Antarctic Science 15 (1): 153–160 (2003) © Antarctic Science Ltd Printed in the UK DOI: 10.1017/S0954102003001159

Is thermobaricity a major factor in Southern Ocean ventilation?

MILES G. MCPHEE

McPhee Research Company, Naches, WA 98937, USA

Abstract: The Weddell Polynya, a large expanse of water that originated over Maud Rise (a bathymetric protrusion centred near 64°30'S, 3°E) and remained open during winter in the late 1970s, may have manifested a mode of deep ocean convection where despite large heat loss at the surface, sustained heat transport from below prevents lasting ice formation. In a different dominant mode (the present one), sea ice forms early in the winter and subsequently provides a thermal barrier that quickly quells incipient deep convection, thus preventing wholesale destruction of the ice cover. A possible mechanism for overcoming the thermal barrier is thermobaricity, the pressure dependence of the thermal expansion factor for seawater. An idealized, two-layer version of actual temperature and salinity profiles from the Weddell illustrates that thermobaric mixing can persist for extended periods in an ice-covered ocean, provided realistic melt rates (controlled by salt exchange at the ice/ocean interface) are specified. This furnishes a possible explanation for transient winter polynyas sometime observed in the ice-covered Southern Ocean. Thermobaricity may provide a trigger for widespread convection with possible climate impact.

Received 13 February 2002, accepted 16 September 2002

Key words: air-sea-ice interaction, deep convection, deep water formation, ocean mixing, Weddell Polynya

Southern Ocean convection modes and the Weddell Polynya

During the late 1970s a large expanse of open water (or low concentration sea ice) persisted for several seasons well within the confines of the winter sea ice limits of the Weddell Sea. The feature, which has become known as the Weddell Polynya, formed first over Maud Rise, then drifted slowly westward into the central basin of the Weddell Sea. Gordon (1991) interpreted it as manifestation of deep ocean convection far removed from the continental margin, a view bolstered by large changes observed in Weddell Deep Water (WDW) properties in the central Weddell (Gordon 1982, Foldvik *et al.* 1985). Gordon suggested that the persistent polynya indicated a mode of air-sea-ice interaction different from the present, one in which ice formation is relatively infrequent and ephemeral, even in winter.

The impact of Weddell Polynya convection on heat content of WDW in the central Weddell Sea is illustrated in Fig. 1. The National Oceanic Data Center online data base (www.nodc.noaa.gov) was perused for all available temperature/salinity profiles (meeting minimum vertical resolution and other quality criteria) in the time span 1965 to 1996 (the last year available), in a geographic region delimited by 70°S to 64°S and 40°W to 20°W. This region is the central Weddell Sea over the deep basin, away from the direct influence of both Maud Rise and the continental slopes. While coverage is relatively sparse, the impact of the Weddell Polynya in the late 1970s is obvious, as well as the decadal recovery to T_{max} values comparable to the decade before the polynya.

The amount of energy involved in water temperature changes of the magnitude indicated by Fig. 1 is huge, as shown by comparing a temperature profile observed in 1977, with nearby temperature profiles observed in later years (Fig. 2). The integrated change in sensible heat from 1977 to 1996 of the water column from 150 m depth (below the summer seasonal mixed layer) to 3000 m, is about 4.6 GJ m^{-2}. The heat extraction needed to cool the 1996 profile to the same temperature as 1977 is roughly equivalent to the amount of heat required to melt 18 m of sea ice. If the cooling from the early 1970s to 1977 (Fig. 1) was mostly from vertical exchange (cooling and ventilation of WDW), it is little wonder that the Weddell Polynya was ice free in winter. Although available hydrographical data are too sparse to assess the horizontal extent of WDW cooling in the late 1970s, satellite imagery suggested that the polynya occupied as much as 10% of the area of the Weddell Sea. The persistence of the polynya for several years indicates that once convection is established, it tends to maintain itself.

The seasonal ice cover of the Weddell Sea remains relatively thin (typically 0.6 m) after rapid initial growth in early winter, mainly because of relatively high heat flux from the underlying Weddell deep water. Over much of the Weddell, the upward heat flux averages 30–40 W m^{-2} during winter, as inferred from changes in upper ocean properties (Gordon 1991, Schlosser *et al.* 1990) and from direct measurements (McPhee *et al.* 1999) during the 1994 Antarctic Zone Flux Experiment (ANZFLUX). Despite its tenuous nature, the ice cover presents a formidable obstacle to the type of deep convection that maintained the Weddell Polynya. Essentially, the buoyancy associated with rapid ice melt is enough to prevent persistent deep convection, driven by surface processes alone. While no opening approaching

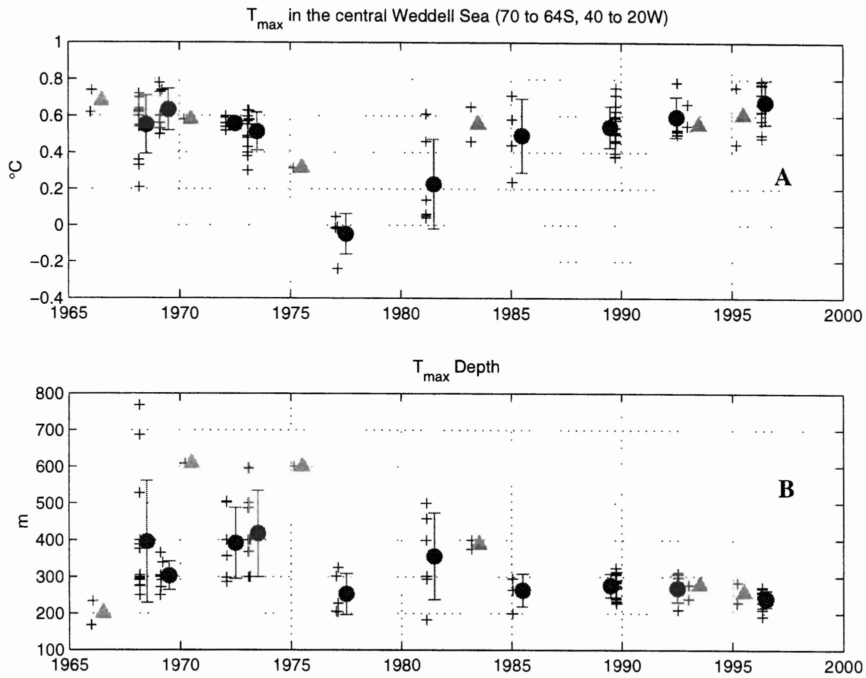

Fig. 1. a. Maximum temperature, T_{max}, and depth at which it occurs for each of 100 casts obtained meeting minimal quality criteria for the time period shown (plus symbols) in a box delimited by 70°S to 64°S and 40°W to 20°W. Circles are annual averages, with ± one standard deviation error bars, when there were three or more profiles, the triangles show the average for years when there were one or two profiles. **b.** Corresponding depth of the temperature maximum.

the extent or duration of the Weddell Polynya has since recurred, numerous examples of extensive ice opening in midwinter have been observed, particularly near Maud Rise (Drinkwater 1997).

Widespread reversion to the type of convection apparently prevalent during the polynya years could have significant climate impact. Indeed, there is evidence that air temperatures inferred from Greenland and Antarctic ice cores during the large climate swings of the Dansgaard-Oescher cycles of the last glaciation (with typical periods of about 1500 years) show temperatures in regions south and downwind of the South Atlantic to be in antiphase with temperature elsewhere on the globe (Alley *et al.* 1999). There appears to be a lag of about four centuries between cooling in the north and warming in the south (R. Alley, personal communication 2001). It is reasonable to conjecture that atmospheric warming in the south resulted from cooling and ventilation of the Southern Ocean. As deep convection in the North Atlantic slowed or stopped, deep waters of the world ocean would warm, with the signal propagating southward, eventually increasing the contrast between near surface and deep water in the Weddell, setting the stage for extensive convection. In this scenario, the Weddell polynya could represent a microcosm of the response of the Southern Ocean to abrupt climate change, with significant short-term impact on southern hemisphere climate, and long term modification of deep and bottom water in the world ocean.

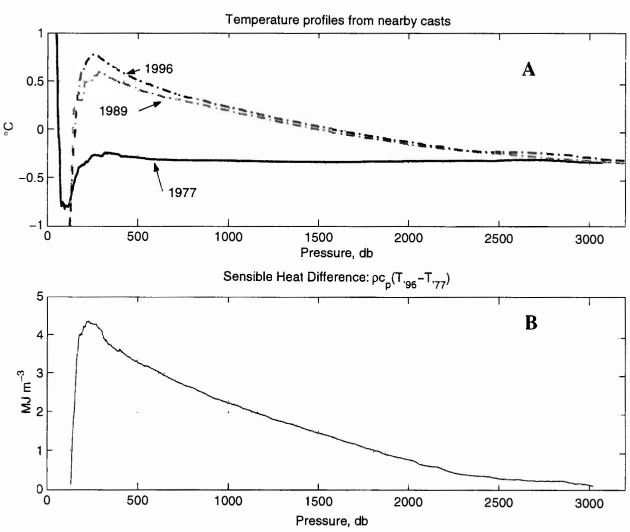

Fig. 2. a. Comparison of three temperature profiles in different years within a 70 km (meridional) x 30 km (zonal) box centred near 67°20'S and 23°W. **b.** Difference in sensible heat between nearby profiles measured in 1977 and 1996.

The intent of this paper is to examine the possibility that an important trigger for deep convection persistent enough to overcome the ice-melt buoyancy barrier, lies with thermobaricity, the depth dependence of the thermal expansion factor for seawater. Recent work has emphasized the possibility of thermobaric instability in the Weddell Sea, from both theoretical considerations (Akitomo 1999a, 1999b) and observational inference (McPhee 2000). The next section discusses both the destabilizing potential of thermobaricity, and the strongly stabilizing tendency of ice melt associated with upward entrainment of heat by convective mixing. It addresses the dynamic balance between them in terms of heat and salt transfer at the

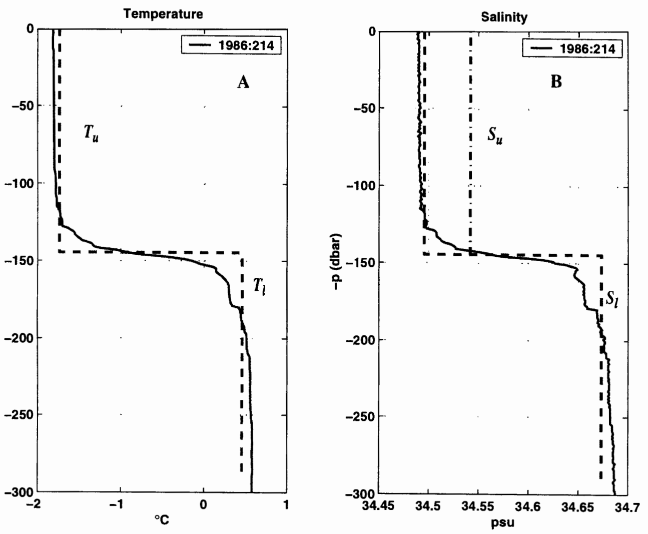

Fig. 3. a. Temperature and **b.** salinity profiles from station 1 taken on 2 August 1986 (day 214) at 65.06°S, 3.8°E (solid). Idealized two-layer upper ocean with mean values of *T/S* in layers from 0 to -145 m and from -145 to -290 m, respectively (dashed). The dash-dot line represents increased salinity in the upper layer due to about 25 cm of ice growth.

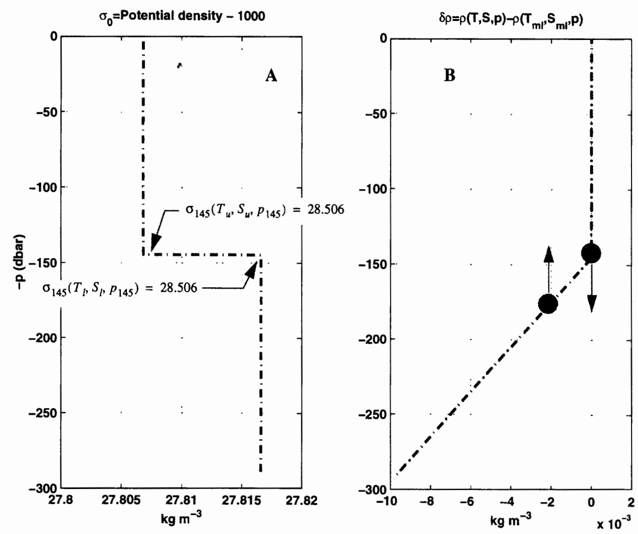

Fig. 4. a. s_0=r *(T, S, p = 0)-1000* for the idealized (salinity enhanced) profile of station 1. Densities (including pressure compressibility) of the upper and lower layer are equal at the thermocline/halocline interface. **b.** Difference between actual density and the density of an upper ocean with mixed-layer *T/S* characteristics. Displacing cooler water downward increases its density relative to ambient, whereas displacing warm water upward decreases density relative to ambient.

ice/ocean interface. The potential for thermobaric instability in historical hydrographic data from the Weddell Seas is then evaluated.

Thermobaricity in Deep Ocean Convection: heuristic arguments

By mid to late winter, static stability of the water column over large areas of the Weddell Sea is weak enough that relatively modest ice growth (sometimes a fraction of a metre) would lower the buoyancy of the mixed layer enough to allow efficient mixing at the pycnocline interface, driven either by shear generated turbulence, or by destabilizing surface buoyancy flux (Martinson & Iannuzzi 1998). But as soon as mixing begins eroding the pycnocline, heating of the mixed layer from below induces ice melt, providing strong positive buoyancy at the surface and quelling deep turbulence. The so-called "thermal barrier" (Martinson 1990) is a powerful negative feedback which limits surface driven convection whenever sea ice is present.

There is a way of extracting energy from the water column to enhance mixing at depth which does not depend on surface buoyancy flux. Cold seawater is more compressible than warm. In most of the Southern Ocean, particularly the Weddell, there is normally a large temperature contrast between an upper, well mixed layer of relatively fresh water, and warmer, more saline deep water. The pycnocline separating the two water masses is often quite thin and weak. If the density of the upper layer approaches that of the lower layer at pressure corresponding to the elevation of the pycnocline, vertical displacement of cold water downward (or warm water upward) can lead to an unstable state in which the displaced parcel is heavier (or lighter) than its surroundings, tending to reinforce the displacement by converting potential energy to kinetic, ultimately dissipated by turbulence. The effect is called thermobaricity, and particularly around Maud Rise, the water column in the Weddell seems conducive to the onset of thermobaric instability (e.g. Akitomo 1999a, McPhee 2000).

Some characteristics of thermobaricity can be illustrated by considering an idealized, two-layer rendition of an actual wintertime (1986) CTD cast from the east slope of Maud Rise (Fig. 3). Other things being equal, growth of about 25 cm of ice would increase mixed layer salinity from 34.50 to 34.54. From the standpoint of potential density (density at surface pressure), there would still exist a weak pycnocline after increasing the salinity by 0.04 psu (Fig. 4a); however, the increment was chosen so that at pressure corresponding to a depth of 145 m, the density of the upper layer water equals that of the lower layer:

$$\rho\,(\boldsymbol{T}_u, \boldsymbol{S}_u, \boldsymbol{p}_{145}) = \rho\,(\boldsymbol{T}_l, \boldsymbol{S}, \boldsymbol{p}_{145})$$

Therefore, as a starting point, we have a system with a sharp interface between layers with different T and S characteristics, but with no density contrast at the interface level. Following Akitomo (1999a), thermobaricity is illustrated (Fig. 4b) by considering the difference between actual density (including the compressibility effect of pressure) and density of an upper ocean with uniform T and S values equal to those of the mixed layer:

$$\Delta\rho = \rho(\boldsymbol{T}, \boldsymbol{S}, \boldsymbol{p}) - \rho(\boldsymbol{T}_u, \boldsymbol{S}_u, \boldsymbol{p}) \qquad (1)$$

Below the mixed layer, $\Delta\rho$ has positive slope, so a parcel of colder water displaced downward will be denser than its surroundings, while warmer water displaced upward will be lighter.

Sophisticated numerical modeling of thermobarically enhanced turbulence (e.g. Garwood *et al.* 1994, Akitomo 1999b, R. Harcourt, personal communication 2001) generally shows the convective exchange occurring in a complicated manner involving large, organized plumes. Nevertheless, it may be useful to develop a simple heuristic model of the two-layer system, assuming that a thermobaric event covers a wide enough area that on average there is a uniform deepening of the upper layer, and that there is enough ambient turbulence in the upper layer to keep it well mixed. Consider first entrainment due only to thermobaric instability at the layer interface (no surface buoyancy flux). If the layer elevation changes by Δz, and temperature and salt are conserved in the ocean (i.e. ignoring the adiabatic lapse rate in both layers), the new mixed layer characteristics are:

$$T_u' = (T_u z_{ml} + T_l \Delta z)/(z_{ml} + \Delta z)$$
$$S_u' = (S_u z_{ml} + S_l \Delta z)/(z_{ml} + \Delta z) \qquad (2)$$

For a downward interface displacement of 25 m, new mixed layer values are $T_u' = -1.415°\mathrm{C}$ and $S_u' = 34.561$ psu. At the level of the new interface (-170 m), the density of water with mixed layer properties is about 0.006 kg m^{-3} greater than water on the other side (instead of zero, as in Fig. 4b), with the thermobaric effect increasing as the layer deepens. The energy for mixing derives from the potential energy of the water column, which decreases from the initial value by about 4 kJ m^{-2} when the two layers have mixed to twice the initial mixed layer depth.

Given that the density contrast becomes more pronounced as the mixed layer deepens, what limits thermobaric mixing once it has begun? An obvious factor is that actual *T* and *S* profiles do not comprise distinct layers with discontinuous boundaries, as discussed further in section 3. Indeed, Akitomo (1999a) stresses that the temperature maximum often found in the deep water of the Weddell Sea makes it more susceptible to "Type II" thermobaric overturn than the Greenland Sea, where the increase in temperature with depth is monotonous. But even with the two-layer idealization, it is important to recognize the impact of the thermal barrier. Suppose, for example, that instead of letting the mixed layer temperature rise, we instead hold it constant by stipulating that the entrained heat go immediately and entirely to melting ice. The change in ice draft (equivalent water thickness), Δh_i, associated with a change in mixed layer thickness is then given by

$$\frac{\Delta h_i}{\Delta z} = \frac{T_l - T_u}{Q_L} \qquad (3)$$

where Q_L is the latent heat of fusion (adjusted for brine volume) divided by specific heat of seawater, with a value of about 66 kelvins for sea ice with salinity 7 psu (McPhee 1990). For no change in the layer temperatures, the ratio is constant (about 0.033 for the Maud Rise profile), meaning that each metre of mixed layer deepening can potentially melt 3–4 cm of ice. The melting introduces freshwater (i.e. upward salinity flux) at the surface tending to increase mixed-layer buoyancy. The change in salinity for the mixed layer, ΔS, is then given by

$$(z_{ml} + \Delta z)\Delta S = \left[(S_l - S_u) - \frac{T_l - T_u}{Q_L}(S_u - S_{ice})\right]\Delta z \qquad (4)$$

where the first term on the right side represents addition of salinity from below, while the second product is the reduction of salinity due to melting. For the parameters of the idealized system, the latter is seven times larger. Thus, for example, if we compare a 1-m deepening (from the initial neutrally stable condition) with no melting to one where all of the entrained heat melts ice instantaneously, the density contrast across the interface is

$$\Delta\rho_{zml-} = \rho(T_l, S_l, p_{146}) - \rho(T_u, S_u, p_{146})$$
$$= -0.3 \times 10^{-3}\ \mathrm{kg\ m^{-3}}\ \text{(no melting)}$$
$$\Delta\rho_{zml-} = +4.3 \times 10^{-3}\ \mathrm{kg\ m^{-3}}\ \text{(complete melting)}$$

where the sign change indicates that the system has gone from unstable (denser water above the interface) with no melting to stable with complete melting. For the latter, further mixing requires external addition of turbulence energy.

Given the strength of the thermal barrier effect, it appears then that the main question regarding thermobaricity changes from "what stops it?" to "does it ever occur when ice is present?" Two critical assumptions in the "complete melting" scenario (in addition, of course, to the "two- layer" assumption) are that (a) mixing in the upper layer is complete, and (b) heat transfer to the ice is instantaneous. Neither is realistic. Rapid melting severely limits turbulence scales in the ocean boundary layer (McPhee 1994), so that unless surface generated turbulence is intense, the freshwater flux is limited to the upper part of the existing mixed layer, and a new, shallower layer will form. Thus thermobaric mixing could continue at the original interface, but the (original) upper layer would no longer be mixed by surface generated turbulence. Attempting to model such behaviour with a simple layer approach is probably fruitless.

Regarding the second assumption: heat transfer at the ice/water interface is not instantaneous, but rather is rate limited by salt transfer (McPhee *et al.* 1987, Notz *et al.* 2002). We have found from observations (e.g. McPhee *et al.*

1999) that the kinematic turbulent heat flux, $\langle w'T'\rangle_0$, is reasonably well described by

$$\langle w'T'\rangle_0 = \frac{H_w}{\rho c_p} = c_H u_{*0}\left[T_u - T_f(S_u)\right] = c_H u_{*0}\delta T \quad (5)$$

where H_w is ocean-to-ice heat flux, u_{*0} is the ice-ocean friction velocity, T_f is the freezing temperature of the mixed layer, and c_H is a bulk heat exchange coefficient, approximately 0.006.

If the layer interface in the "complete melting" problem migrates at a rate $\dot{H} = \Delta z / \Delta t$, then Eqs 4 & 5 combine to provide an estimate of the friction velocity required for the interface heat flux to "keep up" with the entrainment rate, *viz.*

$$u_{*0} \geq \frac{(T_u - T_l)\dot{H}}{c_H \delta T} \quad (6)$$

For a moderate migration rate of $\dot{H} = -1\ \mathrm{m\ h^{-1}}$, combined with the other parameters of the present problem, an ice/ocean friction velocity in excess of 0.64 m s^{-1} is required, more than 20 times the largest value observed in the Weddell during ANZFLUX (McPhee *et al.* 1999). This disparity is consistent with observations that the mixed layer does warm substantially during storms by entrainment of heat from below (McPhee *et al.* 1996); i.e. that melting (and positive buoyancy flux) is severely rate limited by exchange processes at the ice/ocean interface.

The inference that ice melt cannot absorb all of the entrained heat during incipient thermobaric mixing suggests a third exercise of the simple two-layer model: one in which the upper layer in the marginally stable system is kept well stirred (say, by a powerful storm at the surface), but where the heat loss to melting ice is just sufficient to keep the density jump at the interface equal to zero, i.e. neutrally stable. We can then track the heat flux to the ice as a function of the interface migration rate. If the required heat flux is less than plausible exchange at the interface (e.g. from Eq. 5), then the thermal barrier dominates and the mixed layer will "restabilize," otherwise if the ice-ocean exchange cannot keep pace with the required melting, the system will continue to convect. This third model is as follows:

a) for each increment of interface elevation change, calculate the density excess at the upper side of the interface assuming no melting,

b) adjust the salinity of the mixed layer so as to remove the density excess, leaving the interface neutrally stable,

c) calculate the amount of ice melt needed to reduce the salinity by the necessary amount, and

d) remove the required latent heat of fusion from the mixed layer.

Pertinent equations are:

$$T_u', S_u' \text{ as in (2)}$$

$$\delta S = \Delta\rho_{\mathrm{zml-}} / \beta$$

$$S_u = S_u' + \delta S$$

$$\Delta h_i = \frac{\delta S(z_{ml} + \Delta z)}{S_u - S_{ice}}$$

$$T_u = T_u' + \frac{Q_L \delta S}{S_u - S_{ice}}$$

where ß is the saline contraction coefficient (0.81). Like the others, this version is linear in Δz. For 24 m of interface elevation change, about 5 cm of ice melt (water equivalent), requiring roughly 13 MJ m^{-2} of heat extraction, will still allow the bulk mixed layer to be neutrally stable (and thermobarically unstable) at the depth of the interface. If the interface migration (entrainment) rate is -1 m h^{-1} as before, the limiting heat flux is about 150 W m^{-2}. Values this high were measured only for a few hours at the peaks of the most intense storms during the Antarctic Zone Flux Experiment (McPhee *et al.* 1999). Thus, the important implication here is that despite the potentially strong negative feedback associated with the thermal barrier, the rate limiting (by molecular sublayers) nature of the ocean/ice heat exchange may effectively prevent the ice from melting fast enough to restabilize the layer.

The thermobaric tendency for actual *T/S* profiles in the Weddell Sea

From very simple consideration of a two-layer system, it appears that persistent thermobaric convection can occur, even in the presence of melting ice. Real potential density profiles are not discontinuous across the pycnocline, hence the level at which the pycnocline starts is typically above the level where the actual Δρ (computed according to Eq. 1) profile becomes unstable. As a generalization of Akitomo's (1999a) criterion for estimating the susceptibility of upper ocean *T/S* structure to thermobaric instability, a method was developed (McPhee 2000) which

a) computes the additional mass required to densify the water column above the level of the maximum in Dr (i.e. zmax) to thermobaric instability,

b) determines the amount of ice growth needed for additional salinity above zmax, to supply the required mass, after taking into account the density increase associated with cooling the water column above zmax to near freezing, and

c) combines the sensible heat loss from cooling and the latent heat of freezing to compute the total heat extraction needed for thermobaric instability (see fig. 7 of McPhee 2000).

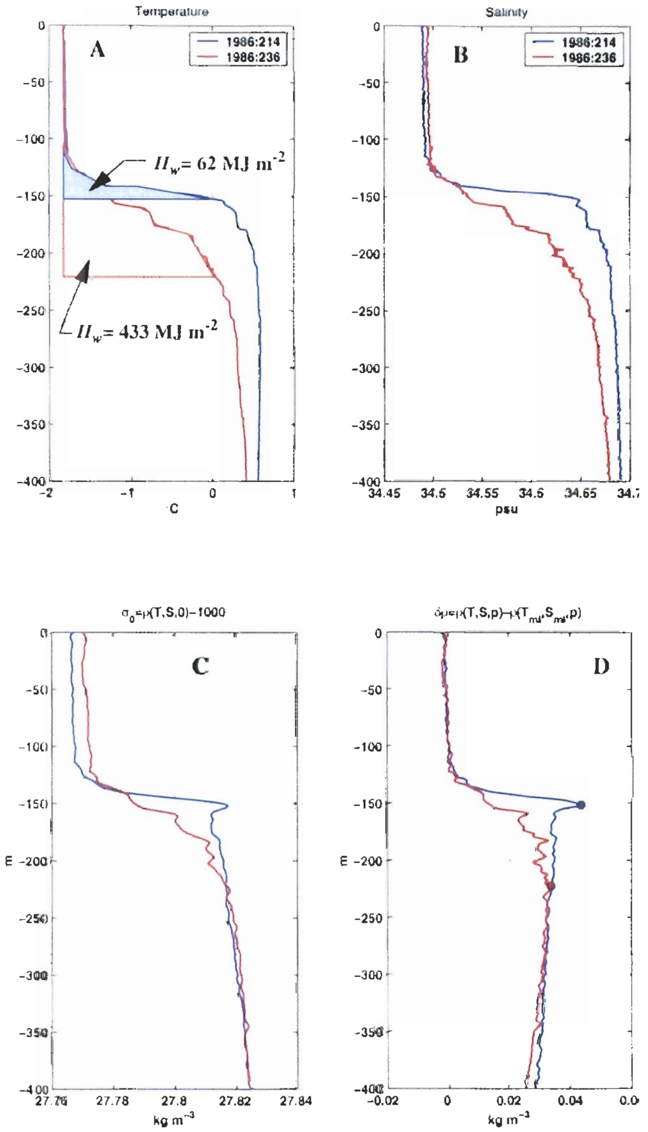

Fig. 5. Comparison of two profiles near Maud Rise: Station 1, from Fig. 3 (blue) and station 2 taken on 24 August 1986, at 64.89°S, 2.49°E (red). **a.** Temperature, with stippled areas representing the sensible heat relative to mixed layer values above the respective $\Delta\rho_{max}$ levels. **b.** Salinity. **c.** σ_0. **d** $\Delta\rho$.

This total heat extraction was called the thermobaric barrier for a particular profile. We found that the thermobaric barrier for a sizable fraction of the "yo-yo" casts measured during ANZFLUX would have been reached by the end of winter, provided the surface heat loss was comparable to that inferred from a nearby drifting buoy, and provided that all exchange processes were vertical.

Two winter profiles from 1986 (Fig. 5) illustrate the method, and also show that susceptibility to thermobaricity depends not only on the "bulk" stability (as measured say, by the difference in potential density across the pycnocline), but also on the shape of the temperature and salinity profiles near the interface. Station 1 (blue) is that discussed in the previous section (Fig. 3), while station 2 was taken on day 236 of 1986 near the centre of Maud Rise, about 70 km west of station 1. Below 200 m depth, potential density is similar, but both temperature and salinity are somewhat lower at station 2. Since the mixed layer salinity at station 2 is slightly higher, its bulk stability is lower. However, even disregarding the spike at 151 m in the σ_0 profile at station 1 (possibly due to conductivity/temperature lag mismatch), it is clear that the slope of its $\Delta\rho$ profile goes positive (i.e. $\Delta\rho$ decreasing with depth) about 71 m higher than at station 2 (Fig. 5d). Increasing the mass (and extent) of the mixed layer at station 1 so that its density (including pressure contraction) matches the density at z = -151 m requires a sensible heat loss of 62 MJ m^{-2} (shaded area in Fig. 5a) plus

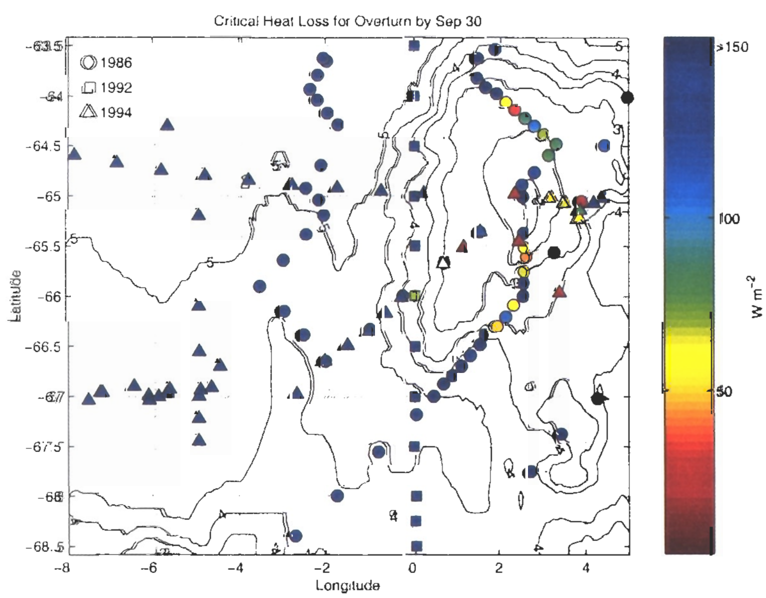

Fig. 6. Map of critical heat flux required to initiate thermobaric instability (in a one-dimensional sense) for stations archived in the online NODC data base for the winter season in the central Weddell Sea, as explained in the text. Profiles requiring in excess of 150 W m^{-2} are coloured blue, but may have much higher values. Red colours (< 50 W m^{-2}) are considered susceptible to thermobaric instability by the end of the winter. Bathymetry contour interval is 0.5 km.

a latent heat loss 65 MJ m^{-2}, associated with ice growth of 0.27 m, for a total thermobaric barrier of 127 MJ m^{-2}. Assuming somewhat arbitrarily, that winter lasts until 30 September, the average heat flux out of the ice/upper ocean system needed to accomplish the required heat removal is about 25 W m^{-2}.

By contrast, since z_{max} for station 2 is considerably lower, the sensible heat removal required to make the mixed layer density match deep water density at z = -223 m is 433 MJ m^{-2} plus 8 MJ m^{-2} from about 3 cm of ice growth, for a thermobaric barrier of 441 MJ m^{-2}. The corresponding required average heat flux from the station time to 30 September is about 138 W m^{-2}.

Our estimate of the average upward heat flux in the sea ice column from a drifting buoy near Maud Rise during a comparable period of the 1994 winter (McPhee *et al.* 1999) was about 27 W m^{-2}. Since this did not include heat loss through leads or from surface flooding, it is entirely plausible that the water column at station 1 would vent enough heat to reach thermobaric instability by the end of winter, if unaffected by advective processes. On the other hand, station 2 would be an unlikely candidate for thermobaric instability, even though its initial bulk stability was less than station 1.

Thermobaric barrier calculations were done for all of the NODC online modern CTD stations made during winter over the deep basin of the Weddell Gyre and in the vicinity of Maud Rise. A "critical heat flux" estimate was then made by dividing the thermobaric barrier by the interval between the station time and 30 September of the station year. Results (Fig. 6) show that most of the stations with critical heat flux estimates of 50 W m^{-2} or less (i.e. those most susceptible to thermobaric instability) occur on the flanks of Maud Rise, in agreement with the assessment from the 1994 data alone (McPhee 2000).

Discussion and conclusions

In the Southern Ocean south of the Polar Front, deep water below a winter mixed layer typically 100–200 m thick contains a tremendous heat reservoir which, at least in the Weddell Sea, has apparently been increasing in temperature since the late 1970s (Fig. 1). Despite weak static stability of the upper ocean, convective upwelling of enough heat to remove or prevent a seasonal ice cover over a widespread area during the entire winter has not appeared since the time of the Weddell Polynya. However, transient polynya-like features with extensive areas of low concentration ice, have been observed during August and September near Maud Rise in many years (Drinkwater 1997), indicating the possibility of a "preconditioning" for sustained deep convection there.

The persistence of the thin seasonal ice cover in the face of weak stratification and warm water in direct contact with the mixed layer (in contrast to much of the Arctic Ocean where warm Atlantic water is separated from the mixed layer by a cold, very stable halocline) is most likely to be due to the strong negative feedback provided by the thermal barrier. As soon as heat upwells in a convecting mixed layer, positive buoyancy from ice melt stratifies the upper layer, cutting off turbulent transfer of energy from the surface into the main pycnocline. Thermobaricity, and other idiosyncrasies of the equation of state for seawater*, furnish a mechanism for circumventing the thermal barrier by extracting mixing energy from the potential energy of the water column. Results from idealized scenarios in section 2 suggest that, in a bulk sense, there may exist a delicate balance between negative buoyancy induced by thermobaricity at the base of the mixed layer, and positive buoyancy from melting at the surface, which is critically rate limited by the transfer of salt at the ice/ocean interface during melting (McPhee *et al.* 1987, Notz *et al.* 2002). Estimates of melt rates (ocean-to-ice heat flux) that would allow persistent two-layer thermobaric instability (with vigorous mixing) were larger than values observed during winter storms in the Weddell, suggesting that thermobaricity can indeed overcome the thermal barrier. If convection is long-lived enough to remove the ice cover over a large enough area during winter, the thermal barrier is eliminated, and convection driven by surface cooling would continue until summer warming, or until enough ice advected into the area to again cap off the active mixed layer by melting.

Away from the Weddell Sea, there are very few wintertime CTD records in the NODC online database. For example, in the Cosmonaut Sea, where satellite observations document another large polynya in some winters (Comiso & Gordon 1996), we were unable to locate any stations between 1 June and 30 October in any year. A late winter cruise penetrated relatively deep into the Ross Sea in October 1994 (Giulivi *et al.* 1999). Profiles from that cruise (NODC cruise no. 10201, 6NP94053) also show a well defined temperature maximum (somewhat warmer than in the Weddell), but the salinity contrast between near surface water and water at the T_{max} level is considerably larger than in the Maud Rise region. We calculated the thermobaric barrier for each of the 1994 Ross Sea profiles using the algorithm of McPhee (2000). Minimum values (least thermobarically stable) were around 2000 MJ m^{-2}, which is two orders of magnitude greater than the profiles considered marginal in the ANZFLUX study. However, the area in which the Weddell Sea is considered marginally thermobarically stable is quite small, associated with the bathymetric anomaly of Maud Rise. Our search emphasizes more the general paucity of winter data in the Southern

*Arguments in previous sections depend on the full UNESCO equation of state (e.g. Gill, 1982) thus include both thermobaric and cabbeling (dependence of thermal expansion on temperature) effects in calculating $\Delta\rho$.

Ocean, rather than suggesting that thermobaricity is unimportant elsewhere. This lack of winter data presents a formidable barrier to increased understanding of important air-ice-ocean exchange processes in the Antarctic.

For profiles with similar bulk stabilities based on potential density, the thermobaric tendency is greatly enhanced by a strong thermocline, i.e. warmer water higher in the water column. Such profiles (warmer colours in Fig. 6) are found predominantly on the flanks of Maud Rise, where a "Taylor-cap" feature sitting over the seamount appears to interact with the surrounding Weddell Gyre waters (Muench *et al.* 2001). The increase of the maximum temperature in the water column over the deep basin to the west of Maud Rise (Fig. 1a), along with a decrease in the depth of T_{max} (Fig. 1b), suggests a more general rebound from the dramatic cooling associated with the Weddell Polynya that, if the trends continue, will be more conducive to the onset of thermobarically driven mixing. It is plausible that widespread warming of the deep water in the Southern Ocean could similarly lead to a mode shift in deep convection, triggered and maintained by thermobaricity. That could significantly decrease Antarctic sea ice extent, with large impact on southern hemisphere temperatures, and acceleration of deep water ventilation.

Acknowledgements

S.S. Jacobs kindly pointed out the winter Ross Sea data mentioned in the Discussion section. He and two anonymous reviewers provided helpful comments for improving the manuscript. Organizers of the Second International Conference on the Oceanography of the Ross Sea are to be congratulated for a well run and productive meeting. Research support for this work was provided by the National Science Foundation under Grants 9725478, 0090992, and 0125406, and by the Office of Naval Research under Contract N00014-02-C0393.

References

AKITOMO, K. 1999a. Open-ocean deep convection due to thermobaricity. 1. Scaling argument. *Journal of Geophysical Research*, **104**, 5225–5234.

AKITOMO, K. 1999b. Open-ocean deep convection due to thermobaricity. 2. Numerical experiments. *Journal of Geophysical Research*, **104**, 5235–5249.

ALLEY, R.B., CLARK, P.U., KEIGWIN, L.D. & WEBB, R.S. 1999. Making sense of millennial-scale climate change. *In* CLARK, P.U., WEBB, R.S. & KEIGWIN, L.D., *eds. Mechanisms of global climate change at millennial time scales. American Geophysical Union Geophysical Monograph*, **112**, 385–394.

COMISO, J.C. & GORDON, A.L. 1996. Cosmonaut polynya in the Southern Ocean. *Journal of Geophysical Research,* **101**, 18 297–18 313.

DRINKWATER, M.R. 1997. Satellite microwave radar observations of climate-related sea-ice anomalies. *Proceedings of a Workshop on Polar Processes in Global Climate, 13–15 November, 1996, Bulletin American Meteorological Society*, 115–118.

FOLDVIK, A., GAMMELSRØD, T. & TØRRESEN, T. 1985. Hydrographic observations from the Weddell Sea during the Norwegian Antarctic Research Expedition 1976/77. *Polar Research,* **3**, 177–193.

GARWOOD JR, R.W., ISAKARI, S.M. & GALLACHER, P.C. 1994. Thermobaric convection. *In* JOHANNESSEN, O.M., MUENCH, R.D. & OVERLAND, J.E., *eds. The Polar Oceans and their role in shaping the global environment. Geophysical Monograph Series,* **85**, 199–209.

GILL, A.E. 1982. *Atmosphere-ocean dynamics*. New York: Academic Press, 661 pp.

GIULIVI, C.F., JACOBS, S.S., GORMAN, E., SUTHERLAND, S. & HELLMER, H.H. 1999. *Oceanographic data beneath South Pacific sea ice. NB Palmer Cruise 9405, Sep–Oct 1994. Technical Report LDEO-99-1.* Palisades, NY: Lamont-Doherty Earth Observatory, 230 pp.

GORDON, A.L. 1982. Weddell deep water variability. *Journal of Marine Research*, **40**, 199–217.

GORDON, A.L. 1991. Two stable modes of Southern Ocean winter stratification. *In* CHU, P.-C. & GASCARD, J.-C., *eds. Deep convection and deep water formation in the oceans*. Amsterdam: Elsevier, 17–35.

MARTINSON, D.G. 1990. Evolution of the Southern Ocean winter mixed layer and sea ice: open ocean deepwater formation and ventilation. *Journal of Geophysical Research*, **95**, 11 641–11 654.

MARTINSON, D.G. & IANNUZZI, R.A. 1998. Antarctic ocean-ice interaction: implications from ocean bulk property distributions in the Weddell Gyre. *Antarctic Research Series*, **74**, 243–271.

MCPHEE, M.G. 1990. Small scale processes. *In* SMITH, W., *ed. Polar oceanography*. San Diego, CA: Academic Press, 287–334.

MCPHEE, M.G. 1994. On the Turbulent Mixing Length in the Oceanic Boundary Layer. *Journal of Physical Oceanography*, **24**, 2014–2031.

MCPHEE, M.G. 2000. Marginal thermobaric stability in the ice-covered upper ocean over Maud Rise. *Journal of Physical Oceanography*, **30**, 2710–2722.

MCPHEE, M.G., ACKLEY, S.F., GUEST, P., HUBER, B.A., MARTINSON, D.G., MORISON, J.H., MUENCH, R.D., PADMAN, L. & STANTON, T.P. 1996. The Antarctic Zone Flux Experiment. *Bulletin of the American Meteorology Society,* **77**, 1221–1232.

MCPHEE, M.G., KOTTMEIER, C. & MORISON, J. 1999. Ocean heat flux in the central Weddell Sea during winter. *Journal of Physical Oceanography*, **29**, 1166–1179.

MCPHEE, M.G., MAYKUT, G.A. & MORISON, J.H. 1987. Dynamics and thermodynamics of the ice/ upper ocean system in the marginal ice zone of the Greenland Sea. *Journal of Geophysical Research,* **92**, 7017–7031.

MUENCH, R.D., MORISON, J.H., PADMAN, L., MARTINSON, D., SCHLOSSER, P., HUBER, B. & HOHMANN, R. 2001. Maud Rise revisited. *Journal of Geophysical Research,* **106**, 2423–2440.

NOTZ, D., MCPHEE, M.G., WORSTER, M.G., MAYKUT, G.A., SCHLÜNZEN, K.H. & EICKEN, H. In press. Impact of underwater-ice evolution on Arctic summer sea ice. *Journal of Geophysical Research,*

SCHLOSSER, P., BAYER, R., FOLDVIK, A., GAMMELSRØD, T., ROHARDT, G. & MUNNICH, K.O. 1990. Oxygen 18 and helium as tracers of ice shelf water and water/ice interaction in the Weddell Sea. *Journal of Geophysical Research,* **95**, 3253–3264.