Elsevier Oceanography Series, 57

DEEP CONVECTION AND DEEP WATER FORMATION IN THE OCEANS

PROCEEDINGS OF THE INTERNATIONAL MONTEREY COLLOQUIUM ON DEEP CONVECTION AND DEEP WATER FORMATION IN THE OCEANS

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ELSEVIER

Amsterdam — London — New York — Tokyo 1991

TWO STABLE MODES OF SOUTHERN OCEAN WINTER STRATIFICATION

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ABSTRACT

The Southern Ocean is responsible for most of the world ocean's deep water characteristics. Cold Southern Ocean water masses are produced as buoyancy is removed by the polar atmosphere, usually in association with sea ice formation. Great amounts of sea ice formed within coastal polynyas are continually swept away by coastal winds, exposing the ocean surface to the cold antarctic atmosphere and further ice formation. Salt rejection forms dense shelf water masses, a key ingredient of Antarctic Bottom Water. In recent years a greater appreciation of winter processes over the deep ocean has developed. The extensive winter sea ice cover of the Southern Ocean is limited in its thickness by mixed layer entrainment of relatively warm and salty deep water. The same processes quickly disposes of the ice cover in the austral Spring. A network of negative feedbacks produces a relatively stable but thin veneer of sea ice. The sea ice cover and ocean static stability are maintained by salinity; this is referred to as the *saline mode*. The saline mode can be upset. The Weddell Polynya of the mid-1970's is a dramatic example of another stable mode. In the polynya condition the ocean stratification is destroyed and vigorous convection persists, eliminating the sea ice cover. This configuration is driven by the temperature instability and is referred to as the *thermal mode*. The conversion of the saline mode to the thermal mode requires the mixed layer salinity to become sufficiently high to force free convection with the deep water. It is unlikely that the Winter can be cold or long enough to force this conversion. Enhanced wind-induced sea ice divergence could accomplish this task, though that is difficult to reconcile a localized atmospheric effect with the repeated occurrences of small site specific polynyas. It is likely that the formation of offshore polynyas is driven by ocean processes, perhaps associated with circulation interaction with bottom topography. The persistent Weddell Polynya of the mid-1970s altered the thermohaline stratification of the water column to 3000 meters. Smaller, recurring polynyas observed since then are likely associated with sporadic convection events. These events slightly alter thermohaline stratification by injecting low salinity water into the mid-water column.

1 INTRODUCTION

Within the antarctic zone, stretching from the continental margins of Antarctica to the oceanic polar front (a span of roughly 35 million Km^2) the thermohaline stratification consists of cold surface water "floating", by virtue of its reduced salinity, over a much thicker stratum of warmer and saltier deep water (Bagriantsev et al. 1989). The top of the deep water is marked by temperature and salinity maxima and an oxygen minimum near 200 to 400 meters depth. The source of the warm/salty deep water mass is the circumpolar deep water, a blend of the world ocean's deep waters, including (but not exclusively composed of) the warm and salty North Atlantic Deep Water (NADW). The vigor of surface and deep water exchange within the antarctic zone to a large measure determines the ventilating powers of the Southern Ocean, the characteristics of the seasonal sea ice cover and the coupling of the polar ocean with the atmosphere. The ventilation potential is strongly coupled to the frequency and size of polynyas, both of the coastal and offshore varieties (Gordon and Comiso, 1988). Polynyas are persistent breaks in an otherwise ice covered region; they are centers of extreme ocean atmosphere interaction (Smith et al. 1990).

Deep ocean vertical exchanges of ocean properties across the pycnocline during the winter are large enough to be a critical factor in sea-air heat exchange, the spring removal of the sea ice cover and in limiting winter sea ice thickness to substantially less than 1 meter (Gordon, 1981; Gordon and Huber, 1990). Static stability of the surface water depends for the most part on the fresh water budget, including the sea ice formation, melting and divergence rates. It is quite weak; small perturbations in the various factors which influence the surface salinity budget can overwhelm the pycnocline stability, inducing vigorous convection. However, it is likely that a powerful network of negative feedback mechanisms preserve the thin veneer of winter sea ice in a more-or-less stable configuration (Martinson, 1990). I refer to this type of stratification as the *saline mode.*

The Weddell Polynya of the mid-1970's (Carsey, 1980) is a dramatic example that another stratification mode can develop when the static stability of the saline mode is removed. The saline mode may be upset by additional salt (density) input to the surface mixed layer. Potential mechanisms for additional salt input are: increased wind induced sea ice divergence; decrease in the precipitation minus evaporation difference; stronger topographic induced oceanic upwelling; or perhaps by a colder and/or longer winter season (this topic discussed in section 4). Another stratification mode occurs within offshore polynyas. In this mode, increase of vertical heat flux accompanying persistent vigorous deep reaching convection, eliminates the sea ice cover, forming the polynya. The removal of the insulating sea ice cover permits enhanced ocean / atmospheric fluxes as the sea surface is exposed to the polar atmosphere. Continued convection ensues as relatively warm deep water carried to the sea surface is cooled by the atmosphere and sinks. I refer to this "polynya" condition as the *thermal mode.*

Vigorous modification of ocean water also occurs over the continental shelf of Antarctica. Satellite data reveal the frequent presence of coastal polynyas which are production centers for enormous quantities of sea ice which is transported seaward, leaving the coastal ocean exposed to the Antarctic air masses (Cavalieri and Martin, 1985; Zwally, et al. 1985). The heat flux into the atmosphere is supported by latent heat of fusion. Coastal polynyas are responsible for production of cold salty shelf water, dense enough to convect into the deep ocean. Thin sheets of dense shelf water are observed convecting down the slope (Foldvik, et al., 1985a,b), forming an essential ingredient of Antarctic Bottom Water (AABW; Carmack and Foster, 1975; Foster and Carmack, 1976; Foldvik, et al. 1985). Portions of shelf water temperatures are depressed by several tenths of a degree below the one atmosphere freezing point due to ocean / glacial ice exchanges (Ice Shelf

Water; Jacobs et al. 1985). This slight cooling may promote AABW formation as cold water is more compressible than warm water and therefore more susceptible to deep reaching convection (Killworth, 1979).

The link between the "ventilation potential" and frequency of the various polynyas types is being explored with observational and modelling methods. In this paper I first present a review of the impact of the Southern Ocean on global ocean characteristics and then discuss circumstances that might lead to offshore polynyas and deep ocean ventilation. Many ideas are offered that hopefully will encourage further consideration by oceanographers studying deep reaching ocean convection. For a complete discussion of the Southern Ocean circulation and water masses, the reader is referred to the review papers: Carmack 1986; Gordon, 1988; Patterson and Whitworth, 1990. Jacobs (1989) discusses the continental margin oceanography, specifically how it pertains to sedimentation. The companion papers of Gordon and Huber (1990) and Martinson (1990) present details of Southern Ocean Winter mixed layer characteristics with a quantitative treatment of observations and by model, respectively.

2. SOUTHERN OCEAN RELATION TO THE GLOBAL OCEAN

2.1 Water Masses

The Southern Ocean directly influences two strata within the global ocean (Fig. 1): there is the chilling effect of Antarctic Bottom Water, AABW and the the freshening influence of Antarctic Intermediate Water (AAIW). In the "classical" schematic of the Southern Ocean circulation and water masses developed by Deacon (1937) deep water drawn from the world ocean upwells to supply mass, heat and salt to the surface layer. The surface water migrates away from the Antarctic Divergence (average latitude of 65°S) towards Antarctica to the south and towards the Polar Front (often called the Antarctic Convergence; average latitude of 53°S) to the north, where it contributes to the sinking and northward spreading of the two basic Southern Ocean water masses- AABW and AAIW, respectively.

(i) Antarctic Bottom Water. The lower 2000 meters of the global ocean is dominated by AABW. About 57% of the deep ocean is colder than 2° C (determined from the temperature-salinity volumetric data supplied by Worthington, 1981). This is the percentage of ocean which must be influenced to some measure by AABW as the other major deep water mass, NADW, has a characteristic temperature near or above 2°C. The more intense influences of AABW occur below 1°C, which marks a stratum of stronger thermal gradients, a benthic thermocline (Fig. 1; and see

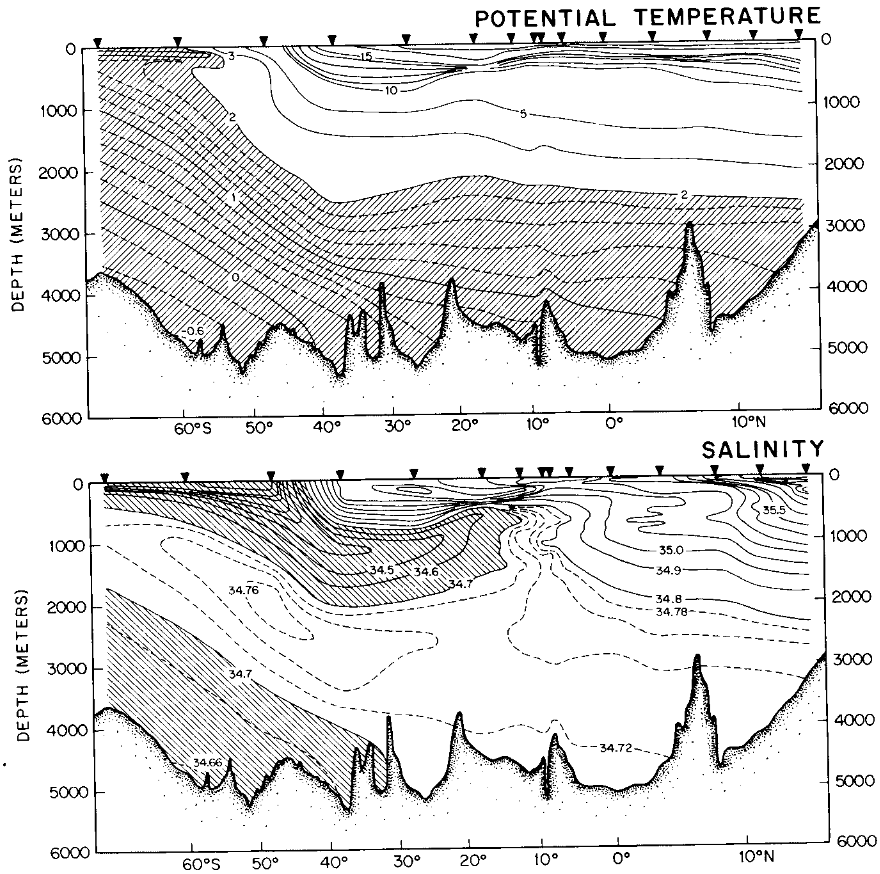

Fig.1. Potential temperature and salinity along the meridional plane in the western Indian Ocean, re-drawn from Spencer, et al., 1982. The hatched regions in each panel are vigorously influenced by Southern Ocean water masses. Southern Ocean water mass intrusion into the Indian Ocean is comparable to that of the Pacific and Atlantic Oceans.

the figures of Mantyla and Reid, 1983). Newly formed AABW is confined to temperatures well below 0°C, so its influence at higher temperatures requires mixing of purer forms of AABW into the overlying ocean. Mantyla and Reid (1983), who provide a review of the global spreading of AABW, conclude that the pure forms of AABW are hemmed in by circum-Antarctic ridges and it is the mixtures of AABW with deep water that invade the global ocean. The most concentrated forms of AABW intrude into each ocean basin along specific routes (Fig. 2); eventually the cooling effects spread throughout the world ocean from these advective features.

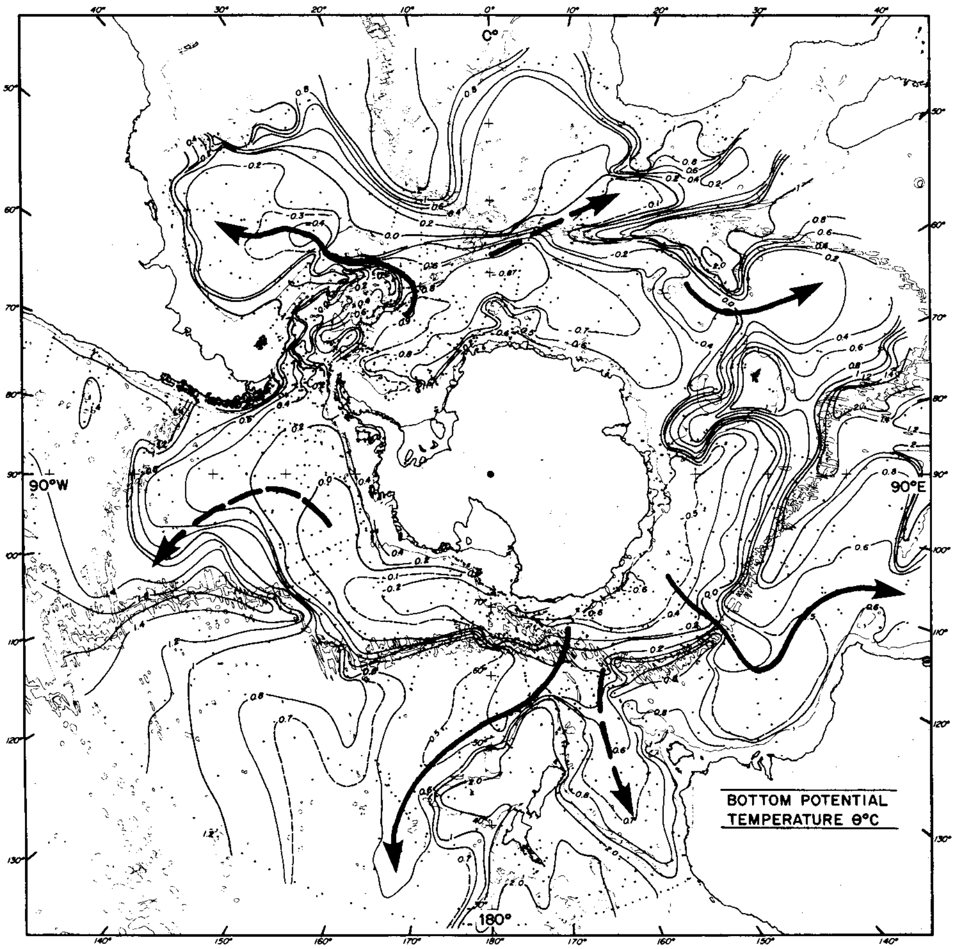

Fig. 2. Bottom potential temperature distribution of the Southern Ocean (Gordon and Molinelli, 1982, Plate 214). The primary escape routes of cold Antarctic Bottom Water into the global ocean are indicated by solid arrows. Northward flow into closed basins are shown with dashed lines.

There are two types of water mass formation leading to AABW: plume convection at the continental margins and open ocean convection.

The bulk of AABW is believed to be formed along the margins of Antarctica. The AABW formation process involves mixing of shelf and slope water across a well defined front, the Slope Front, in the vicinity of the shelf break (Foster and Carmack, 1976; Jacobs, in press). The Slope Front is observed around most of Antarctica, although the primary AABW production sites include the Weddell Sea whose output is a very cold variety of AABW and Ross Sea which contributes a somewhat warmer salty variety (Gordon, 1974; Jacobs et al. 1985). Other AABW formation sites along the continental margins of Antarctica are likely, including some where the convective plumes may not reach the sea floor, in what is called "not-quite-bottom water" by Carmack and Killworth, 1978.

Within the open ocean, there is an upper limit of mixed layer density imposed by the underlying deep water density. As this density is less than that achieved by continental shelf waters, open ocean convection cannot displace the products of the continental margins. On consideration of the cabbeling process, a mixed layer with salinity of 34.53 at the freezing point is sufficiently dense to initiate convection with the underlying deep water (Fofonoff, 1956; Carmack, 1986). Shelf water salinity above 34.7 at the freezing point is common (Gordon, 1974). The denser continental margin products spreading into the deep ocean, may be thought of as forming a "floor" for open ocean convection.

Water masses formed by open ocean convection have freer access to the north by isopycnal spreading and can pass above the ridge crests (Gordon, 1978, 1982). Perhaps it is the open ocean convective products that impose a greater constraint on deep ocean temperatures than continental margin produced AABW? The proportion of open ocean to margin water mass renewal determines the nature of Southern Ocean ventilation of the world ocean. Margin input may be quite steady as a reservoir of shelf water builds up over a period of years "behind" the confining walls of the Slope Front, while open ocean convection may be prone to more rapid changes (once the critical density is achieved convection quickly removes the anomalous surface water) and hence more sporadic in its occurrence. Open ocean convection most likely occurs every year but only in some years is it vigorous enough to massively cool the deep water. Formation of AABW may be strongly linked to the occurrence of coastal and offshore polynyas. Polynyas represent 'holes' in the insulating blanket of sea ice, with ocean/atmospheric heat flux 10 to 100 times that through the ice cover, which cool and elevate the salinity of the ocean (Smith et al. 1990).

(ii) Antarctic Intermediate Water. Salinity along the north-south plane within the western Indian Ocean (Fig. 1) reveals the freshening influence of polar surface water sinking and spreading near 1000-m (sigma-0 = 27.3) below the main thermocline. This is the AAIW (Deacon, 1937; Molinelli 1981; Piola and Georgi, 1982). Northward spreading of AAIW demarcates the lower boundary of the thermocline. The low salinity of AAIW allows cool water to be delivered to a low density stratum limiting the depth of the Southern Hemisphere's thermoclines, by limiting downward migration of low latitude heating. AAIW helps balance the global water budget by transferring polar excess precipitation into the evaporative subtropic regions. It also injects water of high oxygen into the lower thermocline, confining the subtropical oxygen minimum layer to the mid-thermocline.

A less dense variety of AAIW forms in the subpolar region just north of the polar front. It produces a water mass that invades the mid to lower thermocline, called Subantarctic Mode Water (SAMW, McCartney, 1977). It is drawn from deep winter mixed layer formed just north of the polar front zone and does not involve exchanges of water masses across the polar front zone. SAMW cools and freshens the thermoclines of the Southern Hemisphere at levels above that of the AAIW. SAMW injected into the Atlantic is relatively warm, near 14°C. SAMW of the Indian Ocean is about 10-12°C (its effect can be seen as a slight decrease in the vertical temperature gradient near 30°S on Fig. 1). In the Pacific SAMW is near 8°C in the west and 5°C in the east. Might AAIW represent the end-product of SAMW as suggested by McCartney? Georgi (1979) and Piola and Georgi (1982) consider this question, concluding that it is unlikely that the 5°C SAMW of the southeast Pacific flowing into the Atlantic could, with further modification by sea-air fluxes, account for AAIW. They find significant input of surface water from south of the Polar Front is required to achieve the conversion of SAMW into AAIW. However, the broad and multi-water type characteristic of the S-min in the southwestern Atlantic indicates a complex history for AAIW (Piola and Gordon, 1989) and the issue of the origin of the AAIW salinity minimum in the Atlantic and other oceans requires further research.

While the specific water types of AABW and AAIW are confined to a narrow density range, vertical mixing can carry their effects into other density strata. Additionally, the array of water masses of the Southern Ocean margins covers the full range of deep ocean density (Fig. 3), thus there is the potential for direct spreading of Southern Ocean water into the global ocean interior by isopycnal or neutral spreading surfaces, a point discussed in the next section.

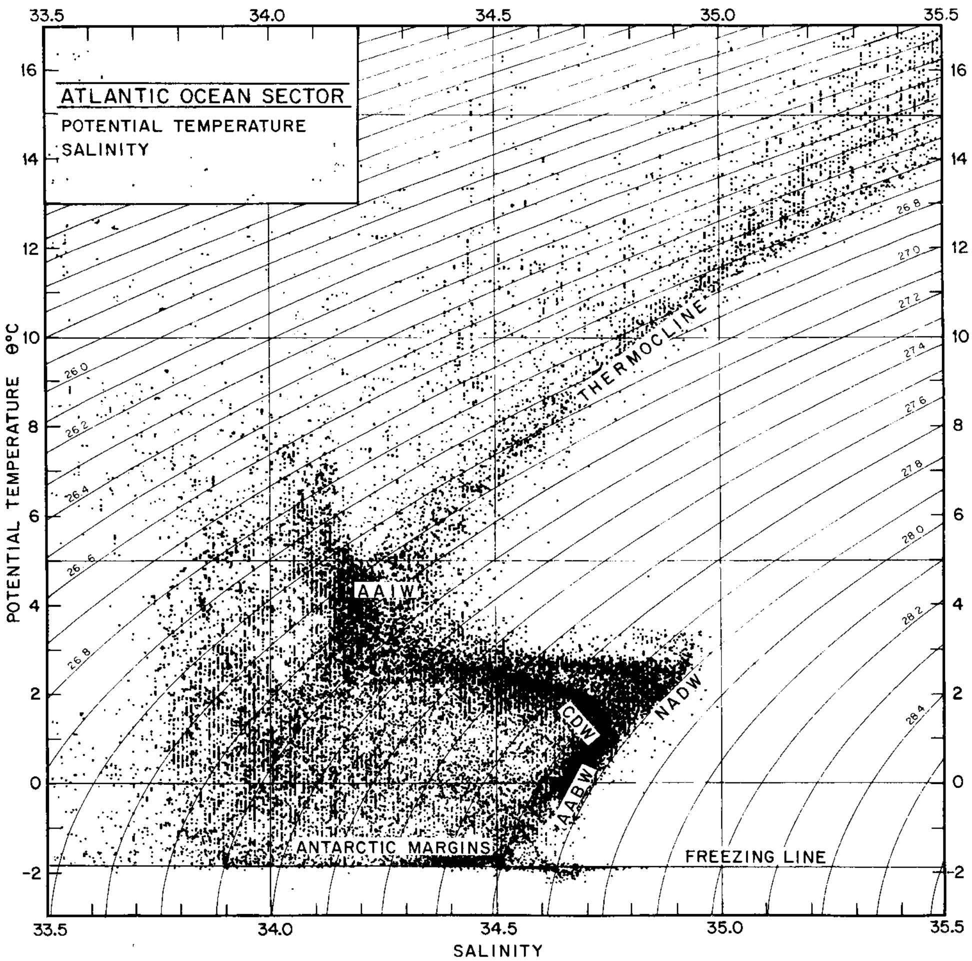

Fig. 3. Potential temperature versus salinity for the Atlantic sector of the Southern Ocean (south of 30°S). The deep waters of the ocean are labelled CDW-AABW (CDW- Circumpolar Deep Water; AABW- Antarctic Bottom Water). The coldest water represents the continental margin region, where waters dense enough to interact with the CDW-AABW density horizons are present.

2.2 <u>Circumpolar Belt</u>

Significant meridional advection of ocean water, outside of the Ekman Layer, requires a western boundary to supply the required vorticity. The lack of continuous western boundary across the deep circum-Antarctic belt disallows advection of warm upper layer waters into the Southern Ocean. Establishment of the deep ocean band with the resident zonally flowing powerful Antarctic Circumpolar Current (ACC), inhibits meridional advection which tends to thermally isolate Antarctica. Sea ice and some continental glaciation probably began 38 million years ago as the circumpolar band was forming, with a persistent glacial ice sheet developing 11-14 million years ago (Kennett, 1977) as the circumpolar belt became fully established. With the cooling of the Southern polar regions came cooling of the deep ocean. Deep ocean temperatures at the end of the Eocene 38 million years ago were well above 10° C, falling below 5°C by mid-Miocene, 13 million years ago to the present day values of near 0°C (Shackleton and Kennett, 1975a, b).

The circumpolar oceanic belt has two competing effects: it isolates the Southern Ocean from the rest of the ocean, which limits its influence, but it also allows what does spread to the north to attain colder temperature increasing its chilling global impact. As the "completeness" of the circumpolar belt decreases with increasing depth, southern ocean water mass access to the ocean north of the circumpolar belt depends on depth or density horizon. At shallow depths meridional spreading may be limited to eddy processes and Ekman Layer transport. At deeper levels topographic channels guide deep and bottom currents across latitudes. At great depth, isolated pools of water without direct access to the north occur; their properties can spread northward only if they can mix vertically to shallower levels. Using a global ocean model, Cox (1989) concludes that establishment of the ACC limits northern penetration of cold antarctic water masses into the northern oceans. Of course, were there no ACC there probably would not be any cold polar water sufficiently dense to influence the deep global ocean. On balance it is likely that global spreading of cold Southern Ocean water masses, albeit hindered by the circumpolar belt, has a fundamental chilling effect on the global ocean.

It is likely that the present day circulation pattern with effective deep ocean cooling is related to the ice sheet reaching the coastal region (Hays, 1967). A fully glaciered Antarctica allows very cold air to reach the coastal ocean, forming cold shelf water that feeds formation of AABW (Gordon, 1971). An additional effect may be the depression of continental margins under the weight of the glacial ice sheet. The Antarctic continental shelf averages 500 m (Jacobs, 1989), significantly greater than the other continental blocks. This allows contact of the cold shelf water with the warmer open ocean deep water at elevated hydrostatic pressure across the base of the shelf-slope front, particularly within deeper shelf channels, such as the Filchner Depression of the Weddell Sea. The greater compressibility of cold water has an interesting effect in the Southern Ocean (Killworth, 1979): cold water not dense enough to convect into the deep ocean at one atmosphere pressure, can do so if it can be depressed in depth by a matter of a few hundred meters. Formation of very cold, dense shelf water coupled with a deep continental shelf may act together to encourage deep convection and cool the global ocean.

While warm upper layer water is prohibited from advecting into polar regions, the geostrophic balance of the ACC allows deep and bottom water to have isopycnal access to the cold polar atmosphere (Fig. 3). Density characteristic of the sea floor north of the ACC is found near the sea surface to the south of the ACC. Exposure of this dense water to the cold antarctic atmosphere produces cold water which can then mix along isopycnal (or more precisely, neutral surfaces as defined by McDougall, 1987 should be applied to the Southern Ocean condition) surfaces into the deep and bottom strata of the world ocean. It is interesting to note that with the present day density field, the baroclinic transport of the ACC, which amounts to about 90 x 10^6 $m^3 sec^{-1}$ (90 Sv) or 70% of the full transport (Whitworth, 1983; Whitworth and Peterson, 1985), is at a maximum. An increase in the ACC baroclinic component would require an increase in the strength of the stratification within the Southern Hemisphere subtropical water column. Without changing the baroclinic field of mass the ACC transport can increase only within the barotropic mode. The similarity of benthic density of the global ocean and the densest water exposed to the antarctic atmosphere represents coupling of the overall wind and buoyancy driven circulation.

3 ANTARCTIC ZONE STRATIFICATION AND POLYNYAS

The sea ice cover of the Southern Ocean acts to decouple the ocean from the atmosphere, limiting cooling of the ocean by the polar atmosphere. The insulating blanket of sea ice protects the ocean from the cold atmosphere. The extreme seasonality and rapid spring melting suggests an ocean heat role in Southern Ocean sea ice budget (Gordon, 1981): the build up of heat within the mixed layer induces melting even before the atmospheric radiation balance and temperatures warm sufficiently to melt ice directly. Ocean heat flux also limits sea ice thickness during the winter (Gordon and Huber, 1984, 1990).

As mixed layer / deep water exchange with associated vertical oceanic heat and salinity flux is responsible for the spring melt and limited winter sea ice thickness, we may consider that the vigor of Southern Ocean ventilation potential is directly related to sea ice seasonality, e.g. year-round constant sea ice cover is indicative of a strongly stratified ocean with small vertical heat flux, whereas a strongly seasonal ice cover is linked to substantial vertical oceanic fluxes, which melts the ice cover as the spring atmosphere heat budget cannot remove the oceanic heat flux. Variations in vertical heat flux are expected to yield interannual changes in ice cover extent and seasonality.

Extreme events do occur, such as the drastic cooling of the water column down to 3000-m within the Weddell Gyre (near Greenwich Meridian and 63°S) during the Weddell Polynya event of the mid-1970's (Fig 4; also see Bersch, 1988). Salinity and density profiles indicates the cooling was accomplished by deep reaching convection. While the large Weddell Polynya was observed only in the mid-1970's, small short lived polynya events are common (Comiso and Gordon,

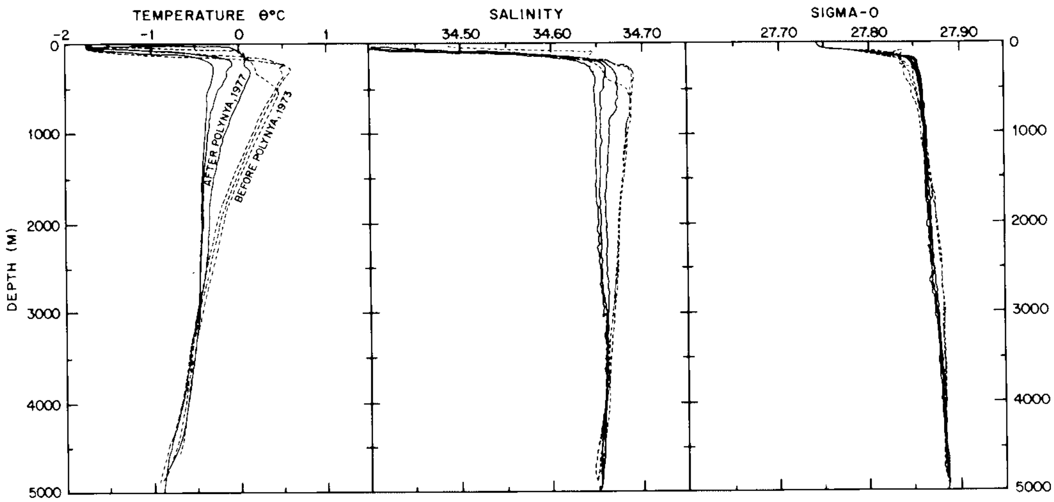

Fig. 4. Potential temperature, salinity and density profiles before and after the Weddell Polynya occurrence of 1974-1976 (data from the central region of the Weddell Gyre, redrawn from Gordon, 1982).

1987). It is likely that these transient polynyas are coupled to sporadic deep reaching convective cells .

There may be signs of sporadic convection in the details of the deep potential temperature (θ)/ salinity (S) relationship. The θ/S between the salinity maximum stratum of the Weddell Deep Water (WDW) and the AABW cold end-member while nearly linear within the Weddell Gyre, does display slight non-linear structure. The θ/S structure is slightly curved, with a salinity deficit relative to a straight line (Fig. 5). The greatest negative salinity anomaly occurs near the -0.2°C level amounting to only 0.005 relative to the reference straight line between warm and cold end-

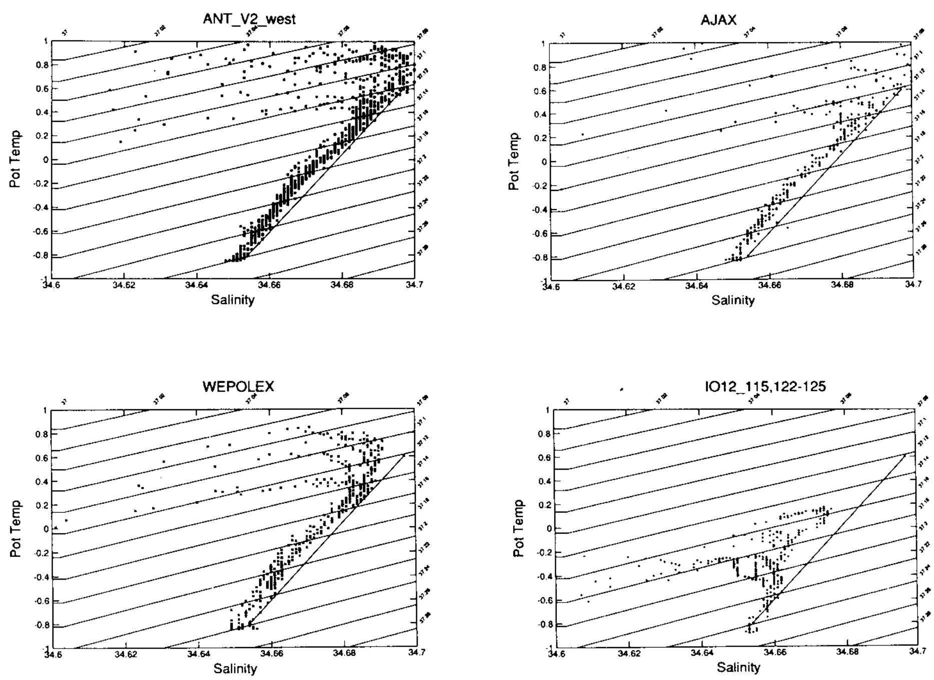

Fig. 5. Potential temperature versus salinity for the deep and bottom water of the central region of the Weddell Gyre. A reference straight line connecting the warmer and colder end-members along their saltier edge of scatter, is placed on the θ/S of representative stations from: ISLAS ORCADAS cruise 12-77 (1977) stations within the region cooled by the 1970's Weddell Polynya (same stations as shown in Fig. 4, with the addition of station 115, which was obtained within a remnant convective chimney of the previous Winter, Gordon, 1978); WEPOLEX, 1981 (Huber et al. 1983); Ajax, 1984 (Ajax Data Report, 1985); and the 1986 data set from POLARSTERN ANT V/2 (Huber et al. 1989).

members. The negative salinity anomaly is barely perceptible but it is present in all of the modern high precision CTD expeditions crossing the Weddell Gyre near the Greenwich Meridian. The ISLAS ORCADAS data set of 1977, obtained after the three year occurrence of the persistent Weddell Polynya, shows extreme low salinity within the deep water (also see Fig. 4). Station 115 was obtained within what was considered a remnant of a convective chimney associated with the Weddell Polynya (Gordon, 1978). The salinity anomaly of 1977 amounting to 0.01 is centered near -0.2°C.

Gordon (1978) suggested that the vigorous convective event of the mid-1970's associated with the Weddell Polynya, drastically modified the deep water, centered at the -0.2°C level, by injecting low salinity surface water into the deep water. In subsequent years, during which there were only brief, small polynyas near the Greenwich Meridian (Comiso and Gordon, 1987), sporadic convection only slightly modified the deep water from a simple straight line mixing curve between WDW and AABW end-members. It is hypothesized that the WDW/AABW θ/S structure is the product of three water types: 1- warmest is the WDW drawn into the Weddell Gyre from the circumpolar deep water; 2- the coldest is AABW formed along the margins of Antarctica; and 3- the middle layer is due to an injection of low salinity surface water by convection. The convection is usually too weak to greatly alter the deep water θ/S , except during persistent large Weddell Polynya events. This hypothesis of sporadic deep convection as the origin of the slight θ/S inflection must be evaluated with high precision data and perhaps with models. It should be evaluated against possible lateral injection of low salinity water, such as the "not-quite-bottom - water" concept of Carmack and Killworth (1978).

4 STAGES OF WINTER AND FORMATION OF OFFSHORE POLYNYAS

Winter observations from the POLARSTERN along the Greenwich Meridian from the ice edge to the Antarctic margin show the mixed layer beneath the winter sea ice cover is "contaminated" with WDW (Gordon and Huber, 1990). WDW introduces heat and salinity into the surface layer which strongly influences the mixed layer stability, sea-air exchanges and sea ice formation processes. The total WDW transfer into the mixed layer along the Greenwich Meridian from the ice edge to the margins of Antarctic averages 45 meters/year (40% of the mixed layer thickness), occurring mostly during the Winter season at a rate as high as 0.4 m/day. The winter heat flux of approximately 40 W/m^2 limits ice thickness to about 55 cm, agreeing quite well with observations (Wadhams et al. 1988). The air temperatures measured during the 1986 POLARSTERN cruise are just sufficient to remove the WDW heat input in the presence of observed ice thickness and concentration. This suggests that the sea ice cover and WDW heat input into the mixed layer are in approximate balance by mid-winter. Late winter measurements indicate that deep water entrainment into the mixed layer continues beyond mid-winter (Gordon and Huber, 1990). Can the WDW deliver enough heat and salt to eventually remove the ice and induce a polynya, i.e. a switch from the saline to the thermal mode?

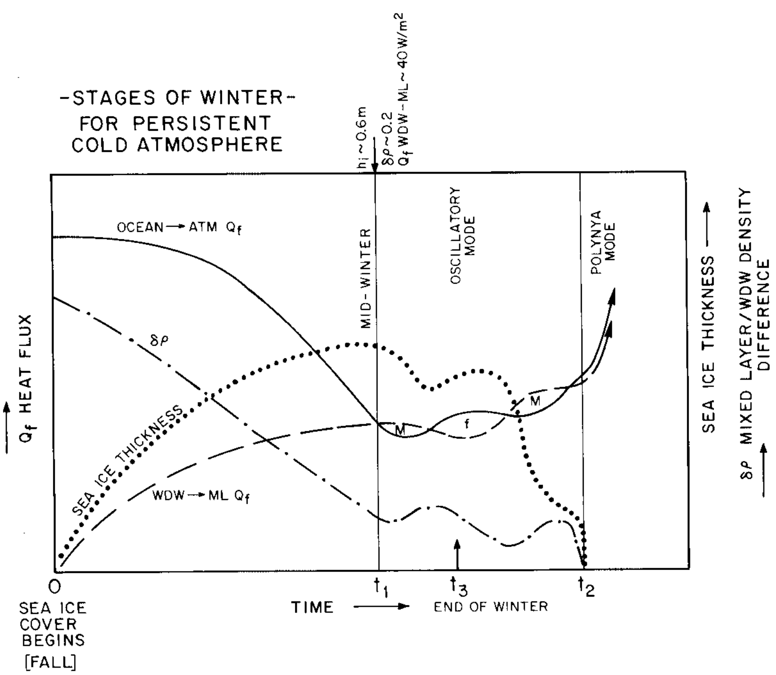

Fig. 6. Schematic of the possible fluxes during the winter period, see text. The schematic depicts expected trends of ice thickness, mixed layer static stability, and vertical heat fluxes for a constant atmosphere. Vertical scales are not shown, it is more the concept that is offered at this point. t_1= time (mid-Winter) when an equilibrium is reached between heat flux, sea ice thickness and air temperature. t_2 = start of an offshore polynya, the switch to a thermal mode of stratification. t_3 = the end of winter, when air temperatures warm.

A schematic representation of the events during winter which might trigger a switch from the saline to thermal mode is presented in Fig. 6. The schematic depicts expected trends of ice thickness, mixed layer static stability, and vertical heat fluxes for a constant atmosphere. Vertical scales are not shown, it is the concept that is offered at this point. Various mixed layer models deal with aspects of this process. See Martinson, 1990 for a model which described to the conditions encountered during the 1986 POLARSTERN expedition.

The 1986 data suggest that an equilibrium is reached between heat flux, sea ice thickness and air temperature by mid-winter (t_1). At this point heat transfer from WDW to the mixed layer balances heat loss to the atmosphere through leads and by conduction through the ice. After time t_1 as WDW to mixed layer transfer continues, the salt build up within the mixed layer decreases the density contrast across the pycnocline, allowing increased entrainment and inducing sea ice melt. Ice melt stabilizes the mixed layer, reducing WDW entrainment. If the air remains cold there can be a re-growth of sea ice, though not to its former thickness. Oscillation between ice growth and melt is expected to occur until the end of winter, t_3, when the air temperature increases, stopping further ice growth.

The point when the build up of mixed layer salt removes the pycnocline and a polynya ensues is marked by t_2. Does t_3 (the end of the winter cold) occur earlier or later than t_2 (start of an offshore polynya)? How long would it take to attain polynya state? A simple calculation based on the 1986 WDW entrainment rates indicates that with a daily entrainment of WDW into the mixed layer of approximately 0.4 m/day, many 100's of additional Winter days beyond the mid-Winter point of t_1 are required to substantially boost the mixed layer salinity and remove the pycnocline. This time is too long to attain a polynya thermal mode state within a single winter. Even with additional entrainment permitted by the diminishing pycnocline strength (mean of 1.0 m/day) and by sea ice divergence (half of the ice thickness is carried away and does not melt locally) it would still require three months of additional "Winter" months to reach convection. Clearly careful modelling of the coupled ocean / sea ice/ atmosphere system is needed, but it appears that unless: 1- the winter becomes much colder or longer, or 2- the wind field can spin-up the baroclinic flow sufficiently to shallow the pycnocline or increase sea ice divergences for specific regions, or 3- as Martinson, 1990 points out- WDW becomes saltier without a matching increase in temperature, the network of negative feedbacks which presently provide stability to the saline mode of stratification cannot easily be broken within the confines of a single winter. A multi-year build up of mixed layer salinity might be required to initiate a polynya condition.

While it seems unlikely that the present saline mode of stratification can be overcome with the present ocean / sea ice / atmosphere coupling, offshore polynyas do occur, so the thermal mode can be achieved (see Martinson, 1991, for further discussion).. Gordon and Huber (1990) suggest that excess salt required to destabilize the stratification is derived from an external source. This source may be enhanced upwelling over topographic features, e.g. Maud Rise.

As a consequence of circulation/topography interaction, the Maud Rise water column stands out as an anomaly relative to the surrounding region, with a significantly more saline and dense mixed layer. The saline mixed layer probably represents a salt build up of more than a single Winter. The Maud Rise mixed layer salinity represents a balance between enhanced upwelling and exchanges with the surrounding region. Below the mixed layer the water column over the crest of the Rise is identical to that over the flanks if the latter water column is upwelled by 400 meters. This uplifting is believed to be a response of the upstream flow encountering the Rise. Increased upstream flow would be expected to increase Maud Rise upwelling and the dependent salinity (density) of the mixed layer (e.g. Hubbert and Bryan, 1978). Slight increases in the mixed layer density could trigger a convective thermal mode and generation of a polynya.

It is hypothesized that spin up of the Weddell Gyre's barotropic circulation induced by an increase of the regional wind stress curl would enhance the probability of polynya development over the flanks of Maud Rise. The Weddell Polynya of the mid-1970's and transient polynyas since then, form over the northern flank of Maud Rise, which is expected to be the primary route for westward flowing water encountering Maud Rise.

The sequence of events leading to a polynya may be as follows-

1- An increase of the regional wind stress curl spins up the barotropic circulation of the Weddell Gyre, with a response time on the order of months. A slower baroclinic response also works towards the thermal mode as gyre spin-up shoals the pycnocline, but with time scales of years or decades;

2- Upwelling over topographic features, such as Maud Rise, increases;

3- The vulnerability to a switch-over from the saline (sea ice covered) mode to the thermal convective (polynya) mode is enhanced as more salt is delivered to the overlying mixed layer by the topographic induced upwelling.

4- The thermal mode continues, until fresh water invades the convective region and damps out the convection (Martinson, et al. 1981). The temporary polynyas are small enough for large storms to 'blow' in enough of the surrounding ice to shut-down convection. For large polynyas individual storms may not be sufficient; the large Weddell Polynya was shut down as it drifted with the mean circulation into a convergent sea ice region of the western Weddell Gyre (Martinson, et al. 1981).

5 CONCLUSIONS

Southern Ocean water masses ventilate the global ocean. They have played this role since the establishment of the deep reaching circum-Antarctic oceanic belt. Formation of Southern Ocean water masses are associated with exposure of the ocean to the frigid Winter atmosphere within coastal and offshore polynyas. The coastal polynyas may be the primary feature responsible for formation of the densest Southern Ocean water mass, the AABW. Offshore, the stratification is occasionally altered from the common salinity stabilized mode to the convectively active thermal mode. The thermal mode produces an offshore polynya. This is achieved by introduction of excess salt into the winter mixed layer. The most likely candidate to accomplish this task is eddy generation and upwelling induced by circulation and topography interaction. Deep reaching convection probably occurs each year, but massive overturning capable of greatly altering the deep water θ/S structure is reserved for the persistent offshore polynya events, represented by the Weddell Polynya of the mid-1970s.

Acknowledgements-

The author's Southern Ocean research is presently supported by the National Science Foundation Division of Polar Program grant DPP 90-08906. Additionally NASA grant NAGW 1752 covers study of polynya occurrences as seen in satellite microwave images. Lamont-Doherty Geological Contribution number 4721.

References-

Ajax Data Report, 1985. Physical, Chemical and In-Situ CTD Data from the Ajax Expedition in the South Atlantic Ocean, aboard RV KNORR Leg I: 7 October - 6 November 1983 Leg II: 11 January - 19 February 1984. University of California Scripps Inst. of Oceanography and Texas A&M Univ. Dept. of Oceanography. SIO Reference 85-24, TAMU Reference 85-4-D.

Bersch, M. 1988. On Deep Convection in the Weddell Gyre, Deep-Sea Res., 35(8A): 1269-1296.

Bagriantsev, N. V., A. L. Gordon, and B. A. Huber, 1989. Weddell Gyre temperature maximum stratum. J. Geophys. Res., 94(C6): 8331-8334.

Carmack, E. C. 1986. Circulation and Mixing in Ice-Covered Waters. In: "The Geophysics of Sea Ice" ed. N. Untersteiner, Plenum Press, pages 641-712.

Carmack, E. C. and T. D. Foster, 1975. On the flow of water out of the Weddell Sea, Deep-Sea Res., 22: 711-724.

Carmack, E. C. and P. D. Killworth 1978. Formation and interleaving of abyssal water masses off Wilkes Land, Antarctica. Deep Sea Res. 25: 357-369.

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

Cavalieri, D. J. and Martin, S., 1985. A passive microwave study of polynyas along the Antarctic Wilkes Land Coast. Oceanology of the Antarctic Continental Shelf, Ant. Res. Ser. 43, Am. Geophys. Un., Wash. D.C., pp. 227-252.

Comiso, J. C. and A. L. Gordon, 1987. Recurring polynyas over the Cosmonaut Sea and the Maud Rise. J. Geophys. Res., 92(C3): 2819-2833.

Cox, M. D. 1989. An idealized model of the world ocean. Part I: The global-scale water masses. Jour. Phys. Oceano. 19(11): 1730-1752.

Deacon, G.E.R. 1937. The Hydrology of the Southern Ocean. Discovery Reports XV:1-124.

Fofonoff, N. P. 1956. Some properties of sea water influencing the formation of antarctic bottom water. Deep Sea Res. 4: 32-35.

Foldvik, A., T. Gammelsrod, and T. Torresen, 1985a. Circulation and water masses on the southern Weddell Sea shelf. In: Oceanology of the Antarctic Continental Shelf, Ant. Res. Ser., 43, Am. Geop. Un., Wash. pp. 5-20.

Foldvik, A., T. Kvinge and T. Torresen, 1985b. Bottom currents near the continental shelf break in the Weddell Sea. In: Oceanology of the Antarctic Continental Shelf, Ant. Res. Ser., 43, Am. Geop. Un., Wash. pp. 21-34.
Foster, T. D. and E. C. Carmack, 1976. Frontal zone mixing and Antarctic Bottom Water formation in the southern Weddell Sea. Deep-Sea Res., 23: 301-317.
Georgi, D. T., 1979. Modal properties of Antarctic Intermediate Water in the southeast Pacific and the South Atlantic. J. Phys. Oceanogr., 9(3): 456-468.
Gordon, A. L. 1971. "Oceanography of Antarctic Waters". In: Antarctic Oceanology I, Vol. 15, AGU Antarctic Research Series, ed., J. L. Reid: pp. 169-203.
Gordon, A .L. , 1978. Deep Antarctic convection west of Maud Rise. J. Phys. Oceanogr., 8(4): 600-612.
Gordon, A. L. 1974. Varieties and variability of Antarctic bottom water, in: Processus de Formation des Eaux Oceaniques Profondes (en particulier en Mediterranee Occidentale). Editions du Centre National de la Recherche Scientifique, Paris, France, No. 215: pp. 33-47.
Gordon, A. L., 1981. Seasonality of Southern Ocean sea ice. J. Geophys. Res., 86: 4193-4197.
Gordon, A. L., 1982. Weddell deep water variability. J. Mar. Res., 40(suppl): 199-217.
Gordon, A. L., 1988 Southern oceans and global climate. Oceanus, 31(2): 39-46.
Gordon, A. L. and J. C. Comiso, 1988. Polynyas in the Southern Ocean. Sci. Am., 256(6): 90-97.
Gordon, A. L. and B. A. Huber, 1984. Thermohaline stratification below the Southern Ocean sea ice. J. Geophys. Res., 89(C1): 641-648.
Gordon, A. L. and B. A. Huber, 1990. Southern Ocean Winter mixed layer. Jour. Geop. Res. 95(C7): 11,655-11,672.
Gordon, A. L. and E. J. Molinelli, *Southern Ocean Atlas* . Columbia University, New York,1982.
Huppert, H. E. and K. Bryan, 1976. Topographically generated eddies. Deep-Sea Res., 23: 655-679.
Hays, J. D. (1967). Quaternary sediments of the Antarctic Ocean. Progress in Oceanography Vol 4: 117-131.
Huber, B.A., J. Jennings, C. Chen, J. Marra, S. Rennie, P. Mele and A. Gordon 1983 Reports of the US-USSR Weddell Polynya Expedition, October - November, 1981, Hydrographic Data. Lamont-Doherty report -83-1.
Huber, B. A., P. Mele and A. Gordon, Report of the Winter Weddell Sea Project, ANT V/2, Hydrographic data. L-DGO-89-1, Lamont-Doherty Geological Observatory, Palisades, N.Y., 1989.
Jacobs, S., R. Fairbanks and Y. Horibe, 1985. Origin and evolution of water masses near the Antarctic continental margin: Evidence from $H_2\,^{18}O/H^2\,^{16}0$ ratios in seawater. In: Oceanology of the Antarctic Continental Shelf, Ant. Res. Ser., 43, Am. Geop Un., Wash., pp. 59-85.
Jacobs, S. S. 1989. Marine controls on modern sedimentation on the Antarctic continental shelf. Marine Geol. 85: 121-153.
Jacobs, S. S. (submitted for publication) On the Nature and Significance of the Antarctic Slope Front. Proceedings Volume of the Brest Symposium, "The Biogeochemistry and the Circulation of Water Masses in the Southern Ocean".
Kennett, J. P. 1977. Cenozoic evolution of Antarctic glaciation, the circum-antarctic ocean, and their impact on global paleoceanography. Jour. Geop. Res. 82(27): 3843-3860.
Killworth, P. D., 1979. On "chimney" formations in the ocean. J. Phys. Oceanogr., 9(3): 531-554.
Mantyla, A. W. and J. L. Reid (1983). Abyssal characteristics of the World Ocean waters. Deep Sea Res. 30(8A): 805-833.
Martinson, D. G., 1991. Open ocean convection in the Southern Ocean. This Volume.
Martinson, D. G, 1990. Evolution of the Southern Ocean winter mixed layer and sea-ice: open ocean deepwater formation and ventilation. *Jour. Geop. Res.* 95(C7): 11,641-11,654

Martinson, D. G., P. D. Killworth, and A. L. Gordon, 1981. A convective model for the Weddell Polynya. J. Phys. Oceanogr., 11(4): 466-488.
McCartney, M. 1977. Subantarctic mode water. In: A Voyage of Discovery, M. Angel, editor, Supplement to Deep Sea Res. 24: 103-119.
McDougall, T. J. 1987. Neutral Surfaces. Jour. Phys. Oceano., 17: 1950-1964.
Molinelli, E. 1981. The Antarctic influence on Antarctic Intermediate Water. Jour. Mar. Res. 39: 267-293.
Patterson, S. L. and T. Whitworth, 1990. "Physical Oceanography" In: Antarctic Sector of the Pacific, ed. G.P.Glasby. Elsevier pages 55-93.
Piola, A. R. and A. L. Gordon, 1989. Intermediate waters in the southwest South Atlantic. Deep-Sea Res., 36: 1-16.
Piola, A. R. and D. T. Georgi, 1982. Circumpolar properties of Antarctic Intermediate Water and Subantarctic Mode Water. Deep-Sea Res., 29(6A): 687-711.
Shackleton, N. J. and J. P. Kennett, 1975a. Cenozoic oxygen and carbon isotope changes at DSDP site 284: Implications for glacial history of the northern hemisphere and Antarctica. In: Initial Reports of the Deep Sea Drilling Project, vol 29, p 801. US Government Printing Office, Washington, D.C.
Shackleton, N. J. and J. P. Kennett, 1975b. Paleotemperature history of the Cenozoic and the initiation of Antarctic glaciation: Oxygen and carbon isotope analysis in DSDP sites 277, 279 and 281. In: Initial Reports of the Deep Sea Drilling Project, vol 29, p 743. US Government Printing Office, Washington, D.C.
Smith, S., Muench, R. and C. Pease, 1990. Polynyas and Leads: an overview of physical processes and environment. Jour. Geop. Res. 95(C6): 9461- 9480.
Spencer, D., W. Broecker, H. Craig and R. Weiss, 1982. Geosecs Indian Ocean Expedition, Volume 6, sections and profiles. National Science Foundation, Wash D.C. 140 pages.
Wadhams, P., M. Lange, and S. Ackley, 1987. The ice thickness distribution across the Atlantic sector of the Antarctic Ocean in midwinter. J. Geophys. Res., 92(C13): 14,535-14,552.
Whitworth, T., III, 1983. Monitoring the transport of the Antarctic Circumpolar Current at Drake Passage. JPO, 13: 2045-2057.
Whitworth, T. III and R. G. Peterson, 1985. The volume transport of the Antarctic Circumpolar current from three-year bottom pressure measurements. J. Phys. Oceanogr., 15(6): 810-816.
Worthington, L.V., 1981. The water masses of the world ocean: Some result of a fine-scale census. Evolution of Physical Oceanography, Scientific Surveys in Honor of Henry Stommel, B.A. Warren and C. Wunsch, Eds., MIT Press, pp. 42-69.
Zwally, J., J. Comiso and A.L. Gordon, 1985. Antarctic offshore leads and polynyas and oceanographic effects. In: Oceanology of the Antarctic Continental Shelf, Ant. Res. Ser., 43, Am. Geop. Un., Wash. pp. 203-226.