A Climatic Freshening of the Deep Atlantic North of 50°N over the Past 20 Years

Abstract. *Observations made in summer 1981 show a significant and widespread decrease in salinity, averaging 0.02 per mil, in deep waters of the subpolar North Atlantic over the past two decades. This implies a relatively rapid response of deep water formation to climatic perturbation.*

The large-scale, predominantly thermohaline circulation of the oceans fulfills at least two critical roles in the maintenance of global climate. First, it transports a significant portion of the meridional heat flux. Second, it provides a long-term geochemical buffering of the atmospheric CO_2 inventory. Inasmuch as both actions are attributes of a global ocean circulation, which is driven in turn by climate, it is important to document and understand the magnitude, mechanics, and temporal variability of these processes.

The Transient Tracers in the Ocean (TTO) program was initiated as a multiyear effort by several institutions to study ocean transport processes by measuring the distribution of man-made (and hence transient) trace substances as they penetrate into the oceans. During the first major field effort of the TTO program in 1981, we carried out a seven-leg cruise occupying 250 geochemical stations covering the North Atlantic from 15°N to Spitsbergen (Fig. 1). Samples were taken for the measurement of tracers produced by atmospheric nuclear weapons testing, such as tritium, its daughter ^{3}He, ^{14}C, and tracers that are by-products of industrial activities, such as ^{85}Kr and trichlorofluoromethane. In addition, precise measurements of temperature, salinity, dissolved oxygen, and nutrient salts were made.

We report here preliminary hydrographic results from the three northernmost legs, which show a systematically different temperature-salinity relation throughout a large portion of the water volume north of 50°N than that observed previously. This difference must be the result of a climatic variation in atmospheric and cryospheric conditions that manifests itself as a net change in the temperature and salinity of components of northern origin that supply North Atlantic Deep Water.

Figure 2A is a scatter plot of potential temperature versus salinity (Θ-*s*) for samples deeper than 100 m from a composite section of TTO stations (bold line in Fig. 1) and for comparable samples taken by the *Erika Dan* in 1962 along 59°N (*1*). Although the sections do not perfectly overlap, an examination of these and other hydrographic data shows that the spatial homogeneity of the Θ-*s* relation justifies such a comparison. There is a systematic offset in salinity between the cruises equivalent to about 0.02 per mil (at constant temperature), a difference well outside both random analytical errors and systematic uncertainties associated with the conductivity standard (*2*).

Since outside the regions of convective overturning and away from boundaries there is a strong tendency for tracers to spread along surfaces of constant density, it is more appropriate to compare the two data sets in density-salinity space. Figure 2B shows the salinity data plotted as a function of σ_2, the potential density anomaly in per mil referenced adiabatically to a pressure of 2000 decibars. The data are presented as averaged values over a σ_2 range increasing in steps of 0.02 per mil, and the bars represent the root-mean-square range of the data points over that interval.

The salinity difference between the two cruises is again seen for the denser water masses, and becomes particularly apparent for $\sigma_2 > 37.10$ per mil, where salinity differences are on the order of 0.02 per mil. West of the Mid-Atlantic Ridge this density range includes a large component of waters that overflow into the northwest Atlantic through the Denmark Strait [Denmark Strait Overflow Water (DSOW)]. At somewhat lower densities, $36.9 < \sigma_2 < 37.00$ ($3.0 < T < 3.5$°C), the TTO salinities are close to or exceed the corresponding earlier values. This density range represents the boundary between the densest Labrador Sea Water (LSW) derivatives and Iceland-Scotland Overflow Water (ISOW). At $\sigma_2 \simeq 36.85$, the TTO data are about 0.05 per mil fresher than the International Geophysical Year data, a feature in the TTO data associated with a large, fresh quasihomogeneous water mass seen in the Labrador basin. This "new breed" of LSW is colder, fresher, and lower in density than the "classical" LSW. We believe that the similar salinities of the two data sets over the density range 36.9 to 37.0 per mil results from the fact that surface climatic conditions are inappropriate for the formation of the denser, classical form of LSW, and that new, fresher, and less dense waters are formed, leaving the older LSW relatively undisturbed. This provides new saline ISOW in the northeast Atlantic with a greater impact on the Θ-*s* correlation, thus accounting for the slightly greater 1981 salinities at $\sigma_2 \simeq 36.98$ in Fig. 2B.

The observed overall changes in salinity are of the same sign as changes over the past three decades in surface salinity of the North Atlantic reflected in weather ship data (*3*). The surface water trends observed spanned a decrease in salinity on the order of 0.1 per mil—perhaps more in the Labrador Sea area (*4*), where the surface salinity balance is dominated by cold, low-salinity water from the Arctic. The surface salinity trend is coupled to the LSW trend: Lazier (*4*) reported an overall decrease of 0.04 per mil in average salinity for the top 1500 m from 1964 to 1973. The convective formation of LSW in the region appears to be impeded by the low-salinity "cap" formation (*4*), but in a climatic sense the waters that are formed during such low-salinity periods bear the stamp of the low-salinity trends.

The other observed deepwater

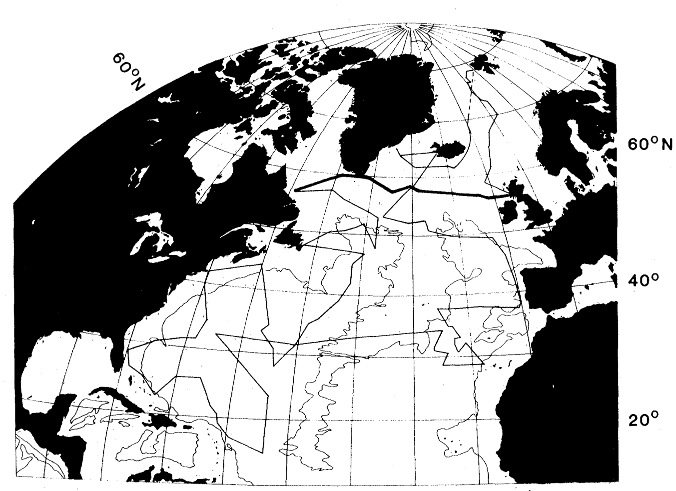

Fig. 1. TTO cruise track during the summer of 1981. The bold line, a composite of sections from legs 4 and 6, corresponds approximately to the *Erika Dan* section (59°15′N).

changes must also reflect changes in surface or near-surface salinities in the various formation regions. The DSOW is formed at or near the sea surface in winter in the Greenland and Iceland seas from a mixture of surface waters of Atlantic and Arctic origin (*5*). Swift *et al.* (*5*) suggested that, because the residence time of this water type in the region is less than a decade, variations in its characteristics may be expected.

If the salinity decreased in only one of the North Atlantic Deep Water components, this would require prior variations in only the region of formation of that water mass. But the broad scope of this deep freshening suggests a linkage with a larger scale climatic change. There are many possible driving agents for the reduction in surface salinity that must precede the deep changes: changes in precipitation, in evaporation, in the rates or characteristics of the advective supply of the low- or high-salinity components of the surface waters, and in the patterns of surface circulation.

Climate-scale atmospheric changes have been recorded over the time span represented by the new and old measurements. For example, Jones *et al.* (*6*) reported that northern hemispheric temperatures have been decreasing since the 1940's to the mid-1960's and have tended to rise since then. This trend has been even more pronounced for the Arctic region (*7*), the source of the low-salinity water carried southward by the East Greenland Current. With a freshwater residence time of about 10 years in the Arctic Ocean (*8*, *9*), cryospheric variations induced by a long-term cooling trend might not have influenced changes in the deep North Atlantic until after the basin-scale surveys there in the late 1950's and early 1960's. We note, however, that there is no clear cause-and-effect relation between only long-term cooling and decreased salinity or increased outflow from the Arctic Ocean. To some extent winter ice formation in the thin, highly stratified layer above the Arctic Ocean halocline does represent a rectification process, producing and maintaining a low-salinity surface layer (*10*). An overall increase in ice formation accompanied by convection penetrating the halocline might possibly decrease the salinity above the halocline, if other major factors remain constant, such as the supply of freshwater to the Arctic basins. Long-term trends for Arctic ice cover (*11*) show, however, that one key region supplying the East Greenland Current may experience an increase in ice cover while another decreases, but evidence of consistent decadal increases over the Eurasian Arctic is weak. In fact, some particularly warm years have been associated with an enhanced outflow of ice from the Arctic in the Greenland Sea (*12*).

Etkins and Epstein (*13*) proposed that global sea level has increased over the past century as a consequence of polar ice cap melting. This has since been challenged (*14*), and it is not possible, on the basis of available data, to directly establish such a reduction in ice volume. It is perhaps oversimplistic to propose a simple causal relation between ice cap melting and deepwater salinity without carefully assessing the oceanographic processes involved.

Indeed, it is not even necessary to invoke a change in the salinity or volume of the low-salinity Arctic Ocean outflow to have this water influence the deep water freshening. For the most part, the low-salinity arctic outflows—the East Greenland, West Greenland, and Labrador currents—are confined to regions near the continental boundaries. For example, Swift and Aagaard (*15*) noted that the East Greenland Current normally has little noticeable impact on the regions of dense water mass formation north of Iceland, where DSOW is formed. A slight net trend toward an increased flux of low-salinity water away from the coast of Greenland (or Canada) into the gyres where the dense waters sink in winter could easily cause a long-term reduction in salinity. We note that the "polar current" nature of the waters north of Iceland in the years 1964 to

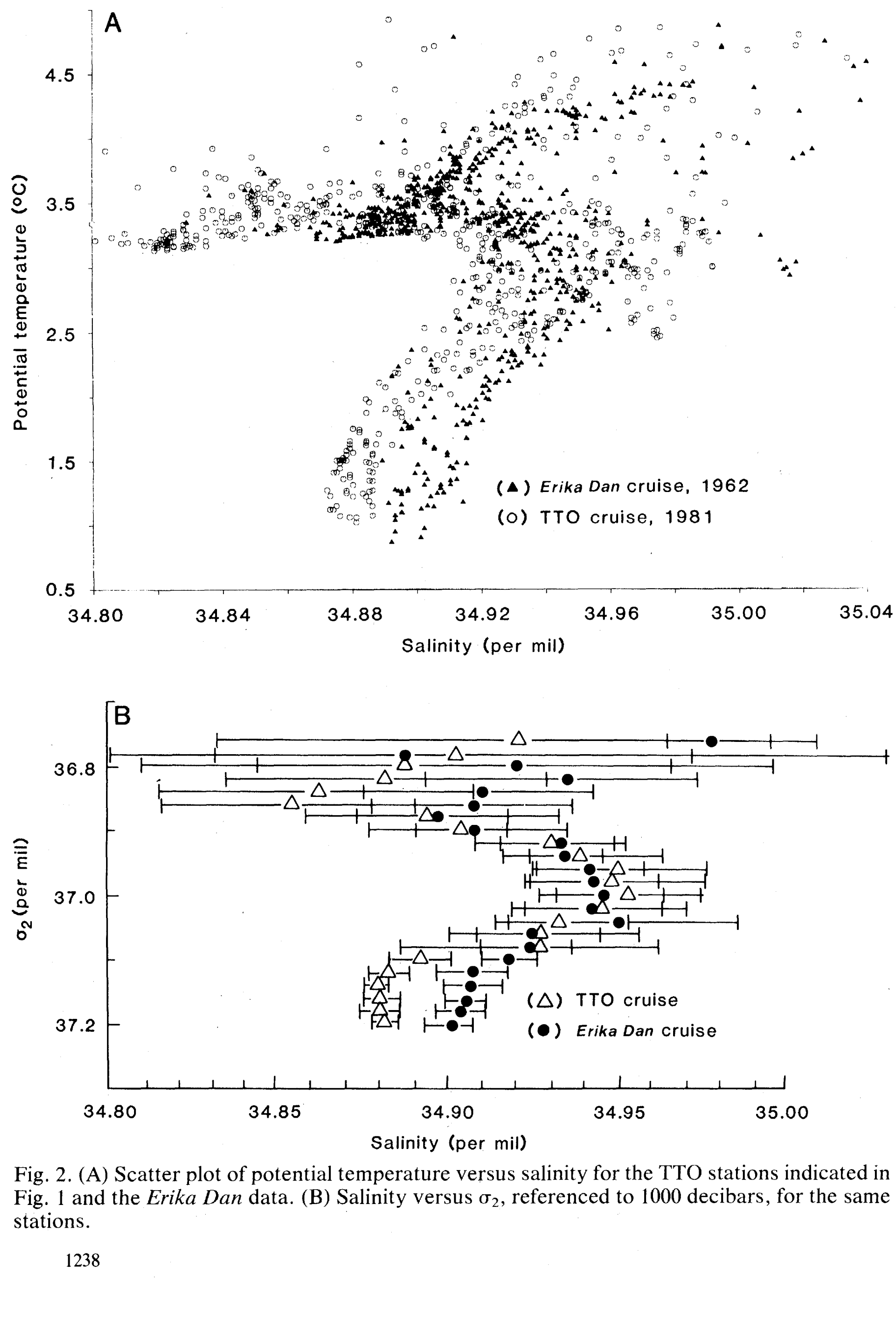

Fig. 2. (A) Scatter plot of potential temperature versus salinity for the TTO stations indicated in Fig. 1 and the *Erika Dan* data. (B) Salinity versus σ_2, referenced to 1000 decibars, for the same stations.

1974, compared to the warmer, more saline waters of the period 1948 to 1963 (*16*), represents in the multiyear mean just such a change in circulation. Similar trends were observed over the same period in the Labrador Sea (*4*). Even though such changes may be subject to short-term reversal (*4, 16*), their long-term effect shows clearly in the salinity of deep waters away from the formation regions, which in a sense are smoothed or averaged during spreading of the new dense components from their sources. In the surface samples from the three northernmost TTO legs, salinities were lower than 1955–1965 values almost everywhere along the track. In the Labrador Sea the 1981 surface salinities were more than 1 per mil below the 1962 values. [Examination of seasonal trends in surface salinity at ocean weather ships (*3, 4*) reveals that as much as half of the observed difference may result from seasonal variations in surface salinity.] It is also noteworthy that the TTO cruise track north of Iceland was changed due to an unexpectedly large ice extent.

In summary, between 1962 and 1981 there was a significant climatic freshening in North Atlantic Deep Waters north of 50°N, and an even larger freshening of shallower waters. We suggest that investigators examine ties of the low-salinity Arctic outflows to long-term atmospheric cooling, or examine shifts in the patterns of circulation that might increase the transport of this low-salinity water away from the boundaries and into the regions of water mass formation. The significant hydrographic response to modest short-term climatic forcing points to the importance of quantifying and understanding the nature of water mass formation in these regions.

P. G. BREWER
Woods Hole Oceanographic Institution, Woods Hole, Massachusetts 02543
W. S. BROECKER
Lamont-Doherty Geological Observatory, Palisades, New York 10964
W. J. JENKINS, P. B. RHINES
Woods Hole Oceanographic Institution
C. G. ROOTH
Rosenstiel School of Marine and Atmospheric Science, Miami, Florida 33149
J. H. SWIFT
Scripps Institution of Oceanography, La Jolla, California 92093
T. TAKAHASHI
Lamont-Doherty Geological Observatory
R. T. WILLIAMS
Scripps Institution of Oceanography

References and Notes

1. L. V. Worthington and W. R. Wright, *North Atlantic Ocean Atlas* (Woods Hole Oceanographic Institution, Woods Hole, Mass., 1970).
2. A. Mantyla, *Deep-Sea Res.* **27A**, 837 (1980).
3. A. M. Taylor and J. A. Stephens, *Oceanol. Acta* **3**, 421 (1980); J. M. Colebrook and A. H. Taylor, *Deep-Sea Res.* **26A**, 825 (1979).
4. J. R. N. Lazier, *Atmos. Ocean* **18**, 227 (1980).
5. J. H. Swift, K. Aagaard, S.-A. Malmberg, *Deep-Sea Res.* **27A**, 29 (1980).
6. P. D. Jones, T. M. L. Wigley, P. M. Kelley, *Mon. Weather Rev.* **110**, 59 (1982).
7. P. M. Kelley, P. D. Jones, C. B. Sear, B. S. G. Cherry, R. K. Tavakol, *ibid.*, p. 71.
8. K. Aagaard and L. K. Coachman, *Eos* **56**, 484 (1975).
9. H. G. Ostlund, *J. Geophys. Res.* **87** (C3), 1035 (1982).
10. K. Aagaard, L. K. Coachman, E. Carmack, *Deep-Sea Res.* **28A**, 529 (1981).
11. J. E. Walsh and C. M. Johnson, *J. Phys. Oceanogr.* **9**, 580 (1979).
12. L. Koch, *Medd. Groenl.* **130**, 3 (1945).
13. R. Etkins and E. S. Epstein, *Science* **215**, 287 (1982).
14. A. Robock, *ibid.* **219**, 996 (1983).
15. J. H. Swift and K. Aagaard, *Deep-Sea Res.* **28A**, 1107 (1981).
16. R. R. Dickson, H. H. Lamb, S.-A. Malmberg, J. M. Colebrook, *Nature (London)* **256**, 479 (1975).
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Identification of the Receptor for Antigen and Major Histocompatibility Complex on Human Inducer T Lymphocytes

Abstract. *Human T cell clones and monoclonal antibodies directed at their surface structures were used to define the receptor for the antigen and major histocompatibility complex on inducer T lymphocytes. The results indicated that the receptor is a single complex consisting of the monomorphic T3 molecule with a molecular weight of 20,000 to 25,000 and a clonotypic disulfide linked heterodimer Ti with a molecular weight of 90,000. Sepharose-bound monoclonal antibodies (anti-Ti_4 or anti-T3) to the receptor could activate clonal proliferation and inducer function for B cell immunoglobulin secretion and thus substitute for the appropriate combination of major histocompatibility complex gene product and specific antigen.*

The initiation of immune responses is critically dependent on the presence of a mature subpopulation of peripheral T lymphocytes termed "inducer T cells." These cells express their individual regulatory programs only upon activation by a specific set of signals consisting of foreign- or autoantigen and autologous class II gene product of the major histocompatibility complex (MHC) (*1*). Since interactions among T cells, antigen, and MHC gene products occur in a highly selective fashion, it has been postulated that each inducer T lymphocyte must possess one or more clonally unique surface recognition receptors. However, there is considerable controversy on whether the T cell receptor reacts with antigen in association with MHC (altered self theory or single receptor model) or antigen and MHC independently (dual recognition theory or two receptor model).

Characterization of inducer T cell receptors would be valuable in understanding the molecular basis for the pivotal roles which these cells play in T-T, T-B, and T-macrophage interactions within the lymphoid system (*2*) as well as their inductive effects on hematopoietic stem cells, fibroblasts, osteoclasts, and other cell types extrinsic to the lymphoid system (*3*). Such information might also provide insights into inducer cell defects in acquired and congenital immunodeficiency states on the one hand or inducer cell hyperactivity in autoimmune disease and atopic disorders on the other (*2*).

To delineate surface recognition structures for antigen or MHC gene products, or both, on inducer T lymphocytes, we generated a series of human inducer T cell clones (phenotype T3+T4+T8−) and subsequently used one of them, termed RW17C, as an immunogen for the generation of monoclonal antibodies. The RW17C clone has been propagated in vitro by stimulation with a combination of ragweed antigen E (RWAGE), autologous antigen-presenting cells (APC's), and interleukin 2 (IL-2) for more than 8 months. As shown in Table 1, RW17C proliferates in response to RWAGE only in the context of autologous APC's as judged by [^{3}H]thymidine incorporation. In contrast, incubation with either autologous or allogeneic APC's alone, or a combination of RWAGE plus allogeneic APC's or autologous APC's plus the unrelated antigen, tetanus toxoid (TT), does not lead to proliferation of RW17C. The same MHC and antigen combination is required to induce helper function for B cells from clone RW17C. Thus, only the combination of RW17C, autologous APC's, and RWAGE results in B cell immunoglobulin G (IgG) production.