The Antarctic Zone Flux Experiment

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ABSTRACT

In winter the eastern Weddell Sea in the Atlantic sector of the Southern Ocean hosts some of the most dynamic air–ice–sea interactions found on earth. Sea ice in the region is kept relatively thin by heat flux from below, maintained by upper-ocean stirring associated with the passage of intense, fast-moving cyclones. Ocean stratification is so weak that the possibility of deep convection exists, and indeed, satellite imagery from the Weddell Sea in the 1970s shows a large expanse of open water (the Weddell Polynya) that persisted through several seasons and may have significantly altered global deep-water production. Understanding what environmental conditions could again trigger widespread oceanic overturn may thus be an important key in determining the role of high latitudes in deep-ocean ventilation and global atmospheric warming. During the Antarctic Zone Flux Experiment in July and August 1994, response of the upper ocean and its ice cover to a series of storms was measured at two drifting stations supported by the National Science Foundation research icebreaker *Nathaniel B. Palmer*. This article describes the experiment, in which fluxes of heat, mass, and momentum were measured in the upper ocean, sea ice, and lower-atmospheric boundary layer. Initial results illustrate the importance of oceanic heat flux at the ice undersurface for determining the character of the sea ice cover. They also show how the heat flux depends both on high levels of turbulent mixing during intermittent storm events and on large variability in the stratified upper ocean below the mixed layer.

1. Background

The cold, abyssal waters of the World Ocean interact directly with the rest of the climate system in limited areas confined to high latitudes, where density structure of the upper ocean permits deep convection. In the mid-1970s, satellite imagery showed a large region in the Weddell Sea, far from both the seasonal ice edge and the continental shelves, that remained virtually ice free for several years (Zwally and Gloersen 1977; Carsey 1980; Martinson et al. 1981). This expanse of open water, called the Weddell Polynya, was maintained by infusion of heat from below, suggesting that transient (by climatic timescales) deep-ocean convection was occurring on a massive scale. Direct evidence of deep convection was observed in the region during the time of the polynya (Gordon 1978), and, indeed, Gordon (1982) estimates that deep-water formation associated with the polynya exceeded typical production in the Weddell Sea (which is derived mainly from the continental shelves and comprises a significant proportion of the total deep-water production) by three to six times. The polynya affected a tremendous transfer of heat from the ocean to the atmosphere: while it was active (1974–76), average water temperature in the upper 2000 m of the central Weddell dropped by over 0.5°C (see Fig. 4 of Gordon 1991). The heat capacity of 2 km of water is about three orders of magnitude greater than that of the entire air column, so there must have

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been a major impact on winter air temperature and atmospheric circulation in the region as well.

Compared with most of the World Ocean, static stability of the water column in the central and eastern Weddell Sea is very weak. It appears that the air–sea–ice system there exists in a precarious balance between two quasi-stable modes that display substantially different sea ice distributions and deep-water formation rates (Martinson et al. 1981; Gordon 1991). The currently active state supports a seasonal sea ice cover of ~8×10^6 km^2 (Zwally et al. 1983) and coastal deep/bottom-water formation (e.g., Gill 1973; Foster and Carmack 1976). The alternative mode (probably exemplified regionally by the Weddell Polynya) is characterized by winterlong deep convection driving strong vertical fluxes of heat, salt, and atmospherically active gases. The convection feeds significant open-ocean deep-water formation (Gordon 1982; Martinson 1991), and the associated vertical heat flux prevents the formation of an insulating sea ice cover (Killworth 1979; Martinson et al. 1981). If the convective mode were to encompass the whole Weddell Gyre, a lack of winter sea ice would almost certainly affect the global climate, contributing significantly to global atmospheric warming (Schlesinger and Mitchell 1985; Rind et al. 1995), with intense cooling and ventilation of the abyssal ocean.

It has been nearly two decades since an extensive area of open water persisted through the entire winter in the central Weddell, yet several recent winter oceanographic expeditions through the region have demonstrated that the ice cover remains relatively thin, the mixed layer is deep by polar standards, and that there is strong inferential evidence (e.g., thin ice, elevation of mixed-layer temperature above freezing, entrainment of geochemical tracers from below) for significant heat flux out of the deep ocean (Gordon and Huber 1990; Schlosser et al. 1990). Away from a relatively narrow region of multiyear ice east of the Antarctic Peninsula, mean ice thickness over much of the Weddell reaches only 0.6–0.7 m or less, even in late winter (Wadhams et al. 1987). At latitudes typical of the Weddell, the thermodynamic balance is such that the ice would be expected to grow to at least twice this thickness in the absence of oceanic heat flux or persistent ice divergence. In many parts of the Weddell, salt rejected by a few tens of centimeters of additional ice growth could initiate deep convection. So in an almost paradoxical way, it appears that anomalously large heat flux into the mixed layer from below prevents additional ice growth, thus forestalling deep convection and much larger heat flux associated with deep-water formation (Martinson 1990).

From the standpoint of upper-ocean physics, sustained high heat flux through the mixed layer in the Weddell poses interesting questions. At high latitudes, Coriolis attenuation of turbulent stress in the boundary layer is strong [e.g., the Ekman stress spiral presented by McPhee and Martinson (1994)], serving to limit the depth of turbulent mixing compared with lower latitudes. Direct upwelling at the base of the mixed layer, or strong diffusion in the underlying pycnocline (sharp density gradient), also tends to keep the mixed layer shallow. The ice apparently reaches its mean thickness during a short period of rapid growth early in the winter and, on average, grows very little thereafter, eliminating brine-driven convection as a major factor in mixed-layer dynamics. At temperatures near freezing, buoyancy flux is determined almost entirely by salinity flux. Thus, several factors seem to work against high heat flux and deep mixed layers in the central and eastern Weddell, yet previous winter expeditions found mixed layers averaging nearly 100 m thick and inferred average upward heat flux during the winter as high as 40 W m^{-2} (Gordon and Huber 1990). Results of numerical modeling studies of the magnitude of oceanic heat flux in the region have been ambiguous; for example, Martinson (1990) suggested that heat flux from below the mixed layer of 20–40 W m^{-2} was necessary to maintain the upper ocean–ice regime, while Lemke et al. (1990) found oceanic heat flux to the ice to be an order of magnitude less in their coupled ocean–sea ice model. McPhee (1994) showed that high heat flux and deep mixed layers could coexist in his numerical boundary layer model, but only if the surface momentum flux was concentrated in storm events with episodes of very intense turbulence. If, in fact, oceanic heat flux plays a pivotal role in maintaining the tenuous seasonal ice cover of the Weddell, understanding the small-scale mechanisms governing vertical flux in the mixed layer and pycnocline is an important first step in developing capability for predicting what environmental changes could again trigger widespread deep convection.

Against this backdrop, a group of scientists interested in air–sea–ice interactions recommended a winter expedition into the Weddell Sea as one component of the international Antarctic Zone Program (ANZONE). ANZONE is a relatively informal but active association of scientists interested in oceanic processes south of the Polar Front in the Southern

Ocean. It has targeted the Weddell Sea for intensive study and provided, for example, the forum for organizing the highly successful U.S.–Russian Ice Station Weddell (ISW; Gordon et al. 1993). The suggested project, called the ANZONE Flux Experiment (ANZFLUX), was designed primarily to measure the response of the upper ocean and sea ice to storm events. It was to be accomplished by driving a research icebreaker deep into the winter ice pack, mooring it to drift with typical floes, and deploying modern instrumentation on the adjacent ice capable of measuring fluxes through the upper ocean, ice, and into the atmosphere. While it was recognized that establishing a research station on thin ice with high probability of gale force winds entailed some risk, recent experience in the western Weddell (ISW) and especially with rapid deployment and evacuation of temporary ice stations during the Leads Experiment in the Arctic (LeadEx Group 1993) led us to believe that the mission was feasible. With acquisition by the National Science Foundation in 1992 of a modern, well-equipped research icebreaker, the R/V *Nathaniel B. Palmer,* an ideal platform for ANZFLUX existed, and the experiment was scheduled for July and August of 1994.

2. The experiment

The layout and science programs for the ANZFLUX ice stations are shown schematically in Fig. 1. The icebreaker provided accommodations for the scientists and power for the station, plus several ship-mounted measurement programs. Insulated shelters, mounted on ski-equipped bases, were assembled onboard, then moved by snowmobile into position along a 300-m power line abeam the *Palmer's* aft working deck, as depicted in Fig. 2 along with a pair of Adelie penguin observers.

Several systems were deployed for measuring mean quantities and vertical fluxes in the air–sea–ice system. In the atmosphere (Guest 1995), wind vector, air temperature, humidity, pressure, and downward longwave and shortwave radiation were monitored continuously aboard ship. During the ice drift stations portion of the experiment, these measurements were replicated on the ice and supplemented by upward radiation and turbulent sensible heat fluxes. Radiosondes were launched twice per day. The ice physics program (Ackley et al. 1995) included ice thickness distribution sampling and mass balance studies, plus automated stations for measuring ice temperature at closely spaced intervals in the ice (J. Wettlaufer 1994, personal communication). Shipboard oceanographic systems included a hull-mounted, 150-kHz acoustic Doppler current profiler (ADCP); a conductivity–temperature–depth (CTD) recorder with rosette water sampler; plus a second CTD system for rapid (yo-yo) sampling (Muench 1995; Huber et al. 1995). For the ice station ocean measurements, the upper ocean was divided conceptually into (i) a near-ice surface ("constant flux") layer, with turbulent fluxes measured directly by covariance techniques, including remote sensing of turbulent velocity by broadband ADCP (Stanton 1995); (ii) the mixed layer, with direct flux measurements at several levels on a mast with turbulence-measuring instrument clusters (McPhee 1995a), plus continuous microstructure profiling (Stanton 1995); (iii) a transition region near the base of the mixed layer, with a second mast of turbulence clusters suspended beneath a high-resolution (600 kHz) ADCP (McPhee 1995a; Muench 1995); and (iv) a pycnocline region extending past the T_{max} level, with a self-recording thermistor mooring supplemented by a second microstructure profiling instrument (Padman et al. 1995). In addition to the moored and profiling systems, an acoustically tracked, autonomous underwater vehicle (AUV) program horizontally sampled salinity and temperature at various depths in a 1-km box surrounding the drift station (Morison 1995). We planned for a minimum of two short-term drift stations during the project, with the aim of monitoring response to at least one energetic storm in different upper-ocean regimes.

The cruise included a complete CTD/geochemical tracer program with stations at predetermined intervals during transits into and out of the ice and between drift stations (Huber et al. 1995). These were accompanied by ice thickness and characteristics surveys, and by an ancillary program to measure microwave backscatter (Lytle and Golden 1995). Longer-term, continuous monitoring of the air–ice–ocean system was provided by unmanned data buoys left behind when the first drift station site was vacated. These included a polar-ocean profiling buoy (Morison 1995) and a meteorological–ocean thermistor buoy. We also deployed an ice and upper-ocean thermistor buoy for the Alfred Wegener Institute (AWI) centered in an array of position/surface pressure buoys, as indicated in Fig. 3 (C. Kottmeier 1995, personal communication).

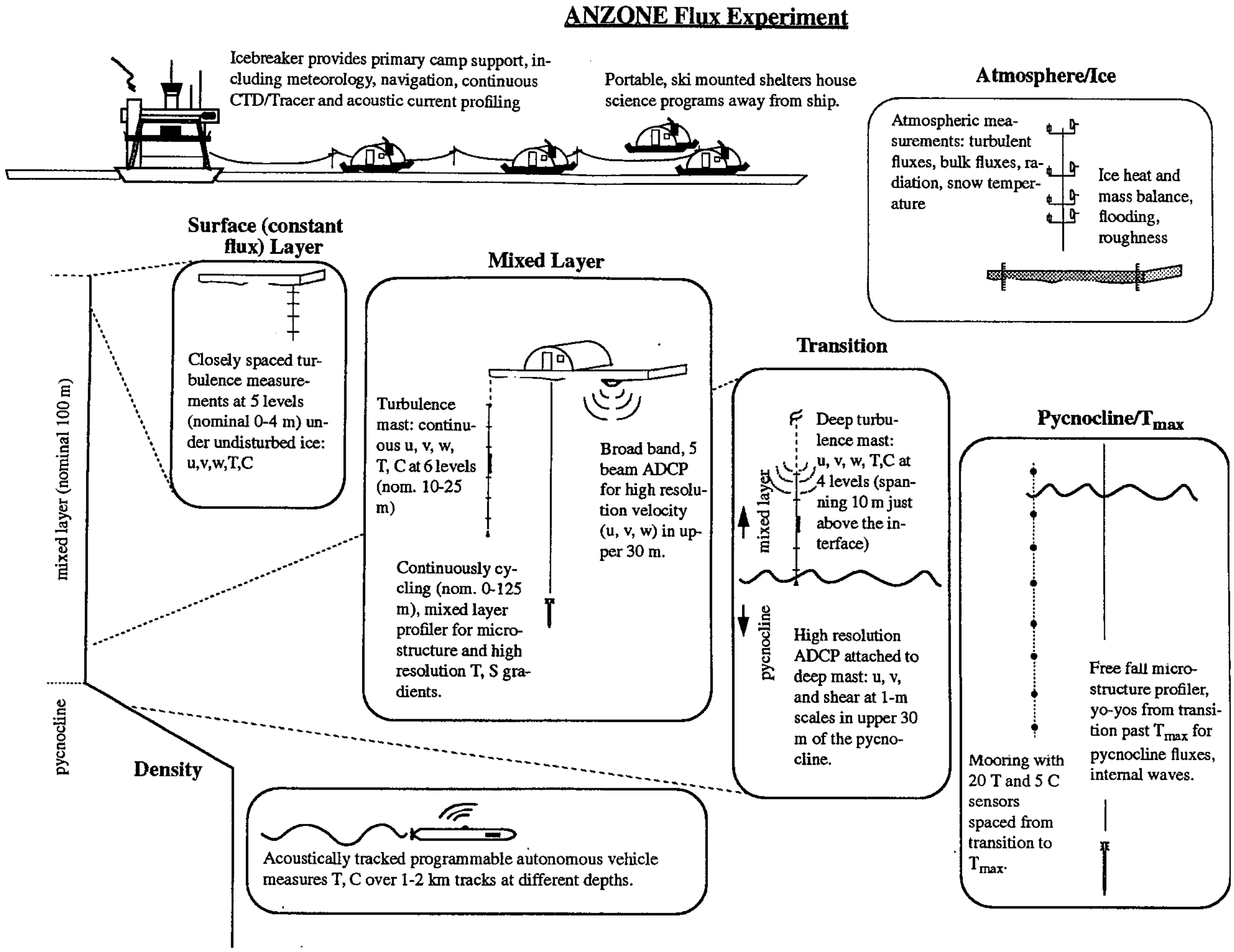

FIG. 1. Schematic of the ice camp layout and science programs during the ANZFLUX drift stations.

The cruise track of the R/V *Nathaniel B. Palmer* after its entry into the ice southeast of the South Sandwich Islands on 14 July 1994 is shown in Fig. 3 (McPhee 1995b). Ice property and CTD stations were made at nominal 100-km intervals on the inbound transect, during which the *Palmer* easily maintained 7-kt headway in the 0.5–0.7-m-thick ice toward the site of the "warm regime" drift, in a region where the maximum temperature in the water column (T_{max}) exceeded 1°C. While it may seem odd to characterize water in which the highest temperature barely exceeds 1°C as warm, the description was used by Gordon and Huber (1984) to distinguish relatively warmer deep water of Antarctic Circumpolar Current origin, spreading southwestward into the Weddell Gyre, from colder, outflowing water to the north. As the warm regime site was approached, the ship traced a cross pattern of more densely spaced stations centered on the prospective site, with AWI position/pressure buoys deployed at the corners of the cross (Fig. 3).

The ship was moored to a floe near the center of the buoy array late on 21 July 1994 (year day 202) to start the warm regime drift. By 0300 the next morning, the first ocean flux data were being recorded, with nearly all of the programs operational by the end of day 203. Although we had encountered several storms on the inbound transect and, in fact, deployed the camp as one cyclone wound down, we were nevertheless surprised by the intensity of the next two storms. The experience is perhaps best described by quoting from the weekly "Science Situation Report" filed with the Office of Polar Programs by M. McPhee (1994, unpublished manuscript) on 25 July 1994 in the aftermath of the first storm:

> At this time (Monday A.M.) we are winding down from one of those intense, short lived polar lows that seem to traverse this part of the world frequently. Yesterday was a blur of blow-

> ing snow in winds with gusts to 50 kt and ice drift speeds as high as 75 cm/s (1.5 kt). I think it fair to say that we are all rather awed by the oceanic response to this event. On Saturday, when winds were calm and ice drift slow, there was rapid freezing: unprotected hydroholes were growing several cm of ice in a few hours, and the ocean mixed layer was near its freezing temperature. By [Sunday] afternoon, previously frozen hydroholes were clear and the mixing shown by our instruments very intense. This morning the mixed layer temperature is 0.3 above freezing! with major excursions in the pycnocline level. Clearly, we are undergoing a massive upward heat flux event. Very exciting science, especially when combined with moving shelters, deploying instruments, and operating offship in full gale conditions. Our safety procedures appear to be working well, with a set schedule of radio checks and in general a high awareness of who is where.

The warm regime drift station was cut short during a second storm by an episode of cracking and pressure ridging near the ship's stern, late on day 208. For a while the ice there was quite active, calling for quick action to rescue threatened snowmobiles in the staging area off the ship's working deck, but then it appeared to stabilize. However, early the next morning, a new lead opened off the bow, prompting the decision to recover the camp and move on to the next drift. The retrieval was accomplished rapidly and without significant equipment loss.

We deployed buoys to be left drifting with the abandoned site and proceeded east to the vicinity of Maud Rise, an extensive bathymetric feature rising almost 3 km above the surrounding abyssal plain (Fig. 3). The region is often distinct from its surroundings, with anomalous hydrography [colder T_{max}, Bagriantsev et al. (1989)], circulation, and ice cover. The Weddell Polynya first appeared near Maud Rise, then drifted slowly westward. It has been conjectured that unique conditions there might again trigger extensive deep convection, and we were eager to investigate how the flux regime might differ from the warm regime to the west. Although the Maud Rise drift was deployed with some trepidation because we were unable to find ice thicker than about 30–40 cm, the camp survived its planned duration uneventfully and was recovered on 8 August. In general, winds and ice drift were less extreme during the Maud Rise drift, peaking at around 15 m s^{-1} and 50 cm s^{-1}, respectively. By comparison with previous experience in the Arctic, these values are still quite energetic.

After completion of the second drift, our plan was to proceed west–northwest into the "cold regime" outflow (eastward) part of the Weddell Gyre, where T_{max} is considerably less than farther south. We anticipated that this region, where the gyre outflow is marked by a much higher density of drifting icebergs, would experience less intense air–sea–ice heat flux than the warm regime. On 11 August, during the CTD/ice properties station transect toward the cold regime drift site, we encountered a very intense cyclone, with winds briefly reaching hurricane force (65 kt) late in the day. Earlier, during an extended CTD station with the ship snugged up against the ice in winds exceeding 50 kt, a flux measuring instrument cluster was deployed over the side, with a 2-h turbulence record obtained from middepth in the 100-m-thick mixed layer. It was almost eerie to operate from a ship in winds strong enough to make standing on deck a struggle, yet with no discernible vertical platform motion. The experience un-

FIG. 2. Maud Rise station on a rare calm day, with the R/V *Nathaniel B. Palmer* in the background. There is one additional shelter just to the right of the field of view. (Photograph courtesy of P. Guest).

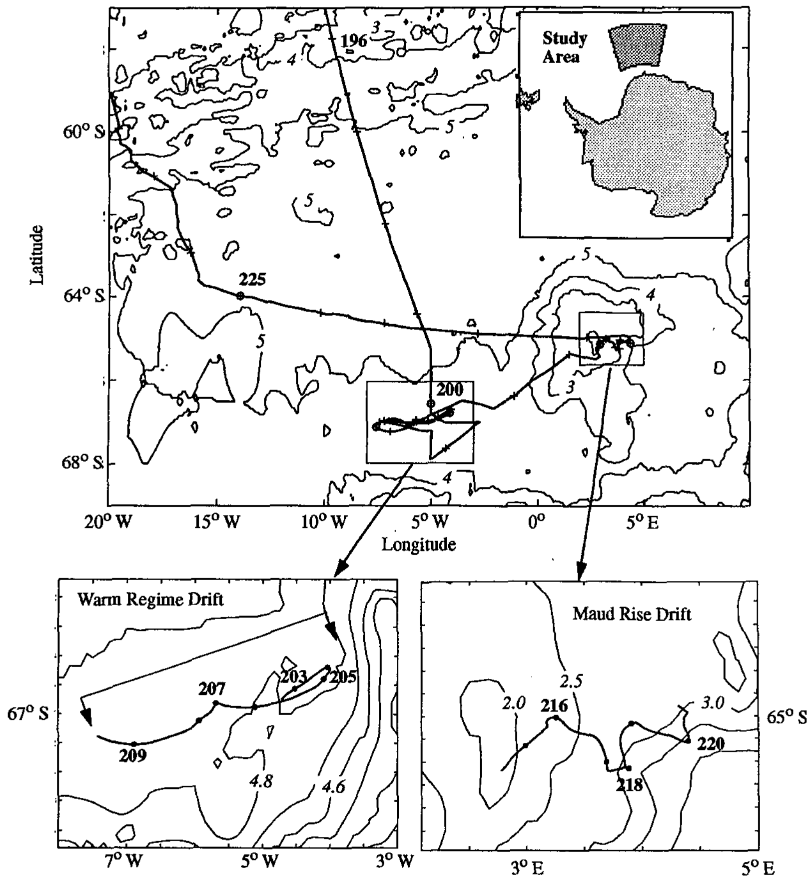

FIG. 3. Cruise track with details showing the two passive drifts, superimposed on the bathymetry contours (depth in kilometers). Bold numbers refer to the position at the beginning of the year day (196 is 15 July 1994). The section in the warm regime detail shows the view along the drift track projected in Fig. 5b.

derscored the notion, probably first advanced by Nansen (Ekman 1905), that sea ice provides an amazingly useful laboratory for studying upper-ocean processes under "controlled" conditions.

The big storm proved to be the undoing of the cold regime drift, however, because fuel consumption increased dramatically in its aftermath as the ship's headway fell to 3–4 kt in the compacted ice cover. To maintain enough fuel reserve to safely exit the ice, the third drift was curtailed to an extended (6 h) station, during which several of the ice camp systems were deployed directly from the ship. After that, we made for Punta Arenas, arriving in port on 24 August 1994 to end the experiment.

3. Selected preliminary results

Initial results from ANZFLUX have been discussed at a meeting of the principal investigators at the Naval Postgraduate School, Monterey, California, in March 1995; presented at the International Association for the Physical Sciences of the Ocean 21st Assembly, in Honolulu, Hawaii, in August 1995; and described in a series of short reports prepared for the *Antarctic Journal of the United States.* A sampling is presented here.

The thin ice cover responded rapidly to wind forcing (Fig. 4). Ice drift during the warm regime station averaged 3.8% of the wind speed referenced to the standard 10-m level (Guest 1995) for times when the wind exceeded 4 m s^{-1}. The direction of ice drift averaged 15.6° left of the wind vector. Over Maud Rise, corresponding values were 3.1%, 15.5° to the left. Martinson and Wamser (1990) found the ratio to be 3%, 23° left during similar drifts in the Weddell Sea during 1986; however, if their wind speed is adjusted to the 10-m level, their average drift ratio was about 3.2%. For contrast, the drift ratio for the much thicker, perennial ice pack in the Arctic rarely exceeds 2% in summer, when ice interaction forces are negligible, and often falls to 1% or less in winter (McPhee 1980). Over most of the Weddell Sea, thin ice and small geostrophic (sea surface tilt) currents ensure that momentum flux associated with the wind stress is often passed with little change into the upper ocean.

The most striking oceanic response occurred during the two gales that dominated the first, warm regime drift. Temperature structure of the upper ocean as observed with a thermistor mooring (Padman et al. 1995) is shown in Fig. 5, as a (panel a) function both of time and (panel b) drift displacement mapped onto a vertical plane aligned with the mean drift. The time plot is clearly dominated by a major upper-ocean warming (labeled *ii*) encountered during the morning of day 206. The 0°C isotherm rose rapidly by almost 100 m and remained within 40–50 m of the surface for the next day and a half. As shown below, upward turbulent heat flux in the mixed layer was intense while the station was over this feature. However, even before the station drifted over the first of the large thermocline displacements (labeled *i*), there was sizable turbulent heat flux deep in the mixed layer in response to the first storm. For several hours before 205.5, when the mixed layer was at least 100 m deep, turbulent heat flux measured with the deep mast (stationed between 63–72-m depth) exceeded 100 W m^{-2} (McPhee 1995a). The drift stalled over the warm-core feature in the calm between the two storms (Fig. 4), tending to emphasize the importance of (*ii*) relative to other thermocline displacements. Plotting contours versus distance (Fig. 5b) illustrates, however, that the horizontal scales of all three events were roughly comparable, although neither (*i*) nor (*iii*) had nearly as much impact on near-surface exchange. It is unlikely that these disturbances were generated locally, because their core temperature was considerably higher than the maximum temperature in the surrounding water; for example, average T_{max} in the half day between 207.0 and 207.5 was 1.28°C versus 0.97°C for the period 205.0–205.5.

Fine structure and direct covariance flux data collected during the buildup of the second storm of the warm regime drift illustrate the complexity of ice–upper ocean exchange during ANZFLUX (Fig. 6). There was considerable temperature structure in the "mixed layer" over the warm-core feature, as illustrated by closely spaced profiles (Fig. 6c) from the mixed layer microstructure profiler (Stanton 1995). Over the first day of the second storm (207–208), ice speed built to around 40 cm s^{-1} (Fig. 6d) above a mixed layer that was as much as 0.5 K above freezing. At the time, this led to much joking about the ice literally melting from beneath our feet, but the concern was not altogether frivolous, as demonstrated by the composite average of turbulent Reynolds stress and heat flux (Fig. 6a) measured at several levels with the upper turbulence cluster mast. Upward heat flux exceeded 100 W m^{-2}, which, roughly speaking, equates to ice melt of about 5 cm per day, or 10% of the total thickness. A bulk formula for oceanic heat flux based on the product of friction velocity and mixed-layer temperature above freezing (McPhee 1992) was applied to conditions observed during the composite average, providing an estimate of 118 Wm^{-2}, within 5% of the average covariance estimate in the mixed layer (113 W m^{-2}). This is a very large vertical heat flux in an ice-covered ocean. For comparison, during the ISW drift in the multiyear pack ice of the western Weddell Sea, the maximum observed heat flux was about 15 W m^{-2} (McPhee and Martinson 1994), and the mean value was around 2 W m^{-2} (Robertson et al. 1995). Indirect evidence for rapid bottom melting, which produces stabilizing surface buoyancy flux, is seen in the strong attenuation of Reynolds stress with depth observed during this time in the upper 20 m.

Fortunately, early on day 208, the ice station drifted off the warm feature, with the mixed layer both deepening and cooling rapidly (a clear indication that the change was advective rather than from vertical mixing). Composite Reynolds stress and heat flux measurements show a much different regime from earlier. Reynolds stress was more than twice as large near the peak of the storm and much more uniform with depth. Turbulent heat flux was *downward* and strongest near the surface. The mixed layer was above freezing (on average 0.05 K during the second composite average) so that the bulk formula still predicted upward flux of about 25 W m^{-2}, mainly because of the exceptionally strong surface stress. Thermal structure in the mixed layer (Fig. 6c) suggests why the heat flux was

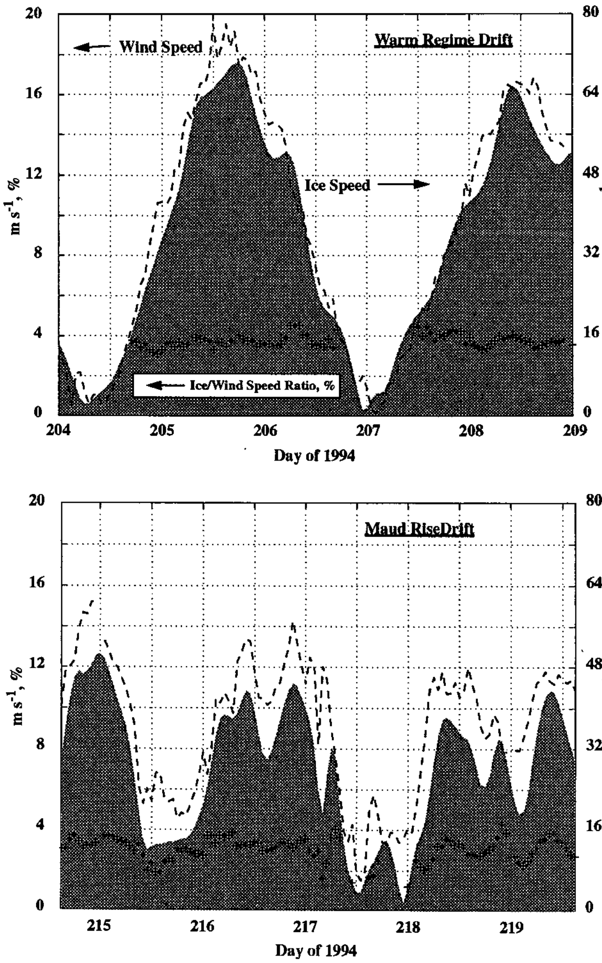

FIG. 4. Wind speed, adjusted to 10 m (dashed, left scale), and ice drift speed (shaded, right scale) for the two drift stations. Plus symbols indicate the speed ratio (%) for all hourly averages in which the wind speed exceeded 4 m s^{-1}.

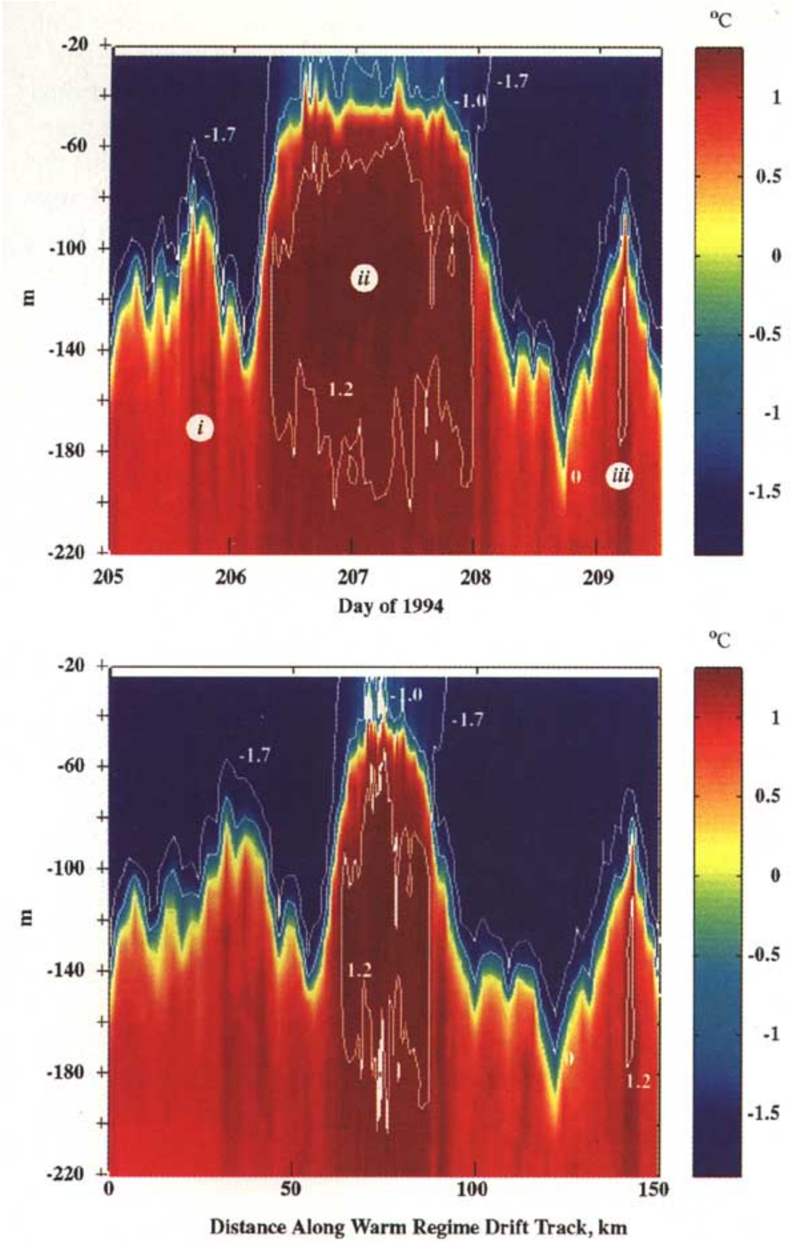

FIG. 5. Contours of upper-ocean temperature during the first warm regime drift, with abscissas representing both (a) time and (b) distance projected onto a plane along the drift track. The station drifted over three significant excursions in the thermocline (lowercase Roman numerals), characterized by relatively shallow mixed layers and warm T_{max} values (adapted from Padman et al. 1995).

downward: given strong shear in the under-ice boundary layer that developed during the storm, warm water near the surface over the eddy was advected preferentially with the ice over colder water. The ensuing positive temperature gradient, coupled with energetic mixing, caused downward turbulent heat flux despite a tendency for extraction of heat from the mixed layer at the ice–ocean interface.

In contrast to the large variation in oceanic heat flux associated with the storms, thermal exchange between the ice and the atmosphere was relatively insensitive to turbulent fluxes. It was dominated instead by the imbalance between downwelling and upwelling longwave radiation, which is primarily a function of cloud cover (Guest 1995). The upper surface is buffered from the large oceanic heat flux events by the thermal-to-latent energy transformation at the ice–ocean interface, as demonstrated by temperature gradients within the ice from one of five self-recording "freeze in" buoys (supplied courtesy of J. Wettlaufer) deployed during the first drift (Fig. 7). Gradients were calculated by finite differencing temperatures from thermistors spaced every 2 cm through the ice column and into the upper part of the ocean. The ice undersurface coincides with the region of sharp gradient above the blue area marked "Ocean." Over the course of the drift, bottom ablation totaled 8 or 9 cm at this site [there was considerable scatter among the sites, but this was representative (Ackley et al. 1995)]. Rapid melting occurred during a relatively short period beginning around 206.5, as the station drifted over the warm eddy. For the first half of 207, near-surface temperature in the ocean was very warm (Fig. 6c), but oceanic heat flux and melting were small because there was little shear-generated mixing. As the intensity of the second storm increased, the ice started melting rapidly again, slowing only after we had drifted back into cold water. Had the mixed layer remained warm for the entire second storm, it appears that the ice might indeed have become dangerously thin.

Despite the large changes at the lower surface, heat flux through the ice did not change drastically. As a rough estimate, conductive heat flux in sea ice (W m^{-2}) is about twice the negative thermal gradient (K m^{-1}). Thus, variation in the thermal gradient at 20 cm (middepth) in the ice column implies variation of 10–15 W m^{-2} in the upward conductive flux, about 10 times less than variation in the oceanic heat flux. Thin ice has little freeboard to begin with, so during storms the combination of shifting snow load and bottom melting often lowers the upper ice surface to below local sea level, with extensive areas of flooding (Ackley et al. 1995). Thus, the ice cover is replenished by freezing of slush at the surface, as well as by bottom growth whenever the oceanic heat flux is less than the conductive flux within the ice.

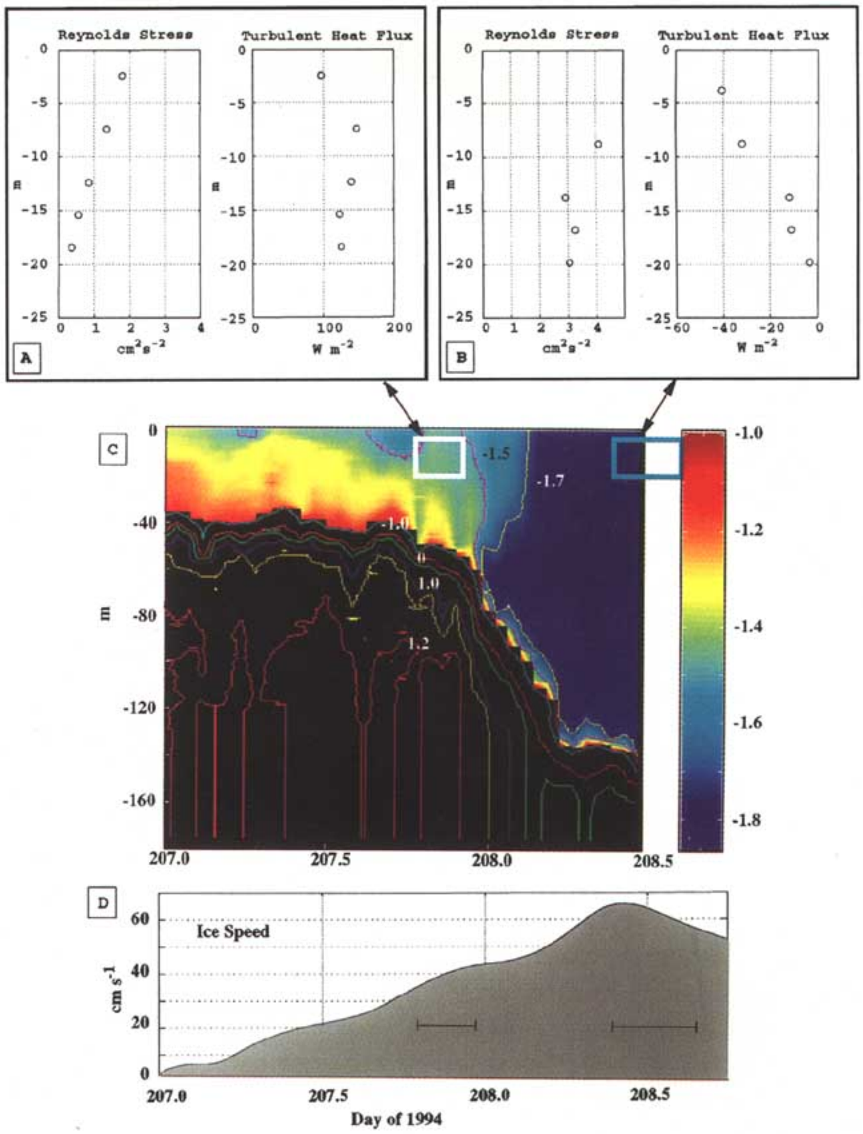

FIG. 6. Response of the upper-ocean system to the onset of the second storm, illustrated by the increase in ice velocity (d), during the warm regime drift station. The central panel (c) is thermal structure from a rapidly cycling microstructure profiler (Stanton 1995), where the color shading has been chosen to emphasize gradation in the mixed layer (temperatures above −1°C are shown by contours against the black background; vertical contour lines are a plotting artifact for depths greater than the depth to which the profiler penetrated). Turbulence characteristics from the instrument cluster mast in the upper mixed layer (McPhee 1995b) are shown by composite averages from five 1-h realizations, both (a) before and (b) after the station drifted off the warm-core feature. The uppermost level for Reynolds stress is missing in (b) because the horizontal current meter component failed during this time (with negligible impact on measurements of vertical heat flux). Flux values are approximate, pending postcalibration of the current meter clusters.

4. Summary

ANZFLUX was the first concerted effort to measure upper-ocean turbulent fluxes in the seasonal winter ice pack of the eastern Weddell Sea. The project was designed to approach the problem of air–ice–ocean interactions from several different points of view and included direct covariance methods for measuring turbulent flux (low wavenumbers in the turbulence spectrum, accomplished using a variety of current measuring techniques); microstructure methods (high-wavenumber measurements); high-resolution measurements of mean velocity, temperature, and salinity time series; an AUV for assessing horizontal variability in the ocean; plus careful monitoring of the ice energy and mass balance along with atmospheric energy and momentum fluxes. The resulting dataset is extensive and complex and at this point far from fully analyzed. Nevertheless, some principles are emerging from our experience during ANZFLUX that may affect how upper-ocean mixing in the Weddell Sea is viewed (and modeled).

First, our observations of several flux episodes indicate that variability in the wind appears to be at least as important as its mean value in determining fluxes in the upper ocean. In essence, at high latitudes it takes very powerful surface forcing to produce strong turbulent transfer near the base of a 100-m-thick mixed layer, so that a series of gales over a given period has

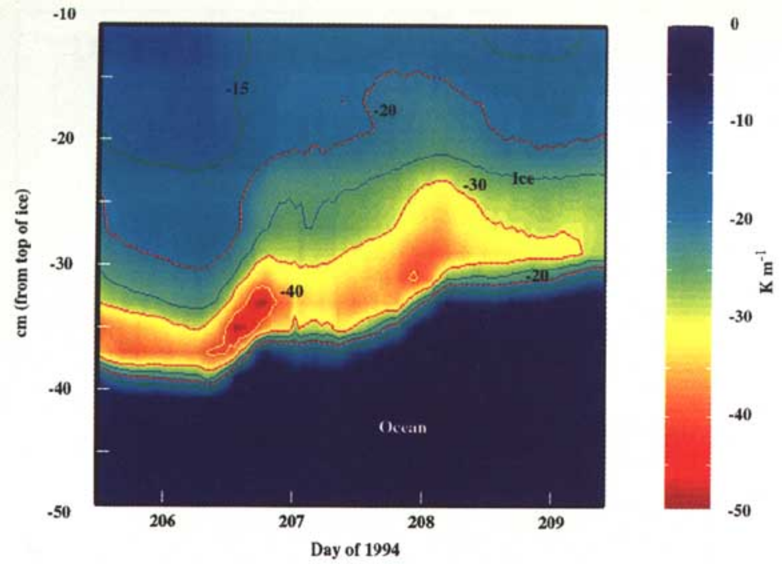

FIG. 7. Shaded contours of temperature *gradient* in the ice column from a vertical array of thermistors embedded in the ice. Large gradients in the snow/slush surface layer (upper 10 cm) have been excluded to emphasize gradients within the central part and near the base of the ice. Contour units are K m^{-1}.

a quantitatively different effect from a steady, stiff breeze with the same mean wind speed. While part of this may be as obvious as considering the difference between mean wind stress and mean wind speed, turbulent exchange between the well-mixed layer and the stratified upper pycnocline is a complex, apparently nonlinear process. In the central Weddell, where storms are frequent and intense, ocean-mixing parameterizations must properly account for conditions of very high stress to simulate the mean ice–ocean fluxes correctly.

Second, variability in the "weather" of the upper ocean underlying the mixed layer may play as important a role as wind in determining heat flux at the ice–ocean interface. The large-scale excursions in the thermocline apparent in Fig. 5 were clearly "advective" in the sense that observed changes in the upper ocean could not have resulted from local one-dimensional mixing processes. As we drifted over them, we observed large changes in mixed layer fluxes. Such events are not peculiar just to the warm regime region: they are a ubiquitous feature in drifting ocean buoy measurements in the central Weddell Sea (Morison 1995; C. Kottmeier 1994, personal communication). Their presence during the second drift over Maud Rise is illustrated by Fig. 8, which shows upper-ocean thermal structure, taken this time from a series of closely spaced yo-yo CTD casts (Huber et al. 1995). The value of T_{max} was considerably less than during the warm regime drift (Fig. 5), indicating a quite different thermal regime for the pycnocline over Maud Rise, yet variability in the elevation of isotherms is similar. The Maud Rise disturbances also had a major impact on the thermal structure of the mixed layer, as shown by elevation of water temperature above freezing (proportional to available heat content) in the upper contour plot. Figure 8 demonstrates the interplay between real weather and upper-ocean variability. At the start of the drift and on the last day, the surface stress was moderately strong (drift speeds in excess of 30 cm s^{-1}), but the mixed layer was deep and available heat small. Days 216 and 218 were also energetic, but the thermocline averaged 30–40 m shallower with direct impact on mixed-layer temperature elevation (which remained mostly uniform vertically, indicating active mixing). In the afternoon and evening of day 217, the mixed layer was still relatively shallow, but ice drift was sluggish and heat content small, presumably because of a lack of active entrainment at the base of the mixed layer.

The goal of ANZFLUX was to measure and understand the processes that maintain the relatively high flux of heat out of the deep ocean during winter in the central and eastern Weddell Sea. For the most part, our measurement program appears to have been very successful, and we are optimistic that a much improved understanding of turbulent exchange mechanisms will emerge from analysis of ANZFLUX data, particularly for high-stress environments. A key issue developing from analysis to date is variability in the pycnocline separating the mixed layer from the underlying ocean. Its source, magnitude, and extent promise to be subjects for future lively discussion and research.

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References

Ackley, S. F., V. I. Lytle, K. M. Golden, M. N. Darling, and G. A. Kuehn, 1995: Microwave backscatter measurements from first year pack ice in the eastern Weddell Sea. *Antarct. J. U.S.,* in press.

Bagriantsev, N. V., A. L. Gordon, and B. A. Huber, 1989: Weddell Gyre: Temperature maximum stratum. *J. Geophys. Res.,* **94,** 8331–8334.

Carsey, F., 1980: Microwave observations of the Weddell Polynya. *Mon. Wea. Rev.,* **108,** 2032–2044.

Ekman, V. W., 1905: On the influence of the earth's rotation on ocean currents. *Ark. Mat. Astr. Fys.,* **2,** 1–52.

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

Gill, A. E., 1973: Circulation and bottom water production in the Weddell Sea. *Deep-Sea R es.,* **20,** 111–140.

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

——, 1982: Weddell Deep Water variability. *J. Mar. Res.,* **40,** 199–217.

——, 1991: Two stable modes of Southern Ocean winter stratification. *Deep Convection and Deep Water Formation in the Oceans,* P.-C. Chu and J.-C. Gascard, Eds., Elsevier Oceanography Series, Vol. 57, Elsevier, 17–35.

——, and B. Huber, 1984: Thermohaline stratification below the Southern Ocean sea ice. *J. Geophys. Res.,* **89,** 641–648.

——, and ——, 1990: Southern Ocean winter mixed layer. *J. Geophys. Res.,* **95,** 11 655–11 672.

——, and Ice Station Weddell Group of Principal Investigators and Chief Scientists, 1993: Weddell Sea exploration from ice station. *Eos, Trans. Amer. Geophys. Union,* **74,** 121–126.

Guest, P., 1995: Surface heat fluxes in the eastern Weddell Sea during winter. *Antarct. J. U.S.,* in press.

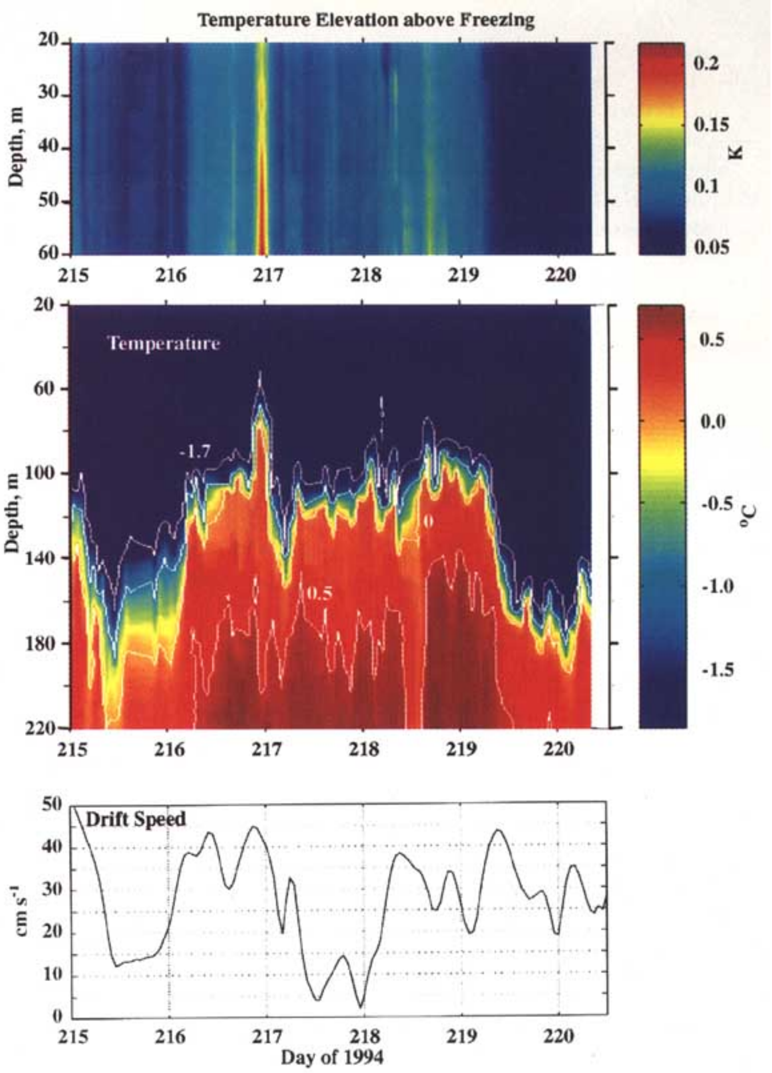

FIG. 8. (Top): $\Delta T = T - T_f(S)$, where T_f is the freezing temperature (at surface pressure) for the mixed layer (20–60 m) during the Maud Rise drift, from the yo-yo CTD program (Huber et al. 1995). (Middle): Shaded temperature contours in the upper 220 m. (Lower): Drift speed.

Huber, B. A., P. Schlosser, and D. G. Martinson, 1995: Thermohaline structure and tracer studies during ANZFLUX. *Antarct. J. U.S.,* in press.

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

LeadEx Group, 1993: The LeadEx Experiment. *Eos, Trans. Amer. Geophys. Union,* **74,** 393.

Lemke, P., W. B. Owens, and W. D. Hibler III, 1990: A coupled sea ice–mixed layer–pycnocline model for the Weddell Sea. *J. Geophys. Res.,* **95,** 9513–9525.

Lytle, V. I., and K. M. Golden, 1995: Microwave backscatter measurements from first year pack ice in the eastern Weddell Sea. *Antarct. J. U.S.,* in press.

Martinson, D. G., 1990: Evolution of the Southern Ocean winter mixed layer and sea ice: Open ocean deepwater formation and ventilation. *J. Geophys. Res.,* **95,** 11 641–11 654.

——, 1991: Open ocean convection in the Southern Ocean. *Deep Convection and Deep Water Formation in the Oceans,* P.-C. Chu and J.-C. Gascard, Eds., Elsevier Oceanography Series, Vol. **57**, Elsevier, 37–52.

——, and C. Wamser, 1990: Ice drift and momentum exchange in the winter Antarctic pack ice. *J. Geophys. Res.,* **95,** 1741–1755.

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

McPhee, M. G., 1980: An analysis of pack ice drift in summer. *Sea Ice Processes and Models,* R. Pritchard, Ed., University of Washington Press, 62–75.

——, 1992: Turbulent heat flux in the upper ocean under sea ice. *J. Geophys. Res.,* **97,** 5365–5379.

——, 1994: On the turbulent mixing length in the oceanic boundary layer. *J. Phys. Oceanogr.,* **24,** 2014–2031.

——, 1995a: Ocean flux measurements during ANZFLUX. *Antarct. J. U.S.*, in press.

——, 1995b: The Antarctic Zone Flux Experiment. *Antarct. J. U.S.*, in press.

——, and D. G. Martinson, 1994: Turbulent mixing under drifting pack ice in the Weddell Sea. *Science,* **263,** 218–221.

Morison, J. H., 1995: Autonomous vehicle and hydrographic buoy measurements during ANZFLUX. *Antarct. J. U.S.*, in press.

Muench, R. D., 1995: Upper layer current measurements in the eastern Weddell Sea during winter. *Antarct. J. U.S.*, in press.

Padman, L., R. A. Robertson, and M. D. Levine, 1995: Pycnocline structure and fluxes in the eastern Weddell Sea during winter. *Antarct. J. U.S.*, in press.

Rind, D., R. Healy, C. Parkinson, and D. Martinson, 1995: The role of sea ice in 2 × CO_2 climate model sensitivity. Part I: The total influence of sea ice thickness and extent. *J. Climate,* **8,** 449–463.

Robertson, R., L. Padman, and M. D. Levine, 1995: Fine structure, microstructure, and vertical mixing processes in the upper ocean in the western Weddell Sea. *J. Geophys. Res.*, **100,** 18 503–18 515.

Schlesinger, M., and J. Mitchell, 1985: The Potential Climatic Effects of Increasing Carbon Dioxide. U.S. Department of Energy, DOE/ER-0237, 81–148.

Schlosser, P., R. Bayer, A. Foldvik, T. Gammelsrod, G. Rohardt, and K. O. Munnich, 1990: Oxygen 18 and helium as tracers of ice shelf water and water/ice interaction in the Weddell Sea. *J. Geophys. Res.*, **95,** 3253–3264.

Stanton, T. P., 1995: Mixed layer structure and turbulent fluxes in the eastern Weddell Sea during the ANZFLUX experiment. *Antarct. J. U.S.*, in press.

Wadhams, P., M. Lange, and S. F. Ackley, 1987. The ice thickness distribution across the Atlantic sector of the Antarctic Ocean in midwinter. *J. Geophys. Res.,* **92,** 14 535–14 552.

Zwally, H. J., and P. Gloersen, 1977: Passive microwave images of the polar regions and research application. *Polar Rec.*, **18,** 431–450.

——, 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 Scientific and Technical Information Branch, NASA SP-459, 206 pp. [Available from Superintendent of Documents, U.S. Government Pronting Office, Washington, DC 20402.]