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DEEP CONVECTION IN THE WORLD OCEAN

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Abstract. A brief discussion of, and a little speculation about, the relevance of the polar regions on climate is given. The main body of the paper gives a survey of the known deep convection areas of the world ocean. There are two distinct types of convection. The first is the classic sinking occurring on continental shelf slope systems, as typified by various locations around the Antarctic coast. The freezing of sea ice, and resulting brine ejection, creates dense salty water on the shelf which descends the slope under a balance of Coriolis, gravity, and frictional forces, entraining the surrounding warm deep water as it goes. The second process is the more recently observed open-ocean convection, occurring in locations such as the Mediterranean, the Labrador Sea, and two locations in the Weddell gyre, and is hypothesized to occur in the Greenland Sea. Open-ocean convection has many overall similarities in all these areas: it occurs in narrow (20-50 km) areas; it forms about 10 m^3 s^{-1} of deep water; it occurs only in regions of cyclonic mean circulation; more than one water mass in the mean circulation is involved; a preconditioning seems to be required; some surface forcing (cooling or sea ice formation) is necessary; a violent breakup of the water mass frequently occurs on time scales of 2 weeks.

1. Introduction

Oceanographers, after many decades of patient observation with ever more sophisticated instrumentation, still need to know more about the deep areas of the world ocean. Most of the information we now possess derives from hydrographic casts and conductivity-temperature-depth (CTD) data, with more recent additions of longer-term moored deployments and some current data. The majority of the deep oceanographic data come from equatorial and mid-latitudes; observing difficulties, especially in winter, mean that there are few reliable data from polar latitudes.

Yet the processes such as deep convection which usually occur in narrow areas in polar regions [Stommel, 1962] are of fundamental importance for ocean climate and the maintenance of a stably stratified world ocean. The vast majority of the world ocean is warmer at the surface than it is at depth. Any downgradient heat flux, from molecular to turbulent, must act to transfer heat downward to the deep ocean. Unless this is balanced everywhere by an upwelling of cold water (and there is evidence that this is not the case; cf. Schott and Stommel's [1979] inverse calculations), the deep oceans should become warmer. This excess heat must be lost somehow, and one of the processes which can do this is vertical convection, typically in polar regions, which allows the heat to be given up to the atmosphere. The poleward oceanic heat flux required for this is visible as a residual in global atmospheric budgets [Oort and Vonder Haar, 1976] and is under active investigation.

Deep convection must also modify climate on long time scales. There is a traditional climatic positive feedback mechanism [Budyko, 1969; Sellers, 1969]. A colder surface of the earth yields more ice, which raises the albedo, which gives less net heat absorption and so a colder surface. This mechanism has an opposite counterpart in the ocean. A colder ocean surface does indeed yield more ice, which by brine ejection (see section 3) gives a cold, saline surface layer. This leads to increased convection, bringing warmer water from depth; the warm water acts to melt some of the ice and decrease the surface albedo, allowing more heat to be absorbed and so warming the surface. There are indications in simple one-dimensional ocean climate models that this negative feedback can be as important as the positive mechanism described above. On shorter time scales, of years to decades (the World Climate Research Program definition of climate), the effects of deep convection are more speculative and will be discussed near the end of this paper.

The purpose of this paper is to review our knowledge of deep convection; Warren [1981] has provided an excellent detailed review of deep circulatory processes, and, with some overlap, it is hoped that this paper will be partly complementary to Warren's. The dynamics of ice-covered waters are covered more fully by Carmack [1982]. There are two main types of deep convection, the physics of which are very different. The first is convection near an ocean boundary. A schematic of the process is shown in Figure 1a. The process involves the formation of a dense water mass which reaches the bottom of the ocean by descending a continental slope. The constraints imposed by the boundary are of importance in the formation process. The second type is open-ocean deep convection, shown in Figure 1b. Here the convection occurs far from land and is predominantly vertical. Neither process need occur regularly; indeed, in some areas, deep convection may be the exception rather than the rule, depending on prevailing atmospheric conditions.

Section 2 will discuss the observations of near-boundary convection, and section 3 attempts to draw together the important features and similarities between the physics of this convection in the different areas of the world ocean. Sections 4 and 5 perform the same function for open-ocean convection. Section 6 discusses the interaction between convective regions and the rest of the ocean, and section 7 concludes with a discussion of future work needed to help our understanding both of climate and of the deep convective pro-

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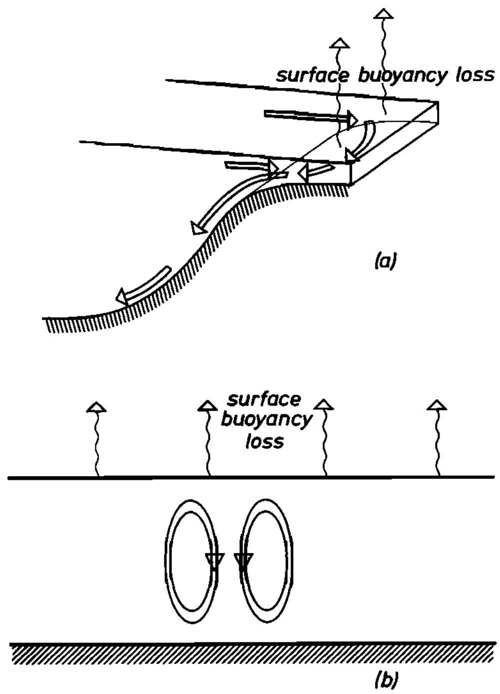


Fig. 1. Schematics of the two main types of oceanic deep convection: (a) near-boundary convection and (b) open-ocean convection.

cesses. All salinities in this paper are quoted as multiples of 10^{-3}.

2. Areas of Deep Convection Near Ocean Boundaries

The areas where deep boundary convection is known to occur are indicated schematically in Figure 2. This section will briefly discuss each area.

The Western Weddell Sea

Antarctic bottom water (AABW) is a dense water mass which comprises, or provides a component of, most of the deep waters of the world ocean. Its primary source is in the western and southwestern Weddell Sea (see Figure 3 for details of Antarctic geography). Observations in summertime (Figure 4) show a layer of cold, fresh water ($\theta < -0.7^{\circ}C$, S = 34.64-66) at the foot of the continental slope, over most of the southwestern corner of the Weddell Sea [Foster and Carmack, 1976]. The layer is up to 200 m thick and is readily distinguished from the warm deep water above it. The data suggest that the layer extends along most of the eastern side of the Antarctic Peninsula (though ice conditions preclude observations [Carmack, 1973], becoming more saline with distance west.

The origins of this Weddell Sea bottom water (WSBW) lie with the cold saline water at the shelf break, at a depth of about 500 m. Closely spaced sections (see Figure 5 [from Foster and Carmack, 1976]) show a complicated interleaving of water masses from on and off the shelf and an indication of sinking down the continental slope. This sinking is accompanied by a movement to the west, by geostrophy; Foldvik and Kvinge's [1974] current meter data show currents westward of about 7 cm s^{-1}. in fact. As a result, the WSBW in Figure 4 must have its origins some way to the east; there are as yet no along-slope sections showing a continuous plume of dense water from shelf break to abyssal depths, however.

Estimates of the rate of production of bottom water in the Weddell Sea vary drastically. Early estimates, based upon an ad hoc definition of AABW, were very high (20-50 x 10^6 m^3 s^{-1}). It is now believed that most of this figure comes from entrainment of surrounding water masses by WSBW after its formation. More recent estimates are based on in situ measurements of WSBW. The most probable such estimates, by Foster and Carmack [1976], are around 2-5 x 10^6 m^3 s^{-1}. This assumes a uniform production rate along the shelf break; intermittency or local topographic effects could yield higher or lower values. For example, A. Foldvik (private communication, 1981) has observed about 0.7 x 10^6 m^3 s^{-1} of supercooled water descending depth contours in a wide canyon at about 35°W.

The Ross Sea

Similar mechanisms to those in the Weddell Sea occur in other parts of the Antarctic. The Ross Sea is probably the most important. It, too, has its origins on the continental shelf, with a downflow along the continental slope of cold, saline water (see Figure 6 [from Jacobs et al., 1970]). Warren [1981] notes that the effect of Ross Sea bottom water on the surrounding deep water is rather less marked than in the Weddell Sea, because the water entrained by the descending slope water is already warmer and more saline. No firm estimates are available for production rates in the Ross Sea.

The Adélie Coast and Enderby Land

A slightly different situation occurs west of the Ross Sea, off the Adélie Coast. Here the shelf possesses a deep depression which apparently acts to trap dense water forming along the shelf (the Weddell and Ross Sea shelves also possess quite strong topographic features, whose role in water mass formation is unclear). Some water from the depression is able to surmount the sill and descend the continental slope in a thin layer (see Figure 7 [from Gordon and Tchernia, 1972]). Evidence is also given by Jacobs and Georgi [1977] for a similar source of bottom water off Enderby Land.

Sinking off Wilkes Land

It now seems likely that there is sinking of dense water around most of the Antarctic coastline. This does not mean that all the water is destined for the bottom, however. Much of the water formed, due to environmental forcing and geographic constraints, may be insufficiently dense to reach the ocean bottom. An example of this is near Wilkes Land, just west of the Ross Sea [Carmack and Killworth, 1978]. Figure 8 shows a tongue of anomalously cold fresh water leaving the continental slope and spreading horizontally westward at about 2000 m. Presumably, this process may occur in many places; it is evident in

Fig. 2. Approximate locations of main areas of near-boundary deep convection, shown solid.

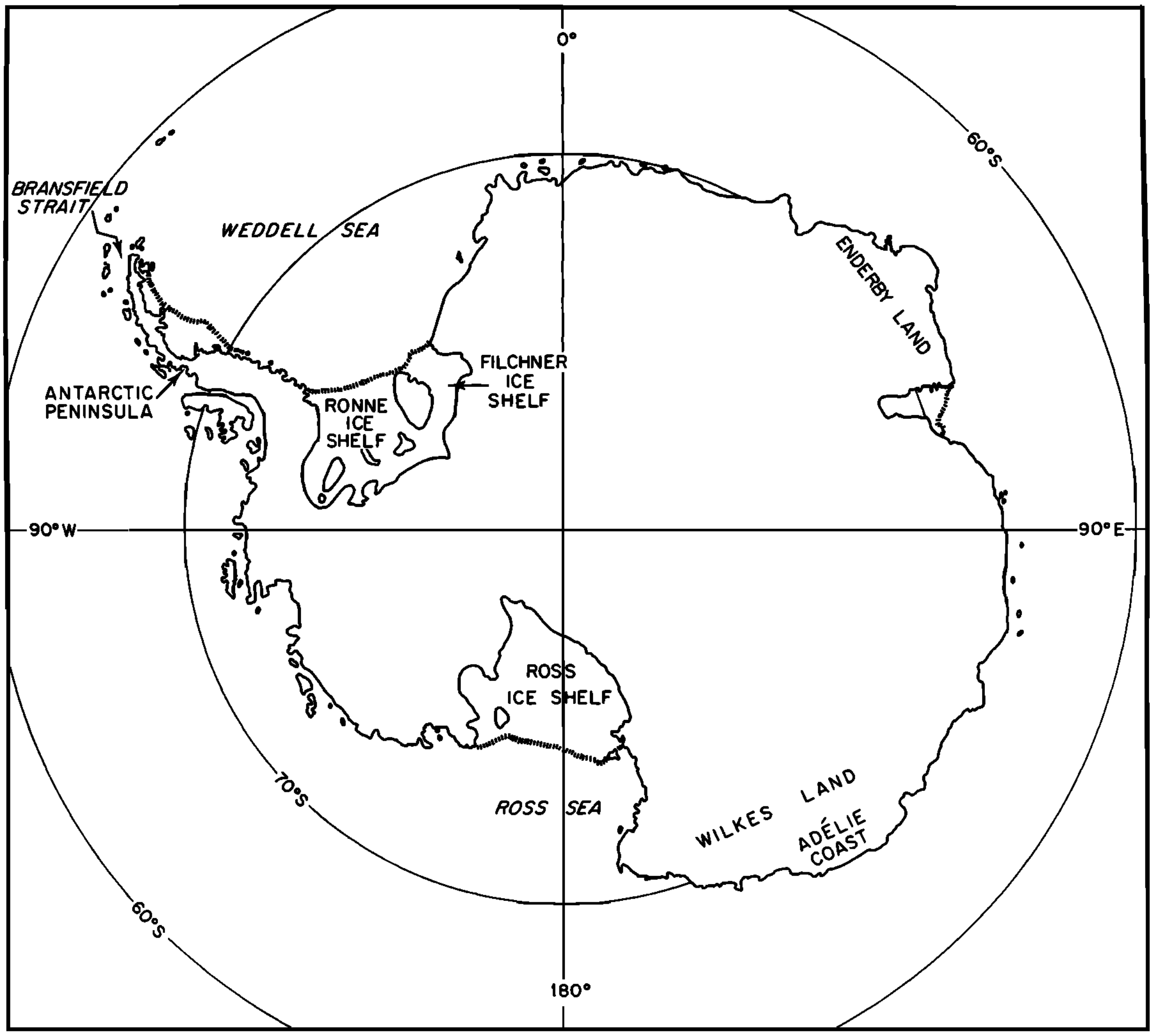


Fig. 3. Geographical locations in Antarctica [from Warren, 1981].

the data off Wilkes Land because the surrounding water is of Ross Sea origin and therefore possesses different properties. It is likely that sinking elsewhere would be concealed due to the surrounding water possessing almost identical temperature-salinity (T-S) characteristics to the water sinking.

Sinking in the Arctic

Aagaard et al. [1981] and Melling and Lewis [1982] have recently suggested that a similar process of sinking from the shelf break occurs in the Arctic, although only to depths of 100-150 m. They are led to this conclusion by the need to explain the source of the Arctic halocline, which is considerably more salty than a vertical mixture between surface and deeper Atlantic waters would allow. There is dense saline water on the wide Arctic shelves which, if it sank to the halocline depth, would be able to supply water of the required salinity. They estimate that a circulation of around 2.5×10^6 m^3 s^{-1} is involved.

Other Areas

There are many other areas of the world ocean where sinking along continental slopes occurs. Predominant are the northern sources of deep water in the Atlantic. These involve movement of water from the Norwegian, Greenland, or Iceland seas over sills into the Atlantic. The best studied is that through the Denmark Strait [Swift et al., 1980], with other passages through the Faroe Bank channel and the Wyville Thompson ridge. The transport of water involved is a few million cubic meters per second [Worthington, 1969]; opinions differ on the precise value. Strong currents can be involved, as Smith [1976] demonstrated. In all cases the water masses descend the continental slope, strongly controlled by topography. They remain recognizable because they differ in T-S characteristics from the water around them.

Similar outflows and sinking occur at other constrictions between oceans: there is a strong (10^6 m^3 s^{-1}) outflow from the Mediterranean into the Atlantic [Lacombe, 1974] which sends a distinct 'tongue' across much of the Atlantic. Indeed, there is evidence [Brass et al., 1981] that in Cretaceous times the Mediterranean was the source of the densest water in the world ocean. The Red Sea and the Persian Gulf also have dense outflows (the former has its own slope deep water formation occurring internally [Manins, 1973]).

The northern Adriatic possesses a wide conti-

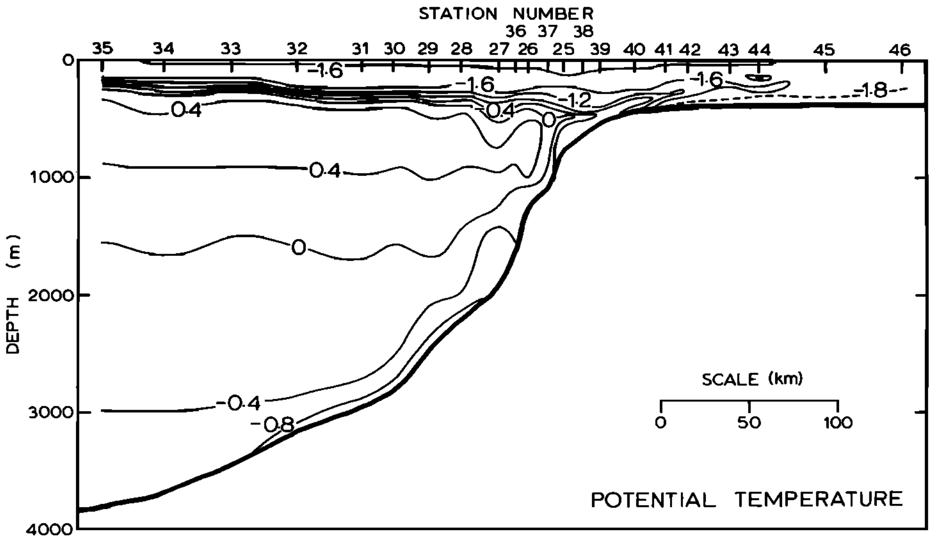


Fig. 4. Section crossing the Weddell Sea shelf break at about 40°W, showing potential temperature (degrees Celsius) [from Foster and Carmack, 1976].

nental shelf, and there is an area of dense water formation at and beyond the shelf break in wintertime, driven by surface evaporation. The dense water descends to depths of about 250 m [Hendershott and Rizzoli, 1976; Zoré-Armanda, 1963]. (See also Franco et al. [1982] for a detailed survey of the Adriatic.) Finally, there are minor locations in which dense water can be seen sinking in canyons on continental slopes: Fieux [1972] gives an example in the Gulf of Lions, in the Mediterranean.

3. The Physics of Deep Convection Near Ocean Boundaries

Five separate ingredients appear to be involved in the production of deep water near oceanic boundaries. The first is a reservoir in which to form dense water. For most of the Antarctic and Arctic production zones the reservoir is the continental shelf. In the Weddell and Ross seas, and in the Arctic, the shelf has the capacity to store dense water due to its great width (some hundreds of kilometers from the coastline). Off the Adélie Coast and Wilkes Land the shelf slopes upward, away from the coast, to a sill; in this case it is the combination of shelf and sill that enables dense water to be stored. The width of the shelf (or more generally the area or volume of the reservoir) seems to be positively correlated with amounts of deep water formed around Antarctica [Gordon, 1974].

The second ingredient is a source of dense water within the reservoir. Over most of the Antarctic and Arctic this source is wintertime ice formation. Sea ice has a salinity of 5, compared

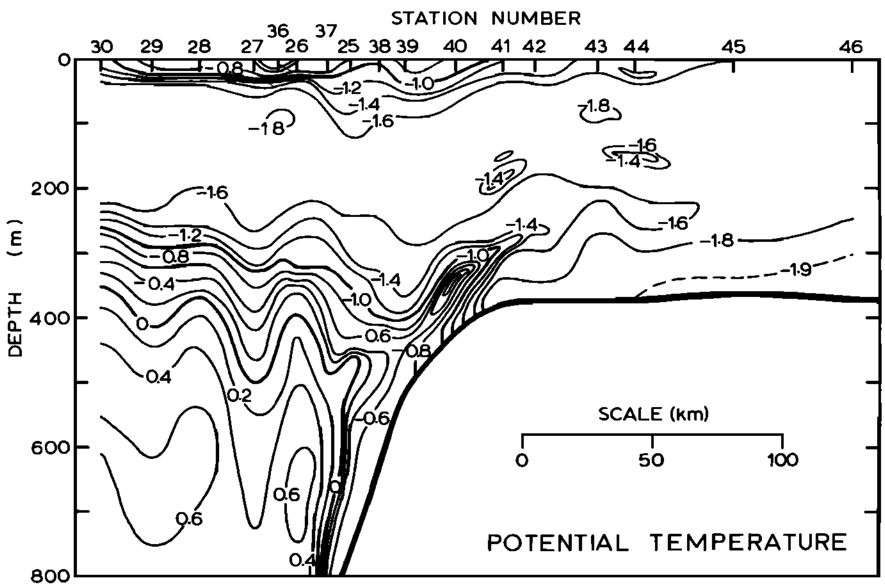


Fig. 5. Detailed section near the shelf break at about 40°W, showing potential temperature (degrees Celsius) [from Foster and Carmack, 1976].

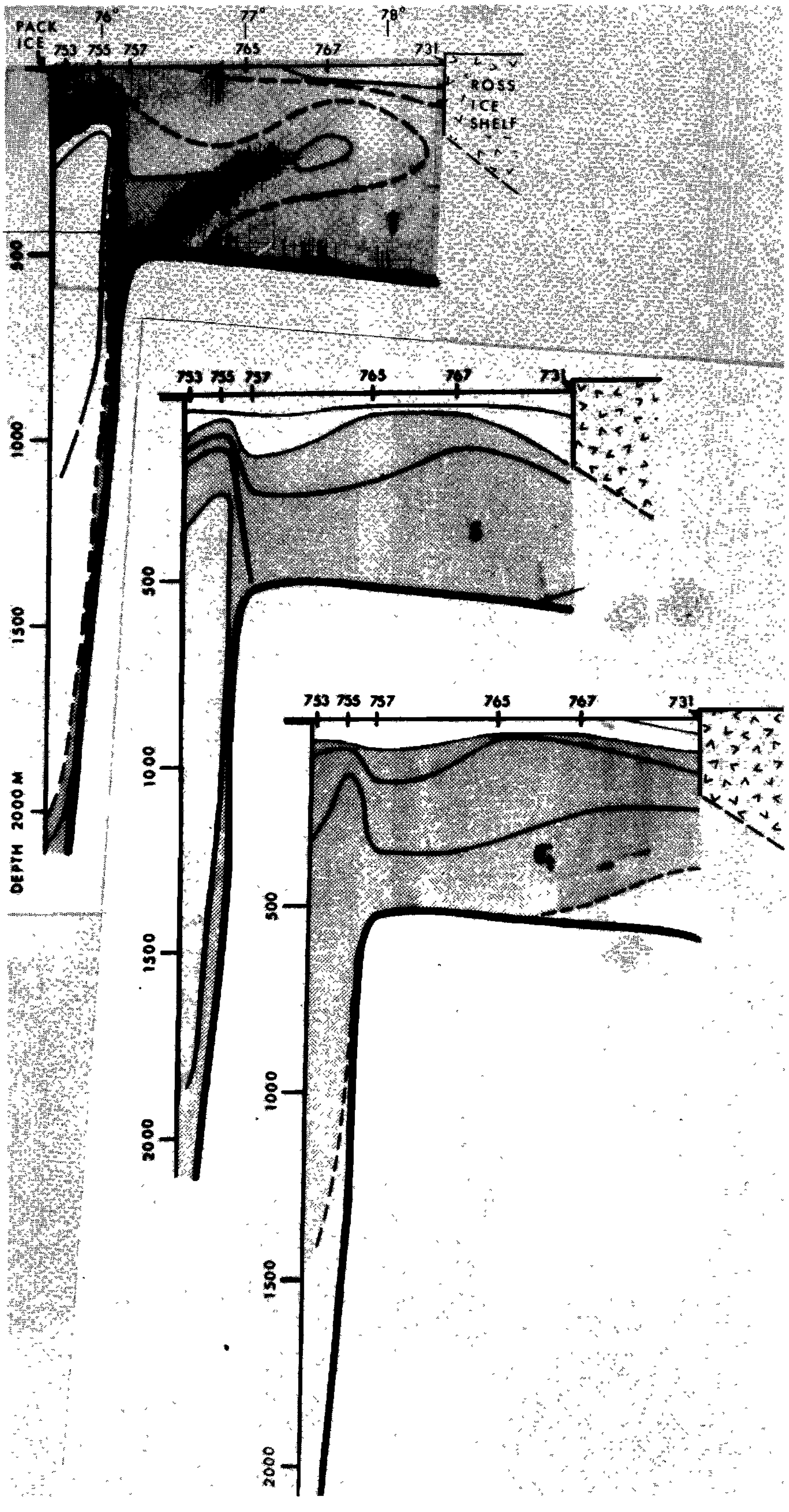


Fig. 6. Vertical cross sections of temperature T, salinity S, and σ_t in the Ross Sea [from Jacobs et al., 1970].

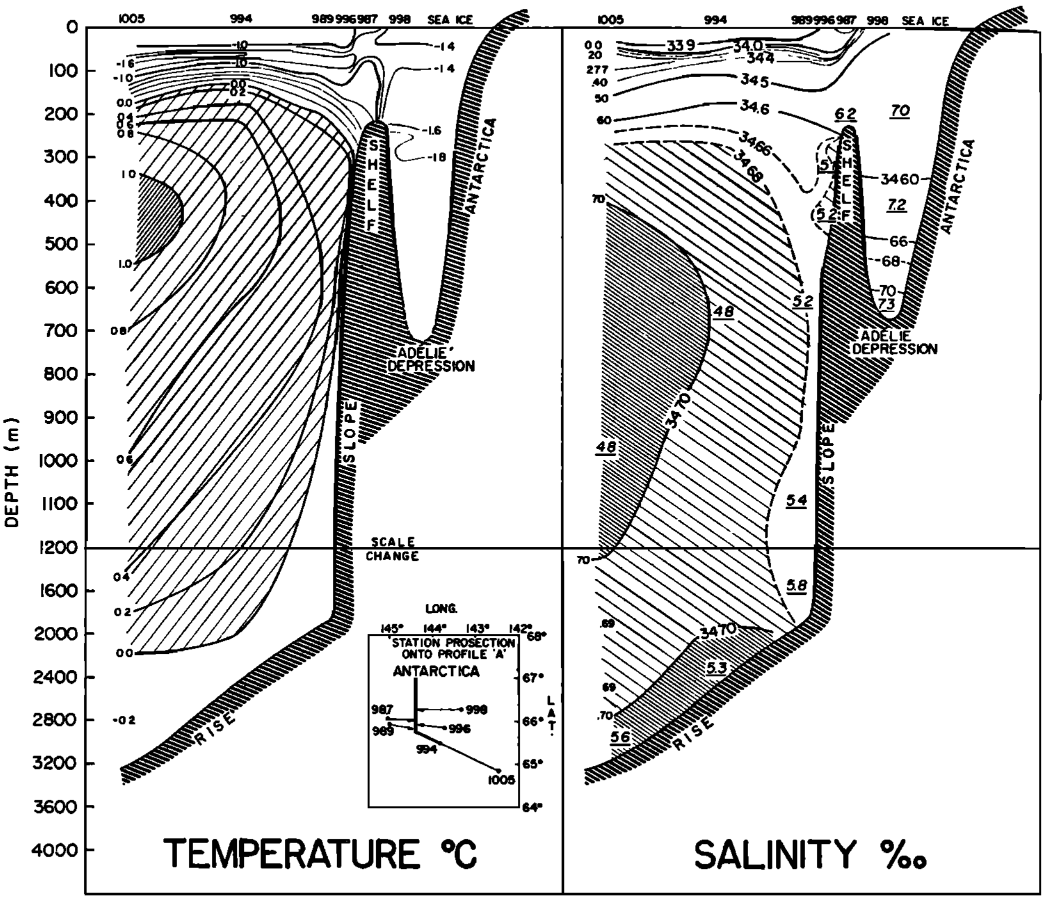


Fig. 7. Section away from Adélie Coast, Antarctica, showing temperature (degrees Celsius) and salinity [from Gordon and Tchernia, 1972].

with typical ocean salinities of 35. The result of freezing of seawater is to induce a drainage of brine from the ice to the water below. Providing the depth of the reservoir is sufficiently shallow (500 m in the Antarctic, 50-100 m in the Arctic), the resulting vertical convection yields cold (near freezing) and salty water at depth on the shelf. The role of ice shelf freezing or melting in this process is still an open question [Gill, 1973; Jacobs et al., 1979], though 2 cm yr^{-1} of freezing below the Ross Ice Shelf is known to occur [Zotikov et al., 1980].

The freezing of sea ice is strongly seasonal. The additional salinity induced by freezing is not simply given up during the summer time melting because of the export of ice away from the coastline [Gill, 1973; Warren, 1981; Aagaard et al., 1981]. Penetrative convection also helps maintain a net salinity increase. Ice formation is not of itself a necessary ingredient for boundary deep convection: in the Red Sea, for example, the source of dense water is the intense surface evaporation in the Gulf of Suez [Manins, 1973], which also acts to increase the salinity within the reservoir.

The third ingredient is a reason for the dense water to leave the reservoir and descend the slope. Early ideas on bottom water formation [e.g., Mosby, 1934] suggested that once the water in the reservoir was sufficiently dense, the water would flow down the slope under the action of gravity. Mosby [1934] quoted a salinity of 34.62, later corrected by Fofonoff [1956] to account for the peculiar equation of state at low temperatures. However, dynamical ideas do not support this view. A dense mass of water at the top of a slope may move outward about a deformation radius (say, 20 km) but will then flow horizontally under a balance between downslope buoyancy forces and upslope Coriolis force (i.e., a simple solution of a geostrophic adjustment problem). Killworth [1977] gives an example of this.

It is necessary therefore for there to be circulation already existing which drives at least part of the dense water off the shelf. Gill [1973] noted the existence of a strong increase in shelf salinity from east to west in both the Weddell Sea and the Ross Sea, which by geostrophy yields a cyclonic circulation at depth on the shelf. Coupled with an onshore surface Ekman flux by the easterly prevailing winds, and perhaps augmented by katabatic winds, which force a return flow at depth [Gill, 1973; Killworth, 1973], the combination seemed capable of moving about 10^6 m^3 s^{-1} of water off the shelf. Killworth [1974] produced a simple model to account for the east-west asymmetries on the shelf. Brine formation decreasing northward induces, by geostrophy, eastward surface and westward subsurface motion, with the most saline water to the south. The necessary upwelling at the western shelf boundary and downwelling at the east leads to an east-to-west den-

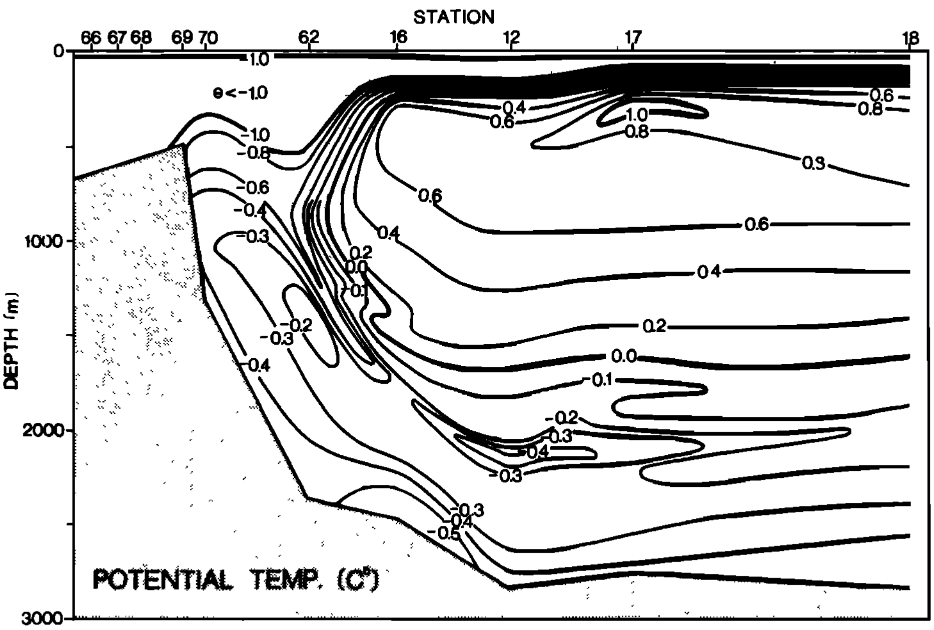


Fig. 8. Vertical section of potential temperature (degrees Celsius) away from Wilkes Land, Antarctica [from Carmack and Killworth, 1978].

sity gradient, approximately as observed. Geometrical variations on the shelf, together with depth variations, can also induce east-west pressure gradients and therefore geostrophic transport on and off the shelf.

The fourth ingredient apparently plays little dynamical role in the process of bottom water formation and yet appears consistently in the observations. It seems necessary that more than one water mass be involved in dense water formation. Sometimes the mixing process can be remarkably complicated. Figure 9 shows Foster and Carmack's [1976] proposed mixing diagram for AABW production in the Weddell Sea. Surface winter water, at or about freezing, mixes, perhaps by cabeling [Foster, 1972], with the underlying warm deep water to produce modified warm deep water. This in turn mixes with Weddell shelf water to produce Weddell Sea bottom water, which gradually mixes with further warm deep water to produce AABW. Such a mixing process may be site-or even season-specific; frontal zone mixing is a complicated process. Upwelling frequently seems to be important in the mixing.

In no area is the simple case of dense water on a shelf running down a slope past a single water mass observed. It is unclear whether this is accidental, or whether the differentiated water masses are required so that the bottom water produced is detectably different from the water around it, or even whether the flows induced by dense water formation help to create and maintain the differentiated water masses.

The last ingredient is frequently overlooked in discussions of deep water formation. We require that the densities, geography, and dynamics involved do actually permit the dense water to sink. Clearly, strategically sited sills, etc., can prevent water from reaching the bottom. If the surface forcing is insufficiently strong, the water-mass will not be dense enough to sink to the ocean floor (as is the case off Wilkes Land [Carmack and Killworth, 1978]). And, most important, the dynamics must allow the water to sink. Dense water has a tendency to flow horizontally, following depth contours, in geostrophic equilibrium. Only a (turbulent) friction between the flow and the slope can induce a third component in a triangle of forces balance to let the flow have a downslope component. Turbulent entraining plume models have been successfully constructed by Smith [1975] for the Mediterranean and Denmark Strait overflows, using the above concepts. However, Killworth [1977] showed that neither two- or three-dimensional plumes nor intermittent thermals could both descend the Weddell continental slope and entrain water at a rate consistent with observations. He found that the equation of state could provide an additional source of internal energy for the plume. In fact, a cold plume which is denser than its surroundings can actually become still denser relative to its surroundings while sinking in a

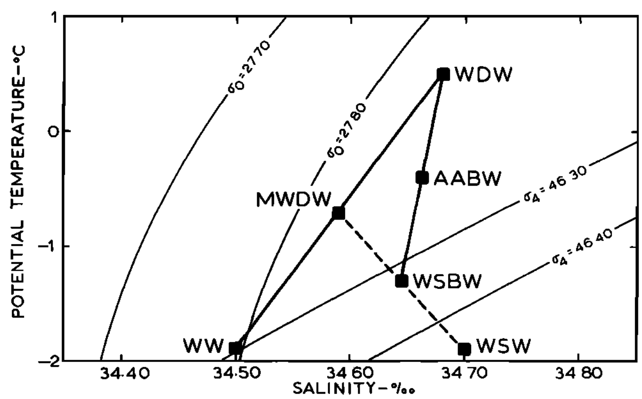


Fig. 9. Proposed temperature-salinity mixing diagram for production of Weddell Sea bottom water (WSBW). Other water masses indicated are winter water (WW), Weddell shelf water (WSW), warm deep water (WDW), Antarctic bottom water (AABW), and modified warm deep water (MWDW) [from Foster and Carmack, 1976].

Fig. 10. Approximate locations of main areas of open-ocean deep convection, shown solid.

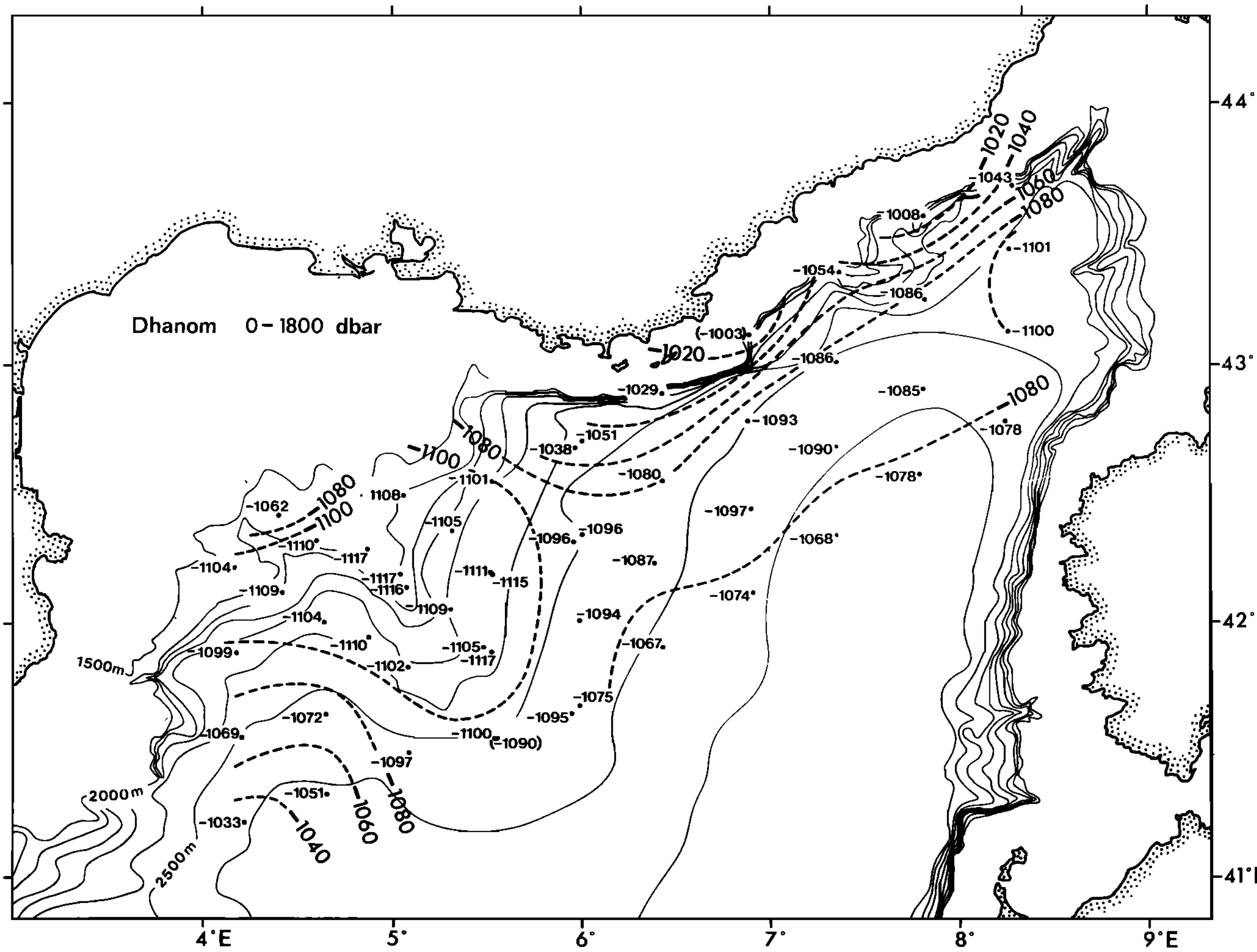


Fig. 11. Dynamic height anomaly relative to 1800 dbar (dynamic millimeters) in the Gulf of Lions, superimposed upon bottom topography [from Swallow and Caston, 1973].

stably stratified fluid. If this effect is included, model plumes can descend the slope in a manner consistent with observations. On the Arctic slope, plumes do not sink to great depths (although occasional incursions of plumes to abyssal depths may be necessary to account for the anomalously salty deep water in the Eurasian Basin [Aagaard, 1981]). Melling and Lewis [1982] use a turbulent plume model to account for the Arctic halocline. In this case the plume sinks to perhaps 140 m.

Such models tend to ignore unsteadiness in the dense water, presumably modeling this by turbulent coefficients of entrainment and friction. Yet the sharp density fronts involved make such flows highly unstable. Smith [1976] reports violent fluctuations in the flow downstream of the Denmark Strait, for example, and models these as baroclinic instability of the flow. Recent theoretical work [Griffiths et al., 1982; Killworth and Stern, 1982] together with laboratory experiments [Griffiths and Linden, 1981a; Stern, 1980] shows that flows with one or two density fronts, and with or without a boundary, are very unstable even if the effect of the lower layer is ignored. It is unclear what role such instabilities can play in the sinking of dense water on slope: recently, Foster and Middleton [1980] found evidence of unsteadiness and interleaving in bottom water formation in the western Weddell Sea.

4. Areas of Open-Ocean Deep Convection

The second major variety of deep oceanic convection occurs far from boundaries. Figure 10 shows the areas where open-ocean convection is known, or hypothesized, to occur. Each area will be discussed in turn.

The Gulf of Lions in the Western Mediterranean

The best documented, and certainly most exhaustively observed, case of open-ocean convection occurs most winters in the Gulf of Lions and is usually referred to as the Medoc area [MEDOC Group, 1970]. The circulation in the gulf is cyclonic (see Figure 11 [from Swallow and Caston, 1973]), which generates a 'doming' of the isopycnal surfaces in the center of the cyclonic gyre. This results in a weakening of the vertical static stability over an area of about 100 km in the gyre center, in the surface layer of low-salinity water. The intermediate warm salty layer of east-

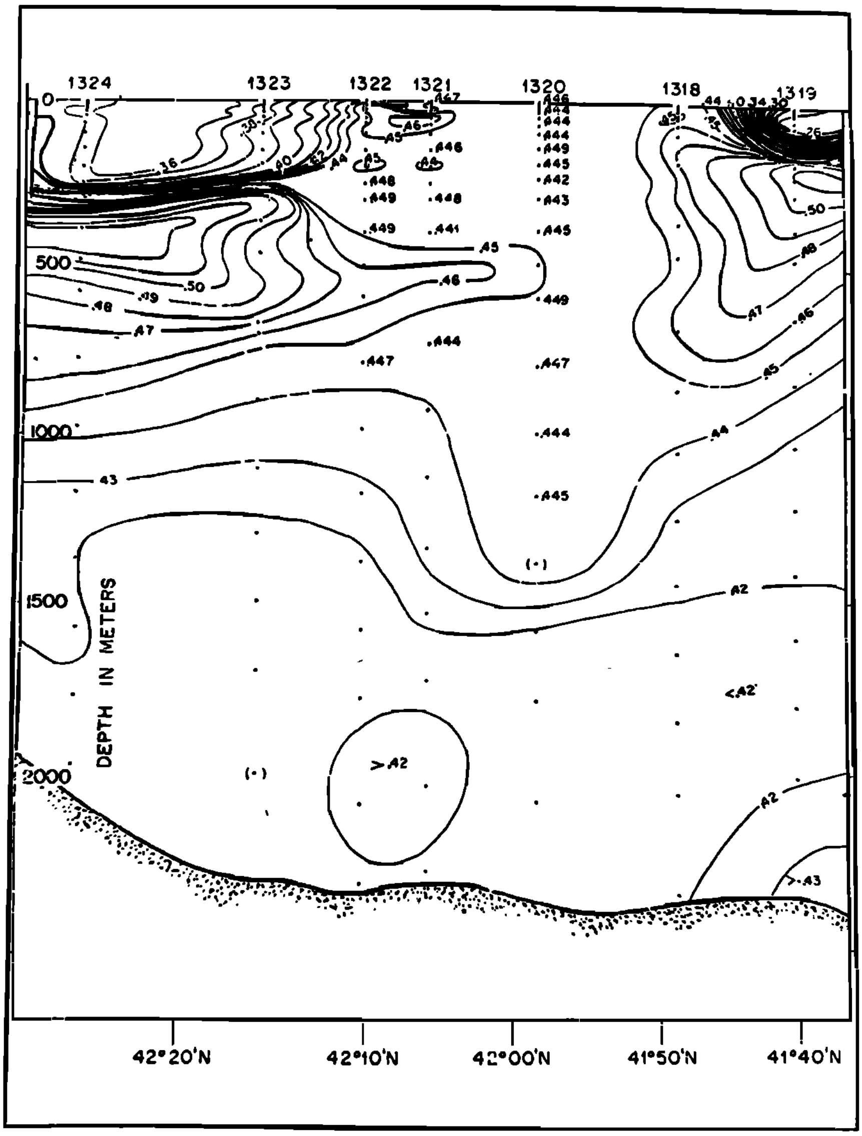


Fig. 12. North-south section through the Medoc chimney, February 1969, showing salinities referenced to 38 [from Anati and Stommel, 1970].

ern Mediterranean water, at depths of 200-500 m, and the deep bottom layer are relatively unaffected. This weakening of vertical stability is known as the 'preconditioning' phase [Stommel, 1972].

Toward February, two intensely cold winds begin to blow from the north or northwest: the Mistral and Tramontaine. Vertical convection begins over a wide area of the Mediterranean as a result of surface cooling (the sea can be 10° warmer than the atmosphere). Over most of this area, however, the convection results merely in a gradual erosion of the thermocline by a deepening surface mixed layer. In the center of the preconditioned dome, convection can achieve more violent results. Several different phenomena can occur, either individually or together, depending on the degree of forcing and the precise shape of the preconditioned area. These phenomena are known collectively as the 'violent mixing' phase.

In 1969, for example, the surface cooling was enough to break through the surface stratification about 1 week, and intense vertical convection, with vertical velocities reaching 2.5 cm s^{-1} [Voorhis and Webb, 1970; Gascard, 1973], occurred over a patch of width 30-50 km, from the surface to at least 2000 m. A vertical section of this mixing is shown in Figure 12 [Anati and Stommel, 1970], and a time-vertical cross section, showing the rapid deepening, in Figure 13 [Stommel, 1972]. In 1975, however, the deep convection took place to one side of a surface front which separated vertically stable and neutrally stable waters [Gascard, 1978].

Accompanying the convection is strong (10 cm s^{-1}) horizontal eddying, on scales as small as 10 km (see Figure 14 [from Gascard, 1978]), usually situated at or near the front dividing overturned and nonoverturned water. The role of these eddies in the formation of deep water is under active investigation; there is no doubt that they form an efficient mixing mechanism between the water masses. Within the column of mixed water, convection tends to swamp the instability [Killworth, 1976], so that eddies tend to be observed only at the boundary of the mixing.

At the end of the period of cold winds, a third

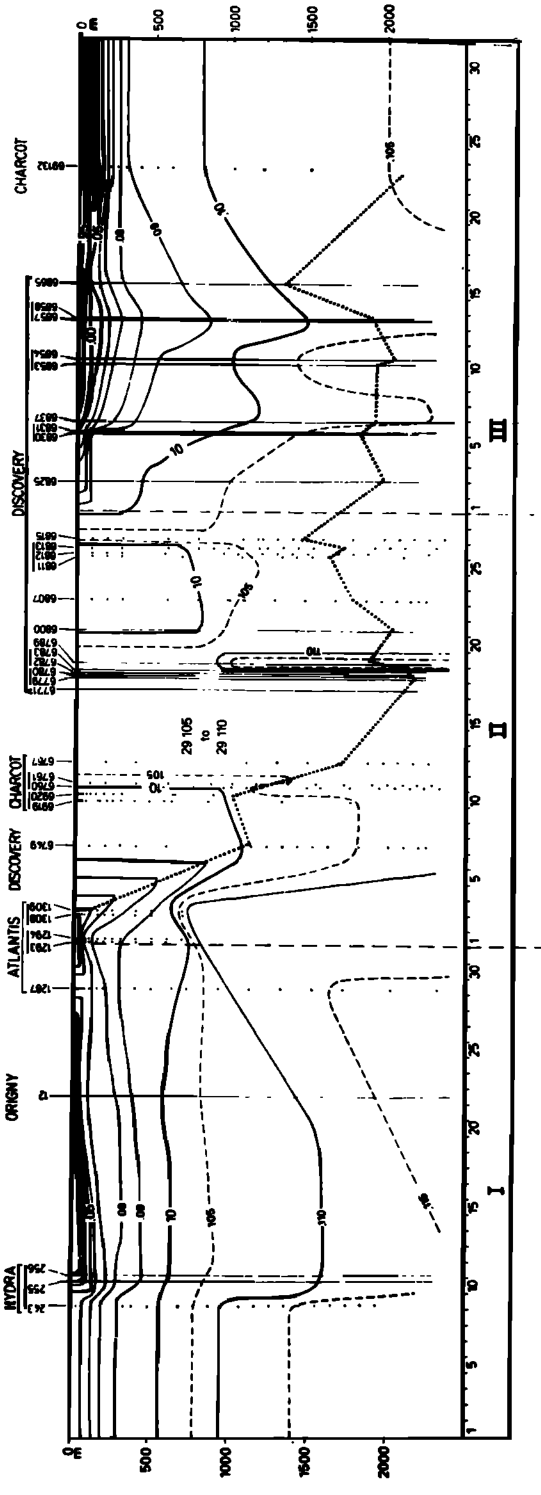


Fig. 13. Time-depth section of sigma-theta at the center of the Medoc chimney. January (I) to March (III) 1969. The dotted line shows the bottom of the mixed layer [from Stommel, 1972].

phase, the 'sinking and spreading phase,' can occur. Observational difficulties at this time limit our knowledge of what occurs. Certainly, there is a rapid restratification of the surface waters within only a few days. Horizontal eddies rather larger than those formed previously (about 20 km) are observed, with speeds as high as 15 cm s^{-1}. The dense water column formed during the violent mixing seems to sink (although not usually reaching the bottom; there is usually a layer of 'old' water below it [Lacombe, 1974]) and fragment by a process of sideways interleaving into small 'pockets' of deep water. These pockets can be tagged by their anomalously high oxygen values (the water was, after all, recently in contact with the atmosphere). The pockets drift away at about 2 cm s^{-1} in a direction which usually has a westward component.

Within a few more days the normal stratification is restored, and there is little evidence of the process remaining. The entire process takes less than a month. Lacombe and Tchernia [1974] and Sankey [1973] give comprehensive discussions.

The Labrador Sea

The situation which exists in the Labrador Sea is qualitatively very similar but has only recently been observed in any detail. Several years' more data will be necessary to achieve the same descriptive level as for the Medoc process. The circulation in the Labrador Sea is again cyclonic, resulting from the northward flowing West Greenland current and the southward flowing Labrador current. There are several distinct water masses present in this circulation: cold, fresh surface water brought in by the West Greenland current; below this a warmer, saltier water from the Irminger Sea; and a deep lower layer of North Atlantic deep water.

Before 1976 there is little direct evidence of deep water formation in the Labrador Sea (although, of course, the presence of the deep water is evidence that formation had to occur somehow). Lazier [1973] did, however, record a time series taken by weather ship Bravo in 1966 which showed deep convection, of unknown horizontal dimensions, occuring to a depth of about 1000 m. In 1976, Clarke and Gascard [1982] found a 50-km-wide area of homogeneous Labrador deep water extending to a depth of 2 km above the North Atlantic deep water. The upper 1000 m of this are shown in Figure 15 [Clarke and Gascard, 1982]. The convection is accompanied by intense horizontal eddying on several scales. The convection itself seems to occur on scales of about 1 km [Gascard, 1982], as would be suggested by laboratory studies [e.g., Keffer, 1979]; eddying occurs on scales from 5 to 50 km. The time for the Labrador stratification to return to normal following the end of winter convection is unknown. It cannot, however, be more than a few days or weeks, or else evidence of wintertime convection woud presumably have been observed prior to 1976, even accounting for the natural year-to-year variation in surface forcing.

The Bransfield Strait

The Bransfield Strait, situated at the tip of the Antarctic Peninsula (see Figure 3), is unique among areas of open-ocean convection in that con-

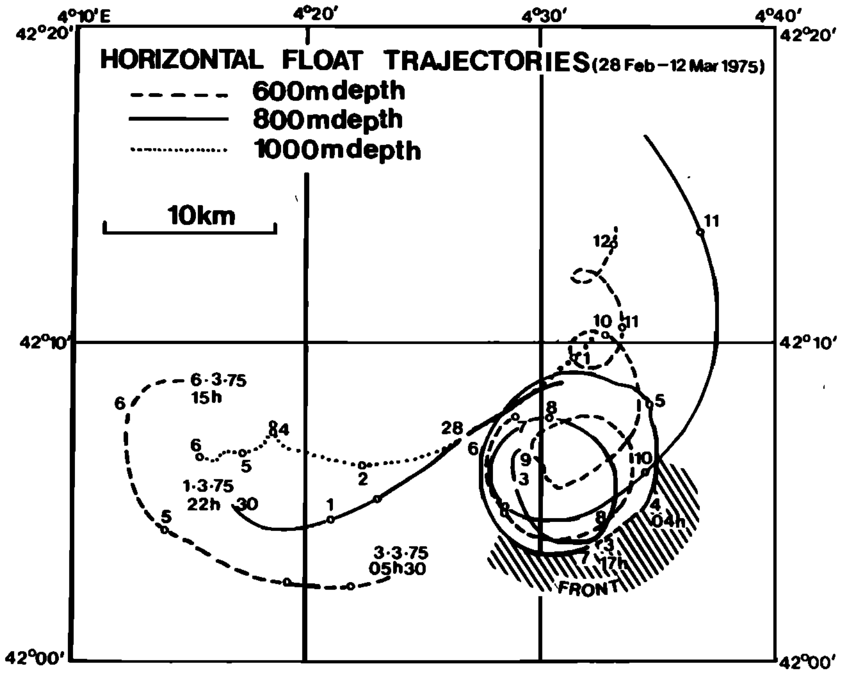


Fig. 14. Horizontal float drifts, February 27 to March 12, 1975, in the Medoc region. Nominal float depths are 600, 800 and 1000 m [from Gascard, 1978].

vection occurs all the way to the bottom. The strait is occupied by a trough whose depth varies from 1100 m to 2800 m; the trough is in turn divided into three basins separated by sills. Figure 16 [Gordon and Nowlin, 1978] shows data along the axis of the strait. The water within the strait is clearly different in character from that surrounding the strait. It is nearly homogeneous, with much lower temperatures, salinities, and silicates and much higher oxygen values. This, together with tritium data, shows that deep wintertime convection has taken place within the strait. However, Gordon and Nowlin [1978] could find little evidence that the dense water could leave the strait, because of topographic constraints.

The Weddell Chimney

In February 1977, Gordon [1978] observed a remarkable oceanic 'chimney' at about 7°W, 67°S, some hundreds of kilometers north of the Antarctic coast but within the main Weddell gyre. Figure 17 shows a section oriented roughly northward away from the coast. A thin column of water, of radius 14 km, is clearly marked in the data. It is anomalously cold and fresh, with high oxygen values, and the cyclonic eddy structure extends to at least 4000 m (but with signs of convection only to σ_2 level of 37.23). The chimney, although vertically stable (Figure 17d), is almost certainly a relic from wintertime convection. Although geostrophic flow tends to be around such a column (this is confirmed by Gordon's [1978] ship drift observations, which show 50 cm s^{-1} cyclonic flow around the column), it is difficult to see how the chimney was able to survive coherently until summer. The weak mean flow in the area (1 cm s^{-1}) may have been important for its survival. Gordon [1978] suggests that there may be several (up to 30) such eddies in the Weddell gyre, as their detection on widely spaced CTD surveys is obviously a matter of chance; he estimates that 0.6 x 10^6 m^3 s^{-1} of deep water can be formed by such chimneys.

The Weddell Polynya

A remarkable feature of the winter sea ice distribution in the (cyclonic) Weddell gyre is a large (3 x 10^5 km^2) irregularly appearing patch of open water, or polynya. It is hypothesized (there are no direct oceanographic measurements) that the polynya is a result of deep wintertime convection (possibly involving chimneys like those in Figure 17d). Typical positions of the polynya are shown in Figure 18 [Martinson et al., 1981], which shows that the polynya lies far from coastlines and so cannot simply be a result of wind blowing ice away from coasts. The polynya occurs irregularly: it was present during the entire winters of 1974-1976; in 1973, 1977, 1979, and 1980, small, short-lived polynyas were seen; and in 1978 no polynya occurred. Since the heat loss from the Weddell gyre to the atmosphere can be increased by as much as 50% when the polynya is present, and the polynya forms a vast 20°C sea surface temperature anomaly (the ice surface temperature averages about -20°C), it is important to understand its origins. Martinson et al. [1981] propose that the deep convection produced via brine release from sea ice can be sufficient to bring up 0°C warm deep water which can melt the ice and form the polynya. (This process is naturally irregular even if the atmospheric forcing is regular.) Such convection should yield deep water or about σ_2 ~37.22-37.23, which is about 0.1°C colder and 0.01 fresher than the surrounding water each time a polynya occurs. To support this, examination of the deep temperatures around Maud Rise (shaded in Figure 18) shows a cooling of 0.3°C between 1973 and 1978, after three polynyas.

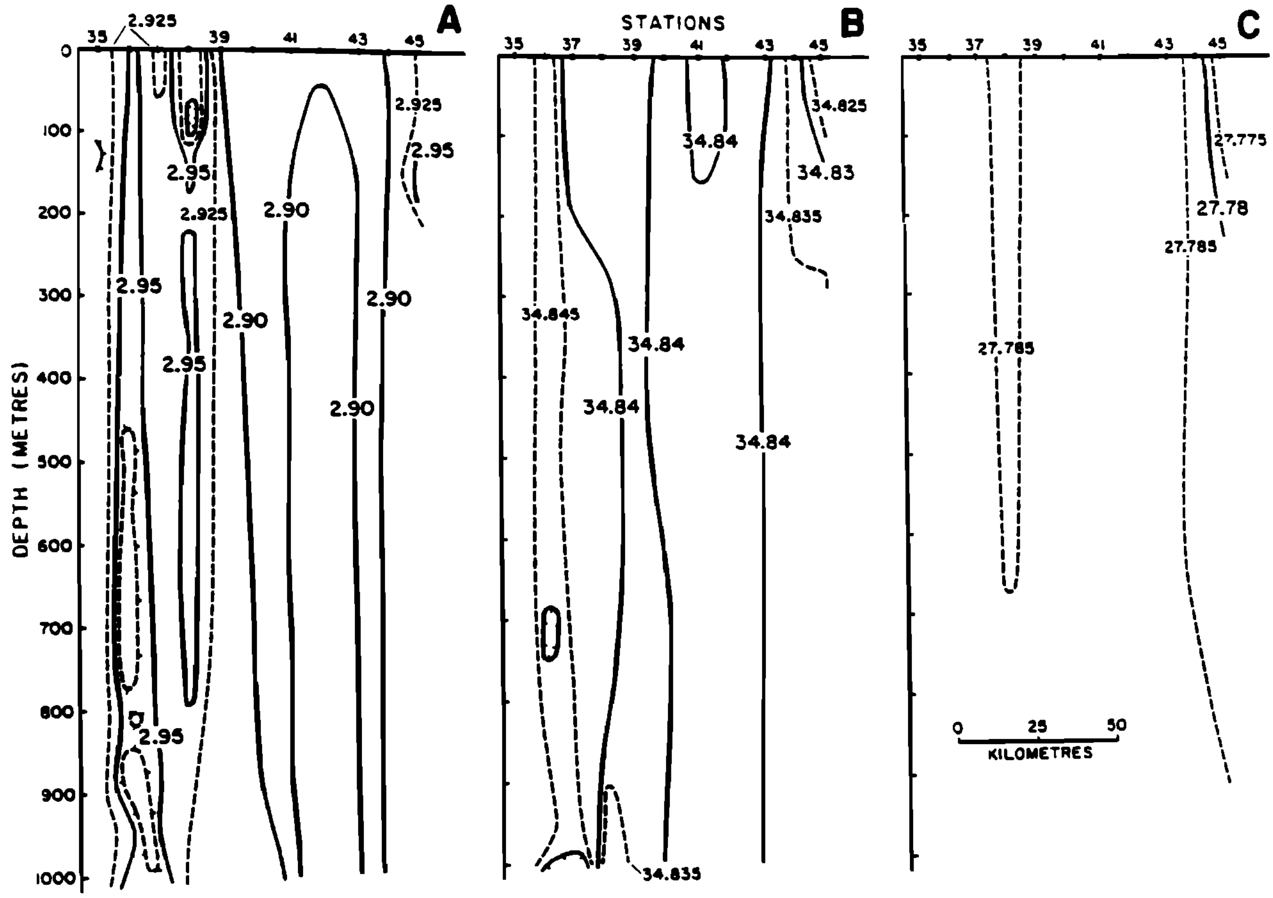


Fig. 15. (a) Potential temperature, (b) salinity, and (c) potential density in the upper 1000 m of the Labrador Sea along the line B-B (shown below), March 1976 [from Clarke and Gascard 1982].

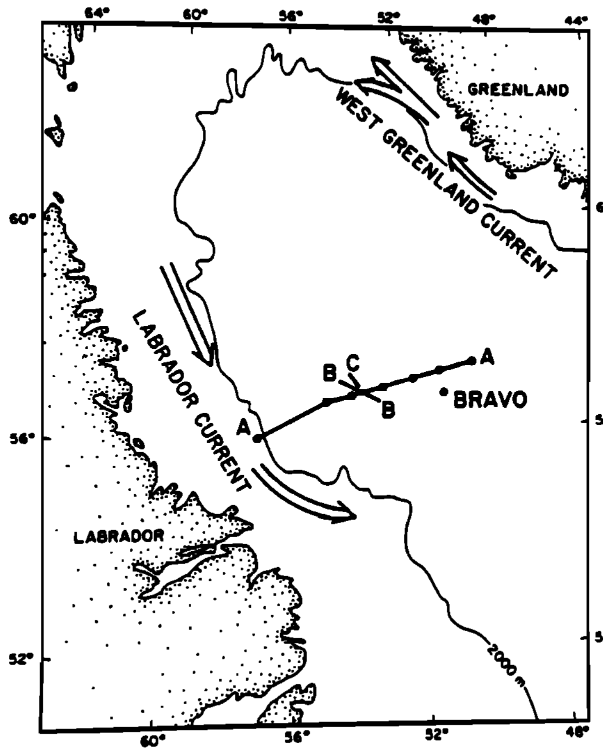


Fig. 15. (continued)

Martinson et al. estimate a lower limit of an annual rate of 10^6 m^3 s^{-1} of deep water formed by this mechanism. Gordon [1982] has increased this estimate to 1.6-3.6 x 10^6 m^3 s^{-1} during 1974-1976, equivalent to a rate of 3.8-7.7 x 10^6 m^3 s^{-1} during deep water production itself.

The Greenland Sea

The existence of deep water in the Greenland Sea has been recognized for over 100 years. It forms a domed structure centered on 0°E at about 75°N and comprises water colder than 0°C with a high salinity of 34.90. The bottom water is yet colder, with temperatures below -1°C (see Figure 19 [from Carmack and Aagaard, 1973]). The circulation around the dome is again cyclonic. For these 100 years the origin of this water mass has remained a mystery. The classic model of Nansen [1902] suggested large-scale wintertime convection extending eventually to the bottom. Metcalf [1955], finding no evidence of this in his winter observations, posited slantwise sinking of dense water along isopycnals.

Carmack and Aagaard [1973] examined all winter hydrographic stations from the area and found no surface waters sufficiently dense to sink to the bottom by any path, vertical or slanting. Many researchers now believe that the process forming Greenland Sea deep water is similar to the other processes in this section, in that formation regions are narrow in both space and time. A calculation by Killworth [1979] estimated that there was a probability of 80% that bottom water formation in the area would not have been observed to date. There are currently (1982) plans by several countries for fine-scale winter observations at or near the ice edge, to attempt to find convection actually occurring. Tracers such as cesium which were introduced into the sea only recently (e.g., from the Windscale nuclear reactor) and which are only now entering the Greenland Sea would be extremely useful for data interpretation; the large water samples involved, however, may preclude their use.

Other Areas and the Subpolar Oceans

The above areas do not exhaust the known areas of open-ocean convection, but they comprise the well-observed areas. There is, for example, evidence of other open-ocean convection in the Mediterranean, the Adriatic Sea [Zoré-Armanda, 1963], the Ligurian Sea (E. Salusti, private communication, 1981), and Baffin Bay. The Dead Sea has overturned completely, or is in the process of so doing, because of surface evaporation and the diversion of inflowing rivers for agricultural purposes [Steinhorn et al., 1979]. The boundary between the Weddell Sea and Drake Passage may occasionally undergo deep convective events [Deacon and Moorey, 1975].

Convection extending to great depths occurs each winter over most of the subpolar oceans, of course. Robinson et al. [1979] give a chart showing (a slight overestimate of) the depth of the surface mixed layer in wintertime. (It portrays depths to the top of the thermocline). There is clear evidence of convection well below 600 m in the entire subpolar area of the North Atlantic, for example. Such convection is believed to form water masses like the 18° water [Warren, 1972] or 14° water [Worthington, 1976]. The horizontal spreading of these water masses after formation forms an efficient way for anomalous wintertime atmospheric features to extend their effect far below the surface layer of the ocean.

A notable omission from this compilation of convection areas is the North Pacific. Warren [1982] addresses the problem of why no deep water is formed in this region. He concluded that the vertical stratification is too large to permit sinking. This seems to be caused by low surface evaporation, which is due to low surface temperature, caused in turn by a relatively weak northward flow of warm water.

5. The Physics of Open-Ocean Deep Convection

The obvservations in section 4 suggest, correctly, that there are various fundamental ingredients involved in open-ocean deep convection. These are rather different from those needed for convection near boundaries. A summary of some of these ingredients, together with other salient facts, is given in Table 1 for some of the major production regions (and some which are hypothesized). These will now be discussed in detail.

The first requirement for open-ocean convection, except in the more confined areas such as Bransfield Strait (which can be thought of as a reservoir along the lines discussed in section 3), is a background cyclonic circulation. This is necessary to form an upward 'doming' of isopycnals in the center of the cyclonic gyre and so to reduce the vertical stability of water columns within the gyre.

The second requirement, to borrow the Medoc term, is a preconditioning. The preconditioning (operating over a period of weeks) creates a region of very weak static stability within the cyclonic dome which will then become eligible for convection if the surface forcing is sufficiently intense. Several mechanisms appear capable of providing a preconditioning. In the Gulf of Lions the topographic feature known as the Rhône fan, sited underneath the Medoc chimney region, can modify the circulation locally so as to reduce the stratification in the area [Hogg, 1973]. In the Labrador Sea, preconditioning is seen as a small cyclonic gyre embedded in the larger cyclonic circulation of the Labrador Sea proper [Clarke and Gascard, 1982]. The scenario suggested by Clarke and Gascard for this small gyre is intense heat loss just south of the ice edge in the Labrador Sea caused by intensely cold winds blowing off the ice. This forms a 200-km-diameter pool of preconditioned water. What role brine release from freezing ice may play here is unknown.

It is not known whether preconditioning occurs in the Weddell and Greenland seas. Killworth [1979] suggests that baroclinic instability can occur within the Weddell gyre due to the weak vertical stratification. This instability will be on scales of the order of 20 km (the deformation radius) and will occur predominantly at some distance from the gyre center, because there is no mean flow (and hence no source of energy for the instabilities) at the gyre center. Cyclonic eddies will have reduced static stability still further and be more prone to convection during wintertime if the surface forcing is strong enough (which makes convection in the northern half of the gyre less likely, for example, as the region is too far north). Certainly, preconditioning by this mechanism, or by topographic features, is necessary; Killworth's [1979] model predicts that most of the gyre would undergo convection without a method for preselecting certain narrow areas. Laboratory tests of this suggestion have been made by Keffer [1979] who drove a two-layer rotating fluid by a differentially rotating lid. After a domed interface had been created, ice was added to cool the lid. After some minutes, narrow chimneys of convective fluid began to appear, not, significantly, at the center of the gyre, but on the slopes of the dome. Keffer [1979] examined the stability of the laboratory situation and showed that the width of the chimneys was set by friction as well as by the deformation radius. (It is interesting that of the 55 eddies observed by the Arctic Ice Dynamics Joint Experiment (AIDJEX) project in the Arctic in 1975-1976, only two were cyclonic [Hunkins, 1979]. Is this an indication of convection destroying cyclonic eddies? Or are the eddies sufficiently nonlinear that centrifugal terms partly balance the Coriolis terms in cyclonic eddies, as can happen in the atmosphere?)

Martinson et al., [1981] note that the (hypothesized) convection within the Weddell polynya is sufficient to precondition the area for the next winter, despite a gradual drift of the water masses with the mean flow. However, an original pre-

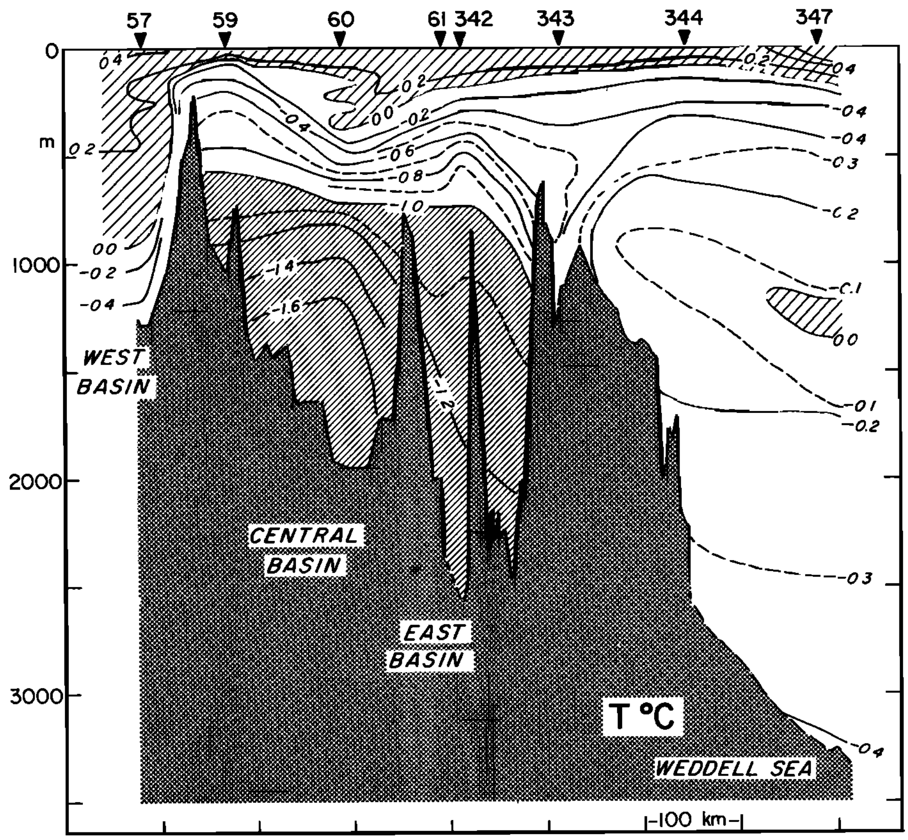


Fig. 16a

Fig. 16. (a) Potential temperature, (b) salinity, (c) oxygen, and (d) silicate along the axis of the Bransfield Strait and adjacent areas [from Gordon and Nowlin, 1978].

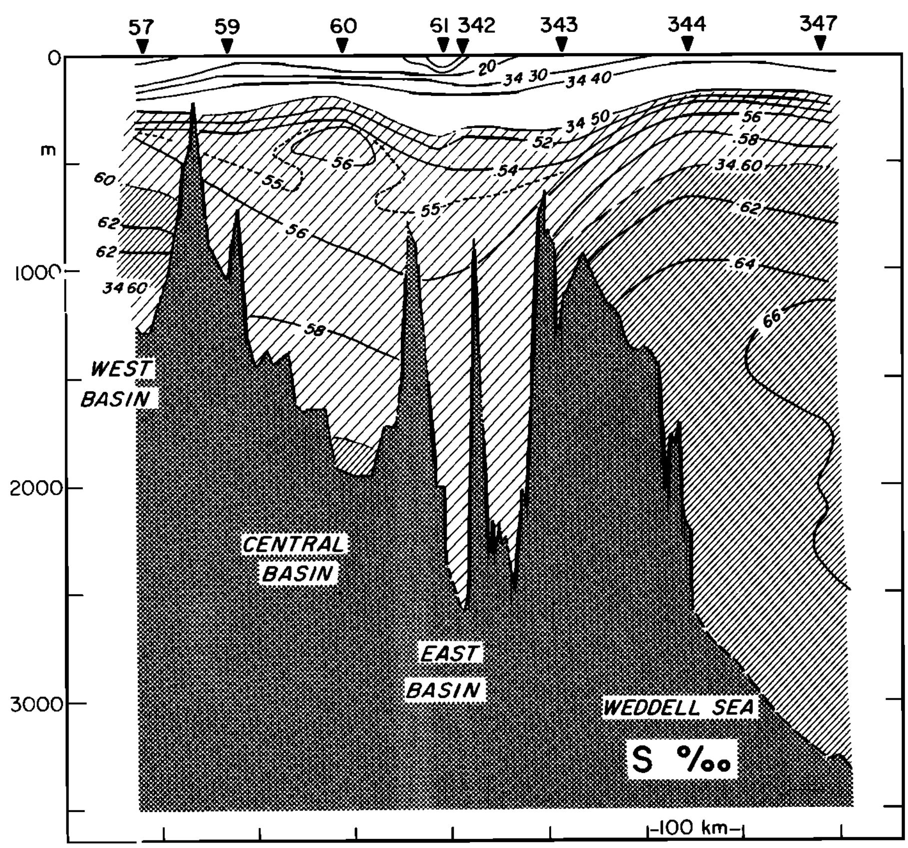


Fig. 16b

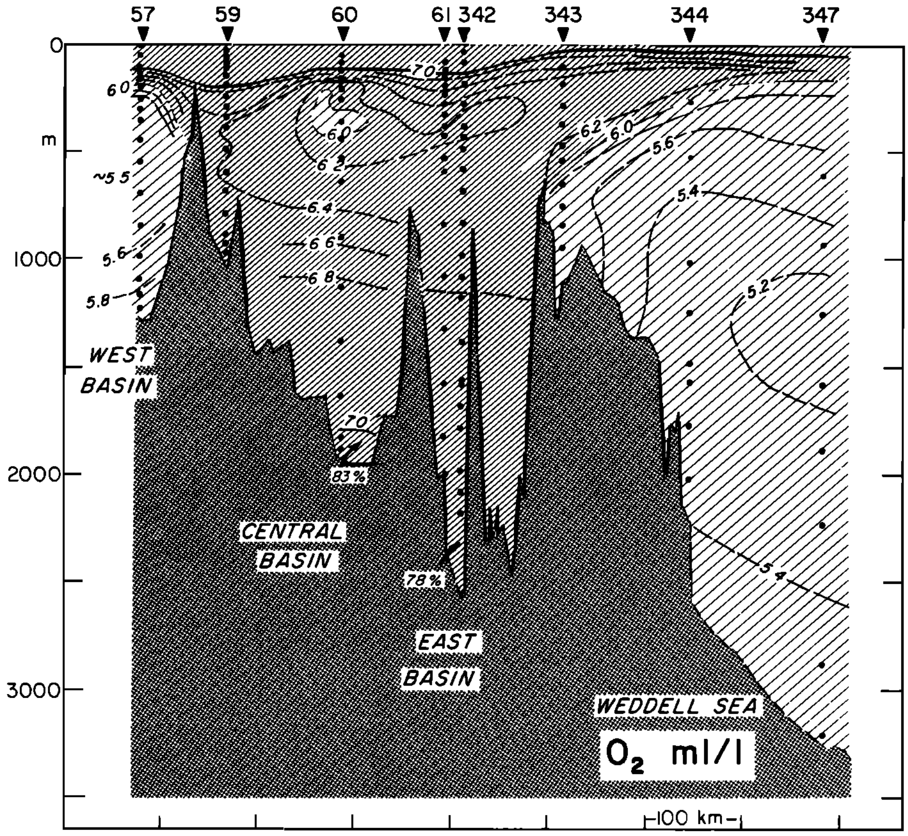


Fig. 16c

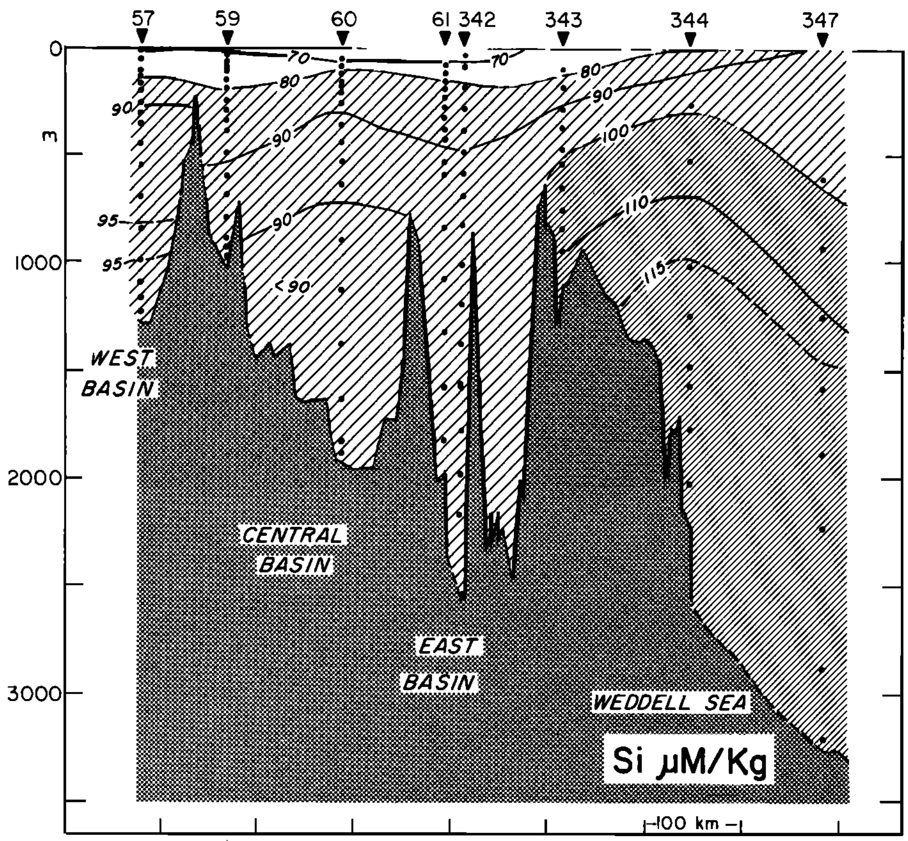


Fig. 16d

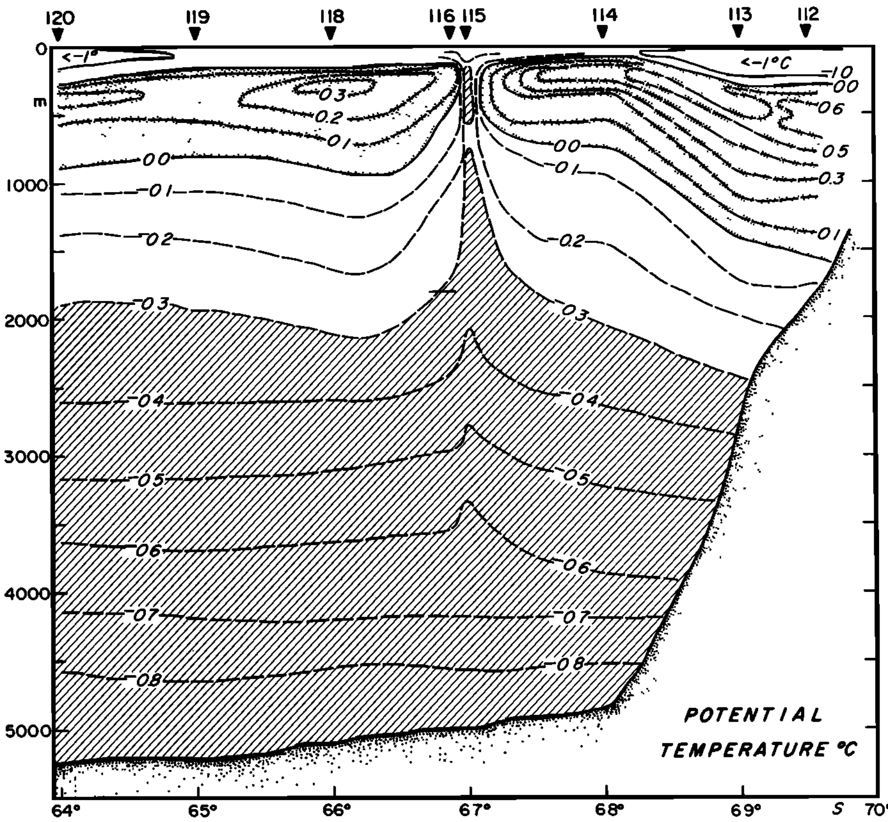


Fig. 17a

Fig. 17. Section northward from the Antarctic coast in the Weddell gyre, showing the 'chimney': (a) potential temperature, (b) salinity, (c) oxygen, and (d) potential density referred to zero, 2000, and 4000 m. A fine-scale survey near station 115 confirmed the width of the chimney [from Gordon, 1978].

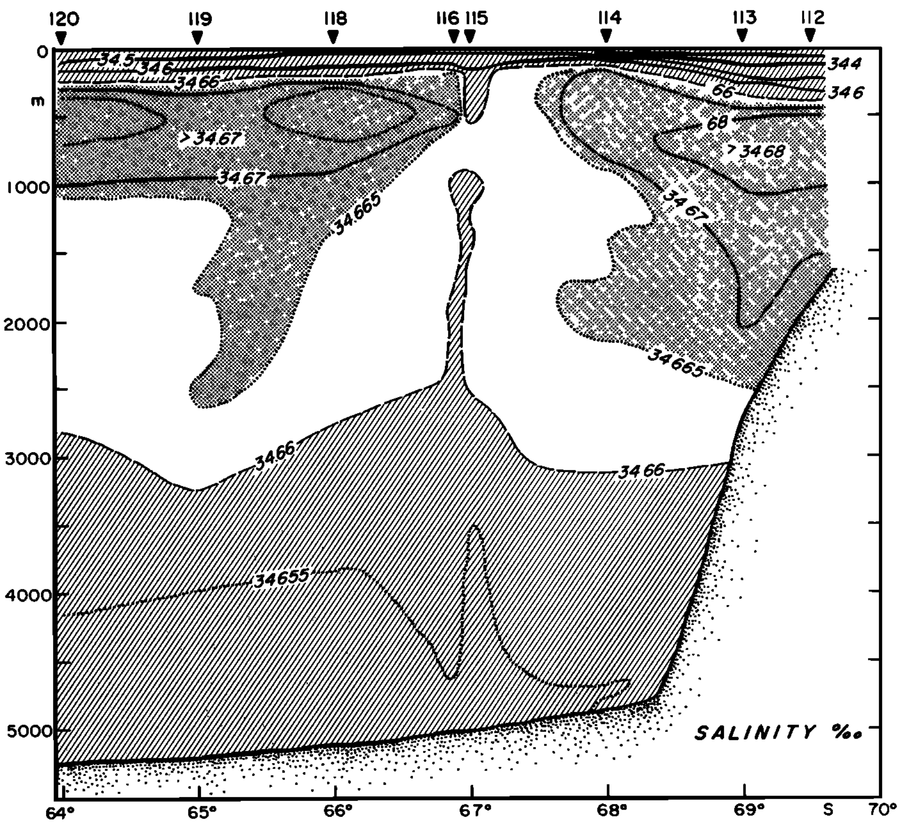


Fig. 17b

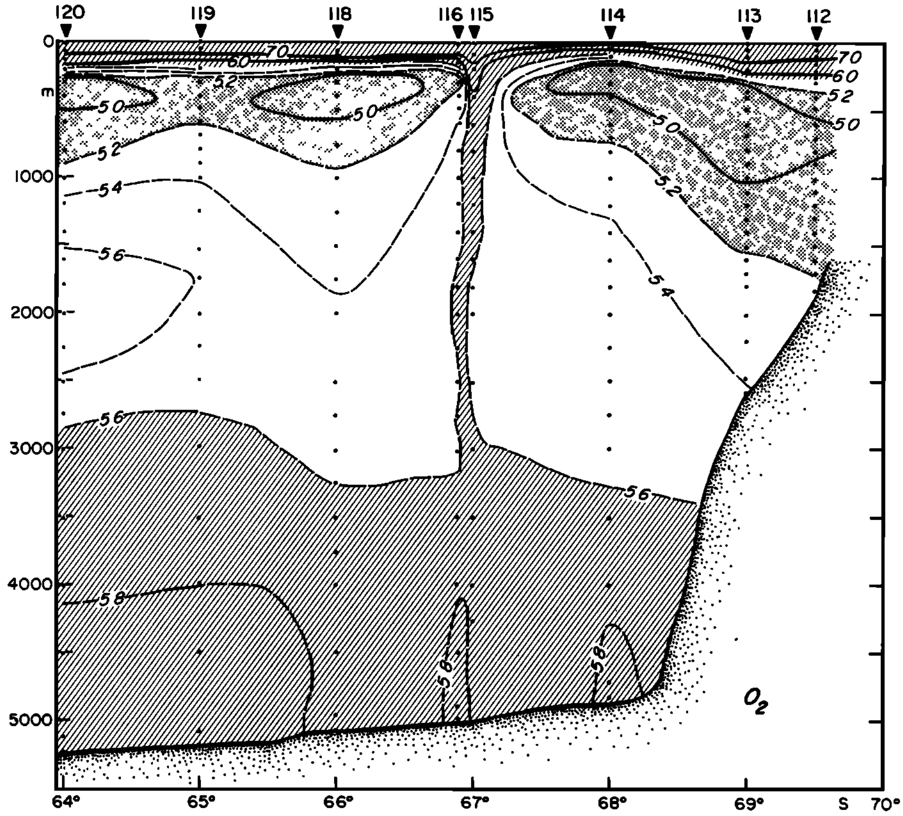


Fig. 17c

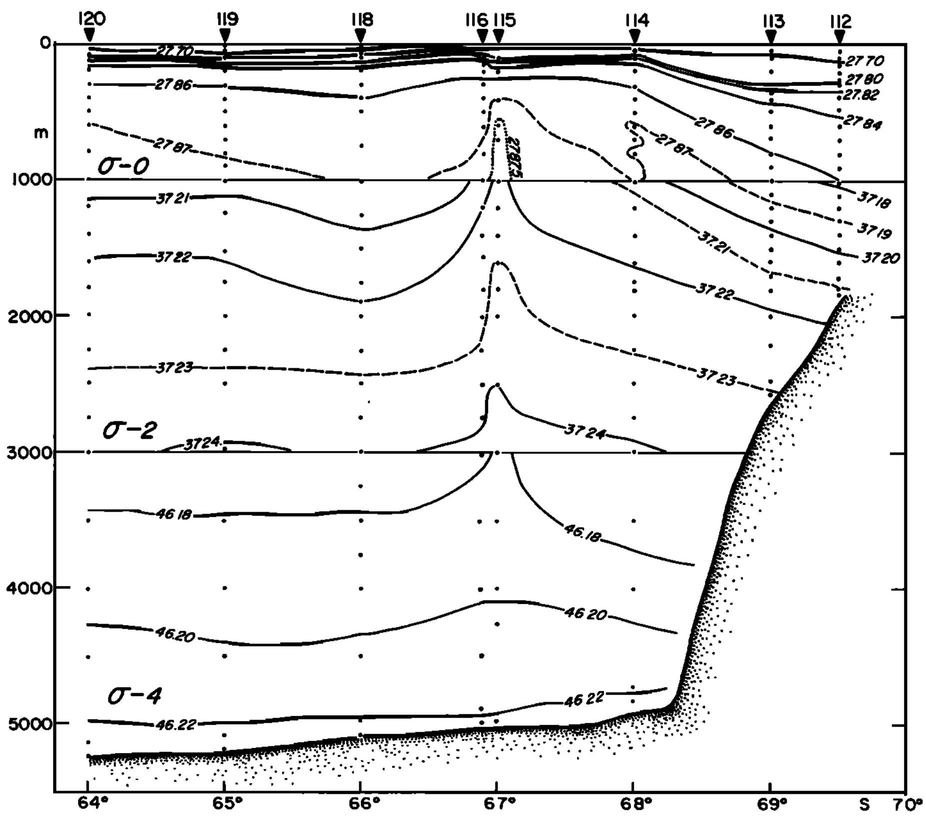


Fig. 17d

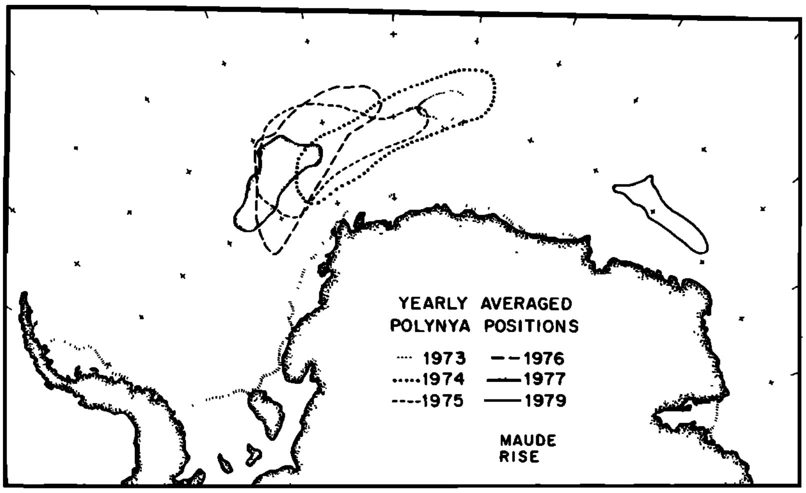


Fig. 18. Yearly averaged polynya positions in the Weddell gyre, from satellite data [from Martinson et al., 1981].

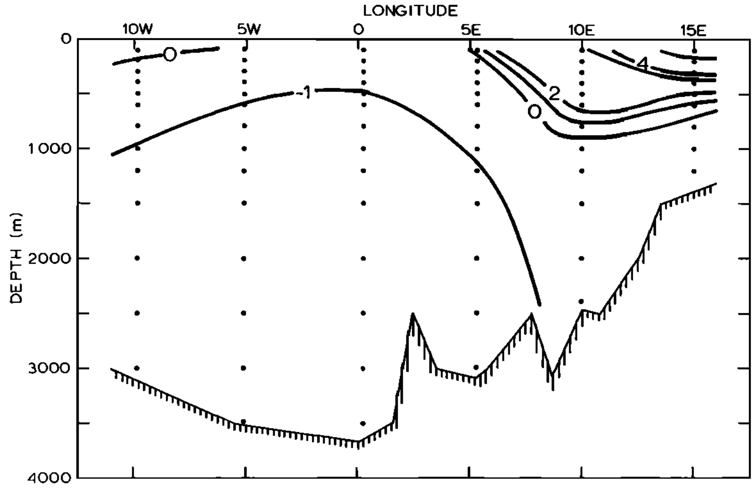
conditioning is necessary to create the convection area in the first place. Because of the great size of the polynya, they propose wind-driven upwelling, together with a preferential depth for their convective mechanism, which acts to form a polynya area on the southern side of the Weddell gyre. Many possibilities, discussed by Gordon [1982], could have initiated the polynya. These include topographic features such as Maud Rise; atmospheric effects, in particular a reduced precipitation or an increase in ice formation; an increased flux of deep water salt into the region; or a fluctuation in the position of the Weddell gyre.

In the Greenland Sea it is now believed that the preconditioning (certainly necessary; Killworth's [1979] chimney model found no evidence of deep convection after 6 months of forcing) produces narrow regions, of the order of 10-20 km, predisposed to convection. The relevant process may involve baroclinic instability, the variable ice front in the northern Greenland Sea, topography, double diffusion, or several other features. More winter cruise data are urgently needed.

As with boundary deep convection, the areas all share the feature that more than one water mass is involved. Unlike those discussed in section 3, however, the differing water masses probably have an important role to play during vertical convection, since salinity and temperature both contribute to the vertical stability of a fluid. The existence of several water masses provides subsurface sources of heat and salt which can be exposed to the surface during convection. Water columns stabilized by heat can be destabilized by cooling; water columns stabilized by salt can have this stability eroded by cooling (or removed directly by ice formation or evaporation). At polar temperatures the equation of state permits an unusual form of one-way convection when several water masses are involved [Killworth, 1979]. If there is mixing to some depth, below which is water with different T-S properties, further surface forcing can mix down water particles which, though neutrally stable within the surface environment, are denser in situ than the lower water mass. The result of steady surface forcing above a fluid comprising two water masses can then be a gradual erosion by the mixed layer, followed, when the lower water mass is reached, by a dramatic increase in depth of the mixing, often by up to 2000 m [Killworth, 1979]. It would be interesting to have CTD winter data from the Antarctic to see if these theoretical predictions can occur or whether the mixing process is far more complicated than these simple models suggest.

All known examples of open-ocean deep convection share the feature that sufficiently intense surface forcing is involved. In the Mediterranean and Labrador seas the mechanism is heat loss by sensible and latent heat to the very cold winds blowing over the area. J.-C. Gascard (private communication, 1981) estimates the rate of heat loss, conservatively, as 480 W m^{-2}, surely one of the strongest surface coolings reported. The Weddell chimney and polynya are believed to be driven by brine ejection from growing sea ice (together, of course, with heat loss into the ice, although this usually plays a weaker role than saline ejection because of the strong dependence of density upon salinity). In the Greenland Sea both cooling and brine ejection may be important.

Concomitant with surface forcing is the violent mixing phase, or rapid vertical convection and mixing. The evidence seems to be that the convection takes place in cellular structures like those seen in the laboratory, with horizontal and vertical scales of similar sizes [Clark and Gascard, 1982]. The mode of convection seems to be nonpenetrative rather than penetrative [Anati, 1970]. In other words, mixing occurs in such a way as to keep the density structure a continuous function of depth. Mixing is not driven further into the fluid (i.e., by wind mixing), which would produce a density step, except possibly early in the process, when the effects of wind or ice keel stirring can reach the bottom of the mixed layer [cf. Kraus and Turner, 1967]. Observed discrepancies in the rate of deepening of the Medoc chimney [Anati, 1970] can be accounted for by a weak cross-frontal circulation which adds stable fluid to the column undergoing cooling [Killworth, 1976]. There is some evidence of penetrative con-



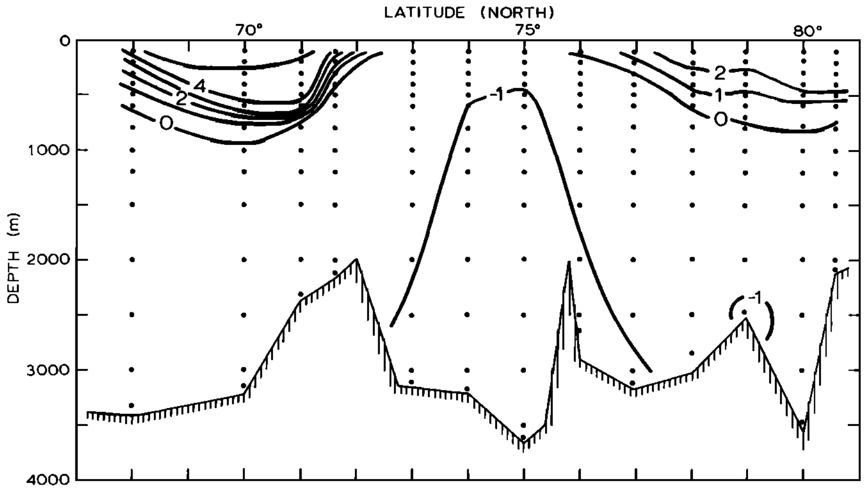


Fig. 19. Temperature sections in the Greenland Sea, along 75°N (top) and along 0°E (bottom), showing doming and bottom water [from Carmack and Aagaard, 1973].

vection in the Arctic mixed layer [Solomon, 1973; Lemke, 1979]. No large-scale ocean circulation model has attempted to deal with the convective process in any but the simplest fashion (of replacing statically unstable columns with well-mixed ones), but the small amount of observational detail currently available precludes the need for more sophisticated modeling.

The final part of the process is a breakup, or sinking and spreading phase. As the Mediterranean observations show, such a process is difficult to observe. Baroclinic instability is certainly important in the removal of the chimney [Killworth, 1976; Gascard, 1978]. However, instability mechanisms alone are insufficient to erase the chimney. P. D. Killworth (unpublished work, 1977) simulated the Medoc chimney within a fine-scale version of the nonlinear Geophysical Fluid Dynamics Laboratory ocean circulation model [Bryan, 1969]. The violent mixing phase and ensuing eddying after termination of the surface cooling were both well simulated and agreed with theory. However, 3 weeks after the end of the cooling the (eight) eddies were still circulating around a central, coherent, neutrally stable column. There was no sign of any breakup or slumping (even by a deformation radius, which might have been expected). This suggests that vertical shear in the global cyclonic circulation, topography, and mixing by processes such as internal waves are probably involved in the breakup of the Mediterranean and Labrador chimneys.

Partial confirmation is found in the anomalous survival into summer of the Weddell chimney [Gordon, 1978], where topographic gradients are locally weak and the vertical shear of the mean circulation is also small (about 1 cm s^{-1} over the whole depth). However, despite the tendency of geostrophic flow to be around such a chimney, it is still surprising to see its survival 6 months later in a coherent fashion.

6. Interaction Between Deep Convective Events and the Remainder of the World Ocean

The discussion in the preceding sections has treated the formation of deep water as an event isolated from the surrounding ocean. This reflects, historically, most of the research on deep convection. But the existence of deep convective

TABLE 1. Some Features of Observed Open-Ocean Formation

Area	Size, km	Average Amount, 10^6 m^3 s^{-1}	Regularity	Background Circulation	Number of Water Masses	Preconditioning	Surface Forcing and Vertical Mixing	Breakup
Gulf of Lions	30-50	1	most winters	cyclonic	> 1	yes	yes	yes
Labrador Sea	10-50	1-4	some winters	cyclonic	> 1	yes?	yes	?
Bransfield Strait	450	0	?	?	1	irrelevant	yes	too big
Weddell chimney	20-30	1	?	cyclonic	> 1	yes?	yes	?
Weddell polynya	600	1(or 10?)	most winters	cyclonic	> 1	yes	yes	too big
Greenland Sea	small?	1-3	?	cyclonic	> 1	yes?	yes	yes?

See text for discussion.

events must in some way modify the circulation of the surrounding water. Is this modification large or small? What time scales are relevant for the modification?

Of course, the dense water has to flow somewhere after its formation; because of its density its route is prescribed to lie along the ocean floor. Laboratory and analytical work by Gill et al. [1979], Gill [1979], and Talley [1979] for two-layer fluids shows that the transformation of light to dense fluid yields outward motion of the new dense water only over a deformation radius. The inclusion of friction (interfacial or bottom) and/or thermal and saline diffusion permits the relaxation of the geostrophic constraint, and the dense water may leave the area of its formation, typically as a boundary current [Talley, 1979]. The constraining effects of geostrophy can also be seen with intrusions into continuously stratified fluids [Gill, 1981], where again, friction is required to allow spreading over more than a deformation radius.

Real fluids can circumvent the geostrophic constraint by dynamics not involving friction. Griffiths and Linden [1981b] studied the laboratory situation of an axisymmetric homogeneous fluid, with a source of water at a density less than that of the fluid. A domed mass of the injected water is produced, with a front between it and the surrounding fluid. When the dome is sufficiently wide (in terms of the deformation radius), it becomes unstable to one of several types of disturbance (in a qualitatively similar manner to the frontal instabilities mentioned earlier). On the large scale, do such instabilities merely act to entrain and mix fluid, or is there a wider dynamical effect?

When the dense fluid begins to occupy more than about 100 km of latitude, the effects of varying Coriolis parameter cannot be ignored. The effects of this are felt most strongly in the vortex-stretching equation, which shows that geostrophic, mass-conserving bottom flows cannot move equatorward and upwell into the rest of the ocean. Warren [1981] discusses the effects of this constraint in some detail. It turns out that dense bottom flows must, if they are to flow equatorward, be bounded on one side by a coastline. This is automatic for any of the overflows of dense water, which are bounded by two coastlines at the sill, which also acts to limit the net flow of dense water [Gill, 1976]. The frontal instabilities mentioned in section 3 may also act as a limit. After passing through such constraints, the bottom flows appear to keep a coastline on their right (left) in the northern (southern) hemisphere. In the course of their progress equatorward, layers may stack vertically. For example, Labrador Sea water moves south around the Labrador coast but rides above the Atlantic deep water, which itself has derived from the Denmark Strait overflow. Warren [1981] surveys these flows and their dynamics; the most detailed attempt to include the deep flows in a circulation scheme is by Worthington [1976] for the North Atlantic. There is biological evidence (D.G. Martinson, private communication, 1981) that the Antarctic bottom water observed in the Atlantic may derive from open-ocean convection with the shelf slope variety constrained to remain in the Weddell gyre.

The problem of how the deep convection affects its environment is largely bypassed in these studies. When thinking about environmental effects, one must specify a time scale. On the flushing time of the oceans (100-1000 years) any change in deep water formation will have had a chance to alter the entire circulation of the ocean. This will, presumably, be at some highly diluted level, as a $1^{o}C$ change in bottom water temperatures, maintained for 10 years at a flux over all sources of 50×10^6 m^3 s^{-1}, yields an average change of only 10^{-2} ^oC when advected through the 1.4×10^{18} m^3 of ocean. Even such a small change may, of course, be capable of changing vertical stratifications significantly in polar regions and of increasing or decreasing the amount of convection possible. On a longer time scale (or for smaller oceanic basins) a source of deep water, even if it varies with time, can create a stratified environment around it by upwelling of dense water produced when the source is at its strongest [Baines and Turner, 1969; Killworth and Turner, 1982]. This could suggest the existence of deep convection around much of Antarctica. The convection could modify its environment so as to minimize the evidence of the convection by a modifcation of the local water masses. On a short time scale it is possible that the creation of deep convection can drastically modify an area of ocean: for example, Aagaard and Coachman [1975] suggest that proposed Russian river diversions could, by removing a surface freshwater source, weaken the vertical stability of the Arctic Ocean and permit wintertime convection. This could mix up warm Atlantic water and conceivably melt the Arctic ice cap, with ramifications for world climate.

On the time scale of years to decades we know very little about the effects of deep convection in the world ocean climate (tritium measurements by Ostlund et al. [1976] show that deep water does indeed spread a long distance through the world ocean on this time scale). Nor do we know even how heat and salt are carried through the ocean, whether by mean flows (in the North Atlantic [Bryden and Hal, 1980]) or by eddies (in the Drake Passage [Sciremmemano, 1980]). (The excellent survey by Bryden [1981] summarizes the various estimates of ocean heat transport.) Certainly, some effects must be important: the 20^o sea surface temperature anomaly produced by the Weddell polynya must have a strong effect on the atmospheic circulation due to the large size of the polynya and therefore back on the ocean elsewhere. Unlike the effects of tropical sea surface temperature anomalies [Rowntree, 1972], this large polar anomaly has not been studied by theoretical meterologists.

It seems likely that the main effect of convection on scales of years to decades will be a series of modifications to the mean and deep flows by horizontal advection of vertically convected water masses. Under normal conditions, water masses like 18^o water spread slowly southward (along isopycnals? or in isolated blobs such as the 'meddies' of McDowell and Rossby [1978]?), each year's supply replenishing the area in front, vacated by the previous year's supply. We can envisage, after an abnormally cold winter, that deeper convection than usual, over a wide area, yields a collection of new water masses whose characteristics differ from the usual ones. They will, for example, be deeper than usual and probably colder. Those in polar regions may be saltier or fresher depending on the details of the forcing and mixing. As these masses spread equatorward on isopycnal surfaces, they will tend to change the local density structure by thickening homogeneous layers [Gill, 1981] and hence, by thermal wind, the mean flow in the areas passed through. Modification of the mean flow will, in turn, alter the properties of wave propagation (as may be occurring in the deep eddy signals described by Dickson et al. [1982] and production and thus react back on the atmosphere by altering the sea surface temperature. As yet, however, there are no estimates of the magnitude of such processes.

7. Discussion of Future Needs

The main feature common to both convection near a boundary and open-ocean convection is the presence of a suitably intense surface forcing. This is either a direct cooling (by latent and evaporative losses) or a salinity enhancement (when sea ice and therefore brine are being produced by a cold air mass above the ice). We still know little about transfers of heat and salt between ocean and atmosphere at polar latitudes, however (even the sign is sometimes in doubt). Improvement of our knowledge of transfers in these regions needs a high priority. As an example of the wide variation possible, consider the following estimates. Best fits to the Bunker and Goldsmith [1979] Atlantic heat flux data give (P. D. Killworth, unpublished data, 1982) net infrared losses to the atmosphere as functions of surface temperature T,

Water

$$IR \simeq 27.4 + 0.9T \quad W\ m^{-2}$$

Ice

$$IR \simeq 26.3 + 1.0T \quad W\ m^{-2}$$

which depend very little on whether the surface is liquid or ice. Fits to latent and sensible heat between atmosphere and surface, however, are rather different and depend on the proportion of ice cover; Vowinckel and Orvig [1970], indeed, find latent and sensible heat directed from air to ice for a fully ice-covered ocean. However, fitting two Legendre polynomials to incoming solar radiation as a function of latitude yields a best fit for incoming radiation at the north pole of 60 W m^{-2}. Even allowing for albedo, a gain of sensible and latent heat by the ice implies a net gain of heat by the ice, which must also be heated from below by the seawater. Because of the presence of leads, ice-covered oceans must be losing latent and sensible heat. An amalgamation of data suggests the formulae

Water

$$L + S \simeq 77 + 2.2T \quad W\ m^{-2}$$

Ice

$$L + S \simeq 9. \quad W\ m^{-2}$$

Assuming albedos of 0.6 for ice and 0.1 for water (both of which are still in doubt after many years of research!), we find, for example, that ice at

$-20^{o}C$ gains 9 W m^{-2}, ice at $-10^{o}C$ loses 1 W m^{-2}, and water at $-1.8^{o}C$ loses 45 W m^{-2}.

To be sure, these estimates may be quite inaccurate, but they help to make the point that whereas the surface heat balance may well be close to zero for ice-covered regions, we are unsure of its size to within an order of magnitude. (It is known that thin ice areas (<80 cm) contribute 96% of the heat loss from ice-covered regions, however, [Maykut, 1977].) Knowledge of evaporation and precipitation is so poor that an estimate of the salinity balance at the surface is almost impossible.

Gordon's [1981] estimates of heat loss from ocean to atmosphere in the band 60°-70° make the same point. Averaged seasonally, an ice-free ocean loses 75 W m^{-2}, a partially ice-covered ocean loses 31 W m^{-2}, and a totally ice covered ocean loses only 8 W m^{-2}. Again there is an order of magnitude difference in heat losses depending on the degree of ice cover. Although satellite coverage is now sufficient to describe the ice cover in polar regions much of the time, we still need much information, such as albedos, to be able to close the heat budget satisfactorily. Vowinckel and Orvig [1970] given an excellent discussion of surface fluxes in the Arctic.

Heat and salt transfers from sea to ice are even less well known. Typically, turbulent heat coefficients are used for heat losses from ocean to ice [Welander, 1977; Martinson et al. 1981], but these are certainly not known to within an order of magnitude. Salt transfer is more straightforward: given the salinity of sea ice, one need merely conserve salt between sea and ice as the latter freezes or melts. Solomon [1973] gives a more accurate version of this. The effects of surface ablation of snow (and rain) are less easy to handle (see Martinson et al. [1981] for a crude attempt). The effect of iceberg melting [Neshyba and Josberger, 1980] either on the heat-salt balance or on the mean stratification is also not well understood.

The list of oceanographic experiments and observations whose results are needed is almost endless, with a growing need for closely spaced stations. The simple physics discussed in this paper needs to become more sophisticated. To do this, we must be able to deduce physical quantities from data, and this in turn needs detailed measurements. Conversely, more work is needed on models of deep convection. On the large scale it is probably unnecessary yet to model convection more accurately than the Bryan [1969] adiabatic instant mixing model. On smaller scales, however, there is a need for the improvement of mixed-layer models to allow convection to depths of the order of 600 m. How do wind and/or ice keel stresses partition themselves throughout the mixed layer? Models of the effects of convection on the surrounding circulation are needed, and a detailed treat of polar eddies and their ability to transport heat, salt, momentum, and vorticity should also be undertaken.

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