Steric sea level variations during 1957–1994: Importance of salinity

John I. Antonov
National Oceanographic Data Center, NOAA, Silver Spring, Maryland, USA
University Corporation for Atmospheric Research, Boulder, Colorado, USA

Sydney Levitus and Timothy P. Boyer
National Oceanographic Data Center, NOAA, Silver Spring, Maryland, USA

Received 8 May 2001; revised 6 December 2001; accepted 7 January 2002; published 23 November 2002.

[1] Spatially averaged (50°S–65°N) temperature and salinity changes in the 0–3000 m layer during the 1957–1994 period resulted in a sea level rise at a mean rate of about 0.55 mm per year. About 10% of this rate is due to a decrease of the volume mean salinity. The magnitude of total steric sea level (TSSL) changes and the ratio of thermosteric and halosteric anomalies to TSSL anomaly are nonuniform geographically. Salinity effects are critically important to the TSSL changes in some regions of the ocean. For example, the thermosteric anomaly is nearly compensated by the halosteric anomaly in the subpolar North Atlantic. This fact will cause erroneous heat content estimates based on altimetric observations from space if a climatological salinity is assumed. Based on the present historical archive of salinity data, a decrease in global mean salinity has occurred. This increase of fresh water would cause sea level rise at a rate of 1.3 ± 0.5 mm/yr if the added water comes from sources other than floating sea ice. *INDEX TERMS:* 4215 Oceanography: General: Climate and interannual variability (3309); 4556 Oceanography: Physical: Sea level variations; 1635 Global Change: Oceans (4203); *KEYWORDS:* temperature and salinity variability, steric sea level, sea level rise, climate change, Labrador sea

Citation: Antonov, J. I., S. Levitus, and T. P. Boyer, Steric sea level variations during 1957–1994: Importance of salinity, *J. Geophys. Res.*, *107*(C12), 8013, doi:10.1029/2001JC000964, 2002.

1. Introduction

[2] Mean sea level variability is an important indicator of changes in the Earth's climate system. Globally averaged sea level increased at an overall rate of 1 to 2 mm/yr during the past century (see reviews by *Gornitz* [1995] and *Warrick et al.* [1996]). The main causes of this phenomenon are attributed to the reduction of land glaciers and the thermal expansion of the world ocean. The thermal expansion theory of the mean sea level rise is based on (1) the fundamental physical property of seawater density to decrease (increase) when heat is added (removed), (2) the ability of the world ocean to store a substantial amount of heat in its deep layers (see *Levitus et al.* [2000] for recent observational evidence), and (3) observed trends in the surface temperature over the 20th century [e.g., *Hansen et al.*, 1999]. Model computations explain more than two thirds of sea level rise by the thermal expansion of the world ocean [*Church et al.*, 1991; *De Wolde et al.*, 1995; *Warrick et al.*, 1996; *Stouffer and Manabe*, 1999].

[3] In addition to the thermal expansion, seawater density (volume) is also a function of its salinity. A higher concentration of salinity will increase the water density (decrease its volume) and vice versa if the mass and temperature of the water sample are constant. This effect, termed "haline contraction," plays a very important role in the annual cycle of steric sea height (combined effect of thermal expansion and haline contraction on sea level) [*Pattullo et al.*, 1955] and in maintaining the ocean thermohaline circulation [*Manabe et al.*, 1991; *Warrick et al.*, 1996].

[4] With respect to the global sea level rise, the effect of salinity changes is assumed to be small [*Warrick et al.*, 1996] (because the globally averaged ocean salt content is assumed to be nearly constant on timescales of centuries if no abrupt changes in the global hydrological cycle occur). Observationally, a significant redistribution of salt between and within ocean basins was reported for multiyear periods (e.g., *Levitus* [1990] among others). Local changes in salt content were reported by *Tabata et al.* [1986]. They studied the interannual variability of steric sea level at North Pacific Ocean Weather Station Papa and found that salinity changes in the upper 1000 m layer governed steric sea level at this station for some years. We will document that long-term changes in salinity affect steric sea level for even larger areas of the world ocean.

[5] In addition to steric effects, changes in salinity can be an indicator of changes in local and global hydrological cycles. We will only present here some general speculations on this topic because the main goal of this paper is to estimate steric sea level changes over the last four decades.

2. Data and Methods

[6] We have used data from the *World Ocean Database 1998* (about 5.2 and 1.4 million temperature and salinity

0148-0227/02/2001JC000964$09.00

profiles, respectively) [*Levitus et al.*, 1998] to prepare objectively analyzed temperature and salinity anomaly fields for 5-year running composites for the 1948–1996 period.

[7] Our analysis of the oceanographic data consists of several steps. First, the climatological monthly mean value [*Antonov et al.*, 1998; *Boyer et al.*, 1998] for each 1° latitude-longitude square at each standard depth level was subtracted from each observed value to reduce the effects of the annual cycle. These anomalies were used to estimate the standard deviation (SD) of data in each 5° square and used to remove any anomaly exceeding 6 SD. This 6 SD threshold for rejecting data was chosen after we found that the conventional "3-sigma rule" eliminated some good measurements [*Levitus et al.*, 2000]. This suggests that statistical distributions of temperature and salinity are deviate from the normal distribution in some regions of the world ocean (especially near the oceanic frontal zones). Thus this formal statistical procedure was considered as a tool to reject only grossly bad data. Our quality control procedures of oceanographic data are described by *Boyer and Levitus* [1994].

[8] Next, mean anomalies (departures from the monthly mean climatologies) were computed for overlapping 5-year periods (analogous to a running-mean average) at each 1° square at each standard depth level from the ocean surface to 3000 m. Finally, to fill data gaps in these fields we applied objective analysis. For each 5-year period, objective analysis starts by assuming a first guess of zero for each grid point. Then, we computed a distance-weighted mean correction for each 1° square using every 1° square anomaly within a specified distance (R, radius of influence) from the center of the 1° square being corrected. This correction field was added to the initial anomaly field forming a first-guess field for the next iteration. We used three iterations (three-pass analysis) with a variable influence radius (888, 666, and 444 km). The distance-related weight function resembles the Gaussian distribution (for more details on our choice of parameters of the objective analysis, see *Levitus* [1982] and *Antonov et al.* [1998]).

[9] We repeat our objective analysis procedure as described above a few times. Each time, we examine maps of objectively analyzed values for the presence of suspicious features such as a "bullseye" or a spatial inconsistency. There are two major causes of such features. The first cause is related to the presence in the database of one or more erroneous observation or observations with erroneous metadata such as errors in geographical coordinates (typically, because of wrong sign of longitude or/and latitude). If it was possible we corrected these location errors, otherwise the suspicious data have been flagged and not used in the objective analysis. Second, in some cases data sparsity forced us to flag even a single supposedly good observation that manifested itself as a "bullseye" due to a lack of surrounding, supporting data. We flagged such observation as non-representative value for studying climatic changes.

[10] Different types of observations (mechanical bathythermograph (MBT), Nansen bottle data, expendable bathythermograph (XBT), and salinity-temperature-depth/conductivity-temperature-depth (STD/CTD)) have been blended together to produce objectively analyzed fields. The objectively analyzed value at each grid point is computed by using all data values within an influence region surrounding each grid point. The standard error of each 1° square mean is reduced by the inverse of the square root of the number of 1° squares containing data within the influence region (see Appendix A). Coarser precision of some data with respect to other data is partly compensated by the larger amounts of less precise data available for analysis. For a few locations (like North Atlantic Ocean Weather Station "C") there are overlapping periods with significant number of different types of observations. Specifically, for this station C, we found no statistically significant differences between MBT and the hydrographic station data for yearly mean anomalies [*Levitus and Antonov*, 1995]. Systematic differences in salinity between selected cruises from the late 1950s and two cruises in 1988 in the eastern Atlantic were found at depths of more than 3000 db by *Mantyla* [1994]. He attributed this shift to the changes in methods of salinity determination from titration to the conductivity salinometers. It is not clear what portion of this shift is due to change of salinity determination and what is due to time (about 20 years) and space separation of selected cruises used in that study. Taking into account that the salinity corrections of *Mantyla* [1994] are applicable only for the deep ocean (depth of more than 3000 m) and they vary significantly (0 to 0.01) from one cruise to another, we assume here that after the data quality control, errors in the objective fields are of a random nature.

[11] Changes in temperature and salinity of a water column cause what are termed "steric" sea level variations [*Pattullo et al.*, 1955]. Here we term the change due to thermal expansion as the thermosteric component (TC) and change due to haline contraction as the halosteric component (SC), which follows the terminology of *Tabata et al.* [1986]. Both are expressed in units of height. Their sum is the total steric sea level (TSSL) change. Note that TSSL scaled by the acceleration of gravity is the anomaly of geopotential thickness between isobaric surfaces which represents specific energy and which is related to the geostrophic circulation of ocean.

[12] We used the 1° latitude-longitude temperature and salinity anomaly fields to compute TC, SC, and TSSL in each 1° latitude-longitude square water column as

$$\mathrm{TC} = \int_{z1}^{z2} \frac{1}{\vartheta} \frac{\partial \vartheta}{\partial T} \Delta T dz \quad \mathrm{SC} = \int_{z1}^{z2} \frac{1}{\vartheta} \frac{\partial \vartheta}{\partial S} \Delta S dz \quad \mathrm{TSSL} = \mathrm{TC} + \mathrm{SC},$$

in which ϑ is specific volume ($\partial\vartheta/\partial T > 0$ for temperature $T > 0°\mathrm{C}$ and salinity $S > 20$ and $\partial\vartheta/\partial S < 0$ for any sea-water T and S), z is depth, z_1 and z_2 are the lower and upper limits of depth of integration, and ΔT and ΔS are the 5-year composite temperature and salinity anomalies at any particular standard depth level. Specific volume ϑ has been computed at each standard depth level in each 1° square as a function of climatological annual mean temperature [*Antonov et al.*, 1998] and salinity [*Boyer et al.*, 1998] fields and pressure using the 1980 equation of state for seawater [*UNESCO*, 1987].

[13] We computed 5-year running composites of anomalies because one of our objectives was to determine the deepest layer affecting sea level change. There is a lack of data for computation of yearly compositing periods. By using 5-year running composites we were able to extend our analysis through 3000 m depth. Data from the International Geophysical Year (IGY) expeditions enabled us to begin our analysis of temperature and salinity anomaly fields in the mid-1950s. Data from the World Ocean Circulation Experiment (WOCE) and Joint Global Ocean Flux Study (JGOFS) programs are critical sources of data for the 1990s. The TC, SC, and TSSL anomalies are relative to the 1948–1994 period.

[14] We present results for 50°S to 65°N, excluding the polar regions (about 15% of the total water volume) of the world ocean because the data coverage is sparse in these regions. However, we did perform computations presented in this paper without these geographical limits. The results obtained are within the confidence intervals for the estimates presented here. If not specified in the text, statistical intervals for any estimate are standard errors (see Appendix A).

3. Results

[15] The importance of salinity changes in decadal variability of steric sea level is clearly seen in Figure 1. Presented are time series of the thermosteric and halosteric components of steric sea level anomaly for different layers of the water column in the Labrador Sea. The halosteric component in Figure 1 is multiplied by −1 for compactness. Thus the degree of positive correlation between TC and inverted SC indicates the tendency of TC and SC to compensate each other. Extremes of SC in the 0–500 m layer occurred around the middle of the 1970s and in the early 1980s and 1990s. They correspond to occurrences of the well-documented “Great Salinity Anomaly” (GSA) events in the North Atlantic [*Dickson et al.*, 1988; *Belkin et al.*, 1998]. The largest changes of the SC and TC occurred in the 1000–2000 m layer which represents Labrador Sea Water. For most of the 1950–1994 period, TC and SC of this layer nearly compensated each other. The decadal range (about 65 mm) of total steric sea level anomaly of the 0–3000 m layer (not shown) is reduced in comparison with the individual ranges of 126 mm and 83 mm for TC and SC, respectively.

[16] The density-compensating changes in the Labrador Sea represent a basin-scale phenomenon. Figure 2 shows the geographical distribution of TC, SC, and TSSL anomaly of the 0–3000 m layer for the Pacific and Atlantic Oceans averaged for the 1978–1994 period. We chose this period because, according to Figure 1, relatively cold and fresh waters occupied the central part of the Labrador Sea during these years. In addition, our compositing period corresponds to the period after the mid-1970s Pacific Regime Shift (also referred to as the Pacific Decadal Oscillation; see, e.g., *Trenberth and Hurrell* [1994]). TSSL (bottom panel) was anomalously high along most of the coasts of the Americas. A TSSL decrease occurred, mainly in the midlatitudes of the western North Pacific and in the western South Pacific, where the negative TSSL anomaly extends southeast from the equator toward the Tuamotu Archipelago (about 15°S, 150°W). Comparing the TSSL anomaly pattern with the TC and SC components reveals that SC enhanced TC in the eastern North Pacific. In the western parts of both oceans the sign of the SC anomaly was opposite to the thermosteric component, but its magnitude was small enough so that it did not significantly affect the TSSL anomaly, with one important exception. In the subpolar region of the North Atlantic, the signs of SC and TC are opposite, and importantly, the magnitudes are nearly equal. This result indicates that estimates of sea surface height changes by satellite altimeter using climatological salinity fields may not detect important joint climatic changes of the temperature-salinity distribution in the subpolar North Atlantic Ocean. If one was to attempt to estimate heat content changes in this region using altimeter data and a climatological salinity field, the resulting estimates of the heat content anomaly would be inaccurate, possibly even with respect to the sign of heat content anomaly. Also, see *Chambers et al.* [1997] for a discussion of errors associated with the assumption of a linear relation between heat storage and sea level anomalies derived from TOPEX altimeter data.

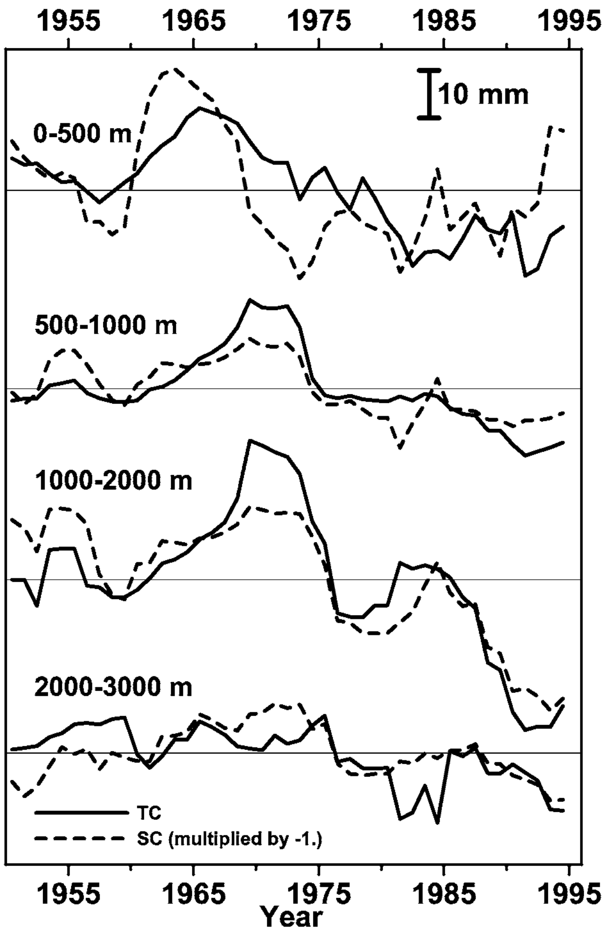

Figure 1. Time series of 5-year running composite of the thermosteric (TC) and halosteric (SC) components (in millimeters) for four layers (0–500 m, 500–1000 m, 1000–2000 m, 2000–3000 m) in the Labrador Sea (56°N, 51°W). Dashed lines are SC components multiplied by −1 (thus for this figure only, the signs of SC and vertically integrated salinity anomaly are the same).

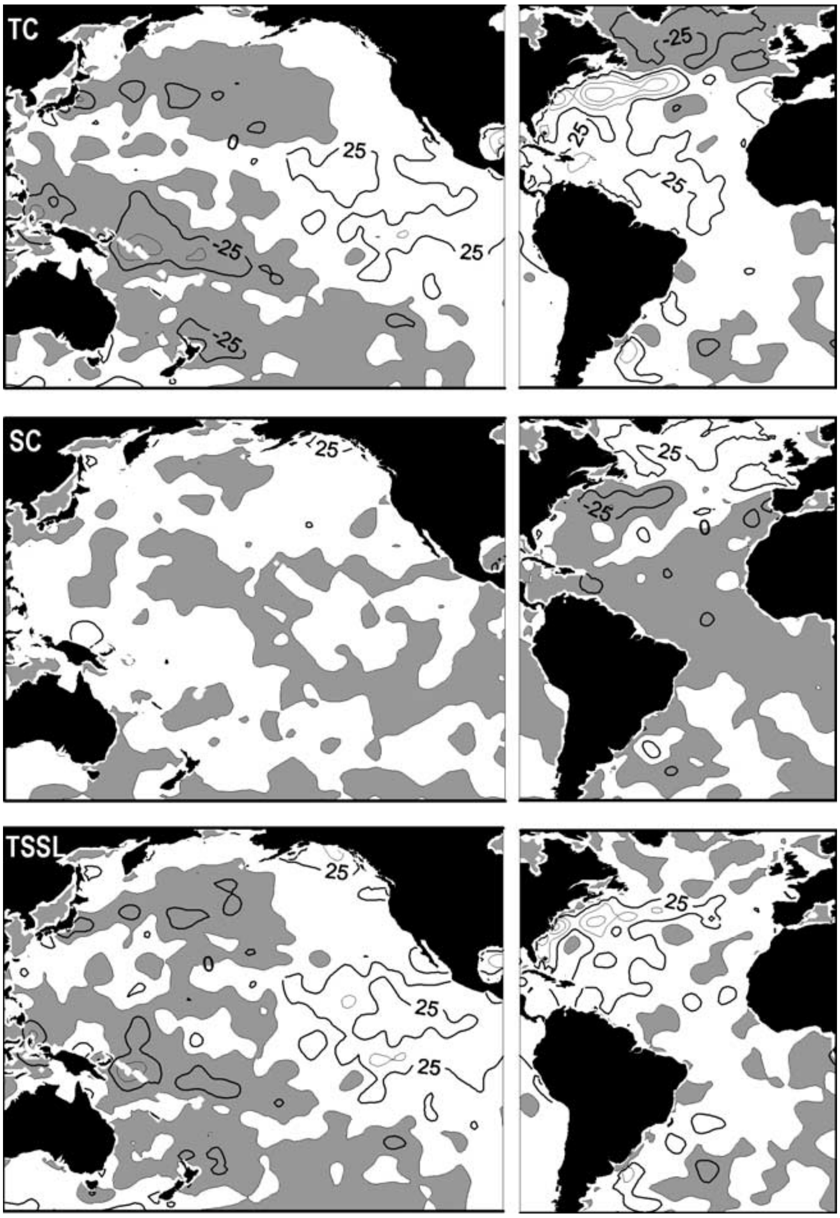

Figure 2. Mean steric sea level anomaly (in millimeters) of the 0–3000 m layer for the 1978–1994 period. (top) Thermosteric (TC), (middle) halosteric (SC), and (bottom) total steric (TSSL) anomaly fields for the Pacific and Atlantic Oceans. The reference period is 1948–1994. The contour interval is 25 mm; negative anomalies are shaded. Thick contour is ±25 mm.

[17] The severe winters of the early 1980s and 1990s in the Labrador Sea along with increased export of fresh water and sea ice from the Arctic [see *Belkin et al.*, 1998] were major sources for the cooling and freshening of the subpolar North Atlantic in 1978–1994. However, a mechanism and relation of these events to global climate change have not been identified.

[18] Except for the subpolar North Atlantic, the mean anomaly of TSSL for the 1978–1994 period mostly resembles the spatial pattern of TC. In the North Pacific, the TSSL anomaly pattern is associated with the mid-1970s Regime Shift of the atmospheric circulation [*Trenberth and Hurrell*, 1994]. Steric sea level rise off the coast of southern California has been estimated to be about 1 mm/yr during 1950–1992 [*Roemmich*, 1992]. The linear trends of TSSL of the 0–3000 m layer for 1957–1994 (not shown) have a similar magnitude of 1–2 mm/yr along the western coast of North America. Although TSSL is only one component that contributes to mean sea level, reports of sea level rise along the eastern coast of North America based on tide gauge data [*Douglas*, 1991; *Ezer et al.*, 1995] are consistent with our results. Also, TSSL time series in the midlatitudes of the western North Atlantic (not shown) are in good agreement with the tide gauge record at Bermuda [*Levitus*, 1990]. The importance of salinity changes for regional long-term TSSL change have been noted earlier for Ocean Weather Station Papa [*Tabata et al.*, 1986] and in the tropical Pacific Ocean along the 165°E transequatorial section [*Maes*, 1998].

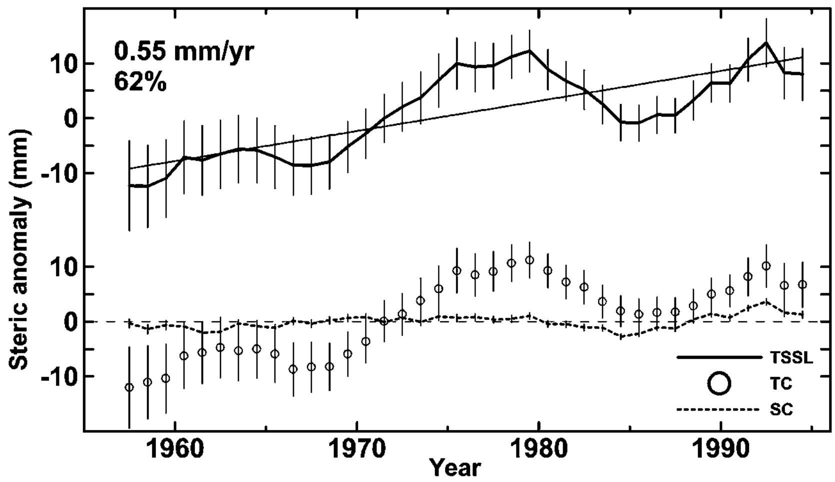

Figure 3. Time series of spatially averaged (50°S–65°N) 5-year running composites of thermosteric (open circles), halosteric (dashed line), and total steric (solid line) anomalies (in millimeters) of the 0–3000 m layer for the 1957–1994 period. Vertical lines represent ±1 standard error of the 5-year mean estimates of steric components. The linear trend is plotted for the TSSL anomaly time series. The trend and the percent variance accounted for by this trend are given in the top left corner.

[19] Spatially averaged (50°S to 65°N) time series of TSSL, TC, and SC anomalies of the 0–3000 m layer for the 1957–1994 period are presented in Figure 3. The total steric change shows that a rise occurred during this period, but the change was not monotonic (similar to changes of the world ocean heat content [*Levitus et al.*, 2000]). Steric sea level rise was slow until 1967 and then increased through 1975 (about 2.0 mm/yr). In the first half of the 1980s, steric sea level dropped, reaching a relative minimum around 1985. After 1985 the steric sea level rise resumed and its rate was very close to the maximum rate for the 1967–1975 period.

[20] The linear trend of the total change of the 0–3000 m layer thickness over the entire 1957–1994 period is statistically significant with 99% confidence and corresponds to a sea level rise rate of 0.55 ± 0.07 mm/yr. This rate of steric sea level rise is a significant contribution to sea level rise estimates based on tide gauge data but less than the contribution of the thermal expansion effect predicted by ocean model simulations [*Warrick et al.*, 1996].

[21] Figure 3 indicates that the interdecadal range of thermosteric variations is 23 mm, which is almost 4 times larger than the range of halosteric variations. The linear trend of the SC is about 0.05 ± 0.02 mm/yr, an order of magnitude less than the linear trend associated with thermal expansion. For the 50°S–65°N average, the volume mean salinity of the 0–3000 m layer decreased slightly, contributing to sea level rise (see the SC curve in Figure 3). Observed changes in the cryosphere over the past several decades such as thinning of the Arctic sea ice cover [*Rothrock et al.*, 1999], a decrease in the Northern Hemisphere [*Parkinson et al.*, 1999] sea ice extent and melting of small glaciers [*Dyurgerov and Meier*, 2000], at least qualitatively, support this salinity decrease. While there is far less salinity than temperature data in many parts of the world ocean, it is of interest to expand on the consequences of the linear increase in the SC we have estimated.

[22] The volume mean salinity change can be converted into an equivalent amount of fresh water added to (or removed from) the world ocean. To explain the linear trend in the SC requires 470 ± 170 km^3 of fresh water to be added to the world ocean every year for the 1957–1994 period. This estimated freshening requires a mean sea level rise of 1.35 ± 0.50 mm/yr, an increase in addition to the steric changes. For simplicity, we assumed that the source of this freshening does not relate to melting of floating sea ice, which would decrease salinity without a significant change in global mean sea level.

[23] Despite the sparsity of salinity data (especially in the Southern Hemisphere), patterns of salinity changes reveal some consistent features that have been predicted by model simulations of anthropogenic changes in climate system. For example, Figure 4 shows zonally averaged temperature and salinity anomaly fields for the 1978–1994 period for the world ocean overlain on the climatological mean distributions of these variables. The largest positive salinity anomaly occurred in the subtropics and tropics of both hemispheres. The characteristic magnitude of these zonally averaged salinity anomalies is 0.02. A freshening of the entire 0–2500 m layer exceeding 0.01 occurred between 50°N and 60°N (with magnitudes of 0.04 and 0.02 in the 0–100 m and 100–1000 m layers, respectively). Antarctic Intermediate Water (AAIW, mean salinity less than 34.7 in the Southern Hemisphere) was fresher by less than 0.01 in the 250–1000 m layer. The zonally averaged temperature anomaly field (Figure 4b) is generally positively correlated with its salinity counterpart. Deviations in this positive correlation between temperature and salinity anomaly fields occurred in the upper ocean layer (the equator to 10°N, 40°N–60°N) and in the intermediate layer occupied by the AAIW waters. Model experiments [*Manabe et al.*, 1991] have demonstrated that a meridional profile of surface water flux change is not uniform in the case of a gradual increase in atmospheric CO_2. This caused an increase of model salinity in the

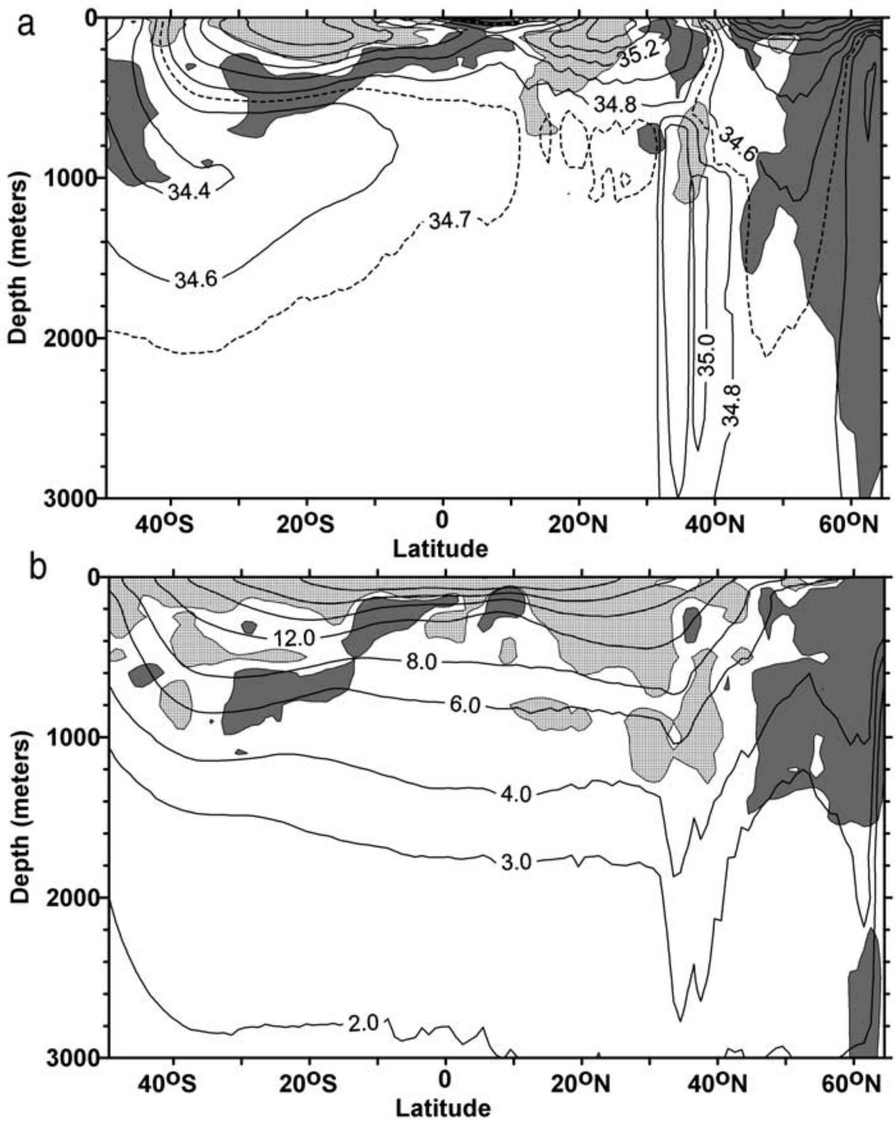

Figure 4. (a) Mean salinity and (b) temperature (in degrees Celsius) anomalies zonally averaged over the world ocean for the 1978–1994 period (darker shaded areas, negative; lighter shaded areas, positive) superimposed on climatological mean salinity and temperature fields (contour lines), respectively. Only anomalies that exceed [0.005] for salinity and [0.025°C] for temperature are shown. The contour interval for climatological temperature varies (from 1°C to 2°C to 4°C). The zonal averages exclude values from the Mediterranean, Black, and Baltic Seas and the Hudson Bay.

tropics and significant freshening in the high latitudes. However, the ratio of salinity change in the tropics and the high latitudes is of an order larger in the model in comparison with the observed changes we document.

4. Concluding Remarks

[24] Our results indicate that although thermal expansion has been largely responsible for the observed increase in steric sea level during 1957–1994, haline effects play a small role. Caution should be taken when satellite data [*Chen et al.*, 1998; *Leuliette and Wahr*, 1999; *Nerem et al.*, 1999] are used to estimate climatic changes in the ocean state. Contributions of nearly equal but opposite sign of temperature and salinity changes to the total steric sea level in the subpolar North Atlantic can mask important climatic events. In addition, heat content in these regions can not be accurately estimated when based on satellite altimeter observations assuming a constant salinity field. A similar conclusion was recently reported by *Sato et al.* [2000] for three selected sites: two in the North Pacific Ocean (near Hawaii and in a region centered approximately at 32°N, 122°W) and near Bermuda (32°N, 65°W) in the North Atlantic. As additional data become available in the future, we will update our analysis and continue our studies of this topic.

Appendix A: Error Estimates of Objectively Analyzed Data

[25] We estimate statistical errors of objectively analyzed data as follows.

A1. Standard Deviation of Analyzed Value at $1^\circ \times 1^\circ$ Square

[26] Standard deviation of the observed means (σ_O) at each $1^\circ \times 1^\circ$ square (ODSQ) at each standard depth level within the area defined by the influence radius (R) is defined as

$$\sigma_O = \sqrt{\frac{1}{N-1}\sum_{i=1}^{N}\left(C_i - \bar{C}\right)^2}, \tag{1}$$

where $C_i = W_i\,(O_i - F_i)$ is the distance-weighted correction, in which $W_i = w_i/\Sigma w_i$ is the weight, O_i is the observed mean, F_i is the first guess at the ith ODSQ within the area defined by R; N is the total number of corrections (i.e., the number the observed means or ODSQs with observations within R). We compute σ_O at the first pass of objective analysis with $R = 666$ km and $\gamma = 0.8$ (these settings will approximate the response function corresponding to the World Ocean Atlas 1998 [*Antonov et al.*, 1998] three-pass analysis).

[27] The standard deviation (or standard error) of the objectively analyzed (σ_A) value at each ODSQ at each standard depth level is defined as

$$\sigma_A = \sigma_O\sqrt{\frac{\sum w_i^2}{\left(\sum w_i\right)^2}}, \tag{2}$$

which is obtained by applying a general formula for error propagation (GFEP) [e.g., *Taylor*, 1997] as follows. If an ODSQ analyzed value is $A = \Sigma C_i$, then its standard error according to GFEP is

$$\sigma_A = \sqrt{\left(\frac{\partial A}{\partial C_1}\sigma_{C_1}\right)^2 + \cdots + \left(\frac{\partial A}{\partial C_N}\sigma_{C_N}\right)^2} \tag{3a}$$

or, in our computations,

$$\sigma_A = \sqrt{\left(\frac{w_1}{\sum w_i}\sigma_{C_1}\right)^2 + \cdots + \left(\frac{w_N}{\sum w_i}\sigma_{C_N}\right)^2}. \tag{3b}$$

Thus formula (2) comes from (3b) assuming that $\sigma_O = \sigma_{C_1} = \sigma_{C_N}$.

A2. Standard Deviation of Values Derived from ODSQ Analyzed Values

[28] Having $1^\circ \times 1^\circ$ fields of σ_A simplifies formal computation of the standard deviation of any value derived (SDDV) from the objectively analyzed fields. This is because we again use the general formula for error propagation. First, we calculate the partial derivatives (PD) of an equation used to compute any value based on our objectively analyzed fields. Second, we multiply PDs by the corresponding σ_A. Third, we add the squares of these products. The square root of this sum is the standard deviation (standard error) of the derived value. This standard deviation is known as the root-mean-square (RMS) error.

[29] There are two major assumptions for computing RMS error this way: all σ_A must be independent and random. If there are any doubts about these assumptions, a safer way is to estimate SDDV as follows [*Taylor*, 1997]. First, calculate the PDs. Second, multiply absolute values of the PDs by the corresponding σ_A. Third, just add these products. This is an arithmetic sum of weighted σ_A (SWE). In any case, RMS is never larger than SWE.

[30] For our error analysis, we compute standard errors as RMS if the derived variable is a function of depth (i.e., for each individual $1^\circ \times 1^\circ$ degree square) and as SWE if the derived variable is a function of longitude and/or latitude (i.e., for basin or zonal mean values).

[31] Standard errors of the linear trend estimates reflect only a deviation of time series from a straight line. Our two-sided confidence test of the linear trend estimate is based on a Student t distribution with 6 degrees of freedom, which is the number of nonoverlapping pentads for the 1955–1996 period minus 2.

References

Antonov, J. I., et al., *World Ocean Atlas 1998*, vols. 1–3, *Temperature of Atlantic, Pacific and Indian Oceans*, 498 pp., U.S. Govt. Print. Off., Washington, D. C., 1998.

Belkin, I. M., S. Levitus, J. I. Antonov, and S.-A. Malmberg, "Great Salinity Anomalies" in the North Atlantic, *Prog. Oceanogr.*, *41*, 1–68, 1998.

Boyer, T. P., and S. Levitus, Quality control of oxygen, temperature and salinity data, *NOAA Tech. Rep. 81*, 65 pp., Natl. Oceanogr. Data Cent., Washington, D. C., 1994.

Boyer, T. P., et al., *World Ocean Atlas 1998*, vols. 4–6, *Salinity of Atlantic, Pacific and Indian Oceans*, 498 pp., U. S. Govt. Print. Off., Washington, D. C., 1998.

Chambers, D. P., B. D. Tapley, and R. H. Stewart, Long-period ocean heat storage rates and basin-scale heat fluxes from TOPEX, *J. Geophys. Res.*, *102*, 10,525–10,533, 1997.

Chen, J. L., et al., Seasonal global water mass budget and mean sea level variations, *Geophys. Res. Lett.*, *25*, 3555–3558, 1998.

Church, J. A., J. S. Godfrey, D. R. Jackett, and T. J. McDougall, A model of sea level rise caused by ocean thermal expansion, *J. Clim.*, *4*, 438–456, 1991.

De Wolde, J. R., R. Bintanja, and J. Oerlemans, On thermal expansion over the last hundred years, *J. Clim.*, *8*, 2881–2891, 1995.

Dickson, R. R., J. Meincke, S.-A. Malmberg, and A. J. Lee, The "Great Salinity Anomaly" in the northern North Atlantic, 1968–1982, *Prog. Oceanogr.*, *20*, 103–151, 1988.

Douglas, B. C., Global sea level rise, *J. Geophys. Res.*, *96*, 6981–6992, 1991.

Dyurgerov, M. B., and M. F. Meier, Twentieth century climate change: Evidence from small glaciers, *Proc. Natl. Acad. Sci. U. S. A.*, *97*, 1406–1411, 2000.

Ezer, T., G. L. Mellor, and R. J. Greatbatch, On the interpentadal variability of the North Atlantic Ocean: Model-simulated changes in transport, meridional heat flux, and coastal sea level between 1955–1959 and 1970–1974, *J. Geophys. Res.*, *100*, 10,559–10,566, 1995.

Gornitz, V., Monitoring sea level changes, *Clim. Change*, *31*, 515–544, 1995.

Hansen, J., R. Ruedy, J. Glascoe, and M. Sato, GISS analysis of surface temperature change, *J. Geophys. Res.*, *104*, 30,997–31,022, 1999.

Leuliette, E. W., and J. M. Wahr, Coupled pattern analysis of sea surface temperature and TOPEX/Poseidon sea surface height, *J. Phys. Oceanogr.*, *29*, 599–611, 1999.

Levitus, S., *Climatological Atlas of the World Ocean*, 173 pp., U.S. Govt. Print. Off., Washington, D. C., 1982.

Levitus, S., Interpentadal variability of steric sea level and geopotential thickness of the North Atlantic Ocean, 1970–1974 versus 1955–1959, *J. Geophys. Res.*, *95*, 5233–5238, 1990.

Levitus, S., and J. I. Antonov, Observational evidence of interannual to decadal-scale variability of the subsurface temperature-salinity structure of the world ocean, *Clim. Change*, *31*, 495–514, 1995.

Levitus, S., et al., *World Ocean Database 1998*, vol.1, *Introduction*, 346 pp., U.S. Govt. Print. Off., Washington, D. C., 1998.

Levitus, S., J. I. Antonov, T. P. Boyer, and C. Stephens, Warming of the world ocean, *Science*, *287*, 2225–2229, 2000.

Maes, C., Estimating the influence of salinity on sea level anomaly in the ocean, *Geophys. Res. Lett.*, *25*, 3551–3554, 1998.

Manabe, S., R. J. Stouffer, M. J. Spelman, and K. Bryan, Transient responses of a coupled ocean-atmosphere model to gradual changes of

atmospheric CO_2, part I, Annual mean response, *J. Clim.*, *4*, 785–817, 1991.
Mantyla, A. W., The treatment of inconsistencies in Atlantic deep water salinity data, *Deep Sea Res.*, *41*, 1387–1405, 1994.
Nerem, R. S., et al., Variations in global mean sea level associated with the 1997–1998 ENSO event: Implications for measuring long term sea level change, *Geophys. Res. Lett.*, *26*, 3005–3008, 1999.
Parkinson, C. L., D. J. Cavalieri, P. Gloersen, H. J. Zwally, and J. C. Comiso, Arctic sea ice extents, areas, and trends, 1978–1996, *J. Geophys. Res.*, *104*, 20,837–20,856, 1999.
Pattullo, J., W. Munk, R. Revelle, and E. Strong, The seasonal oscillation in sea level, *J. Mar. Res.*, *14*, 88–156, 1955.
Roemmich, D., Ocean warming and sea level rise along the southwest U.S. coast, *Science*, *257*, 373–375, 1992.
Rothrock, D. A., Y. Yu, and G. A. Maykut, Thinning of the Arctic sea-ice cover, *Geophys. Res. Lett.*, *26*, 3469–3472, 1999.
Sato, O. T., P. S. Polito, and W. T. Liu, Importance of salinity measurements in the heat storage estimation from TOPEX/POSEIDON, *Geophys. Res. Lett.*, *27*, 549–551, 2000.
Stouffer, R. J., and S. Manabe, Response of a coupled ocean-atmosphere model to increasing atmospheric carbon dioxide: Sensitivity to the rate of increase, *J. Clim.*, *12*, 2224–2237, 1999.
Tabata, S., B. Thomas, and D. Ramsden, Annual and interannual variability of steric sea level along line P in the northeast Pacific Ocean, *J. Phys. Oceanogr.*, *16*, 1378–1398, 1986.
Taylor, J. R., *An Introduction to Error Analysis*, 2nd ed., 327 pp., Univ. Sci. Books, Sausalito, Calif., 1997.
Trenberth, K. E., and J. W. Hurrell, Decadal atmosphere-ocean variations in the Pacific, *Clim. Dyn.*, *9*, 303–319, 1994.
UNESCO, *International Oceanographic Tables*, vol. 4, Paris, 1987.
Warrick, R. A., et al., Changes in sea level, in *Climate Change 1995: The Science of Climate Change*, edited by J. T. Houghton et al., pp. 359–405, Cambridge Univ. Press, New York, 1996.

J. I. Antonov, T. P. Boyer, and S. Levitus, NOAA/NODC, E/OC5, 1315 East West Highway, Silver Spring, MD 20910, USA. (john.antonov@noaa.gov)