The Role of Thermohaline Circulation in Global Climate Change

by Arnold Gordon

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The world ocean consists of 1.3 billion cu km of salty water, and covers 70.8% of the Earth's surface. This enormous body of water exerts a powerful influence on Earth's climate; indeed, it is an integral part of the global climate system. Therefore, understanding the climate system requires a knowledge of how the ocean and the atmosphere exchange heat, water and greenhouse gases. If we are to be able to gain a capability for predicting our changing climate we must learn, for example, how pools of warm salty water move about the ocean, what governs the growth and decay of sea ice, and how rapidly the deep ocean's interior responds to the changes in the atmosphere.

The ocean plays a considerable role in the rate of greenhouse warming. It does this in two ways: it absorbs excess greenhouse gases from the atmosphere, such as carbon dioxide, methane and chlorofluoromethane, and it also absorbs some of the greenhouse-induced heat from the atmosphere. Both these processes tend to forestall the greenhouse effect, and it is possible that without them the global average temperature of the atmosphere would now be 1°-2° C warmer. If this is the case, the question is whether the ocean will continue to retard such warming—and whether the rate of this influence will be reduced or accelerated as the process continues. These very important questions must be answered before we can predict the full extent of global climate change with real confidence. For such confidence we need a much better understanding of ocean circulation.

The Thermohaline Circulation

While surface circulation certainly plays a key role in the climate system, its wind-driven features for the most part only move heat and water on horizontal planes. It is the slower thermohaline circulation, driven by buoyancy forcing at the sea surface (i.e., exchanges of heat and fresh water between ocean and atmosphere change the density or buoyancy of the surface water; cooling and/or increased salt concentration induced by excessive evaporation, form dense water which sinks into the ocean's interior), that on the one hand forces the ocean's deep interior to interact with the atmosphere, and on the other can effectively sequester heat and other properties into the enormous volume of the deep ocean.

This sinking of dense surface water occurs in a few restricted regions, and so initiates the thermohaline circulation. The deepest convection occurs in the northern North Atlantic and around Antarctica. In the North Atlantic, thermocline water (the upper km of the ocean separating the warm surface layer from the colder deep water) with a long history of contact with the atmosphere is cooled and sinks as relatively salty water into the deep ocean. In the Southern

Fig. 1

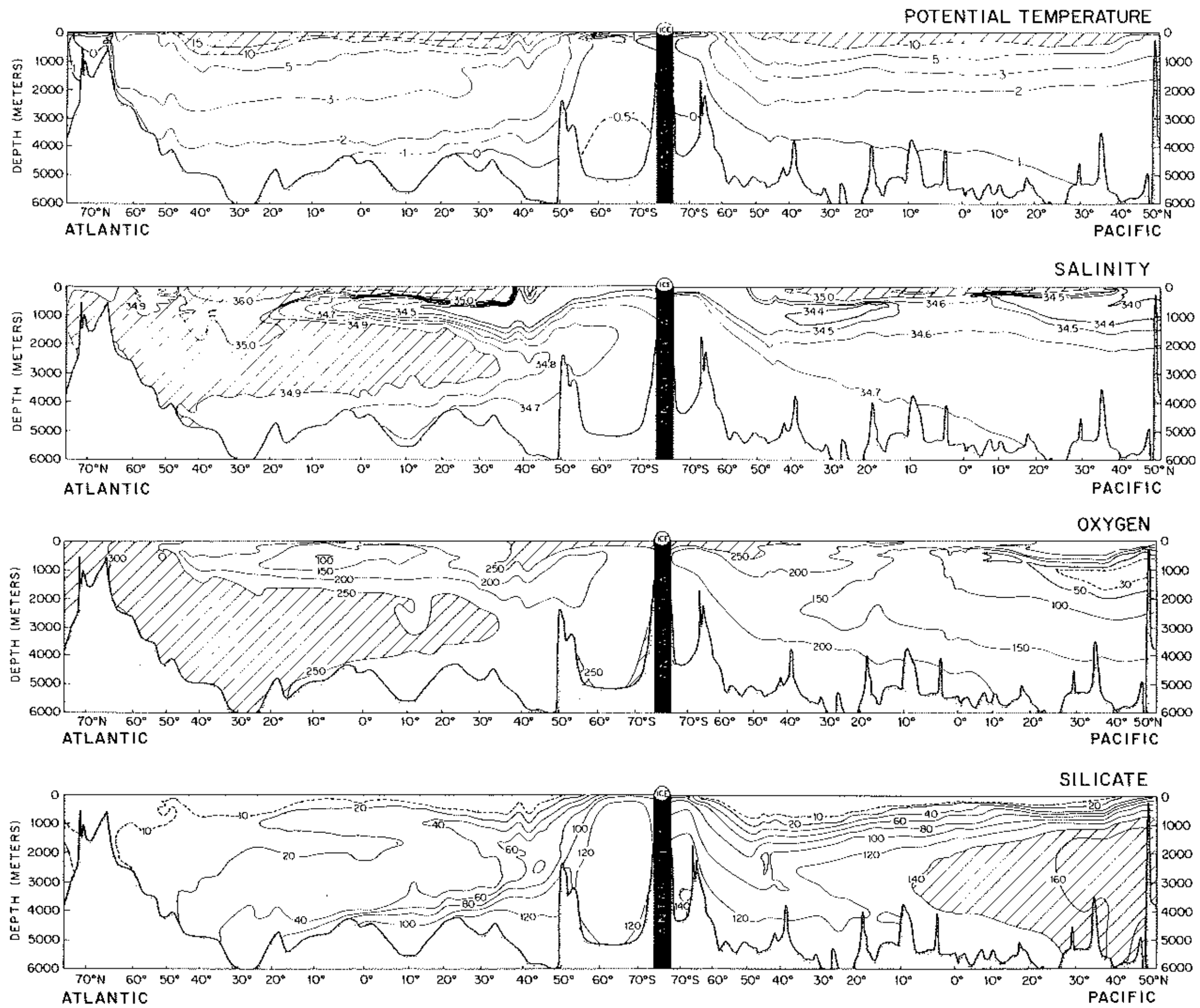


***Fig. 1:** Meridional sections of temperature, salinity, oxygen and silicate across the Atlantic and Pacific oceans. The vertical bar at the center is Antarctica. These sections are constructed from Geosec data (1972 in the Atlantic, 1976 in the Pacific).*

Temperatures generally decrease with increasing depth. This trend is strongest within the upper km of the ocean, in a feature called the thermocline. The 10° C isotherm may be taken as the base of the thermocline. Antarctic Intermediate Water (AAIW) is the low salinity layer near 1000 m depth in the Southern Hemisphere. This feature emanates from the surface waters of the circumantarctic belt. In the Atlantic Ocean it spreads well into the Northern Hemisphere, and traces can be found as far north as 50° N. In the North Pacific a salinity minimum layer at similar depths is derived from the northwestern corner of the ocean, from the Sea of Okhotsk.

The northern North Atlantic lacks a subsurface salinity minimum; it is replaced by high salinity water with high oxygen and low silicate concentrations, sinking from the sea surface to spread southward near 3000 m depth. This is the North Atlantic Deep Water (NADW).

The deep water of the North Pacific is opposite to that of the North Atlantic; it is lower in salinity, lower in oxygen, and higher in silicate— these latter two characteristics are the results of long isolation from the atmosphere. This North Pacific Deep Water (NPDW), formed by the slow process of vertical mixing, spreads to the south near 3000 m depth, and via the Antarctic Circumpolar Current, into the Atlantic Ocean.

Along the seafloor is the very cold, high silicate Antarctic Bottom Water (AABW), derived from the margins of Antarctica, cooled sharply by the cold atmosphere. This dense water slips down the continental slope to the seafloor, under the Antarctic Circumpolar Current (ACC), and into the world ocean.

Hemisphere upwelling deep water, long removed from direct contact with the atmosphere, is quickly converted to cold denser water, and re-enters the deep and bottom layers of the ocean.

These water masses (each with their own characteristic properties) spread throughout the ocean and force a slow but steady upwelling of the "resident" deep ocean water. This resident deep water is composed of older (in terms of time since exposure to the atmosphere) water that has been modified by vertical mixing processes, by organic material descending from the sea surface, and by contact with the seafloor sediments. Eventually, the upwelled water migrates back to the initial sinking regions, to complete a thermohaline circulation cell often referred to (for both the Atlantic-driven cell and for the Southern-Ocean-induced cell) as a "conveyor belt."

North Atlantic Deep Water

Thus one mass, called the North Atlantic Deep Water (NADW), forms in the northern North Atlantic, and flows southward from there into the southern Atlantic, and eventually spreads via the circumantarctic deep ocean belt into the Indian and Pacific Oceans. Total NADW production is estimated as 15 to 20 million cu m/sec. By way of comparison, the Amazon River outflow is only 0.18 million cu m/sec. The NADW production rate would replace all of the deep water of the global ocean in about 2000 years.

NADW has a variety of components; each differs in its thermohaline characteristics, though they all share a common attribute of being relatively warm and salty compared to the average deep waters of the world ocean. All are drawn from roughly the upper kilometer of the ocean, including the thermohaline and intermediate strata. The largest component of NADW, with a formation rate of 13 million cu m/sec, is derived from the Greenland and Norwegian seas. As relatively warm-salty water flows with the Norwegian Current into the Greenland and Norwegian seas, it cools, becomes denser, and sinks into the deep basin north of a submarine ridge that spans the distance from Greenland to Scotland. This water slips through passages across the ridge and cascades into the deep ocean to the south. Smaller contributions are made in the Labrador Sea, and a particularly warm salty constituent is derived from the Mediterranean Sea outflow. Though small in volume flux, the Mediterranean water infuses great amounts of salt into the NADW, branding it with its telltale salinity maximum.

As surface water sinks to form NADW and is exported to the South Atlantic and to the other oceans, a compensating amount of thermocline and intermediate water masses from the world ocean is drawn towards the North Atlantic. As these waters are of higher temperature and (initially) on average lower salinity than NADW, the resultant thermohaline cell, as viewed in the meridional vertical plane, strongly influences the Atlantic's meridional heat and salinity fluxes. It is estimated that if NADW formation were to cease, the compensating flow of warm water into the North Atlantic would diminish, and the surface water of the northern North Atlantic would be as much as 6° C cooler. It is the warm water drawn into this region by NADW formation, not the Gulf Stream alone, that supplies the heat and moisture to the atmosphere that moderates the climates of northern Europe. The cooled surface water, still relatively salty, sinks to continue the NADW formation process.

As NADW spreads across the South Atlantic, it reaches the Antarctic Circumpolar Current (ACC), a strong clockwise current that encircles Antarctica. The ACC is the primary pathway where the three major oceans can exchange large amounts of water, making the thermohaline circulation cells into a global system.

Compensating Return Flow

The export of NADW from the Atlantic Ocean via the circumpolar belt of the Southern Ocean requires a compensating import of upper-layer water into the Atlantic. There are two possible sources for this: Pacific inflow via the Drake Passage of cool low-salinity Antarctic Intermediate Water, or AAIW), and a warm salty inflow from the Indian Ocean's thermocline, which makes its way to the Atlantic around the southern rim of Africa.

Most of the return flow is in the form of AAIW via the Drake Passage. However, recent observations indicate a surprising phenomenon — the Drake Passage AAIW water, rather than flowing directly to the north along the western margins of the Atlantic, may cross the South Atlantic, entering the southwest Indian Ocean, where its mixes with saltier water before flowing back to the southeast Atlantic Ocean. This injection of AAIW into the Atlantic, spiced with extra salt from the Indian Ocean, may be the chief means of balancing the Atlantic export of NADW and of transferring Indian Ocean salt into the Atlantic. The salinity enhancement of the AAIW may boost North Atlantic upper-layer salinity by 0.2 parts per thousand, enough to encourage more NADW formation. Thermocline water from the Indian Ocean also enters the Atlantic. Most of this water seems to return to the Indian Ocean within the upper layer of the ocean (**Fig. 3**), rather than flow into the North Atlantic to feed NADW formation. However, due to mixing processes it may leave behind in the Atlantic some of its excess salt, which eventually spreads to the north, also acting to boost Atlantic salinity. Recent work indicates variability in the Indian Ocean inflow—an energetic influx of Indian Ocean water into the Atlantic occurred during the 1980s. This suggests that the Atlantic's salinity and susceptibility to NADW formation may be influenced by the variability in the Indian Ocean factor, a concept that is being researched.

Tracking of the Indian and Pacific components from the South Atlantic to the northern North Atlantic cannot be done by simple water mass indicators. Mixing of thermocline and intermediate water, particularly in the tropical regions, and the alterations by ocean-atmosphere interaction, obscures these traits. Tracking may be possible by understanding the sequence of regional oceanographic conditions encountered en route. Eventually, with the help of a global-scale numerical model that faithfully portrays the thermohaline circulation, it might be possible to trace the full structure and vigor of the thermohaline circulation.

Stopping the NADW Conveyor Belt

The atmosphere transfers water vapor from the Atlantic to the Pacific drainage basins at a rate of about 0.3 million cu m/sec, mostly across the Isthmus of Panama. This atmospheric flux supports a salty Atlantic and sustains NADW formation while prohibiting deep reaching convection in the North Pacific. However, NADW production draws warm water to high northern latitudes where it enhances evaporation. In other words, the process of NADW formation strongly influences the freshwater balance of the Atlantic, providing excess water vapor for the atmosphere to carry into the Pacific. As the atmosphere

Fig. 2

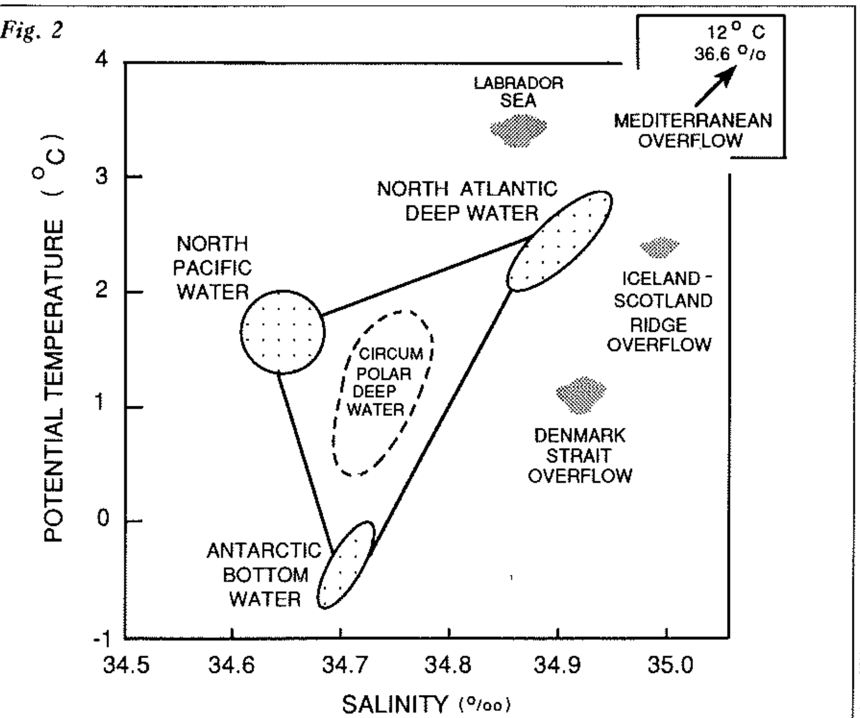


Fig. 2: Potential temperature vs salinity relationship of the three major water masses, or "end-members," that compete for dominance of the deep ocean interior. The North Atlantic Deep Water (NADW) is derived from three sources: there is the warm/salty outflow from the Mediterranean Sea, the cold lower salinity formed within the Labrador Sea, with the bulk of the North Atlantic Deep Water derived from the overflow of dense water across the ridge from Greenland to Scotland. The overflow from east of Iceland is somewhat warmer and saltier than the overflow through the Denmark Straits west of Iceland.

The North Pacific Deep Water (NPDW) is about the same temperature as North Atlantic Deep Water, but is much lower in salinity. Antarctic Bottom Water (AABW) is the coldest of the trio. Circumpolar Deep Water (CDW) forms a voluminous water mass, a blend of the three chief end-members, dominated by AABW. As the formation rates of the three primary water masses vary with time, presumably the position of the CDW in the temperature/salinity diagram shifts, and with it there is a change in heat, water and perhaps carbon storage in the ocean.

process may be a response to NADW it is difficult to sort out cause and effect. Might the ocean circulation in the South Atlantic with its links to the Pacific and Indian Oceans "condition" the Atlantic for NADW formation and perhaps initiate variability of NADW production? Might the distribution of low salinity surface water from the Arctic govern NADW production rates and recipes?

For example, the Atlantic meridional heat and freshwater fluxes are very sensitive to the ratio of these two routes. The Atlantic would become saltier if the Agulhas route is favored, fresher if the Drake Passage input is favored. The more salt that can be obtained from the evaporative Indian Ocean the less is the burden on the atmosphere over the Atlantic to remove the water vapor as required to keep the northern Atlantic salty—and thus the thermohaline conveyor belt in motion.

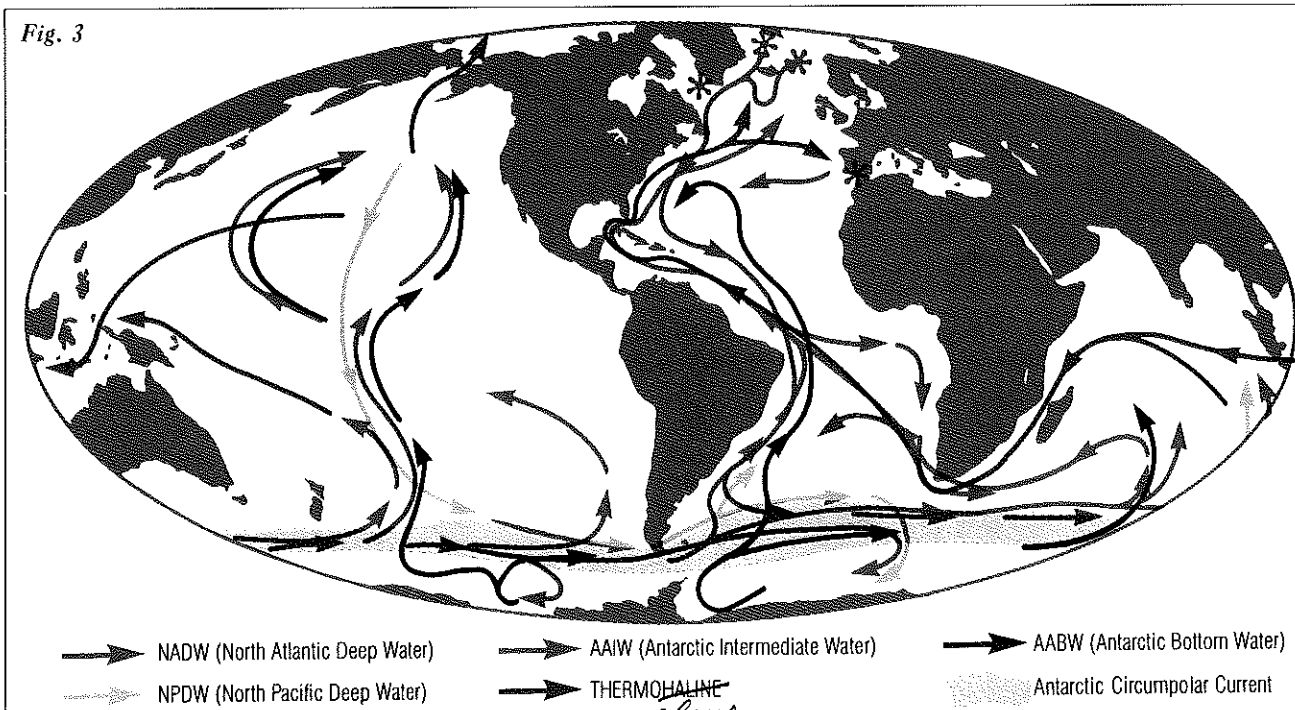
However, the presence of the "right" (warm and salty) surface water makes the NADW formation difficult to stop; it is in effect a positive feedback. But one possible mechanism that might stop NADW formation would be the capping of the surface ocean with low-salinity water. A large source of such water is the Arctic Ocean. River runoff, excess precipitations, and a flow of low salinity surface water from the Pacific (via the Bering Strait), maintains a stable cap of very low salinity water. The intense stratification reduces heat flux from the warmer deep water of the Arctic into the surface layer, thus allowing a persistent lid of sea ice to reside in the Arctic Ocean.

Yet during the last century there have been at least two episodes of low-salinity water outbreaks (referred to as the Great Salinity Anomaly). These large pools, also associated with anomalous expansion of sea-ice cover in the regions of the Iceland and Labrador seas, migrated

Fig. 3: Map view of the two major thermohaline circulation cells or "conveyor belts," driven by the exchange of heat and water between ocean and atmosphere. A relatively warm and saltier cell is induced by formation of North Atlantic Deep water (NADW) at a number of sites within the North Atlantic. As NADW spreads into the rest of the ocean, slow upwelling and return flow to the North Atlantic completes the circulation cell.

Formation of Antarctic Bottom water (AABW) at various sites within the Southern Ocean drives a colder and fresher thermohaline cell. As AABW spreads along the seafloor to the north, it displaces resident bottom water, which returns to the AABW formation sites. A combination of NADW and AABW flows into the North Pacific, where a relatively warm but low salinity water mass, North Pacific Deep Water (NPDW) forms and spreads southward to close both the NADW and the AABW thermohaline cells.

Fig. 3



Ana Maria Alvarez Janet Kroboth-Weber

about the Northern Atlantic, and reduced NADW formation rates. Thus the NADW recipe may be quite variable as the production rate of each component changes, at a variety of time scales, and for a variety of reasons.

The stability of the formation rates of the varied forms of NADW is one of the major concerns in evaluating future climate change. Changes in the 20th century related to the Great Salinity Anomaly are small compared to the suspected changes in NADW formation rates during the swings between glacial to inter-glacial periods. During the disintegration of the ice sheets from the last glacial epoch, a series of events of sudden injections of melt water into the North Atlantic induced low salinity surface water and dramatic reduction of NADW formation. These events indicate that the ocean can change from a condition comparable to the modern vigorous NADW formation state to a low NADW production state characteristic of the last glacial maximum in just 300 years or less. Might we expect rapid changes in NADW during a Greenhouse-induced global warming? If so, will NADW production go up or down? Answers to these questions are crucial if we are to gain a predictive ability of the global climate system.

North Pacific Deep Water

It is often said that no deep water forms in the North Pacific. This is true if we refer to deep water formation by convection, like NADW. The North Pacific surface water is too low in salinity to permit deep convection. Nonetheless, a unique deep water mass is found in the North Pacific, too low in salinity to be a simple mix of NADW and Antarctic Bottom Water (AABW). (AABW forms at the Antarctic continental margin as the very cold Antarctic air mass spreads out over the adjacent ocean. The ocean surface freezes, resulting in a cold dense surface water that sinks to the seafloor, and slips below ther ACC into the world ocean, well into the Northern hemisphere.) With its low oxygen and high nutrient concentrations, it is the antithesis of NADW. This North Pacific Deep water (NPDW) spreads southward and eventually enters the Circumpolar Deep Water (CDW) and can clearly be observed spreading well into the Atlantic Ocean. Presumably it leaves the Atlantic after mixing with NADW.

How does NPDW form? It may be a consequence of vertical mixing, which carries into the deep ocean the properties of the low salinity North Pacific thermocline and intermediate water masses. Deep and bottom water, a blend of NADW and AABW (mostly the latter), flows into the North Pacific, mainly through the Samoan Passage, where it slowly upwells and mixes with low salinity water brought down by a mixing process. Because of the immensity of the North Pacific, just a small degree of vertical mixing can produce great quantities of NPDW — reasonable estimates of the intensity of vertical mixing suggest that NPDW formation can easily be maintained at a level of 10 million cu m/sec.

NPDW does not represent injection into the deep ocean of 10 million cu m/sec of surface water, but rather the modification of existing deep water by slow process of downward diffusion of upper-layer characteristics. As such, it does not "ventilate" the deep ocean in terms of atmospheric gases, but rather it alters the heat and freshwater storage of the deep ocean; it also allows accumulation of carbon, and so has the potential to influence global climate.

While most of the deep water entering the North Pacific eventually exits as NPDW, some deep water of the Pacific must reach the surface layer. This water, diluted by the excess precipitation of the North Pacific, enters the Arctic through the Bering Strait (at about 1 million cu m/sec), and the Indian Ocean through the Indonesian Seas (at approximately 5 million cu m/sec). These flows are an important part of the interocean water budgets, and changes in their strength would influence the global climate system, though the extent and even direction of such change is not known.

While the impact on the complex global climate system of a change in fresher water outflow from the Pacific is not known, a speculation is offered by the following thought experiment. As freshwater outflow from the Pacific via the Bering Straits and Indonesian Seas decreases, the Arctic and Indian Ocean would become saltier. As these waters spread into the Atlantic Ocean, it too will eventually become saltier and more susceptible to NADW formation. The Pacific would become fresher as it accumulates fresh water. Lowering of the Pacific's surface layer salinity increases the stability of the water column and should lower vertical mixing, inducing less NPDW formation. End result: the deep waters of the world ocean become enriched in

NADW, saltier and warmer. Of course, the climate is in an interlocked series of forces, so changing the Pacific outflow would be associated with other changes that may induce other responses. In general, we like to think of our system as difficult to change — that somehow the network of negative feedbacks is difficult to overcome. However, if a system is kicked hard enough, it might suddenly shift into another stable system.

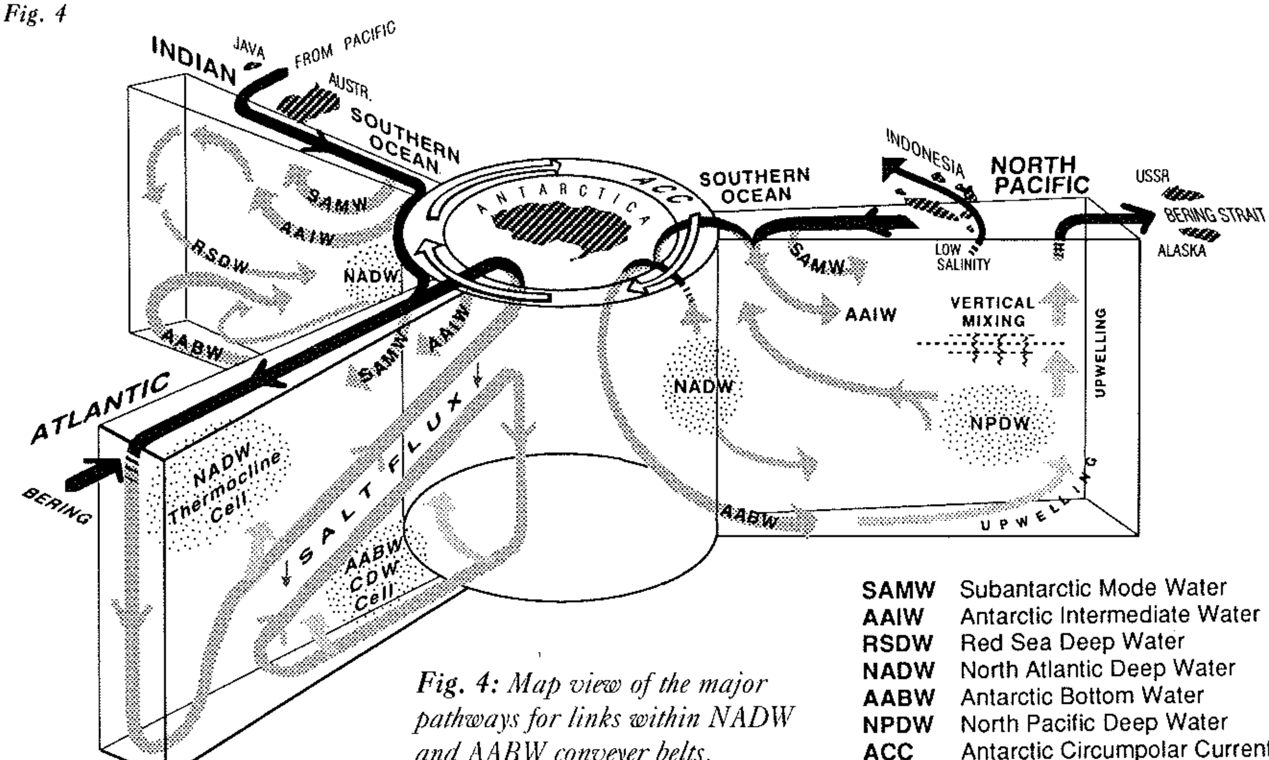
The Southern Ocean

Another thermohaline cell is initiated along the polar edge of the Southern Ocean (i.e. the body of water approximately south of 30° S that connects the three major oceans). Antarctic Bottom Water (AABW) forms at a vigorous rate of 20 to 30 million cu m/sec; this is compensated by the upwelling of deep water known as Circumpolar Deep Water (CDW). As AABW spreads into the world ocean, it slowly upwells and is modified by mixing with less dense water, including NADW and NPDW, to flow back to the Southern Ocean as CDW. Upwelling of relatively warm deep water (about 1° C) is converted into the cold (near the freezing point of sea water — 1.85° C) Southern Ocean surface water, where it resides for only one to two years, before eventually contributing to the formation of AABW.

The sea ice cover of the Southern Ocean acts to decouple the ocean from the atmosphere; the insulating blanket of sea ice protects the ocean from the cold atmosphere, limiting cooling. The extreme seasonality and the rapid spring melting of the Southern Ocean sea ice cover suggest that the heat carried into the surface layer by rapid deep water upwelling is the key to understanding the Southern Ocean sea ice budget. Thus the buildup of heat within the mixed layer under the winter ice cover induces melting even before the Spring atmospheric radiation warms sufficiently to melt the sea ice directly from above. Ocean heat flux also limits sea ice thickness during the winter, to less than a meter, in contrast to the three-meter ice of the more stable Arctic Ocean.

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

Fig. 4



SAMW	Subantarctic Mode Water
AAIW	Antarctic Intermediate Water
RSDW	Red Sea Deep Water
NADW	North Atlantic Deep Water
AABW	Antarctic Bottom Water
NPDW	North Pacific Deep Water
ACC	Antarctic Circumpolar Current

Fig. 4: Map view of the major pathways for links within NADW and AABW conveyer belts.

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Variations in vertical heat flux are expected to yield interannual changes in ice cover extent and seasonality, and in isolated regions, ice-free conditions (polynyas) even in the dead of winter.

There are indications that during the last glacial stage the sea ice around Antarctica was a bit more extensive and did not exhibit the large amount of seasonality characteristic today. Following the arguments stated above, this suggests that during the glacial period the overturning of the ocean below the sea ice cover was less vigorous than it is today. This implies reduced production of AABW. Coupled to the indications that NADW production was also diminished during the glacial period, one concludes that deep ocean ventilation was less efficient during the glacial periods.

Conclusions

The North Atlantic, North Pacific and Southern Ocean each forms, by its own distinctive process, a unique deep and bottom water mass. NADW forming at a rate of 15 to 20 million cu m/sec and AABW forming at a rate of 20 to 30 million cu m/sec, would replace the water of the deep ocean in somewhat less than 1000 years. These water masses differ in terms of temperature, salinity, and nutrients. The trio of water masses battle for dominance of the deep ocean; the ratio of their importance changes with time and with it so does the atmosphere, temperature and chemistry. The Southern Ocean is a cold ocean, and tends to cool the deep and bottom waters of the modern world ocean, whereas NADW and NPDW tend to warm the deep ocean. NADW tends to make the deep ocean saltier, the NPDW fresher. The chilling effect of the Southern Ocean developed as Antarctica became tectonically isolated and grew a persistent glacial ice sheet about 14 million years ago. Before that time the deep oceans were significantly warmer.

What if all of the two major convective deep water-mass sources were turned off? The only communication the deep ocean would then have with the atmosphere would be through the slow process of downward diffusion (as is presently the situation for the North Pacific) and the rain of organic particles. The ocean could either freshen or become saltier, depending on the dominant surface salinity. But since the warm thermocline dominates the ocean's areal extent, the net result would be a downward flux of heat, and the ocean would warm up. The lack of ventilation of oxygenated water by convection would lead to an ocean with lower deep-water oxygen and higher carbon storage. The net effect of the lack of convective deep-water formation may be to forestall global greenhouse warming, though because of the difference in the time scales these processes may not neatly compensate each other. Eventually the warming and density decrease of the deep ocean would enable dense surface water somewhere in the world ocean to re-activate convective ventilation.

One should not think of the two primary thermohaline cells as separate entities. There is exchange of water and ocean properties across the convoluted interface between the cells. The vigor of this exchange may determine the relative size of the cells. For example, the injection of salt from the NADW cell is most likely instrumental in maintaining the vigor of the AABW cell, which would otherwise become fresher, eventually unable to force convection to the seafloor.

In terms of global warming, the capacity of water to absorb heat is high relative to the atmosphere; this means that as the ocean warms less heat is available to warm the atmosphere. If global warming causes the deep ocean to warm, say by a decrease of cold AABW production or an increase in the production rates of NADW (particularly of warmer components—from the Mediterranean and Labrador Seas or NPDW—though the latter, being driven by vertical mixing processes, may be subject to less variability), this would forestall global warming. Of course, warming of the ocean increases its volume due to thermal expansion. For example, a 1° C warming of the 4000 m water column increases sea level by 0.6 m.

Changes in the ratio of importance of the three water masses could also affect the carbon dioxide budget. Of the estimated anthropogenic input of carbon dioxide only 60% remains in the atmosphere, the rest is believed to reside in the ocean, in which case the ocean takes up excess carbon dioxide at a rate of two billion tonnes per year. The ocean carbon storage is about 50 times that of the atmosphere. Small changes in the ocean carbon storage lead to large changes in the carbon burden of the atmosphere.