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Sea-Level Response to Ice Sheet Evolution: An Ocean Perspective

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Abstract

The ocean's influence upon and response to Antarctic ice sheet changes is considered in relation to sea level rise over recent and future decades. Assuming present-day ice fronts are in approximate equilibrium, a preliminary budget for the ice sheet is estimated from accumulation vs. iceberg calving and the basal melting that occurs beneath floating ice shelves. Iceberg calving is derived from the volume of large bergs identified and tracked by the Navy/NOAA Joint Ice Center and from shipboard observations. Basal melting exceeds 600 $km^3 yr^{-1}$ and is concentrated near the ice fronts and ice shelf grounding lines. An apparent negative mass balance for the Antarctic ice sheet may result from an anomalous calving rate during the past decade, but there are large uncertainties associated with all components of the ice budget. The results from general circulation models are noted in the context of projected precipitation increases and ocean temperature changes on and near the continent. An ocean research program that could help refine budget estimates is consistent with goals of the West Antarctic Ice Sheet Initiative.

Introduction

The ocean both forces and responds to ice sheet evolution. It provides the evaporative moisture that nourishes the ice sheet and regulates the sources of that moisture by its waxing and waning fields of sea ice. The ocean alters the position of ice sheet grounding lines and the dimensions of ice shelves by changes in sea level, melting, freezing and mechanical attrition. It exports calved icebergs in the strong coastal currents and polar gyres, and accepts surface runoff and subsurface glacial meltwater in its top and bottom layers. It damps the potential rate of climate change by its large heat capacity, deep vertical mixing in the polar regions, heat transfers associated with melting and freezing, and by its uptake of greenhouse gases. The ocean may also have discrete stable modes of circulation that promote, retard or respond in different ways to ice sheet growth. More than 90% of the earth's ice sheet lies on the Antarctic continent.

From paleoclimate records, it is known that global sea level has risen at an average rate of ~6mm yr^{-1} over the past 20,000 years, including rapid increases that exceeded 24 mm yr^{-1} (Fairbanks 1989). A slower rise over the past century has been attributed in part to an ocean volume increase resulting from a parallel rise in air temperature, and to the retreat of temperate glaciers (Gornitz et al. 1982; Meier 1984). The remainder is unaccounted for or ascribed to changes in surface and groundwater reservoirs and to uncertainties in mass balance of the large ice sheets (Meier 1990a). This balance is unknown, but the Antarctic portion is generally considered to be slightly positive or close to a state of equilibrium (Budd and Smith 1985; Bentley 1989), while Greenland's contribution to sea level could be of either sign (Reeh 1985; Zwally 1989; Braithwaite and Olesen 1990). The slow response time of glacial ice complicates the problem of determinining ice sheet mass balance over the short term, as does the lack of good-quality data, the low signal to noise ratio, and the inherent difficulty of making field measurements in the polar regions.

The average rate of global sea level rise over the past century has been 1-2 mm yr^{-1}, with higher values (2.3 - 2.4 mm yr^{-1}) over the past 50 years supported by glacial isostatic adjustments (Peltier and Tushingham 1989; Barnett 1990). Taking 1.5 mm yr^{-1} as a working number, this addition to the ~360 x 10^6 km^2 ocean surface would equal 540 km^3 yr^{-1} of water, or ~635 km^3 yr^{-1} of ice. For comparison, that volume is roughly one quarter of the present-day annual accumulation on the Antarctic ice sheet, or the approximate annual flow of the Mississippi River. Taking the ice-ocean boundary as a reference interface, this background paper will focus upon the attrition side of Antarctica's ice budget, primarily calving and basal mass balance. Freezing and melting beneath ice shelves will be emphasized because these processes are often discounted, are involved in deep ocean ventilation and may be sensitive to short-term climate change. Sub-ice shelf and general circulation models will be discussed in the context of basal fluxes and possible future ocean changes near Antarctica. Finally, an outline will be given for the oceanographic components of a potential multidisciplinary program to study the relationship of the Antarctic ice sheet to global climate change.

Iceberg Calving

The calving of icebergs from Antarctica is the largest factor in the attrition of the ice sheet, defined here as extending to the ice front perimeter. Calving from floating ice shelves has no immediate impact upon sea level, but reduced back-pressure on the grounded ice should permit faster seaward flow of ice streams across the grounding lines (Hughes 1973). From relatively small changes in the average positions of ice fronts over the past several decades (Markov et al. 1968; Barkov 1985; Zakarov 1988), it can be argued that a rough balance exists between ice sheet accumulation and wastage. However, outflow may not respond rapidly to changes in accumulation, and is likely to be balanced over time by adjustments in calving rate. Similarly, an ice shelf might thicken or thin for some time before major perturbations appeared in its equilibrium calving line. A large part of the attrition may occur by major calving events at periods of several decades, but early ice front positions were often based upon navigational records of low precision, and little of the Antarctic perimeter or its ice velocity has been systematically monitored. Satellite techniques for mapping ice fronts have shown considerable promise (Thomas et al. 1983; Zwally et al. 1987)s and new sensors expected to be in polar orbit late in this decade should allow routine monitoring to begin. Achieving significant results over a short period of time will be difficult, but good spatial and temporal coverage of the ice fronts, ice thickness and ice velocity should aid in the development of ice dynamics and calving models.

Attempts to assess the attrition of Antarctica by iceberg volumetrics from ships of opportunity have not generally been held in the highest regard. The uncertainties of estimating iceberg sizes, ocean area surveyed and typical iceberg lifetimes are compounded by the difficulties of accounting for duplicate sightings, seasonal and areal variability, and wastage between the time of calving and observation. The report "Glaciers, Ice Sheets and Sea-Level" (NRC/DOE 1985) discounted its only relevant contribution on this topic, wherein Orheim (1985) suggested a negative Antarctic ice sheet mass balance based upon a 1977-1984 calving estimate of 2.3 x 10^{15} kg (~2700 $km^3 yr^{-1}$). That study covered a period when the major ice shelf fronts were known to be advancing (Lange & Kohnen 1985; Jacobs et al. 1986) and thus could not have contributed much to the calculated volume. Although the Amery, Filchner-Ronne and Ross Ice Shelves drain more than half of the grounded ice sheet, much of that is the low-precipitation continental interior, so these drainage systems account for only one third of the outflow (Giovinetto and Bentley 1985). Given the high sensitivity of the derived production rate to rough estimates of iceberg lifetimes, a remarkable feature of the Orheim (1985) result and earlier estimates in the same range may be their close approximation to ice sheet accumulation values. Nonetheless, it is worth re-examining the iceberg data in the light of more recent observations.

Much useful information about iceberg distributions, sizes and freeboards can be derived from comprehensive surveys. By extrapolation from a smaller count, the annual number of bergs in the Southern Ocean can apparently exceed 300,000, including all bergy bits >10 m (Orheim 1985). That implies one chunk of ice for each 120 km^2 of the 36 x 10^6 km^2 ocean surface south of the Polar Front, consistent with fig. 8 in Weeks and Mellor (1978). Wadhams (1988) calculated one iceberg for each 164 km^2 of ocean surface for a short observational period in a restricted area of the winter South Atlantic, where survival time was estimated to be < 6 months and it was difficult to detect bergs smaller than 115 m in diameter. Annual production rates inferred from iceberg distributions are inversely proportional to their lifetimes, which Orheim (1985) took to be 4 years. Survival times are longer for icebergs near Antarctica, berg dimensions are larger within the pack ice and numbers of bergs decrease away from the coastline (Orheim 1985; Hult and Ostrander 1973; Morgan and Budd 1978). Grounding is common on bottom shoals of the continental shelf, while larger waves, higher temperatures, stronger currents and less sea ice promote more rapid mechanical and thermal disintegration over the deep ocean. These data suggest that volumetric estimates might be refined by distinguishing between iceberg populations in different areas or size categories.

Orheim (1989) indicated that for the previous several years the annual calving rate of small icebergs (main axis <22 km) was approximately constant both in numbers and total mass, and exceeded the annual mean mass of large icebergs (>22 km) calved during the same period. This covered a period (1986-87) of major iceberg calving in the Weddell and Ross Seas, but we know that large calving events also spawn numerous smaller bergs and that different size classes experience different fates (Keys et al. 1990). For more than a decade, large icebergs (> 28 km) have been identified and tracked from satellite images by the Navy-NOAA Joint Ice Center (e.g., NPOC 1987-88). The annual volume of newly-calved icebergs has been tabulated from their 1979-1989 data in Fig. 1, with the area of each iceberg conservatively approximated by an ellipse. Exceptions are the 1986 Filchner and 1987 Ross icebergs, where we have adopted the Landsat-derived areas of Ferrigno and Gould (1987) and Keys et al. (1990). Average thickness was taken to be 250 m, close to that estimated for the Ross (B-9) iceberg, which was probably thicker than the Larsen calf and thinner than the Filchner ones. From 1982-88, the average large berg volume was ~1520 km^3 yr^{-1}, but for the full 11-year period in Fig. 1 the average was 1279 km^3 yr^{-1}. Depending upon their size limits (<1 km or <28 km) we could get from 839 to 1279 km^3 yr^{-1} for small bergs. Approximating a small berg value from the average of this range (1059 km^3 yr^{-1}) and adding the 1979-89 large berg average, then total calving equals 2338 $km^3\ yr^{-1}$. This is again close to the accumulation value, but we still have to consider basal melting.

Basal Mass Balance

Basal melting of the ice shelves is sometimes ignored in mass balance estimates (e.g., Budd and Smith 1985), for a variety of reasons. It is hidden beneath the ice or sea surface, occurs at temperatures below 0^oC, is believed to be negligible, contrary to or balanced by observed basal freezing, or it is seaward of the grounding line and can thus have no impact on sea level. It is true that melting or freezing at the base of floating ice does not appreciably alter sea level, but most compilations of precipitation on Antarctica include the ice shelves. The total accumulation of Giovinetto and Bentley (1985) drops from 1962.7 x 10^{12} kg yr^{-1} (~2310 $km^3\ yr^{-1}$ at an ice density of 0.85) to ~1730 $km^3\ yr^{-1}$ when ice shelves, ice rises and ice islands are excluded from the accounting. With sufficient satellite data it may be possible at some time in the future to define all grounding lines and monitor the ice flow across them. In the interim it is more practical to include the ice shelves, and more important to investigate their influence upon mass flux off the continent.

Ocean data and models generally point to net basal melting beneath the ice shelves (e.g., Jacobs et al. 1979; Robin et al. 1983; Potter et al. 1988; Hellmer and Olbers 1989). These results might appear contradicted by measured or inferred sea ice accumulations on the ice shelf bases (e.g., Morgan 1972; Neal 1979; Zotikov et al. 1980; Engelhardt and Determann 1987). However, most ice cores and holes have been centrally located on the large ice shelves where most models also show net freezing. Thick sea ice deposits on the large ice shelf bases may be accounted for by an "ice-pump" (Lewis and Perkin 1986), whereby ice is removed from some regions and redeposited in others. This process has important implications for water column properties, basal mass balance and the stability of ice shelf pinning points. In the discussion that follows, basal mass balance will include both melting and freezing.

For several decades, investigators have reasoned from measurements near the major ice shelves that melting in the sub-ice cavities could be on the order of tens of cm yr^{-1} (Wexler 1960; Crary 1961; Thomas and Coslett 1970; Robin 1979). This inference was predicated upon an active circulation beneath the ice shelves, coupled with an oceanic heat source on or north of the continental shelf or a temperature differential afforded by a decrease of the *in situ* freezing point with pressure (Doake 1976; Jacobs et al. 1979). Numerical models subsequently showed that the energy for vertical mixing under the ice, essential to bring heat into the near-freezing boundary layer, could be derived from the tidal and thermohaline circulations (MacAyeal 1984a; 1985). Gravity and direct current measurements documented the active tidal circulations, and closely-spaced hydrographic observations revealed the limited dimensions of inflowing and outflowing shelf waters.

Long-term instrument arrays near the Ross Ice Shelf front in the mid-1980's focused upon a subsurface region of relatively 'warm' water, ~1^{o}C above the in situ freezing point in the austral summer, apparently derived from the continental slope region more than 250 km to the north (fig. 6b in Jacobs et al. 1985). It was hypothesized that this water could supply sufficient heat to melt ~40 km^3 yr^{-1} off the ice shelf base for each cm s^{-1} of net southward flow. Current and temperature records confirmed a persistent southward flow of this water during 1983, averaging ~5 cm s^{-1}. In spite of a seasonal temperature signal, some of the warmest temperatures appeared during the austral winter (Pillsbury and Jacobs 1985). However, a more closely-spaced array during 1984 showed that a major fraction of this ocean heat was recirculated to the open sea a few tens of km to the west of the inflow (Fig. 2). It is not yet known whether similar circulation patterns exist where warm water is drawn toward other ice shelves and glacier tongues, as illustrated by Foldvik et al. (1985) and Jacobs (1989).

The concept of an active circulation and net melting beneath the ice shelves has also been supported by geochemical measurements, from which it is possible to identify the distinctive properties of glacial meltwater, or to calculate the residence time of seawater beneath the ice (Michel et al. 1979; Jacobs et al. 1985; Potter and Paren 1985; Schlosser 1986; Fahrbach et al. 1991; Trumbore et al. 1991). Utilizing a shelf water chlorofluorocarbon (CFC) model, the latter study indicated that the time for High Salinity Shelf Water to evolve into Ice Shelf Water (ISW, with $T<T_{frs}$, i.e., temperatures below the surface freezing point) beneath the Ross Ice Shelf could be as little as 3.5 years. Ocean water properties have also been used to establish the presence of glacial meltwater in Antarctic Bottom Water (Weiss et al. 1979; Foldvik and Gammelsrod 1988; Schlosser et al. 1990). This finding has been extended by some investigators, who believe that ISW is primarily responsible for the coldest bottom water currently observed in the Weddell Sea (Foldvik et al. 1985; Jacobs 1986).

Recent models of the sub-ice shelf circulation and ice shelf mass balance have displayed a number of common features. These include high melting near the grounding lines, low melting or sea ice accumulation over wide areas of the ice shelf base, and high melting near the ice fronts. Integrating a Filchner Ice Shelf transect in the thermohaline circulation model of Hellmer and Olbers (1989) from the grounding line to the northern edge of the accumulation zone (Fig. 3) results in an average melting of ~45 cm yr^{-1}. Similar treatment of a Ronne Ice Shelf transect from the Rutford Ice Stream (fig. 10 in Jenkins and Doake 1991) gives ~65 cm yr^{-1} melting for the region more than 100 km south of the ice front, including a 200 km section where net accumulation prevails. Due to the pressure dependency of *in situ* freezing temperature, the melting rate is directly proportional to depth, leading to > 4 m yr^{-1} melting in the vicinity of the relatively deep Rutford transect grounding line (Jenkins and Doake 1991; Pozdeyev & Kurinen 1987). However, the Ronne Ice Shelf includes a more extensive area of basal freezing east of the Rutford transect (Engelhardt and Determann 1987). For these reasons, 45 cm yr^{-1} will be used as a working value for the entire Filchner-Ronne Ice Shelf region > 100 km south of the ice front, yielding 181 km^3 yr^{-1}. This rate is lower that an estimate of 211-246 km^3 yr^{-1} that can be derived from hydrochemical data in Schlosser et al. (1990), assuming 6-7$^o/_{oo}$ meltwater in an ISW outflow of 10^6 m^3 s^{-1} (Foldvik et al. 1985). That higher range may indicate other sources of meltwater, some perhaps in the inflow (Weiss et al. 1979), or that the model-derived estimate is overly conservative.

Glaciological models for the region well south of the Ross Ice Shelf front have shown equilibrium net melting on the order of 12-17 cm yr^{-1} (Shabtaie and Bentley 1987; Lingle et al. 1990). Oceanographic data and models have suggested similar to higher rates (Jacobs et al. 1979; Scheduikat and Olbers 1990), some of which appear to be at odds with a sea ice core retrieved from the central ice shelf (Zotikov et al. 1980). However, the active ocean circulation beneath an ice shelf means that the products of freezing and melting are dispersed far from their origins. The primary large-scale outflows appear as Ice Shelf Water temperature minima, observed, e.g., in vertical ocean property profiles near the front of the Filchner' Ice Shelf. These features are reproduced by the Hellmer and Olbers (1989) model, from which their depth is shown to be dependent upon the thermohaline characteristics of inflowing High Salinity Shelf Water. Perhaps for this reason and the accompanying changes in circulation strength and melt rate, temperature minima within the Ross Sea ISW outflow tend to cluster at discrete depth intervals, as illustrated by the shaded bands in Fig. 4. Interannual variability in the production and volume of High Salinity Shelf Water or multiple circulation cells beneath the ice shelf could also influence the outflow characteristics.

The salinity difference between High Salinity Shelf Water inflow and the ISW outflow can be combined with current measurements to estimate a basal melt rate for the Ross Ice Shelf. Apparent melting exceeds the heat available, judging by the observed temperature change between High Salinity Shelf Water and ISW. This discrepancy may be accounted for by incorporation of outflow from beneath the grounded ice sheet, by ice crystal formation in the near-freezing water column or by tidal mixing near the ice front (Foldvik and Kvinge 1974; MacAyeal 1984a; Dieckmann et al. 1986). An estimate for salinity at the time of inflow can be obtained from a 1982 profile in fig. 4.1 of Jacobs (1985). Integrating over the region where $T<T_{frs}$ in Fig. 4 and using the long-term average velocity (1.9 cm s^{-1}) of 4 current meters sited as shown, the meltwater in this plume is then ~79 km^3 yr^{-1}. Removing an equivalent volume of ice from the area of the ice shelf base >100 km south of the ice front would correspond to a melt rate of ~20 cm yr^{-1}, or ~88 km^3 yr^{-1}. That is roughly one half the Filchner-Ronne melt rate and may be consistent with less bottom water production in the Ross Sea, which St. Pierre (1989) has attributed to differences between the tidal regimes. It does not include meltwater in the circulation cell responsible for the lower-salinity boundary layer beneath the ice shelf at J-9 (Jacobs et al. 1979).

The George VI Ice Shelf has been accorded a 2.1 m yr^{-1} melt rate (53 km^3 yr^{-1}) from a study of glaciological and oceanographic data (Potter and Paren 1985). Net basal freezing of ~0.6 m yr^{-1} along the central flow line of the Amery Ice Shelf has been calculated by Budd et al. (1982). Since oceanographic measurements near the Amery ice front (Smith et al. 1984) also suggest basal melting, that freezing rate has tentatively been applied to only half of the ice shelf area, resulting in -12 km^3 yr^{-1} (Table 1).

Models and indirect observations suggest high basal melt rates near the ice fronts (references on p. 77 of Jacobs et al. 1985; Fahrbach et al. 1991). These results are consistent with the low temperature and salinity in oceanographic transects perpendicular to the coastlines (e.g., fig. 5 in Jacobs 1989). For the eastern Ross Ice Shelf front, Thomas and MacAyeal (1982) estimated a basal melt rate of 0.7 m yr^{-1}, decreasing to near zero 100 km south of the front. That conservative estimate corresponds to ~3.5 km^3 yr^{-1} for each 100 km of ocean frontage. Excluding the interior Filchner-Ronne, Ross, George VI and Amery Ice Shelves which are covered separately above, most of the remaining 648 x 10^3 km^2 ice shelf area (from Drewry 1983), falls within the 100 km coastal band and nets 227 km^3 yr^{-1}. Laboratory experiments indicate that wall melting exceeds basal melting by a factor of ten (Neshyba and Josberger 1980), which would be ~350 cm yr^{-1} in this case. With a coastline of 31,876 km of which 57% is ice shelves, outlet glaciers and ice streams (Drewry 1983), we obtain another 16 km^3 if the average wall thickness is 250 m.

To the estimates above must be added the small contributions from surface runoff and outflow from beneath the grounded ice sheet. Robin (1987) estimated surface runoff of 36 km^3 yr^{-1} over that 2% of the ice sheet in the ablation area . Melting beneath the grounded ice sheet from geothermal heat flux and glacier sliding may be two orders of magnitude less than beneath the ice shelves and be applicable to half the grounded area (Zotikov, 1963; Budd & Jenssen 1987; Engelhardt et al. 1990). In this case 21 km^3 yr^{-1} will be released along the grounding lines. The sum of all in situ melting components is 610 km^3 yr^{-1} (Table 1), higher than most earlier estimates (e.g., p. 216 in Barkov 1985), but consistent with the Jacobs et al. (1985) appraisal from limited oxygen isotope data.

The model results and data-based mass balances are not well constrained due to uncertainties in the underlying assumptions, scarcity of data, and high spatial and seasonal variability in the coastal ocean. Nonetheless, these conservative estimates of the net basal melting (~610 km^3 yr^{-1}, from Table 1) added to the iceberg volume (2338 $km^3\ yr^{-1}$, from above) indicate that the Antarctic ice sheet was in a negative state of balance for the 1979-89 period. The difference between attrition (2948 $km^3\ yr^{-1}$) and accumulation (2535 $km^3\ yr^{-1}$) is equivalent to a sea level rise of ~1 mm yr^{-1}. A larger difference could result from several other plausible interpretations of the data. While the uncertainties are no less than the net result, the most probable value lies on the negative side of balance. Aside from illustrating the large basal meltwater component, this analysis suggests that we cannot rely upon Antarctica to be a restraining influence on future sea-level rise (NRC/DOE 1985; Bentley 1989; Meier 1990b).

Present Variability and Future Change

The large-iceberg calving rate (Fig. 1) is characterized by a high spatial and temporal variability. We can expect less interannual variability in the basal melt rate, given a steady-state ocean circulation that is buffered near the coastline by large volumes of seawater formed at the sea surface freezing temperature (MacAyeal 1984b). However, some data show large interannual salinity changes in high salinity shelf water at the same site and time of year (0.1-0.15$^o/_{oo}$ in fig. 4.1 of Jacobs 1985). These observations are not from an atypical location, as transects along the full Ross Ice Shelf front at ~8 yr intervals beginning in 1967-68 show a shift to lower salinities and higher temperatures. This does not necessarily indicate progressively more basal melting or less sea ice formation, but does reveal considerable natural variability in the shelf water mass that is important to ISW and bottom water production. Significant annual changes in the properties of bottom water have been observed in transects of salinity and silicate across the continental margin in the northwest Weddell Sea (fig. 2 in Foster and Middleton 1979), possibly related to fluctuations in the characteristics or formation rate of shelf waters. As yet there is little evidence of major variability in the winter atmospheric forcing on the Antarctic continental shelf, but the annual duration of ice cover is known to differ by several weeks in the Ross Sea (Jacobs and Comiso 1989). At present we have only a qualitative understanding of how sea ice cover, thickness and transport relates to brine drainage on the shelf and to the conversion of shelf waters into ISW and bottom water.

The strength of the ocean circulation beneath ice shelves may be highly sensitive to small changes in the properties of inflowing seawater and to the shape of the sub-ice cavity. Hellmer and Olbers (1989) varied input temperature and salinity by .02^{o}C and .02$^o/_{oo}$ and found significant differences between the rates and zones of basal melting and freezing on the Filchner Ice Shelf base (Fig. 3). That variability is small relative to the interannual shifts and seasonal cycles that have been measured along the Ross Ice Shelf front. One sensitive aspect of the cavity shape is the depth of the grounding line, where melting may be proportional to the freezing point depression. This has interesting positive feedback implications if West Antarctic ice sheet stability is linked to grounding line retreat into the deeper interior basins (Hughes 1973). Glaciological models with ocean/ice sheet interactions also suggest that basal mass balance will be sensitive to ocean temperature changes. Most probable bounds, assuming ice shelf thinning of 10% or a basal melting increase of ~1 m yr^{-1}, would contribute 4-30 cm to sea level rise by the year 2100 (Lingle 1985; Thomas 1987). Lingle et al. (1990) show rapid and sustained thinning of the Ross Ice Shelf when the rate of basal melting is increased linearly over 150 years to 2 m yr^{-1}. That is no more than the present-day melt rate of the George VI Ice Shelf (Potter and Paren 1985) where nearly undiluted 'warm' CDW floods the continental shelf. At most other locations CDW access to the ice shelf bases appears to be limited by the presence of higher density shelf water produced by sea ice freezing or lower density shelf water formed by ice melting into westward-flowing coastal waters (Jacobs et al. 1985; Fahrbach et al. 1991). It may also be limited to some extent by the well-developed oceanic frontal zone over the upper continental slope (Jacobs 1989).

Could climate changes cause a shift between a low-melting mode (tens of cm yr^{-1}) associated with High Salinity Shelf Water drainage into the sub-ice cavities, and a high-melting mode (m yr^{-1}) associated with deep water (CDW) intrusions onto the shelf? Warming scenarios linking decreased sea ice formation with less dense shelf water would permit more CDW onto the continental shelf (Parkinson and Bindschadler 1984; Lingle 1985). If precipitation also increased, CDW temperatures could warm by ~0.5°C (Gordon 1983). This would result from a greater density contrast between a warmer or fresher surface layer and the deep waters, perhaps abetted by lower wind speeds lessening the divergence and replacement of the ice and mixed layer. The shift would be toward a more stable Arctic-type pycnocline, through which the vertical heat flux is an order of magnitude less than in the Southern Ocean. Gordon (1982) has used Weddell Sea data to demonstrate the short-term destruction of a marginally stable pycnocline, with widespread impacts on the sea ice cover and deep water properties. CDW temperature increases could be accompanied by greater heat transport across the oceanic slope front and onto the continental shelf. At present there is a direct proportionality between the temperatures of deep water and 'warm' shelf intrusions in the Ross and Weddell Seas, but not at some other shelf locations.

A precipitation change of 3-4 cm yr^{-1}, ~25% above current levels, has been predicted from general circulation models (GCM's) and from accumulation changes on Antarctica since the last glacial maximum (Manabe et al. 1990; Robin 1987). The former derives from model results that show enhanced greenhouse-driven atmospheric warming in the polar regions, particularly during the winter season. Ocean areas near Antarctica have been cited as one of the regions where unambiguous warming would appear earliest (Hansen et al. 1988). The estimated amplitude of surface warming has varied with location around the continent and the interannual variability may be high. In one model a subsurface maximum in ocean warming appeared only near the continental shelf break in the potentially sensitive Amundsen Sea region (Schlesinger 1985). More recent NOAA/GFDL model results have shown an asymmetric response to climate warming, with less of an air temperature increase in the Southern Hemisphere (Bryan et al. 1988; Stouffer et al. 1989; Manabe et al. 1990). This is attributed to the large thermal inertia and greater convective overturning in this ocean-dominated hemisphere, and to the presence of the Antarctic Circumpolar Current. Increased precipitation in high latitudes results in a stronger pycnocline, reduced mixing between the surface layer and CDW, and a decrease in sea surface temperature during the latter part of a 60-year simulation. Surface cooling centered in the Amundsen Sea and higher deep water temperatures are also consistent with the discussion above, but a narrow zonal band of >0.5°C warmer water to depths >2000 m near the Antarctic continent (fig. 2 in Stouffer et al. 1989) may be anomalous. Unless the ocean circulation were to change significantly, that water would be regionally removed northward as bottom water or upwelled into the surface layers and onto the continental shelf. In the latter case, the warmer subsurface shelf waters might increase basal melting (MacAyeal 1984b), a possible counterbalance to more precipitation on the continent.

What next?

The apparent negative mass balance derived above for the Antarctic ice sheet is consistent with the observed direction of sea level change, and does not exceed recent estimates of 2.3 - 2.4 mm yr^{-1} when other contributions are added. It is contrary to recent estimates of Antarctica's negative contribution to sea level rise. The result must be considered preliminary and may be due to short-term variability of iceberg calving and the ocean circulation, the general scarcity of field data, overly primitive models, or the underestimation of present-day accumulation on the ice sheet. The uncertainties associated with these factors could be reduced and some valuable baselines established by a well-focused program of ocean and ice sheet measurements and modeling (Bindschadler et al. 1990). The oceanographic goals of that program should include:

1. Direct measurements of Ice Shelf Water generation and outflow, its spatial and temporal variability, and its fate in the deep ocean. Observations must profile the full water column over several annual cycles and include a mix of geochemical tracers that can yield residence times over relevant time and space scales.

2. Direct sampling and measurement of seawater beneath a major ice shelf by means of holes sited along transects from the ice shelf front to and beyond typical grounding lines. Multiple ice holes and ice cores are essential to resolve regional variability of melting and freezing, particularly near grounding lines, basal crevasses, ice rises and ice fronts. Simultaneous time-

series measurements are needed of the ice shelf thickness, velocity, accumulation, and basal mass balance.

3. Icebreaker penetration and detailed oceanographic sampling of the anomalous and largely unknown Amundsen-Bellingshausen continental shelf region.

4. Modeling of the continental shelf and sub-ice shelf ocean circulations. Models must be sensitized to the measured ranges of temperature, salinity, currents and chemical tracers, and should focus upon regions near the grounding lines, ice fronts and shelf break that may be more highly sensitive to climate change.

5. Establishment of a baseline ice sheet perimeter and elevation, by GPS-controlled shipboard navigation and radar if necessary, against which future satellite mapping may be compared. We assume that polar orbiting satellites will carry altimeters capable of measuring ice extent and thickness within the next decade, but there will be a need for ground-truthing and the siting of weather stations and transmitters for ice velocity measurements.

A schematic figure for portions of such a program (Fig. 6) focuses upon the Ross Sea and Ice Shelf because of its more extensive data base and logistic accessibility. With the AWS/AVHRR/SAR weather/satellite facilities existing or planned for McMurdo, the Ross Sea polynyas can be closely monitored for response to atmospheric forcing. Sediment cores could be obtained at each site, with ice cores and precipitation variability measurements along and across flowlines. Automatic Weather Station or Argos transmitters would provide atmospheric and velocity data between site visits. The proposed ISW transect would cross the Ross Ice Shelf front near its northwest corner, now farther north than at any time in the past 150 years (Jacobs et al. 1986). It could thus provide the opportunity to study the dynamics of an old ice front, and/or a new one, during the term of the project.

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Feature	Area (km^2)	Rate ($cm\ yr^{-1}$)	Volume ($km^3\ yr^{-1}$)
Filchner-Ronne I. S.	403,000	45	181
Ross Ice Shelf	446,000	19	85
George VI Ice Shelf	25,000	210	53
Amery Ice Shelf	19,000	-60	-11
seaward 100km, all ice shelves	650,000	35	227
ice shelf fronts	4,500	350	16
runoff	280,000	13	36
grounded ice sheet	12,000,000	0.35	42
		Total:	629

Table 1. A preliminary estimate of melting around and beneath the Antarctic ice sheet. The seaward 100 km includes the outer 100 km of the Filchner-Ronne, Ross and Amery Ice Shelves. See text for melt rate rationale and literature sources.

Figure Captions

Fig. 1. Annual volume of large icebergs (>28 km) calved from the Antarctic ice sheet, 1979-1989. From NOAA and DMSP satellite data interpreted by the Navy-NOAA Joint Ice Center (e.g., NPOC, 1987-1988). Most iceberg volumes were calculated by assuming an elliptical area and a thickness of 250 m. The other bars show the 1979-1989 large berg average (~1295 $km^3\ yr^{-1}$), and the ice equivalent of 1.5 mm yr^{-1} global sea level change. Accumulation on the ice sheet [1962.7 x 10^{12} kg, from Giovinetto and Bentley (1985) plus 226 $km^3\ yr^{-1}$ for the Antarctic Peninsula from Doake, 1985] is indicated by a dashed line at 2535 $km^3\ yr^{-1}$.

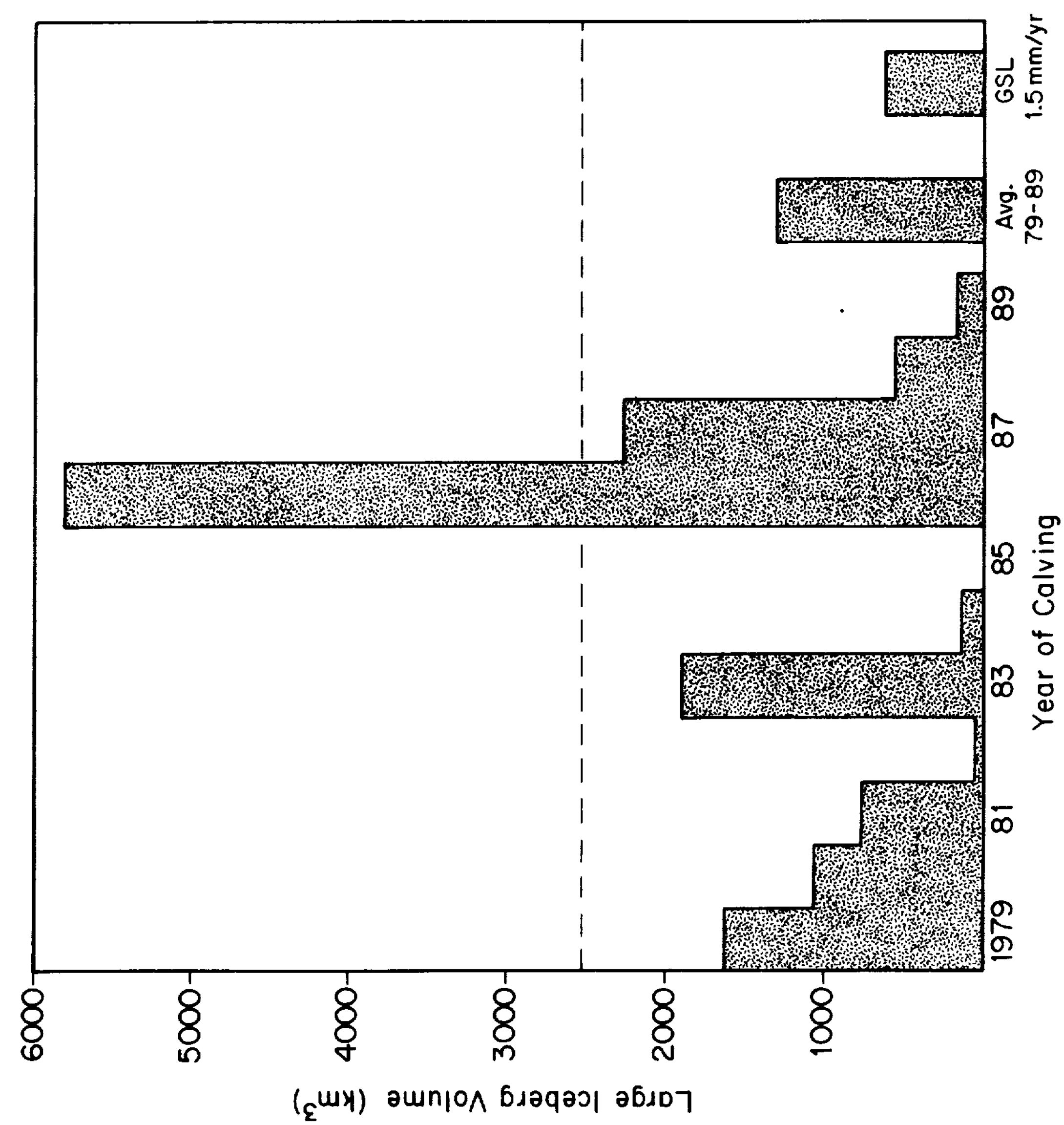
Fig. 2. Annual mean current velocity (top panel) and ocean heat transport (bottom panel), derived from sixteen Aanderaa current meters on nine instrument arrays (A-H) bottom-moored along the Ross Ice Shelf front from February 1984 through January, 1985. Negative contours are shaded and indicate flow into the sub-ice shelf cavity. Data from Pillsbury et al. (1989) and Visser and Jacobs (1987).

Fig. 3. Distribution of melting and accumulation rates at the Filchner Ice Shelf base, from a 2-D thermohaline ocean circulation model by Hellmer and Olbers (1989). The standard simulation, for high salinity shelf water at -1.92°C and 34.72°/oo, is shown by the heavy line. Dashed lines illustrate the altered basal mass fluxes when the shelf water is colder and saltier (circles) or warmer and fresher (triangles), in each case by .02°C and .02°/oo. The lettered bars refer to estimates by previous investigators, some for other ice shelves. A steady-state model of the basal mass flux from glaciological measurements along a 760 km Ronne Ice Shelf flowline shows a similar distribution of melting and freezing areas (fig. 10 in Jenkins and Doake, 1990).

Fig. 4 Ice Shelf Water (ISW) outflow from beneath the Ross Ice Shelf overrides the High Salinity Shelf Water (HSSW) formed by surface freezing in the open Ross Sea. The ISW envelope (heavy dashed line) is defined here by the portions of February 1984 vertical profiles along the ice shelf front where the temperature was below the sea surface freezing point. Data from Jacobs et al. (1989). Small circles denote local temperature minima during several austral summers, with the solid circles (1984 data) joined by light lines. Large squares indicate the locations of long-term (7-12 month) current measurements, from Pillsbury et al. (1989).

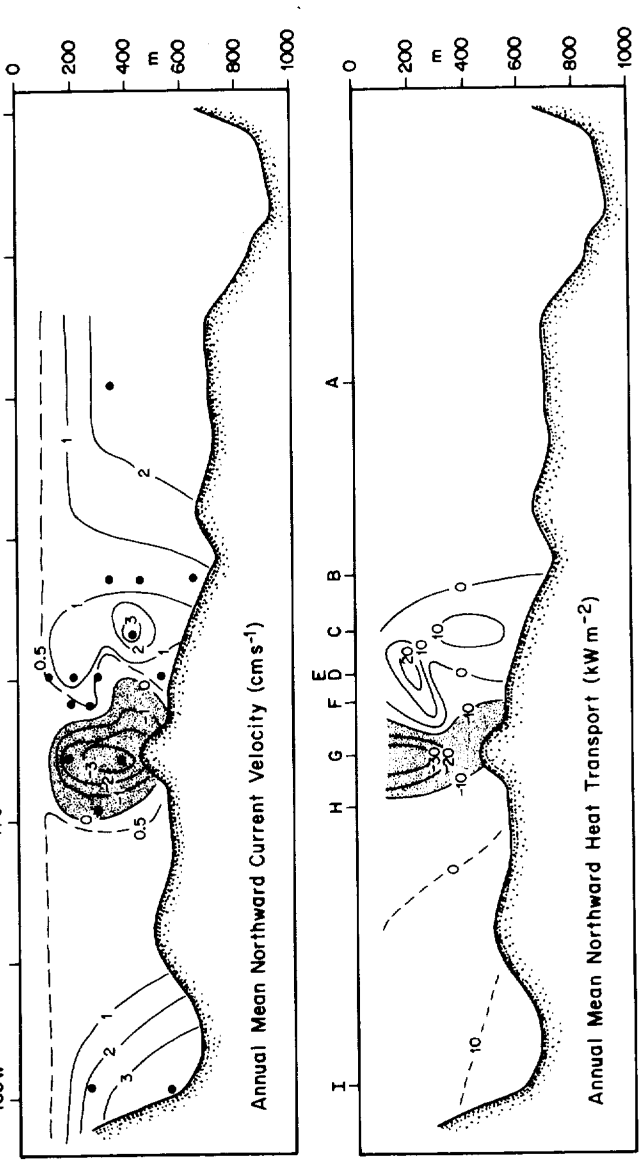
Fig. 5 A simplified representation of sea level change over the past century, and various projections for the next century. In the original sources, most records and predictions are not linear and may have been given as a single value. The 35 cm rise by 2050 of Meier 1990(b) incorporates current and future Antarctic ice sheet growth equivalent to a sea level fall of 0.75 and 5.0 mm yr^{-1}, respectively. The short dashed lines approximate present-day sea level rise (1.5 mm yr^{-1} or 635 km^3 yr^{-1} ice equivalent from Fig. 1), and the largest meltwater pulse from the latest deglaciation, spread over a period of 10^3 yr (24 mm yr^{-1}, derived from Fairbanks, 1989).

Fig. 6 Schematic depiction of a possible SeaRISE transect through the Ross Ice Shelf and across the continental shelf and slope. The vertical lines are drawn to indicate that ocean station profiles and access holes through the ice shelf above the ISW plume should be concentrated near the grounding lines, ice rises, ice front and the continental shelf break. Full sampling for transient tracer and stable isotope geochemistry is essential both across and along the ice front. Beneath the ocean surface, time series recordings of currents, temperatures, and sea ice thickness could be recovered by ship; thermistor chains and other devices in and below the shelf ice could transmit data by satellite. It would be desirable for the ice shelf transect to branch onto both the West and East Antarctic ice sheets, where different water depths and circulation patterns are likely near the grounding lines.



Large Iceberg Volume (km^3)
6000
5000
4000
3000
2000
1000
1979
81
83
85
87
89
Avg.
79-89
GSL
1.5 mm/yr
Year of Calving

Fig 1



160W
170
180
165E
0
200
400
m
600
800
1000
Annual Mean Northward Current Velocity (cm s-1)
I
H
G
F
E
D
C
B
A
Annual Mean Northward Heat Transport (kW m-2)

Fig 2

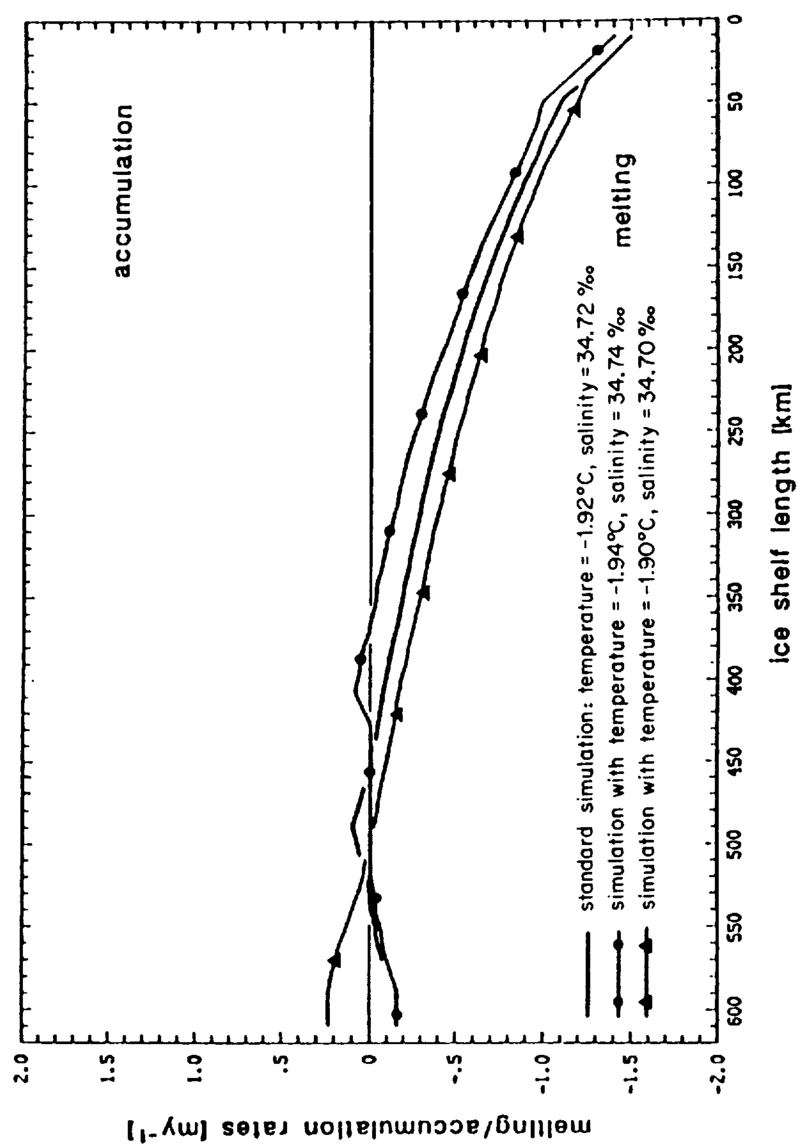
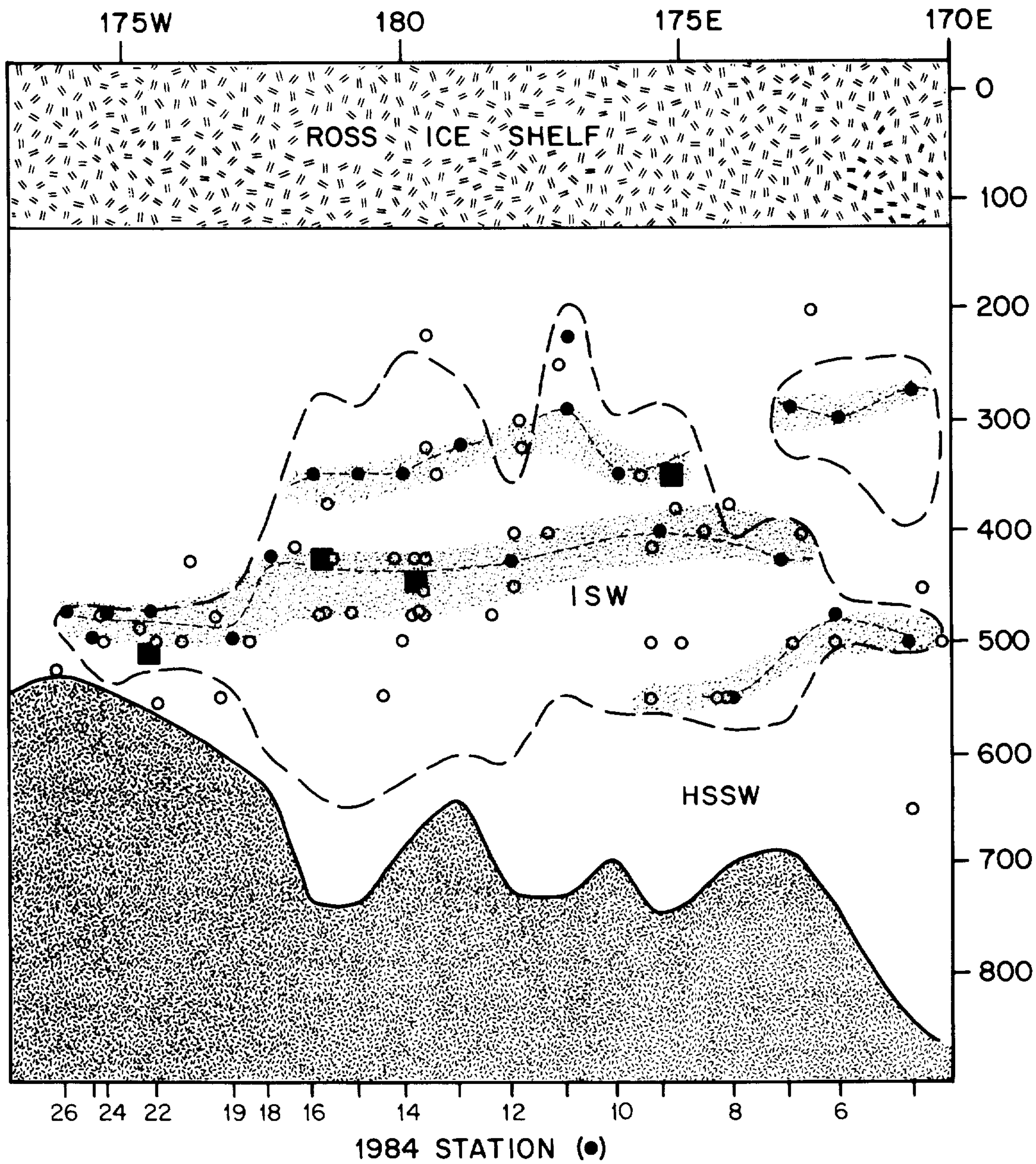
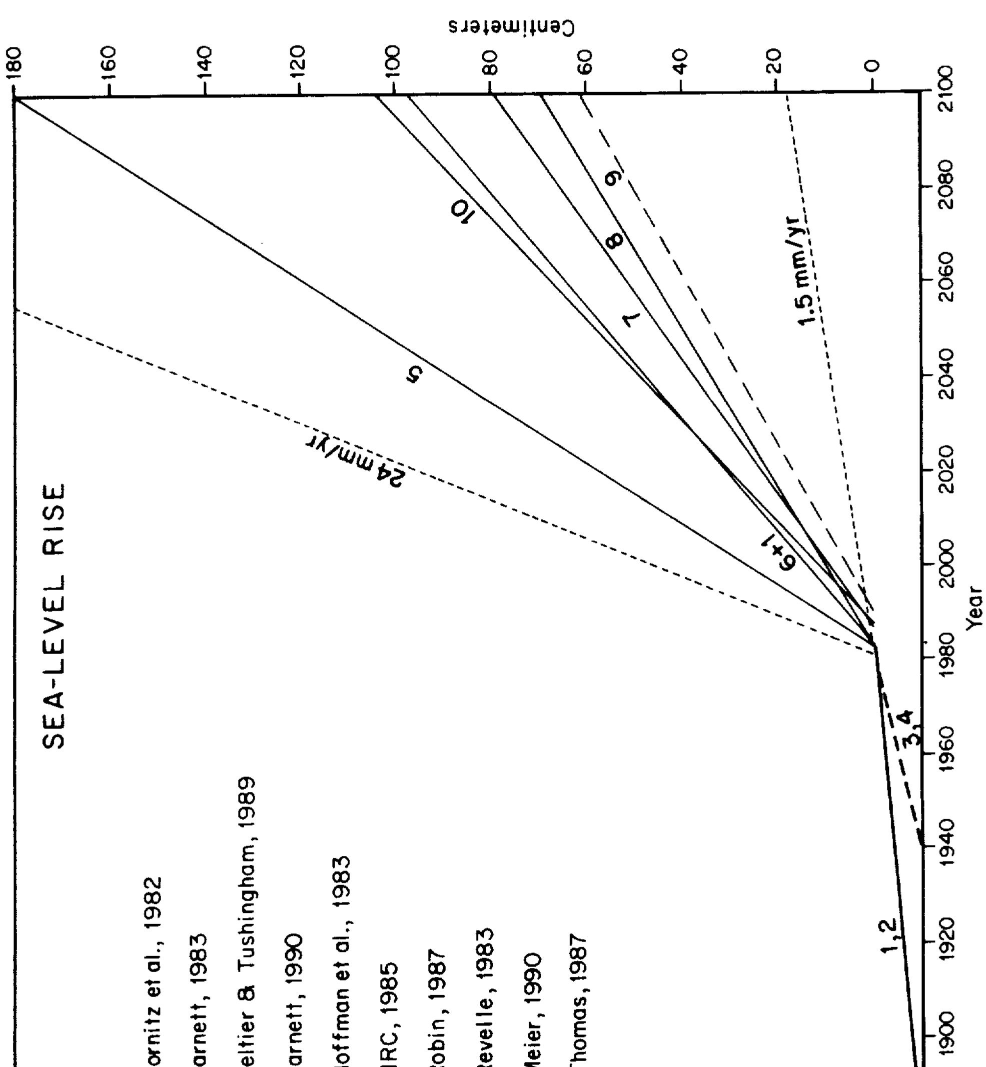

accumulation
melting
standard simulation: temperature = -1.92°C, salinity = 34.72 ‰
simulation with temperature = -1.94°C, salinity = 34.74 ‰
simulation with temperature = -1.90°C, salinity = 34.70 ‰
melting/accumulation rates [my-1]
2.0
1.5
1.0
.5
0
-.5
-1.0
-1.5
-2.0
ice shelf length [km]
600
550
500
450
400
350
300
250
200
150
100
50
0

Fig 3



175W
180
175E
170E
ROSS ICE SHELF
0
100
200
300
400
500
600
700
800
DEPTH (m)
ISW
HSSW
26
24
22
19
18
16
14
12
10
8
6
1984 STATION (●)

Fig 4



SEA-LEVEL RISE
1. Gornitz et al., 1982
2. Barnett, 1983
3. Peltier & Tushingham, 1989
4. Barnett, 1990
5. Hoffman et al., 1983
6. NRC, 1985
7. Robin, 1987
8. Revelle, 1983
9. Meier, 1990
10. Thomas, 1987
24 mm/yr
1.5 mm/yr
1,2
3,4
6+1
Centimeters
Year

Fig 5

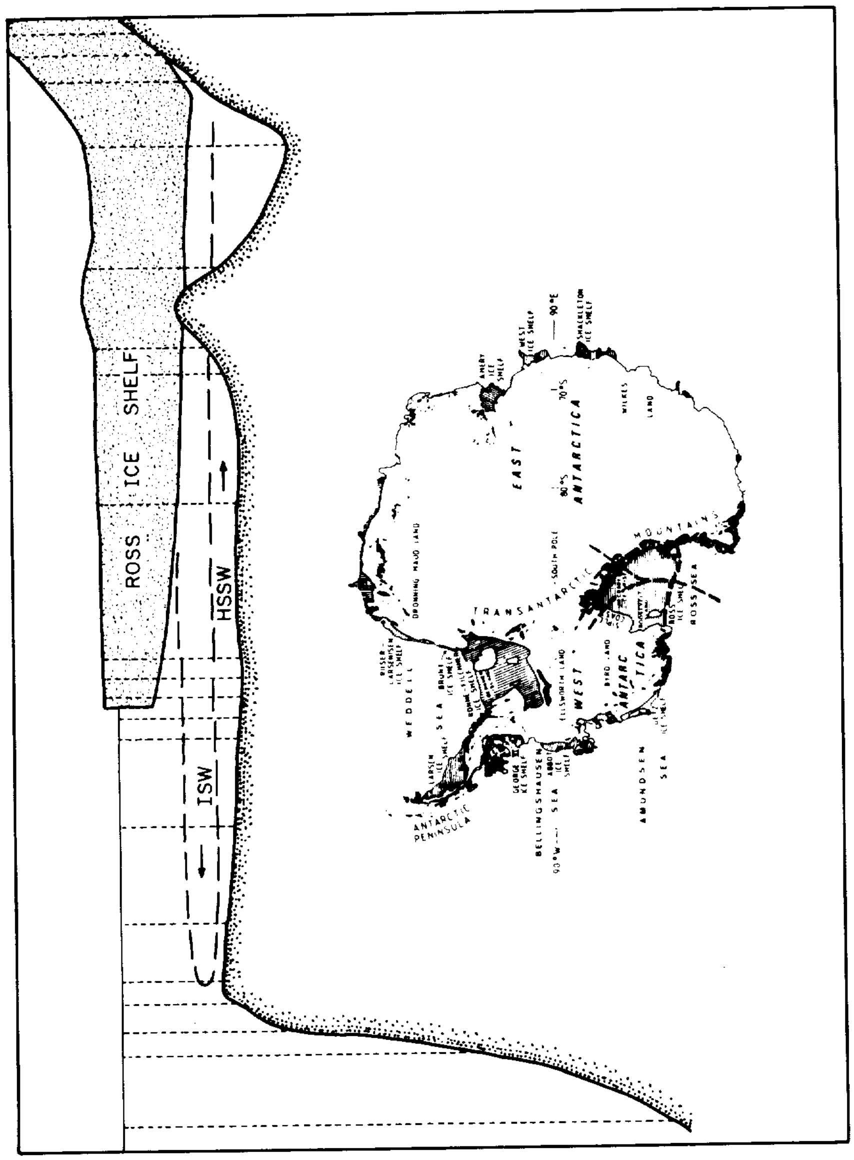

ROSS ICE SHELF
HSSW
ISW
EAST
ANTARCTICA
WEST
ANTARC
TICA
TRANSANTARCTIC
MOUNTAINS
WEDDELL SEA
ROSS SEA
AMUNDSEN SEA
BELLINGSHAUSEN SEA
ANTARCTIC PENINSULA

Fig 6