Temperature and Salinity Staircases in the Northwestern Weddell Sea

R. D. MUENCH
Science Applications International Corporation, Bellevue, Washington

H. J. S. FERNANDO
Dept. of Mechanical and Aerospace Engineering, Arizona State University, Tempe, Arizona

G. R. STEGEN
Science Applications International Corporation, Bellevue, Washington

(Manuscript received 29 December 1988, in final form 25 September 1989)

ABSTRACT

Temperature and salinity data obtained from the northwestern Weddell Sea during March 1986 reveal numerous thermohaline staircases in the thermocline separating warm deep water from the overlying colder, lower salinity winter water. Staircases in the upper, steeper portion of the thermocline were characterized by layers having vertical extents of 1–5 m. Layer thicknesses in the deeper, weaker portion of the thermocline were far greater, sometimes exceeding 100 m. The former staircases are referred to as Type A, and the latter as Type B. Vertical gradients in temperature and salinity decreased abruptly across the boundary between Type A and Type B staircase regions. Mean density ratios R_ρ were 1.52 and 1.36 over the depth intervals containing Type A and Type B staircases, respectively. Type A staircases were present at all sites sampled, whereas Type B staircases were present over approximately the central 50% of the area sampled.

Laboratory-derived results show that the observed time and vertical space scales for the Type B staircases are consistent with the notion that they are maintained by double diffusive processes. These results, combined with temperature-salinity analyses, lead us to suggest that the Type B staircase regime may have originated as a vertically convective feature within which staircases have formed and evolved continually through double diffusion. Laboratory-derived flux laws are used to estimate upward buoyancy flux due to heat flux through the Type B staircase regime of order 2×10^{-9} m^2 s^{-3}, consistent with values derived previously using oceanographic, atmospheric and sea ice data and an order of magnitude greater than computed double diffusive heat fluxes through the Type A staircase regime. The broad areal coverage of Type B staircases, coupled with previous observation of these features at scattered sites throughout much of the Weddell Sea, suggests that they are widespread there and may play a significant role in regional vertical heat transfer.

1. Introduction

Temperature and salinity data were collected in the northwestern Weddell Sea (Fig. 1) during March 1986 as part of the Antarctic Marine Ecosystems in the Ice Edge Zone (AMERIEZ) program, described in Nelson et al. (1989). These data document widespread thermohaline staircases that can be grouped, based upon their physical characteristics, into two distinct categories. The first type, which we will refer to for purposes of discussion as "Type A," consisted of T and S layers having vertical scales of typically 1–5 but occasionally as great as 8 m. These Type A staircases occurred in the steepest portion of the thermocline, between about 100 and 180 m depths, on all casts throughout the approximately 40 000 km^2 study area. Underlying these, over ~50% of the study area, were staircases made up of T and S layers having vertical extents exceeding 10 m and occasionally greater than 100 m. These, which we will refer to as "Type B," occurred in the weaker portion of the thermocline between about 180 m depths and the T maximum at 400–500 m depth.

Both Type A and B thermohaline staircases have been documented previously in the Weddell Sea during summer. (Published winter data are inadequate to determine their presence or absence.) Foster and Carmack (1976) observed both staircase types in the northern central Weddell Sea at a site several hundred kilometers east of the March 1986 field program. Middleton and Foster (1980) reported Type A staircases at a 25-hour time series CTD station in the same area as the March 1986 study region, and observed a Type B structure at a single CTD cast some 300 km to the east. Huber et al. (1981) presented an isolated group of three CTD casts showing Type B structures near 10°E within about

Corresponding author address: Dr. Robin D. Muench, SAIC, 13400B Northup Way, Suite 36, Bellevue, Washington 98005.

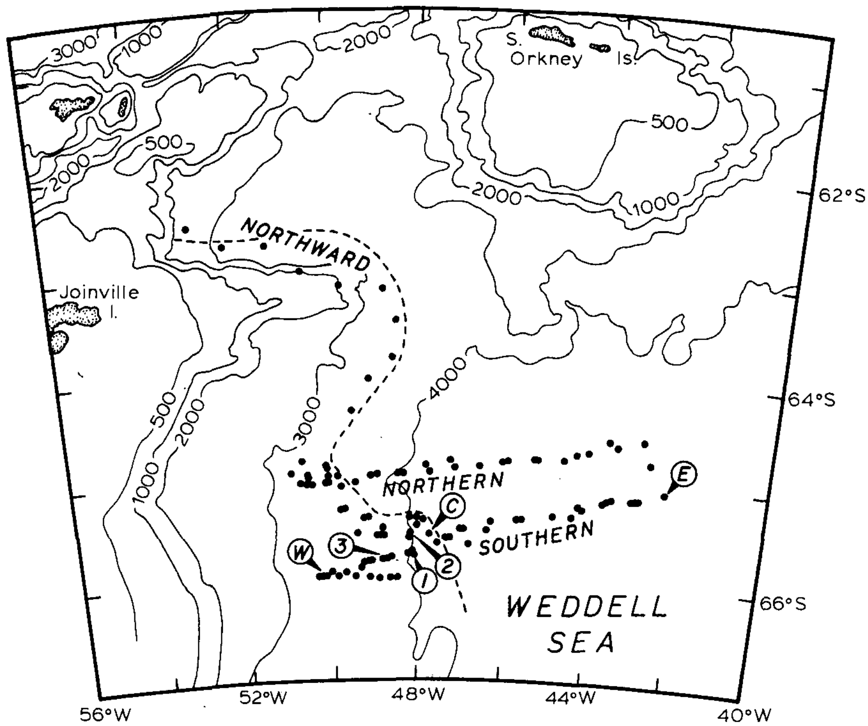

FIG. 1. Geographical location of the study area, showing positions of CTD casts occupied during the March 1986 program. Dashed line indicates approximate location of the sea ice edge. Arrows (W), (C) and (E) indicate positions of CTD casts chosen to represent western (W), central (C) and eastern (E) parts of the study area on Fig. 2. Arrows (1), (2) and (3) indicate locations for T profiles shown on Fig. 5. The "northward," "northern" and "southern" transects are labeled to correspond to usage in the text.

100 km of the Antarctic coast. Foldvik et al. (1985) observed both types of staircase south and east of the March 1986 study region. Foster (1986, personal communication) obtained data in 1980 that showed the presence of Type B staircases east of about 50°W, but not to the west. Conversely, Type B staircases were not observed in the same region during November 1987 (Foster 1988, personal communication).

The primary focus of this paper is upon Type B thermohaline staircases which were observed in the northwestern central Weddell Sea during March 1986. Laboratory based semi-empirical flux laws are used to estimate vertical conductivity in both the Type B and Type A step regions, and speculation is presented concerning their possible origins and roles in the regional oceanography.

2. The March 1986 field program

From 7 to 27 March 1986, 115 CTD (conductivity/temperature/depth) casts were made in the northwestern Weddell Sea (Fig. 1). Forty-eight casts were to 1500 m depth, and all except ten of the remaining casts were to depths between 500 and 600 m. The casts were made using Neil Brown Mark IV profiling CTD systems operated from the vessels R/V *Melville* and USCGC *Glacier*. Lowering speed was limited to about 0.5 m s^{-1} for the uppermost 200 m to reduce time-lag errors (in the worst case, spiking of S values) due to strong vertical T gradients. Lowering speed at greater depths was about 1.0 m s^{-1}.

Both CTD systems were calibrated prior to and following the field program. The *Melville* system was calibrated at Texas A & M University, and the *Glacier* system at the Northwest Regional Calibration Center in Bellevue, Washington. Additional calibration points were obtained at various depths on about every third cast during the field program. Temperatures were obtained using reversing thermometers mounted on Niskin bottles. Salinities were obtained by analyzing samples obtained from Niskin bottles using a shipboard laboratory salinometer. The data were adjusted for the calibration values, and the observations are accurate to within +0.01°C and 0.01‰.

3. Regional oceanographic background

a. The study area

The March 1986 field program extended westward from the deep basin of the central Weddell Sea onto the continental rise off the east coast of the Antarctic

Peninsula (Fig. 1). Bottom depths exceed 4000 m in the eastern portion of the study area. Depths shoal to less than 3500 m at the westernmost stations, and a single northward transect leading away from the primary study area intersected bottom depths of less than 2000 m as it crossed over the rise off Joinville Island at the end of the Antarctic Peninsula. The primary study area (excluding the northward transect, where no Type B staircases were observed) is situated some 100 km east of the continental slope off the Antarctic Peninsula and more than 300 km south of the pronounced bank that surrounds the South Orkney Islands.

Approximately the western 25% of the study area was ice-covered during the 1986 program. This western area is typically covered year-round by multiyear ice (Zwally et al. 1983), which was observed in March 1986 to have thicknesses exceeding 2 m. Active freezing was occurring in leads within the multiyear pack early in the program, and pancake ice formed several kilometers seaward of the ice edge late in the program due to off-ice wind events accompanied by low surface air temperatures. In general, however, there was no significant advance of the ice edge, as indicated both by observations from the *Glacier* and from the Navy/NOAA ice charts for the region, during the field program. The eastern three-quarters of the study area was ice-free, typical for summer. During winter, the entire region is ice-covered.

b. *Regional circulation*

The western Weddell Sea has been characterized, based upon analyses of relatively widely-spaced (compared to the March 1986 dataset) historical data, as a region of net northward flow. This flow is the north-flowing limb of a wind-driven, basinwide cyclonic gyre in the Weddell Sea (e.g., Gordon et al. 1978; Deacon 1979; Gordon et al. 1981). The gyre-associated flow is predominantly barotropic. Geopotential surface heights above the 1000 db level computed using historical data show only very weak baroclinic surface currents (Gordon et al. 1978). The March 1986 data show virtually negligible baroclinic surface currents based upon geopotential surface heights above either the 500 or 1500 db levels (Nelson et al. 1989), consistent with earlier work. The sole exception occurred along the northward transect, where a very weak and poorly defined baroclinic surface current roughly paralleled the bottom topography northward and then turned westward around the tip of the Antarctic Peninsula (Husby and Muench 1988).

Total circulation in the study region is difficult to determine. Carmack and Foster (1975) derived a transport of 97×10^6 m^3 s^{-1} for the Weddell Gyre, based upon a few short-term current measurements. Closure of the wind-driven cyclonic gyre was postulated by Gordon et al. (1981) to require a western boundary current along the Antarctic Peninsula with a northward transport of about 76×10^6 m^3 s^{-1}, vertically integrated current speeds of about 0.08 m s^{-1} and a width of about 225 km at 65°S. The westernmost part of the March 1986 study area would presumably have been near the seaward limits of such a boundary current. A large, multiyear ice floe was instrumented for drift measurements using an Argos-tracked buoy in the westernmost study area near the start of the field program. This buoy showed virtually no net meridional drift and little net zonal drift during March 1986 (Nelson et al. 1989). Winds during this period were light to moderate, random in speed and direction, and generally insufficient to cause significant net wind-driven ice movement. The western boundary current, if in fact it was present, does not appear to have impinged upon the study area which had a very sluggish, if not demonstrably zero, surface circulation.

c. *Regional distribution of temperature and salinity*

The T, S and density characteristics observed in March 1986 typified the Weddell Sea region as discussed in terms of water masses by Carmack (1977). The T, S and density (as σ_t) data from the 1986 field work have been published as vertical sections and briefly discussed in Husby and Muench (1988). These characteristics are summarized for the upper 600 m as vertical profiles from the western, central and eastern portions of the study region (Fig. 2); greater depths are not included because the staircase features of interest here were restricted to the upper 600 m.

The uppermost water mass is the surface water, which varies seasonally in temperature and salinity due to warming and ice melting, cooling and freezing. In March 1986 this layer was 30–50 m thick and vertically well mixed. Temperature varied from the freezing point (about −1.7°C) under the ice cover in the west to a maximum of −0.2°C at the easternmost stations. Salinity was vertically near-uniform and varied laterally between about 33.6 and 33.8‰, where lower values were associated with recent ice melting along the ice edge. The surface water was underlain by a strong S-controlled pycnocline. A small T spike (0.1° to 0.2°C in amplitude), a remnant of the previous summer's warming, was sometimes present just underlying this pycnocline. Beneath the pycnocline lay the winter water, characterized by low T (−1.5° to −1.7°C), a relatively invariant S of about 34.46‰ and representing a remnant of the preceding winter's cold, convective layer. The lateral T and S variations seen in the surface water were not reflected in the underlying waters.

The warm water, defined by $T > 0$°C, underlies the winter water and is separated from it by transition water, which contains the main thermocline extending from about 100 to 180 m during March 1986. Maximum temperatures in the warm water were typically present at 300–500 m, with the shallower depths typifying the eastern study region. The region below the transition zone and above the T maximum was the site of the Type B thermohaline staircases which are the primary focus of this paper. Temperature increased relatively rapidly with depth (about 0.02°C m^{-1}) in the transition zone between about 100 and 180 m depths, whereas below 180 m the T gradient was much weaker (about 0.0015°C m^{-1}). The vertical S gradient was also stronger in the 100–180 m range (<0.002‰ m^{-1}) than from 180 to 500 m (about 0.0002‰ m^{-1}). The change in vertical T and S gradients at the bottom of the main thermocline was abrupt and, where Type B steps were present, extremely so.

Vertical density gradients were extremely small, averaging about 7×10^{-5} (kg m^{-3}) m^{-1} through the warm water layer. In fact, below the uppermost 50 m the total increase in density down to 1500 m was less than 0.12 kg m^{-3}.

4. Thermohaline staircases

a. Description of staircases observed in March 1986

The warm water and overlying transition water from about 100 to 500 m were characterized during March 1986 by staircases consisting of vertical steps in T and S (hence, also density). The staircase structure is illustrated in Fig. 2 (profile C). In the main thermocline, where vertical T and S gradients were relatively large, the staircase layers were small in vertical extent. Below about 180 m, however, the staircases had much greater vertical extents (Type B). The smallest layers seen in the Type B region were of order 10 m vertically, and the largest exceeded 100 m. Strata between the Type B layers varied in thickness from interfaces less than 1 m across (too thin to have been resolved by the CTDs used) to several tens of meters. The layers were in most cases uniform vertically in T and S to within the limits of resolution of the CTD. The transition between strata containing Type A and Type B staircases was clearly defined on the vertical profiles by abrupt changes in the mean vertical T, S and σ_t gradients (see Fig. 2).

Type A staircases were present throughout the March 1986 study region. Type B staircases were not, and their areal distribution was mapped as follows. The criteria for a Type B structure was that T be uniform vertically to within measurement accuracy for 10 m or more. This layer thickness was selected as a cutoff because no identifiable staircases occurred in the Type B strata having thicknesses less than this and most were considerably thicker. Step features meeting this criteria, including between-step interfaces, were enclosed within contours (Fig. 3). This distribution reveals that Type B features occurred in the central portions of each west–east transect and were primarily confined within a single well-defined region. Excluding the single northward-oriented transect, where they were not observed, they were present over about 50% of the approximately 40 000 km^2 area sampled. They occupied a proportionally greater area along the southern west–east transect than along the northern transect. They covered a slightly greater proportion of the northern transect on the second than on the first transect occupation.

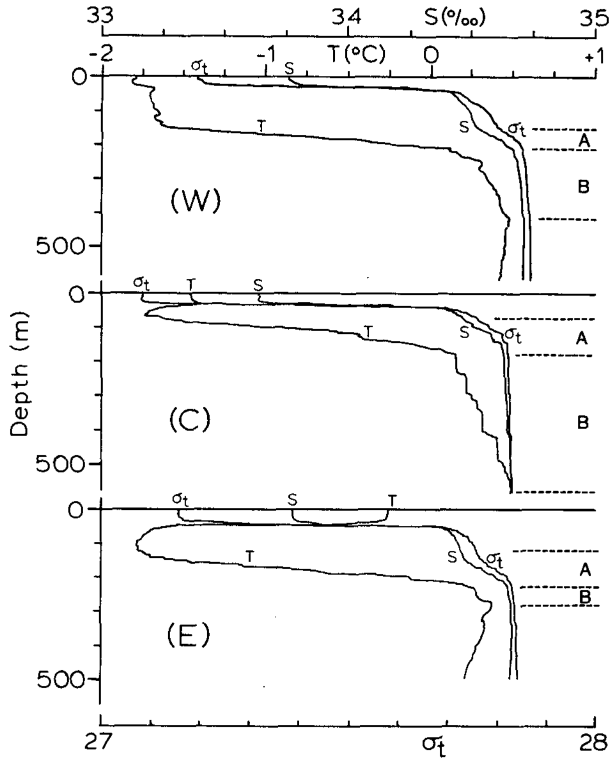

FIG. 2. Representative vertical profiles of T, S and σ_t for western (W), central (C) and eastern (E) portions of the March 1986 study region. Vertical extents of Type A and B regimes are indicated on the profiles. Locations of casts are indicated on Fig. 1.

The Type B staircase layers were reflected in the temperature field as horizontally oriented bands of vertically uniform temperature between about 200 and 400 m (Fig. 4). These layers retained their identities over considerable horizontal distances, particularly in the western half of the southern transect.

Repetitive CTD casts obtained at several locations provide time sequences of data (Fig. 5). The 9–14 March sequence shows variability in the deeper staircases during that period. On both the 12–18 March and 9–14 March sequences, small steps can be seen superimposed upon the larger layers (e.g. D, I). It is improbable that these variations were advective in origin. Horizontal temperature gradients varied about zero up to $\pm 10^{-6}$ °C m^{-1} and were typically near the small end of this range. They were smallest near 200 m depth and were greater in the meridional than in the zonal direction. Zonal temperature gradients were generally smaller than meridional gradients but, otherwise, the data were inadequate to detect any pattern to the variation. Lateral advective changes of the observed magnitude would have required currents of or-

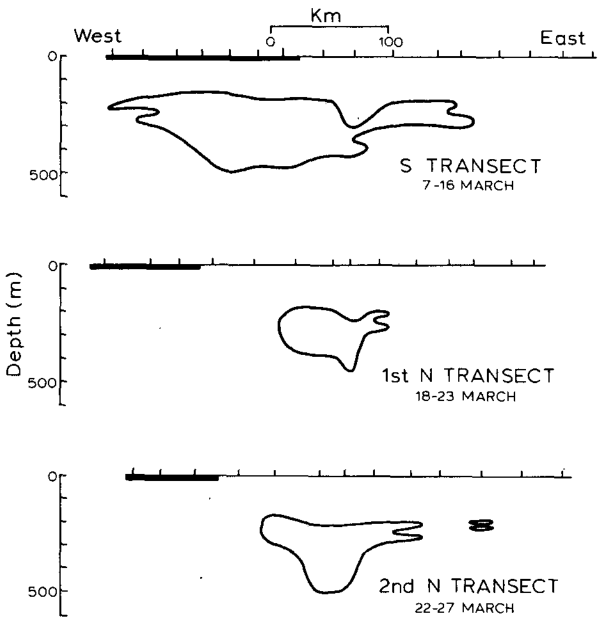

FIG. 3. Areas enclosed by solid contours indicate extent of large step structures, defined as described in the text, along the southern (one occupation only) and northern (two separate occupations) west–east transects. Heavy line at surface indicates ice cover.

der 0.05 m s^{-1} acting in concert with the largest observed gradients. The mean advective field was, as discussed above, immeasurably small during the field program. In view of this observation, and given the very small horizontal gradients, we do not believe it probable that the observed variations were advective in origin. We maintain, rather, that they were due to vertical processes and, in particular, to those of a double diffusive nature.

b. *Estimation of some pertinent parameters*

Laboratory experiments have provided convincing evidence that double diffusion can play a significant role in oceanic thermohaline staircase formation (see Turner 1973, and section 5 below). Staircases containing diffusive interfaces are formed when lower T and S water overlies warmer, more saline water. Layers in such staircases have been observed in the ocean with vertical extents from less than a meter to several tens of meters (Federov 1970). Because of the dependence of these features upon molecular processes acting across density interfaces, they are found in regions where turbulent mixing processes are weak. In general, such regions are remote from either physical boundaries (coastlines or pronounced bottom topographic features) or hydrodynamic boundaries (current or frontal systems). These criteria appear to be well satisfied in the Weddell Sea study area.

The density ratio R_ρ, defined as $\beta\Delta S/\alpha\Delta T$ where α is the coefficient of thermal expansion, β is the coefficient of expansion due to S changes and ΔT and ΔS are the T and S differences between any two adjacent layers, is used as a parameter in describing double diffusive processes (Turner 1973). Density ratios R_ρ were computed separately for the Type A and B regimes using the T and S differences across each regime at each station. These were computed as "bulk ratios" using the formula $\beta\bar{S}_z/\alpha\bar{T}_z$ where $\bar{S}_z$ and $\bar{T}_z$ are the mean vertical gradients of S and T across the Type A and B step regions at each CTD cast. For computing R_ρ, the Type A region was defined on its upper surface by the abrupt increase in T gradient underlying the winter water (about 100 m deep) and on its lower surface by the abrupt change in vertical T and S gradient at the bottom of the main thermocline (about 170–200 m deep). The Type B region was defined by the latter abrupt gradient change on its upper surface, and by the T maximum at 300–500 m for its lower surface. Values for R_ρ were computed regardless of whether or not Type B steps were present. Mean in situ values for α and β were computed for each regime using the 1980 equation of state for seawater (UNESCO 1981).

For the Type A regime the areawide averaged value for R_ρ was 1.52, slightly larger than Foster and Carmack's (1976) value of 1.39. Their value was an average computed across individual steps through this same regime at a single 1973 station located some 600 km east of our study area. The standard deviation for the March 1986 data was only 0.09, reflecting the lateral near-uniformity in T and S conditions within the regime. Middleton and Foster (1980) computed a similar R_ρ value of 1.396, averaged over a time series of CTD casts at a site in the northwestern Weddell Sea where only Type A staircases were present. The average value of R_ρ in the Type B regime for the March 1986 data was 1.36 with a standard deviation of 0.27, slightly but

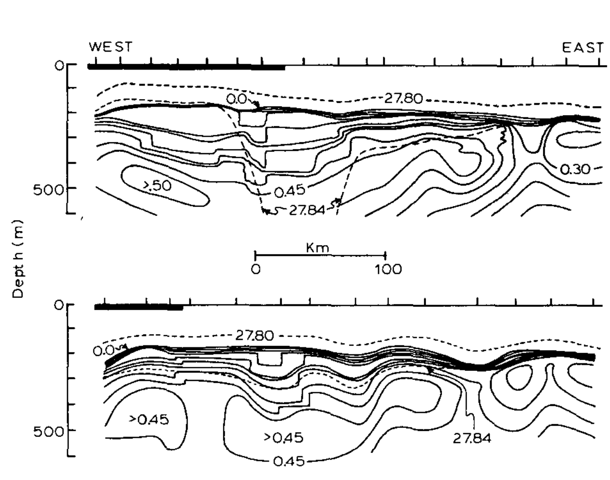

FIG. 4. Vertical distribution of T (°C) plotted with a contour interval of 0.05°C along the southern (upper) and first of two northern (lower) west–east transects. Dashed lines indicate the σ_t = 27.80 and 27.84 isopycnals. Isotherms shallower than the 0°C isotherm have been omitted for clarity.

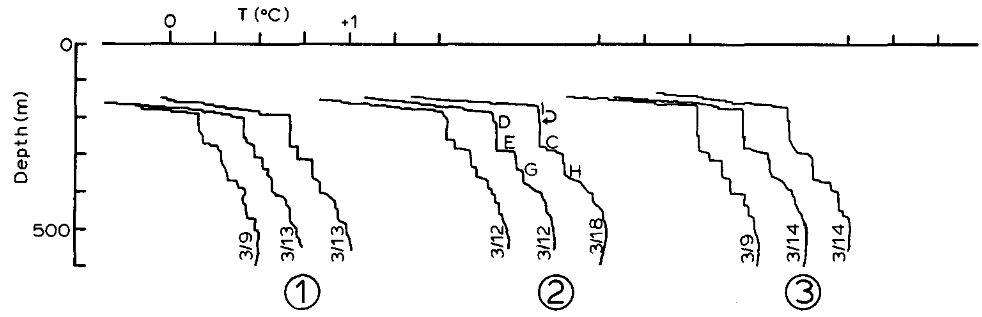

FIG. 5. Three sequential sets of T profiles, each set obtained from the same location, illustrating persistence over time of the thick layers and showing superposed smaller-scale perturbations. Upper case letters and arrows adjacent to profiles indicate step features which are referred to in the text. Geographical locations of the profiles are indicated on Fig. 1. Temperature scale for the left-hand 3/9 station is shown at the upper left, and profiles for each sequence are separated by 0.25°C.

not significantly smaller than for the Type A regime. Foster and Carmack (1976) had computed an R_ρ of 1.03 for the Type B regime using individual steps for the computation. The reason for the difference between their computed value and that for the 1986 data is unknown. It may represent natural variability in the system, or may simply reflect the different modes of computation; Foster and Carmack's (1976) values were computed within the staircases only, whereas our values were averaged over profiles irrespective of whether or not staircases were present. There was no discernable horizontal pattern of variability in R_ρ in either the Type A or B regimes during March 1986. These values all exceeded 1, ensuring static stability of the staircase structure.

Huppert (1971) analyzed the stability of a series of double diffusive layers and showed that the diffusive interface becomes unstable when $R_\rho < 2$. His values suggest that this instability might lead to a layer breakdown having a time scale t_b defined as

$$t_b = 3R_\rho^2 \, \mathrm{Ra}^{-1/3}(h^2/k_h)$$

where Ra is the Rayleigh number $g\alpha\Delta Th^3/k_h\nu$ based on layer thickness h, thermal diffusivity k_h and kinematic viscosity ν. Our R_ρ values are less than 2 for both Type A and B staircases. Huppert's results provide a means for estimating time scales for persistence of the steps. For parameters typifying Type A staircases ($g\alpha\Delta T = 3.2 \times 10^{-5}$ m s^{-2}, $R_\rho = 1.52$, $h = 5$ m) we obtain $t_b = 6$ days, whereas for Type B staircases ($g\alpha\Delta T = 6.9 \times 10^{-5}$ m s^{-2}, $R_\rho = 1.36$, $h = 80$ m) we get $t_b = 60$ days. These results need to be viewed with caution, however, because they are valid only if there is an innate horizontal homogeneity of the convective flow. This may be the case in laboratory experiments, but there is no reason why it should be true for the ocean.

The time scales estimated in the preceding paragraph lead us to expect significant variations in small step structure during the periods 9–13 March, 12–18 March and 9–14 March, whereas during these same periods the large steps should remain relatively invariant. We have not examined the Type A staircase regime with respect to temporal variations because vertical resolution of the data was inadequate for this purpose. Middleton and Foster (1980) noted, based upon a time-series CTD station in the northwestern Weddell Sea, that significant variations occurred in the Type A staircases on an hour-to-hour basis, shorter than the 6-day time scales computed above. Our 1986 field data showed that the larger features within the Type B staircase regime persisted for at least 4–6 days, equal to or exceeding the estimated lifetime for Type A staircases. These values contrast with those estimated by Foster and Carmack (1976) who surmised, based upon observed changes in step structure between downcast and upcast data, that the thick layers were either short-lived or limited in horizontal extent. Their conclusions were based, however, on observations, which were limited compared to the March 1986 data. Our data are inadequate to compare persistence of the Type B staircases with the predicted 60-day duration. Smaller steps were frequently observed superposed on the Type B layers. The time sequences on 9–14 and 12–18 March (see Fig. 5) showed significant variability in the structures of these smaller steps over periods of less than a day. This time scale is consistent with the above prediction for t_b in that smaller steps appear to have shorter time scales.

5. Staircase evolution and heat fluxes

We have observed thermohaline staircases in the Weddell Sea whose occurrence and temporal variability are consistent with double diffusion. We do not, however, know the origin of these staircases. Pertinent laboratory experiments can be used to predict newly evolved states given knowledge of the previous state, and can aid in identifying physical processes which control this evolution. Ideally, we would like to be able to use the laboratory results to guess at the initial state of the water column prior to staircase formation, in other words, to estimate the genesis of the features. There are, however, vast differences between one-dimensional laboratory experiments and oceanic conditions, with the result that extrapolation of laboratory results to the ocean must be done with caution. We assume, in such an extrapolation, that the dominant oceanic processes are primarily one-dimensional and that double diffusive convection is a primary mechanism for vertical buoyancy transfer.

The following sections discuss staircase generation, generation of smaller steps through layer splitting, and heat fluxes by comparing past laboratory results with our field data. In addition, a simple laboratory experiment is carried out to demonstrate the possibility of layer splitting, a phenomenon which is consistent with the field data.

a. Layer evolution, thickness and splitting

Previous laboratory experiments have clarified the processes involved in evolution of thermohaline staircases from an initially stably salt-stratified fluid. Laboratory experimental results also can provide insight into the processes controlling layer thicknesses and generating small steps which are superimposed on the Type B layers.

1) LAYER EVOLUTION AND THICKNESS

Turner and Stommel (1964) demonstrated formation, in the laboratory, of convecting layers separated by diffusive density interfaces when a stably salt-stratified fluid is heated from below. Turner (1968) proposed that new convecting layers are generated through breakdown of a "thermal boundary layer" which develops above already existing layers. In extending this concept to the ocean, Federov (1970) assumed the thickness of the convecting layers to be determined by their thicknesses at the time of initial formation of the layer aloft. However, Fernando (1987) showed that layer thicknesses are determined not by the above criteria, but by a balance between the kinetic and potential energy of turbulent eddies within the convecting layers. Fernando's (1987) results have been extended to the ocean by Fernando (1989a): this model predicts

$$h \approx 14\left(\frac{q_h}{N_s^3}\right)^{1/2} \frac{1}{(1 - R_\rho^{-1})^{3/4}}, \quad R_\rho \ll 10 \qquad (1)$$

where h is the layer thickness in the ocean, q_h is heat flux through the layers and $N_s^2 = g\beta\bar{S}_z$ is the buoyancy frequency based upon salinity stratification. It was noted that (1) is valid only when the staircase structure is in a "quasi-stationary" state. Available field data, with heat flux estimates derived from sources independent of the laboratory-based flux laws, were found to agree satisfactorily with (1) (Fernando 1989a).

An alternative expression for layer thickness in oceanic staircases has been proposed by Kelley (1984) in the form

$$G = G(R_\rho) \qquad (2)$$

where $G = h/(k_h/N)^{1/2}$ and N is the buoyancy frequency based on bulk properties. Since heat flux is generally not known, (2) offers a convenient way to predict h and thus has advantages over (1).

Values for $G\,(=h/(k_h/N)^{1/2})$ and R_ρ were calculated for the Type A and B staircases. The results for five different casts, selected to represent the full range of layer thicknesses, are shown in Fig. 6 and are compared with data compiled by Kelley (1984). Note that the present data for both small and large steps do not agree with Kelley's prediction that there can be a universal functional dependence between G and R_ρ.

Equation (1) can be used to compute thicknesses for the Type A and B layers provided that they are quasi-stationary. We assume Gordon's (1981) estimated upward heat flux $q_h = 2.9 \times 10^{-9}$ m^2 s^{-3}, which was based upon surface fluxes, as a typical value for both Type A and B regimes. The computed thickness for Type B staircases, where $N_s = 10^{-3}$ s^{-1} and $R_\rho = 1.36$, is $h = 65$ m, in reasonable agreement with the observations. Applying the same computation to Type A staircases, where $N_s = 3.1 \times 10^{-3}$ s^{-1} and $R_\rho = 1.52$ yields $h = 10$ m, which is larger than the 1–5 m typical layer thickness (though similar to the maximum ob-

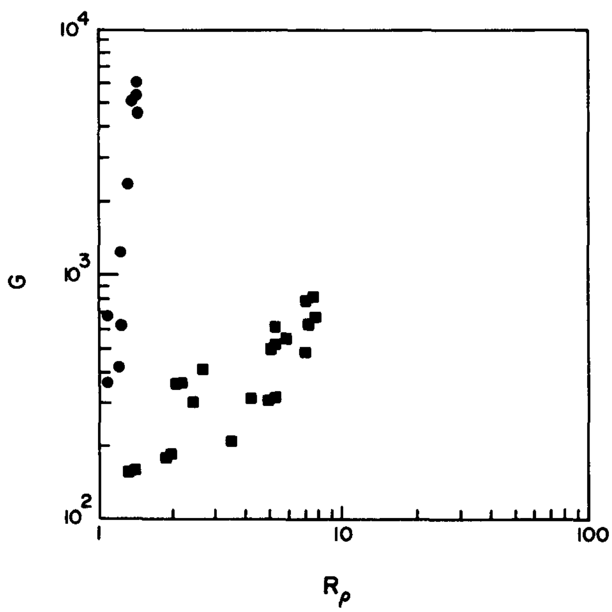

FIG. 6. Plot of nondimensional convecting layer thickness G versus stability number R_ρ computed using March 1986 Weddell Sea field data (circles) and compared with Kelley's (1988) data (squares). Average Type A and Type B layer thicknesses, as derived from the field data, were used for the calculations. Points computed using the 1986 field data do not coincide with those from Kelly (1988), suggesting that the relation between G and R_ρ is not a universal one.

served 8 m layer thickness). The disagreement between observed and computed Type A layer thicknesses suggests that the Type A staircases are nonstationary, an observation that will be referred to in the discussion of heat fluxes.

2) INTERFACE SPLITTING

Kelley (1986, 1988) examined CTD data reported by Neshyba et al. (1972) and concluded that, under certain conditions, the diffusive interfaces can split in two to generate new convecting layers. He further argued that when the convection Richardson number given by

$$\mathrm{Ri}_c = \mathrm{Pr}^{4/3}(R_\rho - 1)^{4/3}/c_3 G^{4/3}, \tag{3}$$

where

$$c_3 = 0.0086 \exp\{4.6/\exp[0.54(R_\rho - 1)]\},$$

falls below a critical value of order one, layer splitting occurs. For oceanic cases, Ri_c was postulated to be between 0.1 and 1.0, with an uncertainty factor of 2. Interfaces that are on the verge of splitting have low Ri_c, whereas following splitting Ri_c was supposed to increase to values near 1.

Features were identifiable in the March 1986 field data that resembled interfacial splitting. Such an event is depicted in Fig. 5, for example, by evolution of the profile between points designated E and C. For this case, we estimated Ri_c using (3) and values for R_ρ and G which characterized layers DE and EG. This estimation yielded $\mathrm{Ri}_c = 1.25 \times 10^{-4}$ and 3.06×10^{-3} respectively, which differ greatly from the values suggested by Kelley (1988) for oceanic staircases. For the subsequent layers IC and CH, presumably after interface splitting had occurred, values for Ri_c were 2.52×10^{-4} and 4×10^{-4}, respectively. If Kelley's (1988) criterion is correct, the low Ri_c values at both E and C suggest that the interface was susceptible to further splitting even after splitting to form the additional layer at C. Such interfacial splitting events are consistent with occurrence of active double-diffusive convection in the Type B staircases.

In order to verify the possibility of interfacial splitting, we carried out a series of preliminary laboratory experiments. These utilized a square plexiglass tank heated from below, and details are provided in Fernando (1989b). Two-layer stratification was generated in the tank by slowly feeding a saline water layer underneath a freshwater layer. Experiments were started by initiating upward heat flux through the tank bottom, and interfacial events were monitored over a shadowgraph.

It was observed during experiments that in cases where the initial interface was somewhat diffuse, the interface did indeed split into two to form an intermediate convecting layer. Evolution of this layer splitting was documented photographically (Fig. 7). Computation of Ri_c before and after splitting did not confirm Kelley's (1988) prediction that splitting leads to a substantial increase in Ri_c. For example, for the case shown in Fig. 7 Ri_c based on lower layer thickness and average N, where $N^2 = (g\beta\Delta S - g\alpha\Delta T)/H$ and H is the water column height, increased from about 1.1 before splitting to 1.2 after splitting. Due to the strong convective activity in the lower layer, which was directly atop the heat source, the interface between the lower layer and the newly-formed intermediate layer moved upwards and finally merged with the upper interface, in this process destroying the intermediate layer.

Layer splitting, as observed in the laboratory, provides a possible explanation for occurrence of the small steps that were superposed on the Type B layers in the field data. The field data were, however, inadequate to resolve the behavior of the small steps sufficiently to allow their comparison with the laboratory results.

b. Heat fluxes

A large number of laboratory experiments have addressed the topic of buoyancy fluxes due to heat q_h and salt q_s. Turner (1965) generated a single double diffusive interface by heating from below a two-layered, salt-stratified fluid. Buoyancy fluxes due to heat and salt across the interface were measured as functions of R_ρ. According to Huppert (1971), Turner's heat (buoyancy) flux data can be expressed by an equation of the form

$$q_h = 0.323\left(\frac{k_h^2}{\nu}\right)^{1/3}(g\alpha\Delta T)^{4/3}R_\rho^{-2}. \tag{4}$$

Later experiments by Marmorino and Caldwell (1976) demonstrated that (4) is valid only for the range $2 < R_\rho < 5$ (see also Taylor 1988). Their measurements showed agreement with a flux law of the form

$$q_h = 8.58 \times 10^{-3}\left(\frac{k_h^2}{\nu}\right)^{1/3}(g\alpha\Delta T)^{4/3} \times \exp\{4.6 \exp[-0.54(R_\rho - 1)]\}. \tag{5}$$

In addition to heating from below, Marmorino and Caldwell (1976) employed cooling from above in order to achieve a quasi-stationary interface.

Based on laboratory experimental results, Fernando (1989a,b) proposed a parameterization for heat fluxes through diffusive interfaces. He argued that, when turbulent eddies in the convecting layers are insufficiently energetic to penetrate the interface, the buoyancy transfer is fully diffusive. The ensuing heat flux was shown to be governed by

$$q_h = 7 \times 10^{-2}[k_h^3(g\alpha\Delta T)^6/h^2]^{1/5}. \tag{6}$$

When interfacial stability falls sufficiently that turbulent eddies can penetrate the interface, i.e., when

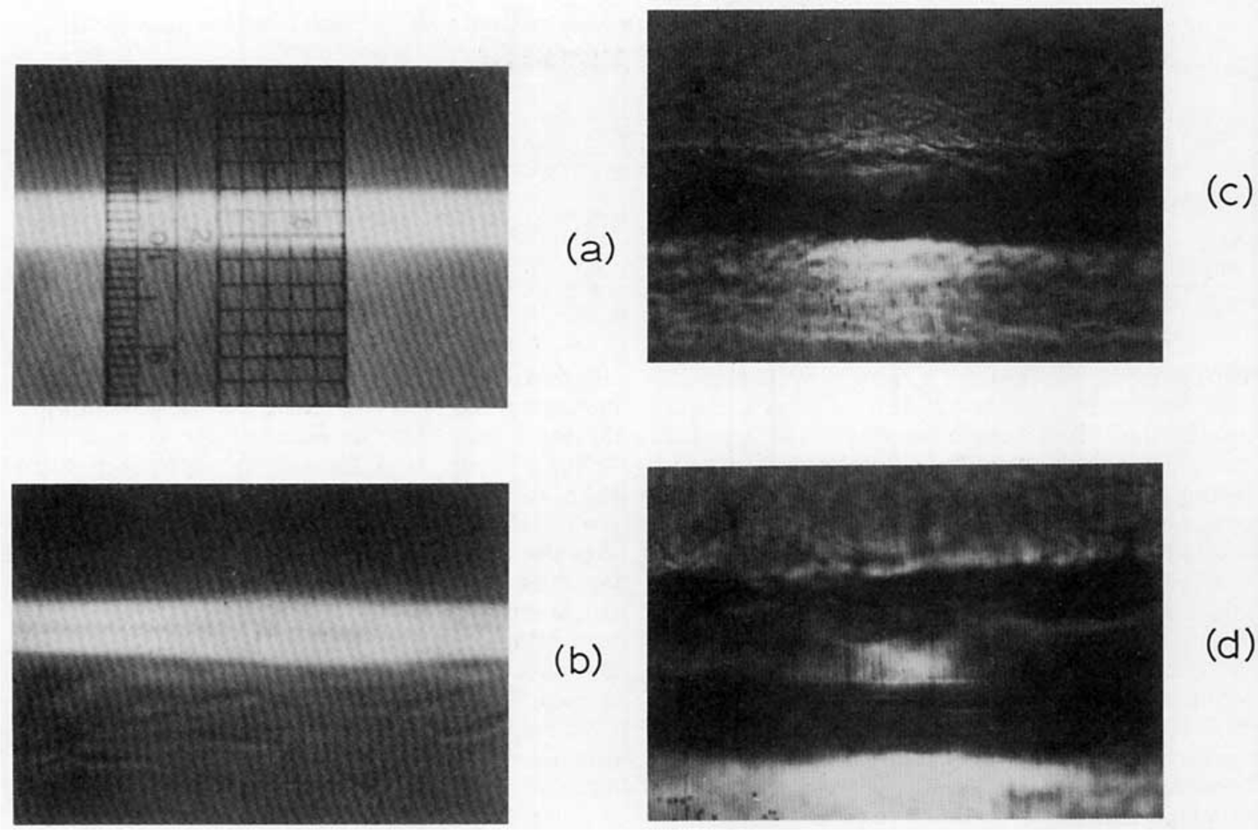

FIG. 7. Sequence of shadowgraph photographs showing the splitting of an initially diffuse double-diffusive interface: (a) initial interface; (b) further diffusion of interface after onset of convection; (c) initial signs of interfacial splitting, with convection underway in both layers; and (d) final state showing formation of an intermediate layer where interface has split in two.

$$R_\rho \mathrm{Ra}^{2/15} \mathrm{Pr}^{2/15} \tau^{1/3}(1 - R_\rho \tau^{1/2})/(1 - R_\rho^{-1}\tau^{1/2})^{1/5} < 0.15 \mathrm{Nu}^{2/3},$$

the flux law was shown to be governed by

$$q_h = 4.7 \times 10^{-4}(g\alpha\Delta T)^{3/2}h^{1/2}, \quad (7)$$

where $\mathrm{Nu} = q_h/[k_h(g\alpha\Delta T)/h]$ is the Nusselt number, $\mathrm{Pr} = \nu/k_h$ is the Prandtl number, $\tau = k_s/k_h$ is the Lewis number and k_s is the molecular diffusivity of salt.

By considering the ice melting rates and heat inputs from the atmosphere and oceanic upper layers, Gordon (1981) estimated a heat flux of 18 W m^{-2} (2.9×10^{-9} m^2 s^{-3}) upward through the Weddell Sea pycnocline. Using parameter values appropriate to the Type B staircases of $g\alpha\Delta T = 6.9 \times 10^{-5}$ m s^{-2}, $R_\rho = 1.36$, $N_s = 10^{-3}$ s^{-1} and $h = 80$ m we find heat (buoyancy) fluxes of 2.26×10^{-9}, 1.04×10^{-9} and 2.40×10^{-9} m^2 s^{-3}, using the formulae of Marmorino and Caldwell (1976), Huppert (1971) and Fernando (1989a,b), respectively. The estimates based on Marmorino and Caldwell (14 W m^{-2}) and Fernando (15 W m^{-2}) are in reasonable agreement with Gordon's (1981) estimate. Huppert's formula underestimates the heat flux (6.5 W m^{-2}), but our R_ρ values exceed the range of validity for his formula.

In addition to estimating heat fluxes, Gordon (1981) also assumed an average buoyancy gradient $g\alpha dT/dz = 1.76 \times 10^{-5}$ and computed an eddy diffusivity of heat through the thermocline of $k = q_h/(g\alpha\Delta T/h) = 1.5 \times 10^{-4}$ m^2 s^{-1}. Values of k were computed for the Type B regime using heat fluxes computed on the basis of (4), (5) and (7) and the March 1986 field values of $g\alpha\Delta T$ and h. These results are shown, along with the heat flux estimates, in Table 1. This table also shows, for comparison purposes, values for q_h and k computed for the Type A regime using typical values $h = 5$ m, $R_\rho = 1.52$ and $g\alpha\Delta T = 3.16 \times 10^{-5}$ m s^{-2}.

The laboratory-based heat flux estimates through the Type A regime are smaller than those through the Type B regime. This suggests a heat flux convergence, due to double diffusive processes alone, at the boundary between the two regimes. Such a convergence would, in the absence of other processes to remove the heat, lead to a temperature increase at the boundary between the Type A and Type B regimes. Visual inspection of both vertical temperature profiles and temperature–salinity curves shows, however, no detectable differences in temperature at this boundary between regions with and without Type B steps (see, also, Fig. 2). We noted above that vertical heat transport through the

the interfaces and convection in the homogeneous layers driven by a heat source below the layer and a heat sink above. The heat source is present in the form of a warm core of the warm water, and the sink is present as the cold winter water.

3) Laboratory-based double diffusion theory provides a heat flux estimate through the Type B staircase regime of order 2×10^{-9} m^2 s^{-3}, which is consistent with that derived from regional scale heat balance considerations using oceanographic and sea ice data. Laboratory-based heat fluxes for Type A layers were an order of magnitude lower, suggesting that a significant fraction of the heat flux in Type A layers occurs through processes other than double diffusion.

4) It is suggested that the observed Type B staircases represent advanced stages of evolution (or, perhaps, decay would be a better term) of winter thermohaline convective features which were formed either in the eastern Weddell Sea or in the Antarctic Coastal Current.

Acknowledgments. We acknowledge the help of David Husby and John Gunn in obtaining and processing the 1986 Weddell Sea field data. Laurie Padman of Oregon State University, Doug Martinson of Lamont–Doherty Geological Observatory and two anonymous reviewers have provided useful comments on an earlier draft of the manuscript. This work has been carried out under National Science Foundation (NSF) Grants DPP-8420646 and DPP-8715979 to Science Applications International Corporation (SAIC) (RDM and GRS), Office of Naval Research (ONR) Contract N00014-82-C-0064 with SAIC, (RDM and GRS), and ONR Contracts N-00014-87-K-0423 and N-00014-88-K-0250 with Arizona State University (HJSF). HJSF acknowledges also support from the NSF Presidential Young Investigator Program in Fluid Mechanics and Hydraulics.

REFERENCES

Bersch, M., 1988: On deep convection in the Weddell Gyre. *Deep-Sea Res.*, **35**, 1269–1296.

Carmack, E. C., 1977: Water characteristics of the Southern Ocean south of the Polar Front. *A Voyage of Discovery.* M. Angel, Ed., Pergamon Press, 15–41.

——, and T. D. Foster, 1975: On the flow of water out of the Weddell Sea. *Deep-Sea Res.*, **22**, 711–724.

Deacon, G. E. R., 1979: The Weddell Gyre. *Deep-Sea Res.*, **26A**, 981–995.

Federov, K. N., 1970: On the step-like structure of the temperature inversions in the ocean. *Izv. Atmos. Ocean. Phys.*, **6**, 1178–1188.

Fernando, H. J. S., 1987: The formation of a layered structure when a stable salinity gradient is heated from below. *J. Fluid Mech.*, **182**, 525–541.

——, 1989a: Oceanographic implications of laboratory experiments on diffusive interfaces. *J. Phys. Oceanogr.*, **19**, 1707–1715.

——, 1989b: Buoyancy transfer across a diffusive interface. *J. Fluid Mech.*, **209**, 1–34.

Foldvik, A., T. Gammelsrod and T. Torresen, 1985: Hydrographic observations from the Weddell Sea during the Norwegian Antarctic Research Expedition 1976/77. *Polar Res.*, **3**, 177–193.

Foster, T. D., and E. C. Carmack, 1976: Temperature and salinity structure in the Weddell Sea. *J. Phys. Oceanogr.*, **6**, 36–44.

Gordon, A. L., 1978: Deep Antarctic convection west of Maud Rise. *J. Phys. Oceanogr.*, **8**, 600–612.

——, 1981: Seasonality of Southern Ocean sea ice. *J. Geophys. Res.*, **86**, 4193–4197.

——, E. Molinelli and T. Baker, 1978: Large-scale relative dynamic topography of the Southern Ocean. *J. Geophys. Res.*, **83**, 3023–3032.

——, D. G. Martinson and H. W. Taylor, 1981: The wind-driven circulation in the Weddell-Enderby Basin. *Deep-Sea Res.*, **28A**, 151–163.

Huber, B. A., S. E. Rennie, D. T. Georgi, S. S. Jacobs and A. L. Gordon, 1981: Islas Orcadas Cruise 12 Jan.–Feb. 1977; Hydrographic Stations 53–126. Tech. Rep. CU-2-81-TR2, Lamont-Doherty Geological Observatory, Palisades, NY.

Huppert, H. E., 1971: On the stability of a series of double diffusive layers. *Deep-Sea Res.*, **18**, 1005–1021.

Husby, D. M., and R. D. Muench, 1988: Hydrographic observations in the northwestern Weddell Sea marginal ice zone during March 1986. NOAA Tech. Memo. NOAA-TM-NMFS-SWFC-106, 33 pp. [Available from SW Fish. Ctr., NOAA/NMFS, Monterey, CA 93942.]

Kelley, D., 1984: Effective diffusivities within oceanic thermohaline staircases. *J. Geophys. Res.*, **89**, 10 484–10 488.

——, 1986: Interfacial migration in thermohaline staircases. *J. Phys. Oceanogr.*, **17**, 1633–1639.

——, 1988: Explaining effective diffusivities within diffusive oceanic staircases. *Proc. 19th Int. Liege Coll. on Ocean Hydrodynamics.* Elsevier Oceanogr. Ser., **46**, 481–502.

Killworth, P. D., 1979: On "chimney" formations in the ocean. *J. Phys. Oceanogr.*, **9**, 531–554.

Marmorino, G. D., and D. R. Caldwell, 1976: Heat and salt transport through a diffusive thermohaline interface. *Deep-Sea Res.*, **23**, 59–67.

Martinson, D. G., P. D. Killworth and A. L. Gordon, 1981: A convective model for the Weddell Polynya. *J. Phys. Oceanogr.*, **11**, 466–488.

Middleton, J. H., and T. D. Foster, 1980: Fine structure measurements in a temperature-compensated halocline. *J. Geophys. Res.*, **85**, 1107–1122.

Nelson, D. M., W. O. Smith, Jr., R. D. Muench, L. I. Gordon, C. W. Sullivan and D. M. Husby, 1989: Particulate matter and nutrient distributions in the ice edge zone of the Weddell Sea: Relationship to hydrography during late summer. *Deep-Sea Res.*, **36**, 191–209.

Neshyba, S., V. T. Neal and W. W. Denner, 1972: Spectra of internal waves: in-situ measurements in a multiple-layered structure. *J. Phys. Oceanogr.*, **2**, 91–95.

Taylor, J., 1988: The fluxes across a diffusive interface at low values of the density ratio. *Deep-Sea Res.*, **35**, 555–567.

Turner, J. S., 1965: The coupled turbulent transports of salt and heat across a sharp density interface. *Int. J. Heat Mass Trans.*, **8**, 759–767.

——, 1968: The behavior of a salinity gradient heated from below. *J. Fluid Mech.*, **33**, 183–200.

——, 1973: *Buoyancy Effects In Fluids.* Cambridge University Press.

——, and H. Stommel, 1964: A new case of convection in the presence of combined vertical salinity and temperature gradients. *Proc. Natl. Acad. Sci.*, **52**, 49–53.

UNESCO, 1981: Tenth report of the joint panel on oceanographic tables and standards. UNESCO Tech. Papers in Mar. Sci. No. 36, 24.

Zwally, H. J., J. C. Comiso, C. L. Parkinson, W. J. Campbell, F. D. Carsey and P. Gloersen, 1983: Antarctic Sea Ice, 1973–1976: Satellite passive-microwave observations. NASA SP-459, National Aeronautics and Space Administration, Washington, D.C.