce the salinity to the observed average of 34.287‰ requires 20 cm of sea ice (with a salinity of 5‰, *Ackley et al.*, [1982]). A meteoric source of water is unlikely during the winter since the snow on the ice is not available for release into the ocean until the spring melt.

Melting would be encouraged by the heat flux discussed above; the mixed-layer temperature slightly above freezing may be a direct evidence of this process. Sea ice melting of 20 cm requires about 20% of the winter heat flux, the rest would be lost to the atmosphere in leads and thin ice. It is probable that much of the ice melting occurs in October, when the ocean heat flux into the atmosphere is reduced from the winter maximum [*Gordon*, 1981], but the entrainment of WDW (which is due primarily to ice movement relative to the ocean; *Gordon and Huber* [this issue]) continues. The dynamic-thermodyanmic model of *Hibler and Ackley* [1983, Figure 8] suggests a net ice advection within the *Somov* region of +10 cm of accumulation in the south to 25 cm of net melting in the north. The above calculations favor the net melting value. Net melting has the effect of transporting freshwater into the area, even in the presence of Ekman divergences.

Assuming there is no heat conduction through the sea ice and that leads make up 5% of the area, the average winter heat loss in the leads amounts to about 350 W/m^2. This value is similar to estimates for Arctic Ocean heat loss within leads [*Maykut*, 1978]. It is noted that leads, though effective in removing heat, and perhaps freshwater, by evaporation, would not be effective in enhancing oxygen exchange or freshwater input by precipitation.

5. DISCUSSION

The potential temperature-oxygen scatter (Figure 3*a*) shows that for all conservative linear mixtures of WDW end members with a freezing point, surface water end members require the surface water to have oxygen saturation as low or lower than that observed from *Somov*. Conservative behavior for oxygen in the Southern Ocean is suggested by *Weiss et al.* [1979] and *Edmond et al.* [1979], so it is not likely this relationship is due to in situ oxygen consumption. Deep water colder than 0°C possesses a slightly steeper slope, indicating that the Antarctic bottom water (AABW) requires a freezing point surface water component with oxygen of 6.8 ml/l or 80% saturation (point D, Figure 3*a*), suggesting that a 35% contribution from WDW might be more appropriate for regions further poleward of the *Somov* region.

The salinity-oxygen (S/O_2) distribution (Figure 3*b*) shows that the mixed-layer points have a negative slope such that the mixed-layer water with higher salinity has slightly lower oxygen. This relation is expected from the WDW origin of the reduced mixed-layer oxygen.

Calculations of the oxygen consumption rate have frequently been made with the use of the measured apparent oxygen utilization (AOU) value and the estimated age of the water [*Jenkins*, 1980]. A hidden assumption in the calculation is that the AOU is near zero when the water was last at the surface. Obviously, this assumption needs to be modified for AABW or any water that has a component of the winter Antarctic surface water. Similarly, caution must be exercised when one calculates the age of water or the oceanic circulation rates by using AOU and the independently estimated oxygen consumption rate.

Acknowledgments. Funding for the US-USSR Weddell Polynya Expedition was received from the Division of Polar Programs of the National Science Foundation, grant DPP 80-05765 for the physical oceanography, and from the Department of Energy for the geochemistry support of Chen 81 EV 10611. Comments from Hsien Wang Ou, Bruce Huber, and John Marra are appreciated. Lamont-Doherty Geological Observatory contribution 3549.

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C. T. A. Chen, Oregon State University, Corvallis, OR 97331.

A. L. Gordon and W. G. Metcalf, Lamont-Doherty Geological Observatory of Columbia University, Palisades, NY 10964.

(Received April 28, 1983;
revised August 25, 1983;
accepted September 8, 1983.)