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POLAR OCEANOGRAPHY

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I. Introduction

Oceanographic progress pertaining to polar waters for the 1979-1982 period is briefly presented within three research areas: Southern Ocean; Arctic; and ocean-ice interaction, after a discussion of general review articles and atlases. A conclusion is added which presents the author's view of future research direction in a general sense. Emphasis is placed on published U.S. work, though key non U.S. publications are included. The bibliography contains articles not cited in the text. Polar snow and ice, meteorology and climate are covered elsewhere in the IUGG review, and mention of these subject areas within this article is made only if they are of direct relevance to oceanography. This article is not meant to be a complete synthesis of research, but rather to provide a list of publications with an associated narrative of the recent U.S. polar oceanographic research.

II. Polar Review Articles and Atlases

A major objective of polar oceanographic research is directed towards understanding the climatic relevance of the polar regions [Baker, 1979; Goody, 1980; Polar Group, 1980; Saltzman and Moritz, 1980; Ackley, 1981a; Fletcher, Radok and Slutz, 1982]. This involves study of exchange rates of heat and water over a variety of spatial and temporal scales for horizontal and vertical processes. During the 1979-1982 period a remarkable degree of progress has been made along these lines. Four reviews are particularly useful: Carmack [1982] discusses our oceanographic knowledge of ice covered oceans; the impact of polar ocean water mass modification on the world ocean is discussed by Warren [1981]; while Killworth [1982] provides a review of field and theoretical considerations of ocean convection, with a natural polar ocean emphasis; and Bryden [1982] reviews eddies in the Southern Ocean.

A selection from the extensive Southern Ocean hydrographic data set is presented in atlas formats: subjective contouring [Gordon and Molinelli, 1982] and in objective analysis with a microfiche section [Gordon and Baker, 1982]. Distributions of temperature, salinity, density, oxygen and nutrient on horizontal and vertical sections, core layers, density surfaces, volumetric temperature-salinity, and profiles are included, as are maps of relative dynamic topography and geostrophic currents. An atlas of the Bering Sea has been prepared by Sayles, Aagaard and Coachman [1979], and Kinder [1981] reviews Bering Sea oceanography.

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III. Southern Ocean

A. Drake Passage Antarctic Circumpolar Current and Associated Fronts

Significant advances in understanding the dynamics and thermohaline zonation of the Antarctic Circumpolar Current (ACC) have been made within the Drake Passage region by the International Southern Ocean Studies (ISOS) project [Neal and Nowlin, 1979]. The 1979 ISOS current meter and hydrographic data reveal a 117 to 134 × 10^6 m^3/sec eastward volume transport in the Drake Passage, which is in general agreement with the 124 × 10^6 m^3/sec value determined from the 1975 ISOS data [Whitworth, Nowlin and Worley, 1982] and estimates of 127 and 139 × 10^6 m^3/sec using an objective analysis technique applied to the 1975 data set [Fandry and Pillsbury, 1979]. The latter authors attribute 27 × 10^6 m^3/sec to the barotropic and 100 × 10^6 m^3/sec to the baroclinic transport, with a total transport range of 220 × 10^6 m^3/sec within the 35 weekly estimates. However, as the authors point out, the large mooring separation relative to the correlation scale of the velocity fluctuations (about 40 km) [Sciremammano, Pillsbury, Nowlin and Whitworth, 1980] may cause substantial overestimations of the time variability. The net ACC transport through the Drake Passage may be steadier than previously thought. Time variability of the ACC was also studied by Wearn and Baker [1980] using fluctuation in pressure at the 500 meter level on either side of the Drake Passage. They find large variability (80 × 10^6 m^3/sec if barotropic) with fluctuations of time scales larger than a month being highly correlated with average circumpolar wind stress. The ACC variation lags the wind by only nine days, suggestive of a barotropic response. This is consistent with the steady baroclinic transport [Whitworth, Nowlin and Worley, 1982]. Chelton [1982] suggests that this high transport-wind stress correlation of Wearn and Baker [1980] is partially a consequence of a strong semi-annual signal in both data sets.

Estimations of the mean and fluctuating kinetic energy level across the Drake Passage and within the water column using the ISOS 1975 to 1978 current meter data [Nowlin, Pillsbury and Bottero, 1981] show north to south increase in the mean kinetic energy, at the 2500 meter levels from 3 to 15 cm^2/sec^2, and upward increase from 10 cm^2/sec^2 at 2700 meters in the central Drake Passage to 300 cm^2/sec^2 at 300 meters at the same site. The fluctuation kinetic energy is partitioned between high (2 hours to 2 days), intermediate (2 days to 50 days) and low (greater than 50 days) frequencies, with about half of the energy occurring at intermediate frequencies.

The fluctuations are nearly an order of magnitude more energetic in the northern Drake Passage relative to the central and southern Drake Passage. The magnitude of the fluctuation component relative to the mean kinetic energy is greatest both at abyssal depths and near the surface. These kinetic energy levels are much above values typical of the ocean interior and closer to those of the Gulf Stream. The authors caution that their results for the Drake Passage may not typify the entire ACC. This caution should be noted when any of the ISOS results are applied to the full circumpolar belt.

Fluctuations at tidal frequencies in the Drake Passage were studied by Nowlin, Bottero and Pillsbury [1982]. They show a predominantly semidiurnal tide in the north, and mixed type in the central and southern region. On average the tides account for 50% (highest in the southern Drake Passage) of the high frequency fluctuating kinetic energy, which contains 28% of the total fluctuating kinetic energy, [Nowlin, Pillsbury and Bottero, 1981], with counterclockwise rotating currents from inertial frequency to 5 cpd are significantly greater than clockwise rotation.

The study of Drake Passage thermohaline zonation and fronts continued with further refinement of definitions and descriptions [Whitworth, 1980; Nowlin and Clifford, 1982]. Seven Drake Passage zones and fronts are defined, from north to south: Subantarctic Zone; Subantarctic Front; Polar Frontal Zone; Polar Front; Antarctic Zone; Continental Water Boundary, and the Continental Zone. The frontal boundaries are characterized by large meridional thermohaline gradients with cores of strong currents, with widths scaled by the vertical stability (about twice the Rossby radius of deformation). The fronts account for 75% of the baroclinic transport but only 19% of the Drake Passage width. Zonation within the entire circumpolar belt seems probable, though some variation from the Drake Passage situation is expected [see Deacon, 1982, for a brief review of circumpolar zonation and its relationship to biology].

Patterson and Sievers [1980] discuss the unique thermohaline structure of the Weddell-Scotia Confluence which is found within the southern Scotia Sea. The Weddell-Scotia Confluence is a thermohaline zone formed at the confluence of the Weddell Gyre and ACC. It is marked by weak vertical stratification and the relatively cold fresh deep water more closely related to continental margin stratification and believed to be a product of more vigorous vertical processes, such as winter convection or mixing at the lateral boundaries of the Antarctic Peninsula and South Scotia Ridge.

B. Drake Passage Mesoscale Features and Poleward Heat Flux

The fronts and zones often display mesoscale features. Understanding of these features within the Drake Passage was also advanced during the 1979-1982 period with research by Joyce, Patterson and Millard [1981] and Peterson, Nowlin and Whitworth [1982]. Joyce et al. [1981; also see Patterson and Sievers, 1979/1980] discuss the formation of a 100 to 120 km diameter cyclonic cold core ring or eddy in 1976 within the Polar Frontal Zone. Peterson et al. [1982] discuss a similar eddy formation in 1979. Such eddies can extend to depths greater than 2500 meters [Sciremammano, 1979]. Though surface speed of above one knot occurs, the Drake Passage eddy is significantly less energetic than Gulf Stream rings. Fine scale intrusions associated with the rings may be as significant in transferring heat poleward across the Polar Front, though they are less effective in regard to poleward salinity flux.

The full impact of eddies to meridional heat and salt flux cannot be fully evaluated until the ultimate fate of the rings is determined. The cyclonic cold rings move with the ACC at 5 to 10 cm/sec. Eventually they must encounter the Scotia ridge, which has a sill depth less than the depth penetration of the eddies. It is here that further work may determine the fate of the rings. Peterson et al. [1982] believe the 1979 ring, formed at the Polar Front, crossed the Polar Front Zone and Subantarctic Zone, and it is unlikely that it rejoined its parent water mass. However, Joyce et al. [1981] believe that even if the Drake Passage rings were typical of the full circumpolar belt, and all rings fully decay, without returning to the parent water masses, it is still unlikely they (or fine scale intrusions) [Toole, 1981b] can fully account for the required meridional heat/salinity flux. Since mean circumpolar geostrophic motion also is inadequate as a meridional transfer mechanism of heat or salinity [deSzoeke and Levine, 1981], the large scale meridional thermohaline balance may be more closely associated with ACC baroclinic instability [Wright, 1981]. This process induces large poleward heat flux, 0.7 to 2.8 W/cm^2 in the Drake Passage in a more or less event oriented fashion dominated by 5 to 60 day or longer time scales [Bryden, 1979, using the 1975 ISOS current meter data; Sciremammano, 1980, using 1975-1977 ISOS data]. The rings can be considered to be a consequence of the baroclinic instability. While caution is suggested in the extrapolation of Drake Passage eddy heat flux values to the circumpolar belt, it is noted that such values are close to what is considered the total poleward heat flux requirement. Conditions for baroclinic instability are met everywhere in the Drake Passage, with barotropic instability conditions met adjacent to the three front zones [Peterson et al., 1982]. Lutjeharms and Baker [1980] discuss general distributions of mesoscale features in the ACC, which are most intense near the ACC axis. A useful summary of ACC rings in the Drake Passage and other sections of the circumpolar belt is given by Bryden [1982].

C. Circumpolar ACC Volume, Heat and Fresh Water Transport

Using a simple Sverdrup model an ACC poleward transport south of 50°S is 173 to 190 × 10^6 m^3/sec is determined [Baker, 1982]; presumably the northward flow in the southwest Atlantic is the primary compensating western boundary current. The Sverdrup transport is about 40 to 60 × 10^6 m^3/sec higher than the Drake Passage transport, but in view of the uncertainties one may conclude that Sverdrup balance may apply to the

ACC and such models deserve more detailed consideration.

The baroclinic and barotropic response of the ACC to broad spectrum wind forcing is studied by Clarke [1982] with a non-linear model. The response to wind variations of periods less than a few years is found to be barotropic and consistent with the results of Wearn and Baker [1980] in regard to correlation with circumpolar wind stress. Baroclinic fluctuations occur at longer periods, particularly for periods greater than 70 years. Clarke suggests that such variations may be related to long period sea surface temperature fluctuations discussed by Fletcher, Radok and Slutz [1982]. The barotropic mode is dissipated by interaction with bottom topography, while the baroclinic mode spins down by baroclinic instability. Clarke questions the validity of the application of Sverdrup dynamics to the ACC.

Georgi and Toole [1982] investigate the interocean heat and salinity flux accomplished by the ACC, which provides the primary conduit for interocean exchange. Using geostrophic currents relative to the sea floor (and to 2500 db) and the associated temperature-salinity distribution between each southern hemisphere continent and Antarctica, they find the ACC cools significantly (-33.5×10^{13} W) and freshens (0.103×10^6 m^3/sec of fresh water) in transit across the Atlantic sector, and cools (-31.7×10^{13} W) and slightly freshens (0.004×10^6 m^3/sec) across the Pacific sector, and gains a compensating amount of heat (64.8×10^{13} W) and reduction of fresh water (-0.107×10^6 m^3/sec) across the Indian sector of the Southern Ocean. While an error analysis yields relatively large uncertainty in these figures, they are much less than values based on global ocean summation of sea-air exchanges, particularly in regard to the interocean heat exchange. Using ACC transport relative to the bottom and 2500 db and Southern Ocean sea-air heat exchange from the literature, Georgi and Toole [1982] determine the heat and fresh water fluxes across 40°S. They find northward heat flux in the Atlantic as do other investigators, which is balanced primarily by southward heat flux in the Indian Ocean. The fresh water fluxes across 40°S are not conclusive.

D. Circulation South of the ACC

Baroclinicity is very weak within the nearly homogeneous water column characteristic of the region south of the ACC, and there are only a few direct current measurements, so we know surprisingly little about the mean circulation of this region [Gordon, 1980]. Much of what is known is derived from drift of ice-locked ships, icebergs and buoys.

The Weddell Gyre [Deacon, 1979] is modeled using the wind stress and simple Sverdrup dynamics [Gordon, Martinson and Taylor, 1981]. The wind-driven circulation is thought to be primarily barotropic with low characteristic velocity, though substantial transports. The western boundary current, adjacent to the Antarctic Peninsula has a characteristic velocity of only 5 to 10 cm/sec which is consistent with ship and buoy drift data [Ackley, 1981b], but with a total volume transport of 76×10^6 m^3/sec.

Direct current measurements in the Weddell Gyre are reported by Foster and Middleton [1979] and by Middleton, Foster and Foldvik [1982]. Foster and Middleton [1979] report on a year long current meter record 50 m above the seafloor at 66°29.3'S, 41°02.6'W, where the average velocity is 1.3 cm/sec directed towards 006°T. They discuss greater variability in summer and dominance of the planetary wave mode of the Weddell Basin. Middleton et al. [1982] discuss low frequency variability measured by a number of multi-year current meter records obtained along the shelf and slope of the southern Weddell Sea. They detect the existence of wind induced shelf waves with periods of 3 to 60 days, with anticlockwise rotation for periods less than 5 days, clockwise for periods longer than 8 days. Energy levels are higher over the continental slope, perhaps due to topographic trapping.

Coastal fronts and circulation of the Ross Sea is discussed by Jacobs, Gordon and Ardai [1979], and Jacobs, Gordon and Amos [1979] with emphasis on ocean-glacial ice interaction (Section VI). The Ross Sea shelf-slope front and sea ice distribution relative to sea-bird population is discussed by Ainley and Jacobs [1981].

The likely dominance of barotropic circulation south of the ACC means that geostrophic flow relative to some zero reference level does a poor job of defining the circulation. A greater effort is required to obtain direct measurement of currents from moorings or by drifters before the circulation of such major features of the Southern Ocean regime as the Weddell Gyre, Ross Gyre, coastal currents, Antarctic Divergence can be defined. Hopefully the FGGE drifter data set will prove useful in defining Southern Ocean circulation within the ACC and poleward.

E. Water Masses

Water mass studies were directed towards questions of Antarctic Bottom Water (AABW) and Antarctic Intermediate Water (AAIW) formation and spreading as well as the circumpolar attenuation of Circumpolar Deep Water (CDW) characteristics.

Warren [1981] provides a very useful review of AABW production and spreading. Much of the AABW formation research is associated with the waters of the Weddell Gyre [Deacon, 1979] which is composed of the coldest-freshest water of the deep ocean regime [Gordon and Baker, 1982]. Continental margin produced AABW forms in the southwest region of the Weddell Sea [Foster and Middleton, 1979; 1980]. They find variability in bottom water characteristics between 1975 to 1976 in the western part of the Weddell Basin, and suggest significant changes in the mixing ratios of the component water types producing AABW, perhaps with intermittent formation. They further show evidence that some of the convective products do not penetrate to the sea floor, but rather spread laterally on isopycnal surfaces just above the seafloor.

Weiss, Östlund and Craig [1979] use the geochemical parameters of tritium and stable oxygen isotopes to decipher the AABW recipe, and provide rates for the general ventilation of the Weddell Sea. Edmond, Jacobs, Gordon, Mantyla and Weiss et al. [1979] study the build up of silicate concentrations by dissolution of silica from bottom sediments. These studies suggest that the oxygen and nutrient distribution below the immediate surface layer is a consequence of conservative mixing and that sea ice forms an effective inhibitor of ocean-atmosphere gas exchange. Chen [1982] investigates the apparent ineffectiveness of the Southern Ocean in the transfer of anthropogenic CO_2 to the abyssal ocean.

Water mass modification by deep reaching convection within the central part of the Weddell Gyre is studied by Killworth [1979] and Gordon [1982]. This modification is believed to be associated with the occurrence of a 200 to 300 $\times 10^3$ km^2 ice free region now called the Weddell Polynya observed during the austral winters of 1974 to 1976 near 66°S and the Greenwich Meridian by micro-wave sensors aboard a satellite [Carsey, 1980]. No Weddell Polynya has been observed since 1976.

Based on summer hydrographic observations it is highly probable that the heat required to maintain the winter Weddell Polynya is derived from the warm Weddell Deep Water by convective exchange with cold surface water [Martinson, Killworth and Gordon, 1981]. The observations show very significant cooling and freshening of the Weddell Deep Water between the pre-polynya year of 1973 and the post-polynya year of 1977, which represents convective sinking of freezing point surface water of up to 7.7×10^6 m^3/sec during the three winters [Gordon, 1982].

The initiation and termination of the polynya is not understood, though Gordon [1981, 1982] and Martinson et al. [1981] suggest that increased influx of warmer-saltier deep water into the Weddell Gyre with subsequent Ekman induced upwelling (without additional buoyancy compensation by fresh water introduction to the Weddell region) would weaken the pycnocline and encourage more vigorous vertical exchange. Hydrographic data prior to 1970's reveal only one earlier polynya occurrence (in 1961), though the data set is far from complete [Gordon, 1982].

Spreading of AABW to positions north of the Antarctic margins and across the Polar Front Zone is studied using distribution of near bottom temperature, salinity and oxygen in the Australian sector of the Southern Ocean by Rodman and Gordon [1982]. They deduce a clockwise flow of bottom water in the basin immediately north of Antarctica (South Indian Basin) fed by saline AABW produced in the Ross Sea and lower salinity AABW from the Adelie Coast. Bottom water moves northward from this basin within the Antarctic Discordance, a saddle in the mid-ocean ridge near 120°E, and within the Balleny Fracture Zone. Georgi [1981] studies bottom water distribution in the southwest South Atlantic using temperature distribution at 4000 m and the topography of the 0°C isothermal surface. He finds a number of circulation gyres in the Argentine Basin with the most dominant carrying Weddell-produced AABW (entering the Argentine Basin via the East Falkland Channel) northward with the western boundary current, at a rate of 2×10^6 m^3/sec.

Northward spreading of AABW from the Argentine Basin via the Vema Channel is studied by Hogg, Biscaye, Gardner and Schmitz [1982]. Using hydrographic stations and direct current measurements they find a mean northward AABW flux of slightly above 4×10^6 m^3/sec. North of the Vema Channel in the Brazil Basin AABW must diffuse upward crossing isopycnals with a mixing coefficient of 3 to 4 cm^2/sec. Whitehead and Worthington [1982] trace AABW movements near 4°N, finding a flux of 1×10^6 m^3/sec, and also determine a mixing coefficient of about 3 cm^2/sec, as required to maintain thermal balance in the abyssal waters consistent with their flux values.

The circumpolar variations of CDW were effectively discussed by Georgi [1981b], using volumetric analyses of temperature, salinity, oxygen and silicate as a function of density. The North Atlantic Deep Water (NADW) characteristics are rapidly mixed laterally into the Pacific Ocean CDW entering the Argentine Basin with the ACC. This mixing is marked by intense fine structure and a lateral mixing coefficient of 4×10^5 cm^2/sec. The CDW enroute around Antarctica slowly loses NADW characteristics by lateral mixing with Indian and Pacific deep water which is colder, fresher, lower in dissolved oxygen and higher in silicate.

Production and spreading of AAIW was addressed by a number of investigators [Georgi, 1979; Molinelli, 1981; and Piola and Georgi, 1981, 1982]. Molinelli [1981], using relative geostrophic flow within the ACC between isopycnals 27.2 to 27.3 for the full circumpolar belt, finds a relatively strong (3×10^6 m^3/sec) transport of relatively cold-fresh water into this layer within the southeast Pacific. This water is derived from Polar Front and Antarctic Zones, thus involving cross front transfer of water types, casting doubt on a purely Subantarctic Zone origin for the AAIW (as the coldest end-member of Subantarctic Mode Water, SAMW) [McCartney, 1982], though such a mechanism can account for slightly less dense water type ventilating the ocean immediately above AAIW levels, and perhaps the primarily intermediate water of the southeastern South Pacific [Georgi, 1979]. Piola and Georgi [1982] discuss AAIW and SAMW spreading in the 40°-45°S zone. They find evolution of the SAMW begins in the Indian Ocean, ending in the southeast Pacific, with the Atlantic sector receiving AAIW, 'spiked' with water derived from the vicinity of the polar front. Apparently sea-air heat exchange in the Drake Passage is insufficient to alter the southeast Pacific SAMW into AAIW characteristics [Piola and Georgi, 1981]. Piola and Georgi [1982] do not believe Atlantic AAIW freely spreads into the Indian Ocean, as it seems 'blocked' by the Agulhas Current System, though lateral mixing may be intense.

F. Wind and Waves

Satellite altimeter observations from SEASAT allow for a unique opportunity to provide synoptic ocean wind and wave maps for the Southern Ocean for the austral winter of 1978 [Mognard, Campbell, Cheney and Marek, 1982]. They include mean monthly maps of wind speed, significant wave height, and swell, showing high wind and wave fields in the Indian Ocean sector relative to the Pacific in August and September 1978, though the Atlantic sector has extensive high wind regions in July. Maximum wind and wind gradients in the Indian Ocean are also known from ground based data [Baker, 1982].

IV. Arctic Circulation and Water Masses

A. Circulation

Much of Arctic physical oceanographic research was directed toward the renewal of the Arctic halocline by continental shelf water mass formation [Aagaard, Coachman and Carmack, 1981; Melling and Lewis, 1982] and the distribution and mixing of ice melt water [Hanzlich and Aagaard, 1980; Tan and Strain, 1980; Paquette and Bourke, 1981]. Subarctic oceanography of the Bering Sea is summarized in a series of articles by Kinder [1981], Kinder and Schumacher [1981a, b]; for the Iceland-Greenland Seas by Swift and Aagard [1981]; and for the Labrador Sea by Lazier [1982].

A much needed discussion of Arctic deep water is provided by Aagaard [1981]. He makes a distinction between upper and lower deep water, separated at 1500 m. The upper layer is derived from the Greenland or Norwegian Sea while the lower layer's higher salinity is produced within the Arctic by an as yet unidentified process. The deep water circulation is cyclonic, counter to the surface layer circulation. Aagaard [1981] also presents one month long near bottom current meter records from two moorings on the flank of the Lomonosov Ridge. The currents are typically of order 1 cm/sec, with 2 to 4 cm/sec episodes, at 5 to 10 day intervals. The tide is semidiurnal with 2 to 3 cm/sec amplitude. Aagaard suggests the existence of bottom trapped oscillations.

Östlund [1982], using tritium concentration deduces a residence time of 11 years for the fresh water (mostly river runoff) input to the Arctic Ocean. He finds the fresh water and Atlantic waters mix nearly linearly to produce the stratum just below the surface water influenced seasonally by the ice cover [Morison and Smith, 1981] but above the pure Atlantic water. This layer ranges from 10-60 m in the Nansen Basin to 150-170 m west of Lomonosov Ridge and in the Canadian Basin. Hanzlick and Aagaard [1980] determine a residence time for river water in the Kara Sea of 2.5 years. The Östlund residence time is for the entire Arctic Ocean and represents the time between first introduction of river water at the shelf and its emergence from the Arctic in the East Greenland Current. The residence time in the Canadian Basin, most remote from the Arctic primary inflow-outflow sites is as large as 17 years. Östlund detects relatively high tritium down to 1200 m west of the Lomonosov Ridge, presumably carried into the Arctic with the Atlantic water.

B. Water Masses

The feasibility of Arctic halocline renewal by large scale seaward spreading of shelf water is a particularly exciting advance in Arctic water masses. Aagaard, Coachman and Carmack [1981] believe winter freezing conditions in coastal areas characterized by sea ice divergence, produce salty freezing point shelf water which then spreads isopycnally into the Arctic halocline at a rate of 2.5×10^6 m^3/sec, which is similar to the influx of the warm-saline Atlantic water. The cold shelf water mixes with warmer Atlantic water, thus shielding the Arctic surface water from upward heat flux [also see McPhee, 1980]. Melling and Lewis [1982] further the discussion with (early) winter period observations in the Beaufort Sea and with a convective tube model. They point out that thermohaline alteration of surface coastal water by winter freezing, even in ice divergence regions, may be insufficient to produce the required salinities for convection into the halocline. They suggest the probable importance of a more direct alteration of relative warm-saline water which might upwell onto the shelf in early autumn, accompanying an acceleration of Beaufort Sea circulation. Upwelling of warm-saline water within Barrow Canyon is reported by Garrison and Paquette [1982].

These studies clearly show that Arctic continental margins are as important in understanding Arctic stratification as are the Antarctic margins to understanding Southern Ocean stratification.

Swift and Aagaard [1981] discuss the transition zone from Arctic to Atlantic water in the Greenland and Iceland Seas and the formation of Arctic Intermediate Water (AIW) in that region. This water mass is the primary contribution to Denmark Strait overflow into the Atlantic Ocean [Swift, Aagaard and Malmberg, 1980]. The Greenland Sea AIW is warmer and saltier than the Iceland Sea produced AIW.

V. Ocean-Ice Interaction

The most unique feature of polar oceans is the sea ice cover, for at least part of the year, and also the occurrence of direct glacial ice-ocean contact at ice shelves and icebergs. Research advances in this area are obviously hindered by logistical difficulties. But, with support by satellite observations [Zwally et al., 1979], moorings [Middleton, Foster and Foldvik, 1982], ice island bases [Morison and Smith, 1981; Manley et al., 1982], ships [Gordon and Sarukhanyan, 1982] and drilling of access holes [Clough and Hansen, 1979], the last four years did see significant achievements in understanding ocean-ice-atmosphere interaction.

A surprising amount of work was directed towards a much neglected area of ocean-glacial ice interaction. Laboratory results [Huppert and Josberger, 1980; Huppert and Turner, 1980] and field data [Neshyba and Josberger, 1980; Josberger and Neshyba, 1981;Jacobs, Huppert, Holdsworth and Drewry, 1981] are used for investigation of fine scale thermohaline stratification features produced by melting glacial ice in a stratified ocean. Melting of the deep draft icebergs in the Southern Ocean involves direct use of the deep water (sub-pycnocline) heat which may be significant in regard to climatic considerations and water mass modification [Jacobs, Gordon and Amos, 1981].

Interaction of the ocean ice shelves and glaciers was studied theoretically by Griesman [1979, 1980] and Gade [1979]. Jacobs, Gordon and Ardai [1979] using hydrographic data obtained along the Ross Ice Shelf Barrier and sub-ice shelf ocean data obtained through the RISP (Ross Ice Shelf Project access hole) [Clough and Hansen, 1979] investigate the role of ocean heat in basal melting of the Ross Ice Shelf, which may be a major factor in the glacial mass balance. They suggest a mean Ross Ice Shelf basal melting rate of 25 cm/year, identifying the primary heat source as the relatively warm water type observed at the ice shelf Barrier, at 200 meters near 175°W. Geochemical studies [Weiss, Östlund and Craig, 1979] are particularly useful in detecting the glacial melt water component within sea water.

Sea ice studies of direct relevance to oceanography [see Carmack, 1982] involve the Weddell Polynya phenomena as discussed in Section IV; vertical heat and salinity flux in ice covered seas; and ice edge processes.

The relatively weak Southern Ocean pycnocline [Middleton and Foster, 1980] and general Ekman induced upwelling within the seasonal sea ice zone of 60°S-70°S allows for substantial (31 W/m^2) annual upward heat flux [Gordon, 1981]. This is about ten times the Arctic value, where a much stronger pycnocline exists. The Ekman effect is convergent and the water below the haline supported pycnocline is near the freezing mark [McPhee, 1980] greatly limiting vertical heat flux. Toole [1981a] presents a sea ice model for the Southern Ocean which successfully predicts the thermohaline characteristics of the surface water and sea ice distribution. Though the required vertical heat flux is less than given by Gordon [1981], Toole [1981a] points to the uncertainties in the southern region (Antarctic divergence) where further field work is required to more fully assess vertical transfer processes.

Gordon [1981] considers the Southern Ocean pycnocline to be only marginally stable due to the nearly compensating buoyancy effects of ocean to atmosphere heat flux and fresh water introduction. The poleward heat flux across 60°S is 54 to 65 $\times 10^{13}$W [Gordon, 1981; Hastenrath, 1982]. While this heat apparently is transferred across the ACC near 50°S by eddy processes [Bryden, 1982] the continued transfer southward across 60°S and ultimately vertically to the sea surface or ice boundaries may be accomplished by other mechanisms.

Ocean to atmosphere heat flux in ice covered seas has gained much attention, [McPhee, 1980, 1981; Allison, 1981; Ackley, 1981a; Stigebrant, 1981; Maykut, 1981; Carmack, 1982; McPhee and Untersteiner, 1982]. Further understanding the sea ice dynamics [Hibler, 1979; McPhee, 1979] and attempts to couple atmospheric and ocean general circulation in an ice covered ocean [Washington, Semtner Meehl, Knight and Mayer, 1980] has yielded improved sea ice modelling results. The Weddell Gyre sea ice cover seasonal variability is effectively modelled by a dynamic and thermal model by Hibler and Ackley [1982].

There is increased attention to the mesoscale, wave and boundary layer oceanographic conditions at the sea ice edge, as these would influence advance or retreat of the ice edge [Alexander and Niebauer, 1981; Bauer and Martin, 1980]. Eddies, fine structure and Ekman induced upwelling/sinking events have been observed [Paquette and Bourke, 1979; Buckley et al., 1979; Røed and O'Brien, 1981; Wadhams, Gill and Linden, 1979].

In the Southern Ocean winter period observations are confined to near coastal stations [Allison, 1981] and from ships which were unfortunately ice bound. In order to understand the full extent of water mass modification and sea-air exchanges in the Southern Ocean, observations within the seasonal sea ice in winter are clearly needed. A start in this direction was taken in October-November 1981 when a joint US-USSR expedition [Gordon and Sarukhanyan, 1982] penetrated 300 nm into the end of winter sea ice cover along the Greenwich Meridian obtaining a comprehensive physical and biological data set.

VI. Conclusions

Progress in further defining the currents and water mass renewal in polar oceans in the 1979-1982 period, is impressive. Continuation of the research in the Southern Ocean circulation should focus on the ACC variability at sites other than the Drake Passage, and on the circulation patterns to the south. We must now determine how representative is the Drake Passage of the ACC dynamics, zonation and mesoscale features of other Southern Ocean segments. Further definition of the rates and processes accomplishing poleward heat and salinity flux across the ACC, and the Antarctic Zone, and onto the Antarctic continental shelf is required. The role of the Weddell Gyre and other gyres south of the ACC in meridional heat/salinity is a particularly relevant research topic.

Arctic circulation below the surface layer is very poorly known, as is the basic stratification and renewal of Arctic deep water. Observations within the deep Arctic basins are required.

Concepts of water mass formation and spreading in both the Arctic and Southern Ocean have now been developed. Specific field projects are required to evaluate these concepts and formulate more quantitative understanding of the formation rates and extent of ocean ventilation. Study of stable and unstable isotopes seems a fruitful approach for these studies.

To fully understand the role of polar oceans to global climate research, the extent of ocean-glacial ice-sea ice-atmosphere interaction should receive more emphasis.

As logistical requirements for polar ocean research are difficult to meet, advance planning and efficient use of science support are particularly important.

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POLEWARD HEAT TRANSPORT BY THE OCEAN

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Introduction

Poleward heat transport and its relation to the thermohaline circulation of the World Ocean have long been neglected subjects, but a perceived need for a better understanding of the Earth's climate have given an impetus to studies of the ocean's role in the global heat balance. Oceanographers have a new appreciation of the existing hydrographic and geochemical data base, and ambitious plans are being made to measure new sections in all the major oceans. It is best to consider ocean heat transport in the context of fresh water, oxygen, silica and other geochemical transports. From the standpoint of climate, however, heat transport is of primary interest. To keep the scope of this review within manageable limits, attention will be restricted to this one aspect of a broader ocean transport problem. Research over the last four years has brought about some important changes in outlook. We will return to this point in the summary.

This review draws heavily on several summaries that have recently appeared (Bryden, 1982; Bryan, 1982a; Hastenrath, 1982; "CAGE Experiment: A Feasibility Study," Dobson et al. 1982). As these summaries indicate, poleward heat transport by ocean currents remains one of the most poorly understood elements of the global heat balance. Estimates based on several different principles all indicate that the poleward heat transport in the ocean is of comparable magnitude to poleward heat transport in the atmosphere, but the processes appear to be quite different. An important indicator of different mechanisms is the observation that the atmospheric heat transport is greatest where the atmospheric poleward temperature gradient is greatest, while the ocean transport appears to be a maximum at low latitudes where the north-south temperature gradients in the ocean are rather weak.

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Heat Balance Methods

The requirement for poleward heat transport by the Earth's fluid envelope arises from the fact that the greater part of direct solar insolation is received in low latitudes, while back radiation to space is more evenly distributed over the globe. A steady state balance can only be maintained by a poleward energy transport. There are two heat balance methods