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Bottom water production and its links with the thermohaline circulation

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Abstract: For more than a century it has been known that the abyssal basins of the world ocean are primarily occupied by relatively cold and fresh waters that originate in the Southern Ocean. Their distinguishing characteristics are acquired by exposure of surface and shelf waters to 'ventilation' by the polar atmosphere and to the melting and freezing of ice over and near the Antarctic continental shelf. Subsequent mixing with deep water over the continental slope results in 'Bottom Water' that forms the southern sinking limb of the global 'Thermohaline Circulation.' Over recent decades, oceanographers have wrestled with a variety of bottom water and thermohaline circulation problems, ranging from basic definitions to forcing and formation sites, source components and properties, generation processes and rates, mixing and sinking, pathways and transports. A brief review of these efforts indicates both advances and anomalies in our understanding of Antarctic Bottom Water production and circulation. Examples from ongoing work illustrate increasing interest in the temporal variability of bottom water in relation to climate change.

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Key words: Antarctic, seawater density, slope front, Southern Ocean, water masses, water properties

The global thermohaline circulation

First measured during the *Challenger* expedition of 1872–76, the frigid temperatures of the deep, interconnected basins of the world ocean clearly indicated a polar origin (Warren 1981). Since that time, the cold, low-salinity characteristics of abyssal bottom waters have repeatedly been traced back to the Southern Ocean, while the warmer, saltier overlying deep waters have been shown to form in the North Atlantic (Mantyla & Reid 1983). The densest Antarctic Bottom Water (AABW) does not spread far into the global ocean without first mixing into overlying deep water, some of which is also formed around Antarctica (Wust 1933, Orsi *et al*. 1999). North Atlantic Deep Water reaches the sea floor at fewer locations, but provides the identifying properties of the Circumpolar Deep Water (CDW) that mixes with AABW. The distinction between 'bottom' and 'deep' waters is thus a matter of definition, as is 'Thermohaline Circulation' (THC), which can also be considered the separate circulation of temperature and salt (freshwater), given their different 3-D distributions and surface boundary conditions (Wunsch 2002).

The THC is more commonly taken to be the "*buoyancy-driven flow-field associated with water cooled (or heated) by contact with cold (warm) air, or modified by sources and sinks of fresh water. [It] may also include flows whose characteristics are significantly altered by upwelling and/or mixing. Water sinking at high latitudes tends to return equatorward in relatively strong, narrow deep western boundary currents*" (Schmitz 1995). The diagram accompanying that definition (Fig. 1) has several interesting features, such as an anomalous BW source near 60°S in the South Pacific, and the lack of any bottom water (BW) transport through the Drake Passage, a topic that will be addressed below. Bottom water pathways extend well into the northern oceans, including 2 Sv entering the western North Atlantic. Following Wust (1933), Amos *et al*. (1971) identified bottom water at 30°N in that region by its thermohaline characteristics, finding < 20% original AABW near the sea floor. That initial signature may have been acquired at a number of southern locations, and in a variety of ways.

Bottom water formation

Investigating the composition of BW, Gill (1973) plotted potential temperature vs. salinity from three vertical ocean profiles taken over the continental slope and shelf in the southern Weddell Sea (Fig. 2). Thermohaline properties of the bottom water near B appeared to be accounted for by a mixture of cold, salty shelf water between K and K' from the shelf station and warmer, fresher water near D and D' between ~200 m and the T/S maxima on two widely separated slope stations. This basic recipe was subsequently employed in several studies by Foster & Carmack (e.g. 1976) for whom it is sometimes named. It provided a sink for the dense, high salinity shelf water generated from brine drainage as sea ice is formed and exported from the continental shelf. The salt from sea ice may not be a major influence on AABW salinity (Toggweiler & Samuels 1995), but modified salty deep waters intrude onto the continental shelf in various strengths at many locations. The mixing scenario depicted in Fig. 2 would combine light and heavy

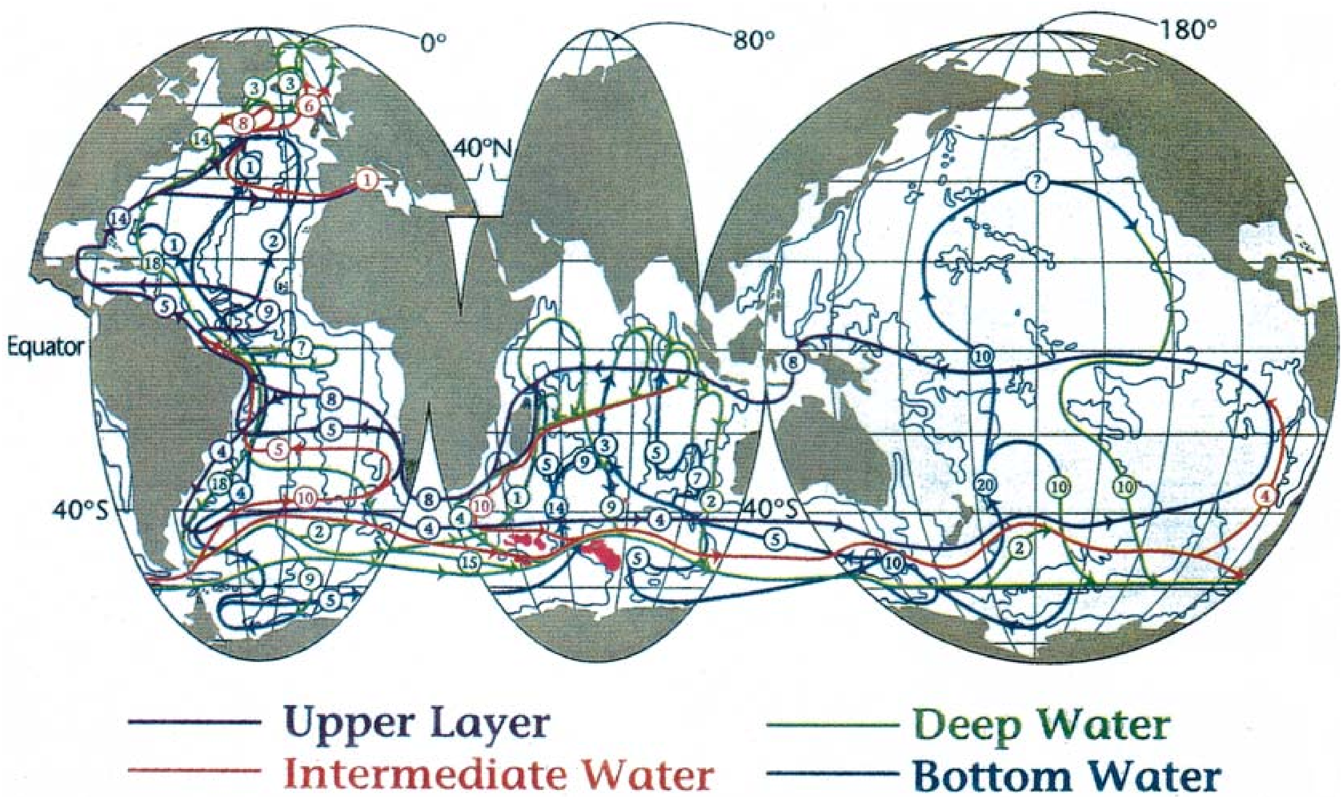

Fig. 1. A four-layer representation of lateral transports in the global thermohaline circulation (THC), by Schmitz (1995). Volume transport estimates in Sverdrups (1 Sv = 10^6 m^3 s^{-1}, sustained over a full year) are in circles associated with the key path segments, along which colour changes indicate water mass conversion. Widely cited wiring diagrams of this sort can be difficult to interpret and overly simplified in their depictions of bottom water. Reproduced by permission of American Geophysical Union (copyright 1995).

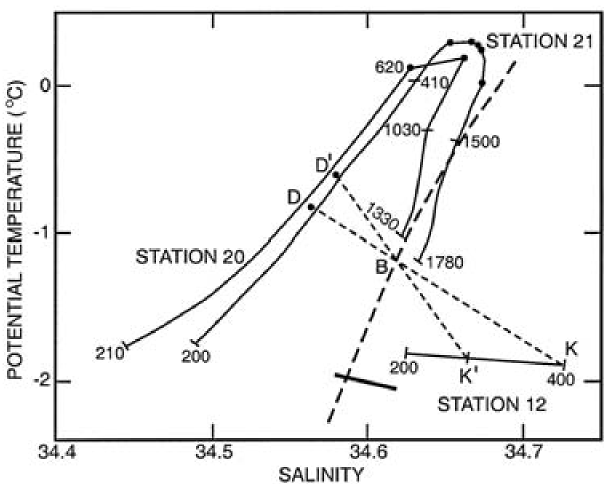

Fig. 2. Potential temperature/salinity diagram for three 1968 USCGC *Glacier* stations in the southern Weddell Sea, with '12' on the continental shelf and '20/21' over the continental slope. The dashed lines postulate mixing between shelf (K) and slope (D) waters to yield the observed bottom water (B). Modified from fig. 9 in Gill (1973). The curved dashed line is an isopycnal (line of constant potential density), and the short solid line shows the thermohaline range of Ice Shelf Water near the mouth of Filchner Trough, from fig. 7 in Foldvik *et al*. (1985).

components, which requires more energy than mixing waters of similar density, and ignored another important water mass, now typically referred to as Ice Shelf Water (Jacobs *et al*. 1970).

Ice Shelf Water (ISW) is generated by basal melting and freezing under ice shelves and glacier tongues (Fig. 3), and is identified by an *in situ* temperature colder than the sea surface freezing point. In the large Ross and Weddell embayments it can reach the continental shelf break and spill down the slope (Foldvik *et al*. 1985, 2004, Rubino *et al*. 2003). ISW production provides another sink for the dense high salinity shelf water, which drains into the deep ice shelf cavities, carrying sensible heat that can melt the basal ice. ISW also has the right thermohaline properties for a more isopycnal mixing with the deep water (Fig. 2). While providing an alternative to the Gill/Foster/Carmack recipe, ISW may not partake directly in much more than ~3 Sv of BW production, a small percentage of the Southern Ocean total (Jacobs *et al*. 1985). More than half of net ice shelf melting appears to be driven by the warmer deep water (Jacobs *et al*. 1996), resulting in a lighter product that upwells into the near-surface layers that contribute less to BW. Where basal ice shelf melting is sufficiently strong, that process may even suppress BW formation (Fahrbach *et al*. 1994).

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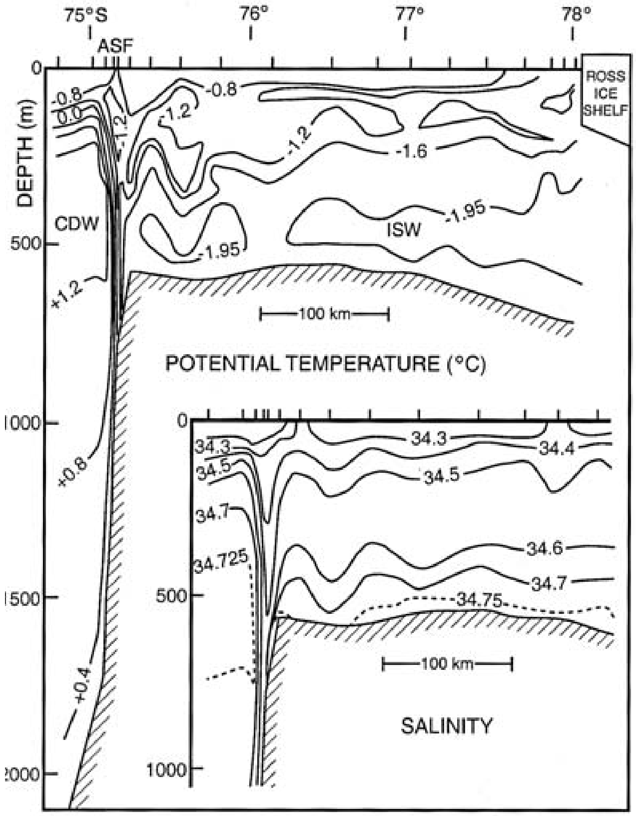

Fig. 3. Potential temperature and salinity (inset) vs. depth from measurements at USCGC *Polar Sea* stations 106–124, taken in the central Ross Sea in February 1984 (section 25 in Fig. 6). On this transect Ice Shelf Water (ISW) is defined as the region with temperatures colder than -1.95°C and Circumpolar Deep Water (CDW) has maximum temperatures above +1.2°C.

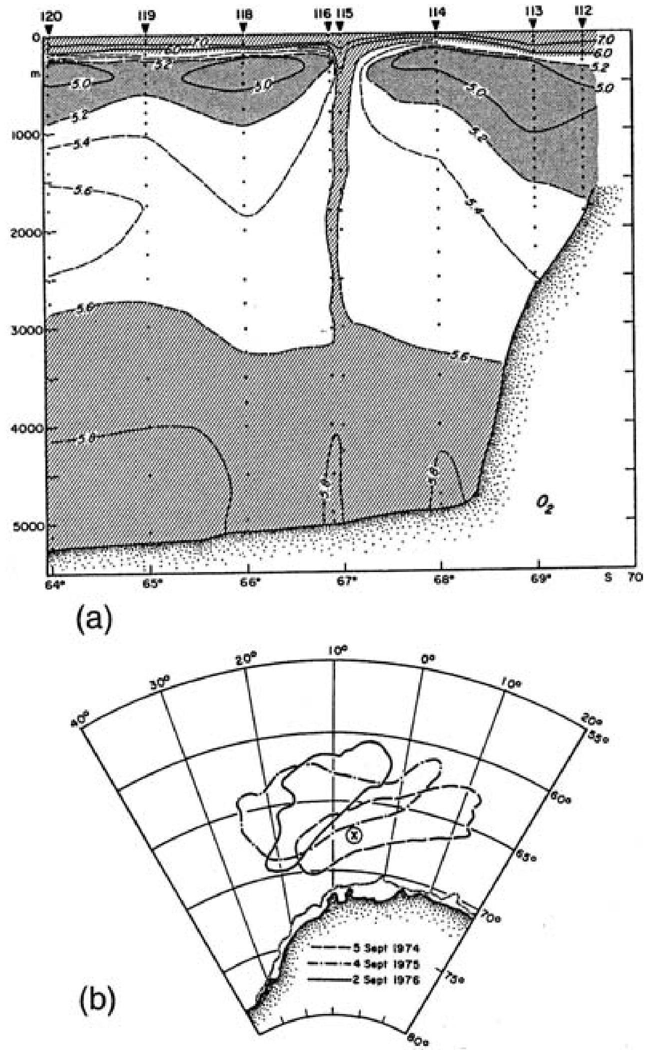

Fig. 4. a. Dissolved oxygen vs. depth from water bottle samples at ARA *Islas Orcadas* stations 112–120 along a NW–SE transect in the eastern Weddell Sea in February 1977. Stations 115/116 are near the 'X' in **b.** showing the outline of the Weddell Polynya in early September of 1974–76. From Gordon (1978). Copyright AMS.

Deep water that eventually becomes bottom water in the global ocean can be produced seaward of the Antarctic continental margin, at least in the Weddell Sea, as evidenced by a remnant convection 'chimney' discovered by Gordon in 1977 (Fig. 4a). There the temperature and salinity structure, along with density and dissolved oxygen, had clearly been perturbed over > 3000 m of the water column at stations 115/116. Perhaps only one of several chimneys formed at the time (Killworth 1979), this finding was interpreted as supporting the idea that deep convection occurs within the Weddell Gyre (Wust 1933). Often associated with the well-known Weddell Polynya that appeared during the winters of 1974–76 in the passive microwave sea ice field (Fig. 4b), this mode of open ocean deep convection has excited persistent interest, particularly as many large-scale, low-resolution numerical models have relied on it to generate BW (Goodman 1998). However, its deep water formation rate may have been no larger than 1.6–3.2 Sv (Gordon 1982), and the large accompanying polynya only appeared during the first 10% of the three-decade satellite sea ice record.

Gill (1973) also observed that BW appears to be formed along the shelf edge west of ~30°W in the Weddell Sea, a concept typically credited initially to Brennecke (1921). He showed a V-shaped, double-sided frontal region over the continental slope (Fig. 5), where shelf, deep and surface waters can readily mix and descend into the deep ocean. Neither requiring nor precluding a role for high salinity shelf water, or ISW, this formation method simply utilises what is available from both sides of the front, from above and upstream. Nonlinearities in the equation of state of seawater can increase the density of the resulting mixture (Fofonoff 1956). The shelf break/continental slope region has relatively low vertical stability (Whitworth *et al*. 1998) and is sensitive to thermobaric effects (Killworth 1977, McPhee 2003) that can also enhance density and sinking.

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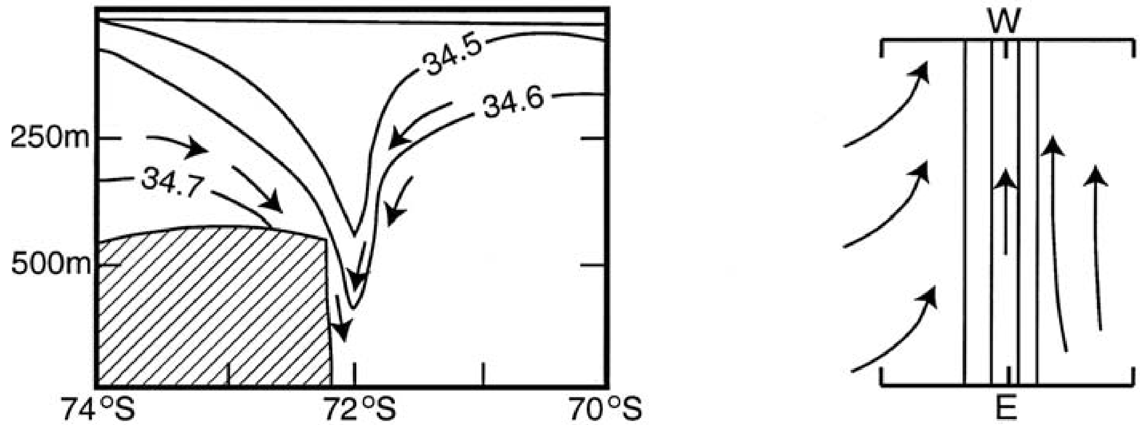

Fig. 5. Salinity vs. depth from February 1968 USCGC *Glacier* stations near 50°W across the outer continental shelf and deeper Weddell Sea, with arrows indicating probable circulation in the vertical plane (left) and near 400 m in the horizontal plane (right). Modified from Gill (1973), who used data near section 50 in Fig. 6.

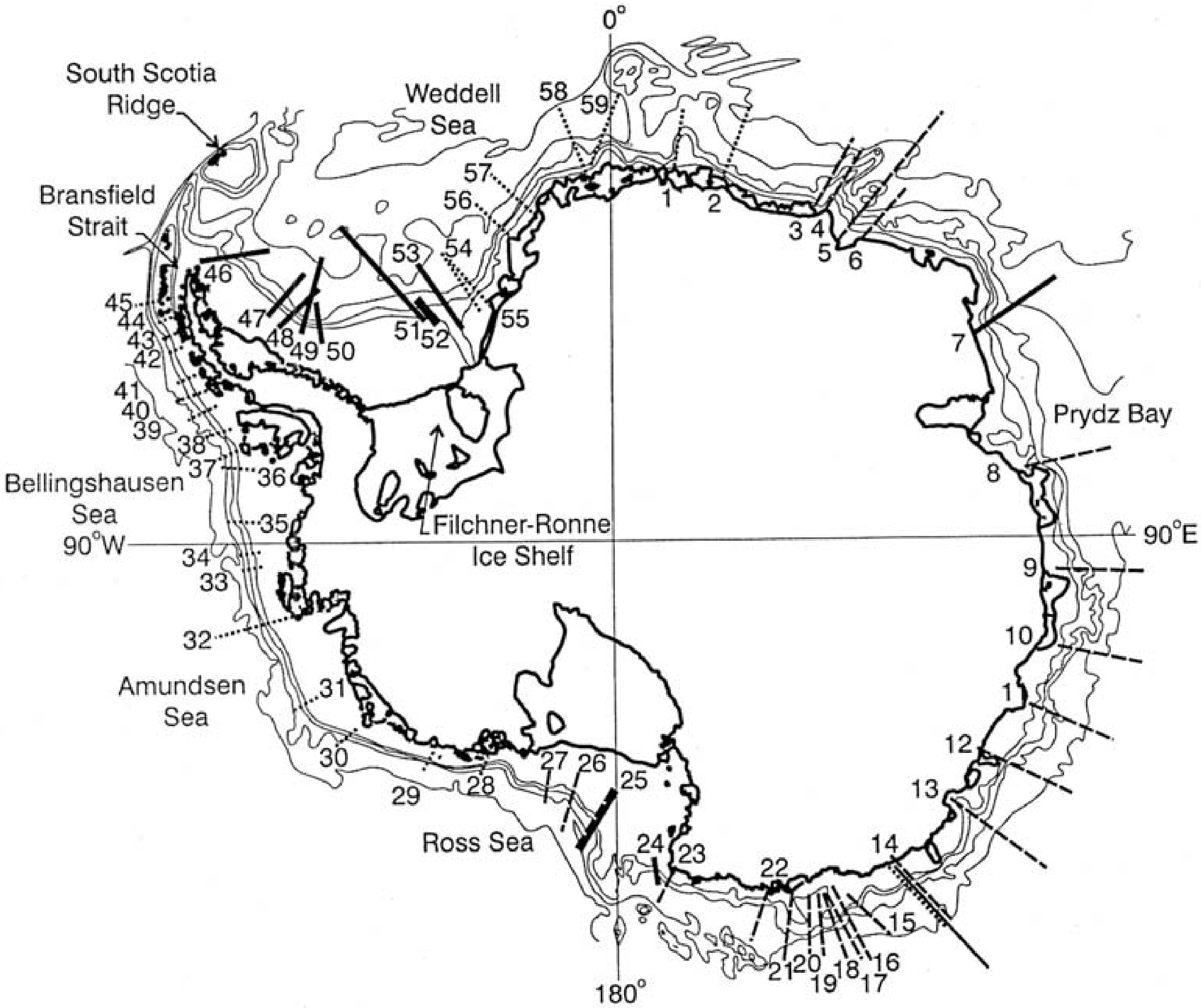

Fig. 6. The locations of sections of water properties (mostly temperature and salinity) taken around Antarctica, approximately perpendicular to the continental slope, that reveal the presence or absence of downslope flow. Depth contours are at 1 km intervals increasing away from the shoreline. Solid lines: downslope flow appears to be occurring at the time of observation; dashed lines: water near the bottom of the slope appears to have come from the shelf, implying downslope flow nearby; dotted lines: no evidence of downslope flow. From Baines & Condie (1998). Reproduced by permission of American Geophysical Union (copyright 1998).

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Tidal energy for mixing is high near the shelf break (Beckmann & Pereira 2003, Padman *et al*. 2003) and can reduce local sea ice concentrations (Robertson *et al*. 2003). Icebergs that are entrained in the swift slope front current (Keys *et al*. 1990) can reduce the local sea ice cover, enhancing the effects of winter atmospheric forcing. BW properties and the intensity of production near the slope front vary seasonally (e.g. Fukamachi *et al*. 2000), but the role of submarine canyons in plumes and other downslope flows (Fig. 6) is not yet clear.

A broad frontal region appears seaward of the George V Coast (Gordon & Tchernia 1972), while temperature and salinity sections in the Ross Sea (Figs 3 & 7) often indicate a narrow, complex front over the outer shelf and upper slope. In addition to the features identified by Gill (1973), Fig. 7 suggests sharp lateral gradients, a temperature minimum imbedded in the upper levels of the shelf-break current and strong CDW upwelling onto the shelf, a less-studied but equally important component of the global THC. Sinking along the Antarctic Slope Front (Jacobs 1991) can also ventilate the deep CDW by the interleaving of less dense varieties of BW (Carmack & Killworth 1978), and some relict convection features over the deep ocean may result from eddies that originate along the continental margin (Bersch 1988). Potential BW can be simply advected off the continental shelf, as into the basins of Bransfield Strait and along the Weddell–Scotia Confluence (Whitworth *et al*. 1994), if not north of Elephant Island (Nowlin & Zenk 1988, Meredith *et al*. 2003) where the persistence and relative strength of downslope and alongslope currents remain to be determined. A slope front may thus not be essential to the formation process, even though its strength in the productive Weddell and Ross seas suggests a close connection. But a frontal region is common along much of the lengthy continental margin, and since the formation processes noted above are not all site specific, BW is probably formed at many sites around Antarctica (Baines & Condie 1998, Whitworth *et al*. 1998).

Bottom water sources

At this point, classical BW theory encounters an awkward problem. In spite of similar water masses, continental shelf and ice shelf areas comparable to the Weddell Sea, persistent winter coastal polynyas and sea ice export, and a strong slope front mixing region, the Ross Sea is often thought to produce little BW. This may result in part from the long history of BW studies in the Weddell Sea, where both formation mechanisms and properties have initially been defined. Volumetric analyses (Carmack 1977, Rintoul 1998) have thus resigned the Ross Sea to a minor role, assuming that its output falls in a narrow high-salinity range, and that all BW must have a potential temperature lower than 0.0°C. Much of the deepest water in the south-east Pacific Basin is warmer than this, however, and clearly

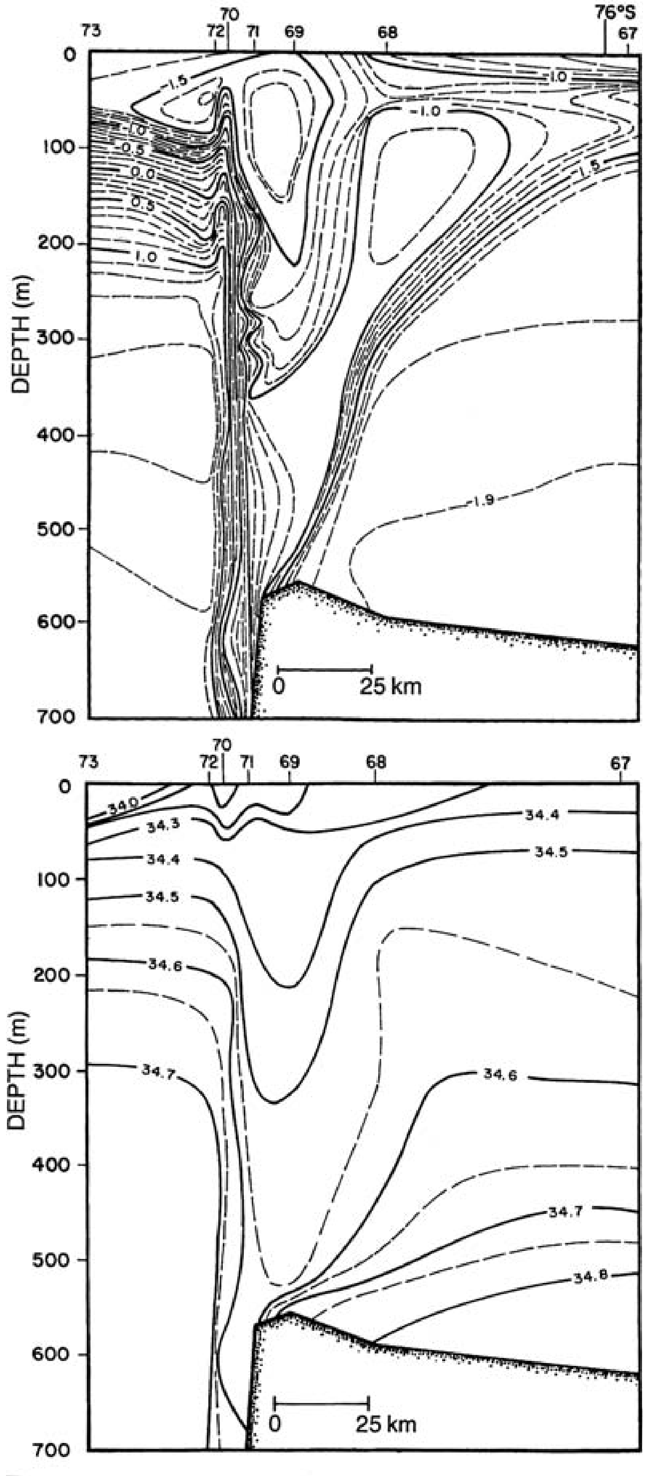

Fig. 7. Temperature and salinity vs depth from late December 1976 USCGC *Northwind* stations near 175W across the outer continental shelf and deeper Ross Sea. While these stations are sequential, 69–72 were occupied over an 8-hr period, and property gradients may have been distorted by the tidal regime.

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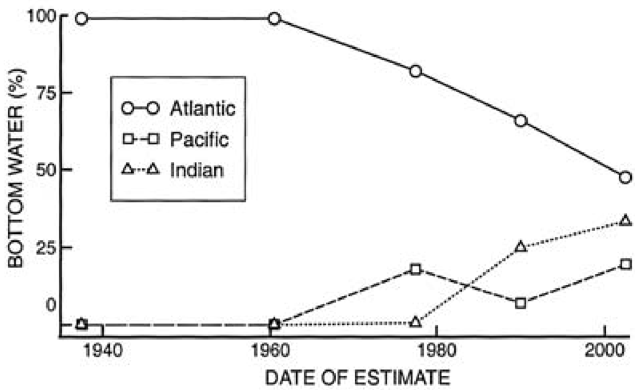

Fig. 8. Evolving estimates of bottom water source regions around Antarctica, with Atlantic representing the Weddell Sea, Pacific the Ross Sea and Indian mainly the George V Coast. From Deacon (1937), Stommel & Arons (1960), Carmack (1977), Rintoul (1998), Orsi *et al*. (2002) and Nakano & Suginohara (2002). The 1998 publication utilised a 1981 volumetric analysis, and is plotted in 1995; the 2002 estimates are averaged.

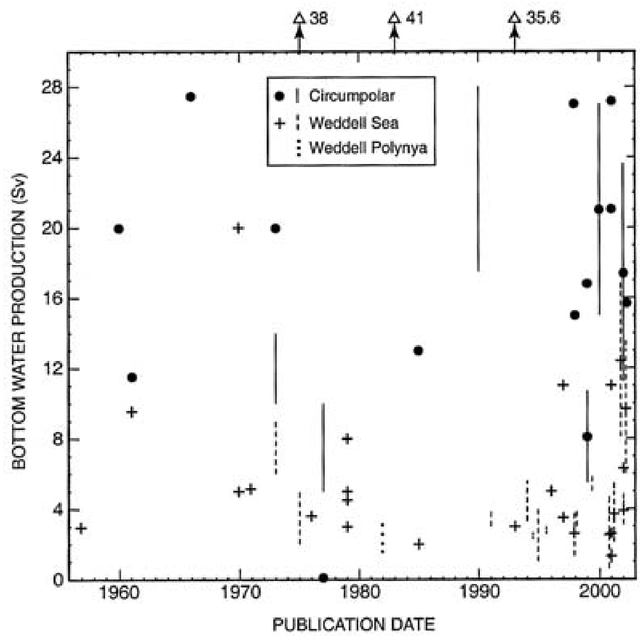

Fig. 9. Evolving estimates of bottom water production rates in the Weddell Sea and circumpolar Southern Ocean. From tables in Solomon (1974) and Naveira Garabato *et al*. (2002), supplemented by Stommel & Arons (1960), Munk (1966), Worthington (1977), Gordon (1982), Jacobs *et al*. (1985), Gordon & Huber (1990), Fahrbach *et al*. (1991), Broecker *et al*. (1998), Goodman (1998), Campin & Goose (1999), Ganachaud & Wunsch (2000), Hellmer & Beckmann (2001), Nakano & Suginohara (2002) and Orsi *et al* (2002). Vertical bars indicate ranges or stated uncertainties, with off-scale values of 38, 41 and 35.6 Sv from Gordon & Taylor (1975), Stuiver *et al*. (1983) and England (1993), the last being the average of 14 simulations with a general circulation model.

emanates from the Ross Sea (Gordon 1966). Some of the Ross Sea outflow is above zero because BW properties reflect those of its deep water component (Fofonoff 1956), roughly a degree warmer over the Ross Sea continental slope than over the Weddell slope (compare Figs 2 & 7). BW from the Ross Sea also has a much wider salinity range than is commonly assumed, and that range probably varies over time (Jacobs *et al*. 1970, 2002).

Speculation about major bottom water source regions around Antarctica does illustrate some interesting trends (Fig. 8). Of course the thermohaline-based distributions are not necessarily an indication of BW provenance, but the Ross sector has typically received a low rating. The Weddell Sea has been slipping steadily since first being identified as the one and only source of any circumpolar consequence. The main beneficiary of that decline appears to be the Indian sector, or perhaps only a short reach along the George V Coast (Rintoul 1998). Nakano & Suginohara (2002) were forced to model lower air temperatures and a higher reference salinity than observed in order to produce any BW along that coastline. They also assumed BW produced in the Ross Sea had no clear path to the Australian Antarctic Basin, in apparent conflict with earlier findings (Gordon & Tchernia 1972, Mantyla & Reid 1983). Other observational and modelling evidence also exists for BW formation in the Indian and Pacific sectors (Orsi *et al*. 1999, Meredith *et al*. 2000, Hellmer & Beckmann 2001). Indeed, Heywood *et al*. (1999) reported 20 Sv of Antarctic BW entering the Weddell–Enderby Basin through the gap south of the Kerguelen Plateau. That estimate may be high (M. Meredith, personal communication 2004), but these and other studies are slowly eroding the notion that most BW has a Weddell Sea origin.

Bottom water production rates

A related BW problem is the production rate, for which many estimates have been made over more than four decades (Fig. 9). The high variability results from a variety of BW definitions and techniques, including boundary current transport, heat, salt and mass budgets, geochemical tracers, and numerical and inverse models. Most Weddell Sea rates are between 2 and 5 Sv, including a Foldvik *et al*. (2004) estimate post-dating the figure, while circumpolar rates have shown a recent tendency to converge near the venerable 20 Sv of Stommel & Arons (1960). Although Stommel & Arons supported the Deacon (1937) interpretation of the Weddell Sea as a sole source (Fig. 8), their value is here labelled circumpolar because the Stommel (1958) figure shows the Weddell point source is supplied from as far east as the Pacific. Even in the more recent estimates, overall uncertainties are too large to allow more than very general inferences about temporal variability in the amount of new BW being added to the global THC.

The low BW production rates repeatedly measured in the

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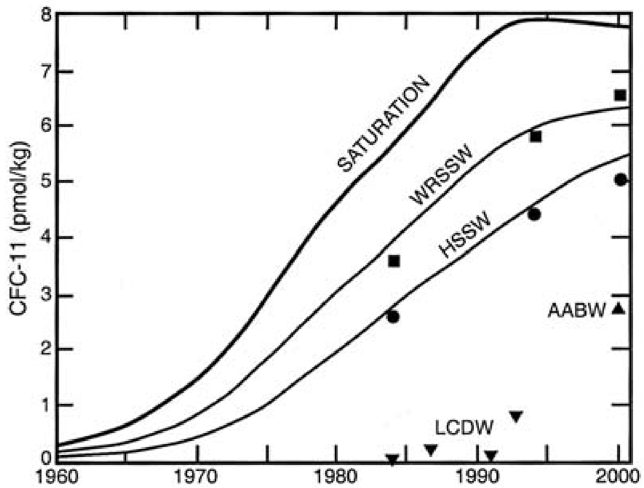

Fig. 10. Chlorofluorocarbon (CFC-11) content vs time for atmospheric equilibrium (Saturation) and for Southern Ocean water masses, modified from a figure in Smethie & Jacobs (2005). The curves labelled WRSSW and HSSW are modelled CFC levels for surface and high salinity shelf water in the southern Ross Sea, with the solid symbols in 1984, 1994 and 2000 showing repeat measurements near the front of the Ross Ice Shelf. Lower Circumpolar Deep Water (LCDW) and AABW values are from Orsi *et al*. (2002).

Weddell Sea, combined with the longstanding belief that most BW forms there, contrast with the need for much larger amounts in global ocean chemistry and numerical models (Broecker *et al*. 1998). However, hypotheses about BW variability often ignore the major role of entrained CDW in the resulting volume, and the highly undersaturated atmospheric gas levels of the surface and shelf waters (Fig. 10). The chlorofluorocarbon (CFC-11) saturation level of ice shelf water is often near 50%, and newly formed bottom water has been closer to 35%, mainly due to its 'old' near-zero CDW component. Reasons for undersaturation of surface and shelf waters include a nearly perennial sea ice cover, residence times that can be very short at the sea surface but several years beneath ice shelves, and the entrainment of deep water components in surface mixed layers (Smethie & Jacobs 2005).

Redefining bottom water

One implication of the above discussion is the need for an

Table I. Southern source water ingredients in Sv ($10^6 m^3 s^{-1}$) for the three neutral density layers in Fig. 12, from Orsi *et al*. (2002). The two deeper layers receive ~27% Antarctic Surface Water (AASW), 31% Shelf Water (SW) and 42% Lower Circumpolar Deep Water (LCDW).

Layer	Atlantic Sector	Indian–Pacific Sector
Top	1.7 AASW	1.9 AASW
Middle	2.9 AASW + 2.9 LCDW	1.8 AASW + 1.8 LCDW
Bottom	3.3 SW + 1.6 LCDW	2.1 SW + 1.1 LCDW

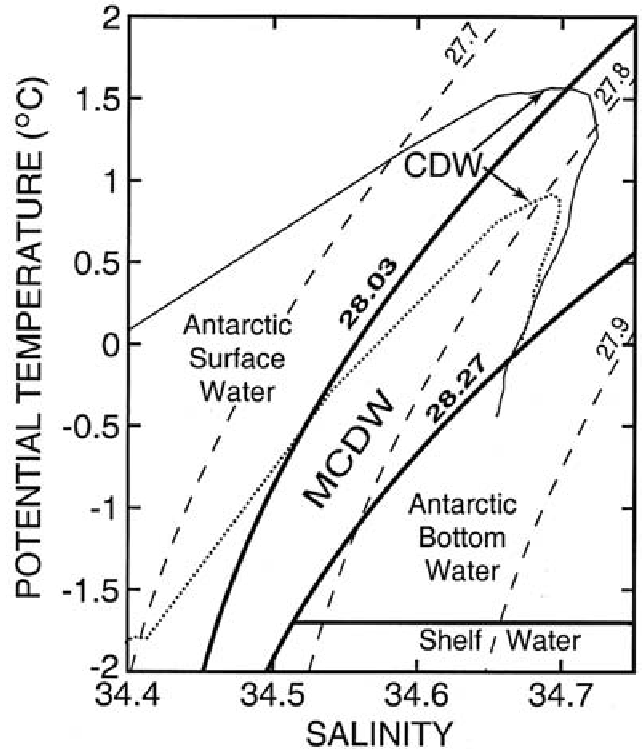

Fig. 11. Water masses near the Antarctic continental margin, defined in relation to potential temperature, salinity, neutral density anomaly (solid curves at 28.27 and 28.03), and potential density anomaly (dashed lines from 27.7 to 27.9). MCDW = Modified Circumpolar Deep Water, with examples from two vertical profiles illustrating part of the CDW range. Adapted from fig. 1a in Whitworth *et al*. (1998). Reproduced by permission of American Geophysical Union (copyright 1998).

improved BW definition, a task that has recently been taken up by Whitworth *et al*. (1998) and by Orsi *et al*. (1999, 2002). Utilizing the neutral density formulation of Jackett & McDougall (1997), they first identified the lightest density surface that did not traverse the Drake Passage. Waters heavier than that 28.27 density anomaly were labelled AABW, or Shelf Water where the temperature fell below -1.7°C (Fig. 11). Shelf water extends to lower values than were once considered critical (e.g. Fofonoff 1956), and with the lack of a minimum depth, typically 2000–2500 m, AABW even appears on the outer continental shelf and upper slope in some sectors (Whitworth *et al*. 1998). The near-bottom distribution of neutral density in the world ocean shows that most of the densest AABW class is restricted to the Antarctic basins (Orsi *et al*. 1999), but in the next density interval, BW can transit the Drake Passage, sometimes under a different name, and it extends northward in the western Atlantic, western Pacific, and on both sides of the Indian Ocean, as in Mantyla & Reid (1983).

Carrying this process one step further, Orsi *et al*. (2002) integrated the moles of chlorofluorocarbon (CFC-11) within three density classes along six N–S transects, one of which is shown in Fig. 12. The two denser classes account for most

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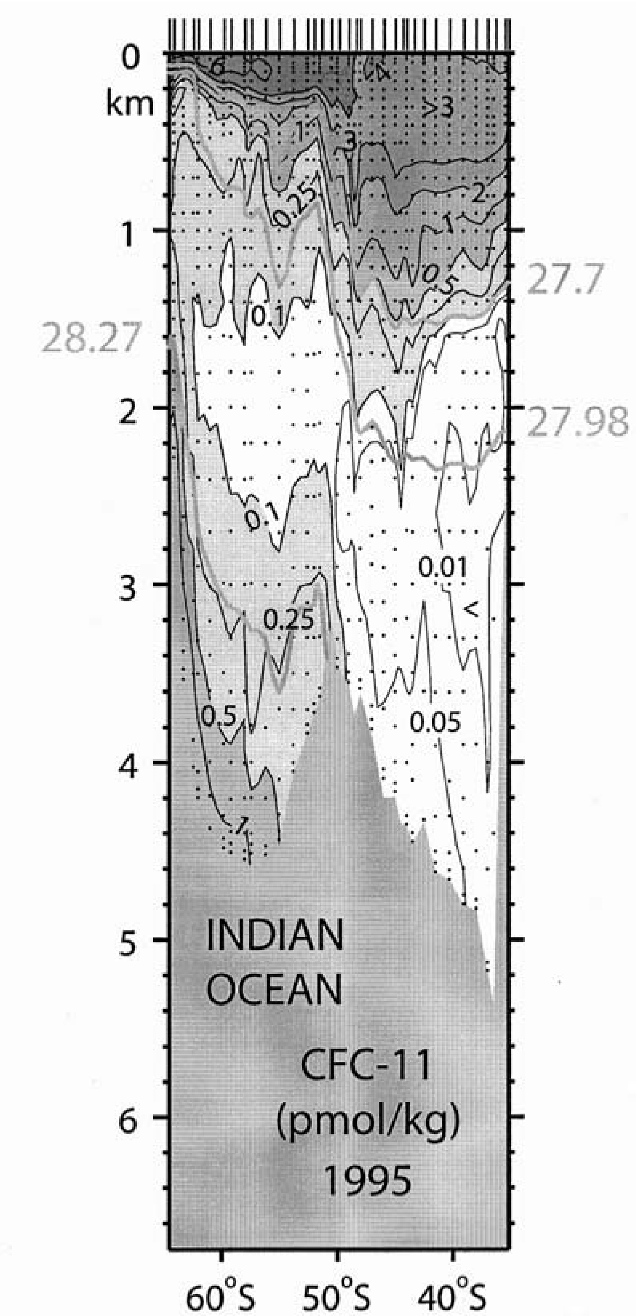

Fig. 12. Vertical distribution of chlorofluorocarbon (CFC-11 in pmol kg^{-1}) from western Australia to Antarctica in 1995, from Orsi *et al*. (2002), near section 12 in Fig. 6. Neutral density anomaly surfaces (27.7, 27.98 and 28.27) define the limits of the three deep layers in Table I. Reproduced by permission of American Geophysical Union (copyright 2002).

of the global ocean area in contact with the deep sea floor, and indicate that ~17.5 Sv of Antarctic waters are being added to the deep ocean (Table I). This provided an independent measure of the ventilation rate of the deep global THC, and revealed remarkably similar inputs from the northern and southern hemispheres. That is consistent with other recent estimates, including the application of an inverse model to the World Ocean Circulation Experiment observations (Ganachaud & Wunsch 2000). From typical water mass mixing recipes, Orsi *et al*. (2002) found that roughly 60% of the Southern Ocean source water ingredients came from the partially-ventilated surface and shelf waters, and 40% from the poorly-ventilated lower CDW (Table I). Their results are preliminary, and it must be assumed that the measured CFC-11 levels and normalizations are representative for this evolving anthropogenic tracer. BW formation in numerical models using CFC and other tracers can be adversely impacted by summer-biased observations and small-scale processes near the continental margin (e.g. Doney & Hecht 2002).

Changing bottom water properties

Reports of annual to decadal variability in BW properties are increasingly common, particularly in the more frequently visited Atlantic sector (e.g. Coles *et al*. 1996, Hogg & Zenk 1997, Andrie *et al*. 2003). One of those findings has been challenged on the grounds of observed offsets in salinity measurements and standardizations between cruises (Gouretski & Janke 2001). However, BW salinity change over short time intervals has been documented elsewhere, and even temperature measurements must be repeated with care, particularly at choke points like the Vema Channel where the range is low and the lower water column structure is spatially variable. Flow reversals also occur at lesser depths near that channel (McDonagh *et al*. 2002), potentially halving the transport indicated in Fig. 1. Similar gaps with restricted flows and countercurrents, such as through the South Orkney Ridge (Naviera Garabato *et al*. 2002), have important implications for BW transport, mixing and monitoring.

Some bottom water in the Atlantic is derived from upper AABW and lower CDW mixing west of the Drake Passage, where it remains to be documented that the resulting properties are constant over time. But the best evidence for BW variability may lie in the Weddell Sea, where its thermohaline properties are known to change on seasonal to decadal time scales (Foster *et al*. 1979, 1987, Fahrbach *et al*. 2001, Visbeck *et al*. 2001). BW temperature increases could result from the observed warming of deep water in the Weddell by ~0.1°C/decade since the early 1970s, perhaps resulting from inflow changes induced by the Antarctic Oscillation or Southern Annular Mode, and/or the recovery from cooling by the Weddell Polynya (Robertson *et al*. 2002, McPhee 2003). Coincident shoaling of the temperature maximum could also signify responses to altered surface forcing. Over seasonal and longer periods, changing wind fields also impact sea ice extent, which in turn may modify temperatures deep beneath the sea surface (Meredith *et al*. 2003).

In the thermohaline range of seawater properties important for BW formation and mixing, seawater density is

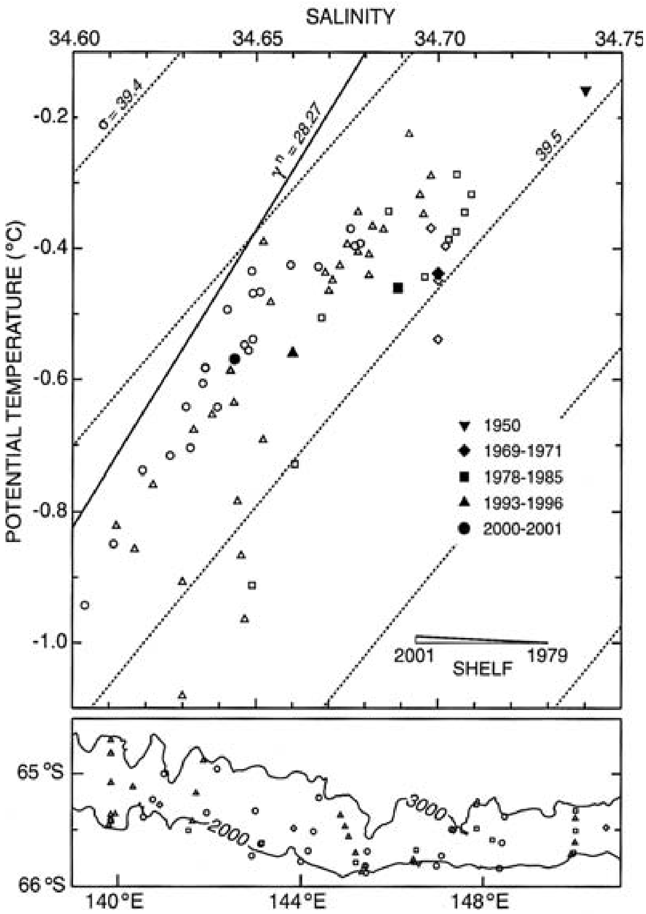

Fig. 13. Potential temperature/salinity (upper panel) near the sea floor in water depths of 1900–3000 m (lower panel) seaward of George V Coast. The open symbols show individual stations and the solid symbols their temporal averages, except for the single 1950 observation. The dotted sloping lines are density surfaces relative to 2500 dbar and the solid line is the AABW neutral density threshold, from Whitworth *et al*. (1998). The arbitrarily positioned 'Shelf' inset shows the mean salinity and temperature change from 1979–2001 from paired CTD stations on the adjacent continental shelf.

reduced as much by a salinity decrease of 0.01 as by a temperature increase of 0.2°C. Salinity reflects shifts in precipitation, melting, sea ice volume and ocean dynamics, all elements of the hydrological cycle believed to be sensitive to climate change. A salinity decrease of as much as 0.02 has been reported since 1994 in BW with potential temperatures < -0.4°C in the Australian Antarctic Basin (Whitworth 2002). While that might seem like a spatial artefact, since the more recent measurements are closer to BW sources, observations taken over a half century in one of those source regions support the salinity decline. Near-bottom temperature/salinity measurements on the continental rise from 140–150°E were aggregated into five time intervals in Fig. 13. The station distribution displayed in the lower panel is less than ideal, but a pronounced drift is evident toward lower salinity and temperature. In this case the cooling is more than balanced by the freshening, with the density decline even approaching the new AABW neutral density threshold. The low bottom salinities in this and other continental slope regions may reflect more melting and lower salinity shelf and surface waters, vs freezing and the deeper, saltier components.

It has long been known that salty, westward BW flow along the southern margin of the Australian Antarctic basin is diluted by local bottom and deep water production (Gordon & Tchernia 1972, Carmack & Killworth 1978, Rintoul 1998). However, it is quite likely that inflow from the Ross Sea has also freshened over the same period, given the 0.12 salinity decrease in its shelf water component (Jacobs *et al*. 2002). The average salinity difference between closely spaced station pairs on the George V continental shelf decreased by 0.036 from 1979 to 2001 ('Shelf' insert on Fig. 13). This may result from upstream influence, local forcing, or both. But if the fresh mode in the Australian Antarctic Basin recurs periodically and is under-represented in the historical data base (Whitworth 2002), then the period of haline variability must be longer than 50 years. Alternatively, this part of the Southern Ocean could be trending toward a reduction in the BW production rate, by analogy with modelled and observed changes in the North Atlantic, or at least a change in its BW properties.

The bottom line

The densest waters in the world ocean THC are formed in the Southern Ocean, primarily along the Antarctic continental margin. Several interrelated mechanisms involve the addition of 'negative buoyancy' by cooling and salinization of partially ventilated surface and shelf waters, and by enhancement of mixing with warmer and saltier deep water. Polynyas, high salinity shelf water, and meltwater from continental ice play supporting roles, but may not be essential to the formation process. With more than half of the circumpolar BW product reported to be in the Atlantic sector, one might question Mosby, who in 1966 wrote (Mosby 1968) *"in my opinion there is no evidence for considering the Weddell Sea as the main source of Antarctic bottom water."* But as BW formed elsewhere now appears to be imported into the Weddell Sea from the east along the continental margin, and into the Atlantic via the Drake Passage, it may turn out that Mosby was correct. The BW formation rate is not well constrained, although geochemical tracers and model results continue to point toward a production in the vicinity of 20 Sv. Increasing evidence for interannual to decadal variability in BW properties includes freshening in the Ross Sea and near the George V Coast that is comparable to the Great Salinity Anomaly (Dickson *et al*. 2002) of the North Atlantic. Possible reasons for the Southern Ocean freshening include changes in winds and precipitation, evolving sea ice volume, and increased melting of the Antarctic Ice Sheet

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(Jacobs *et al*. 2002). Possible changes in BW production on millennial time scales is a matter of increasing interest, but exceeds the scope of this brief review.

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