CITATION

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Ocean Density Change Contributions to Sea Level Rise

ABSTRACT. Ocean warming contributes to global mean sea level rise by reducing the density of seawater, thus increasing its volume. Freshening of seawater also reduces its density, and this effect contributes to regional sea level variations. However, the effect on global mean sea level rise of ocean freshening by land-ice melt is overwhelmed by the effect of the addition of that melted ice to the ocean. After a brief introduction on the importance of the ocean in the current planetary energy imbalance and the consequences for sea level rise, this article reviews the changing mix of ocean measurements of temperature and salinity, and evolving sampling patterns since the 1950s. Estimates of global sea level rise due to warming and the effects of temperature and salinity on regional sea level change patterns are discussed. The article concludes with an examination of the prospects for accomplishing more global temperature and salinity observations by extending autonomous measurements into the currently poorly sampled ice-covered ocean and the half of the ocean volume below the current 2,000-dbar pressure limit of the Argo array of profiling floats.

INTRODUCTION

The ocean expands as it warms, like mercury in a thermometer. Thus, a warming ocean will drive "thermosteric" sea level rise (SLR) even without any addition to ocean mass from land-ice melt. Seawater density is also a function of salinity, as seawater expands when it becomes fresher and contracts when it becomes saltier—this contribution is termed halosteric SLR. The combined thermosteric and halosteric contributions are called the steric change.

The ocean has a vast thermal inertia, with a heat capacity roughly 1,000 times that of the atmosphere. It also has very low albedo (~ 0.06 compared to a global average of ~ 0.3) and thus a propensity to absorb a large fraction of incoming energy from sunlight. These properties, together with the ocean's high ability to radiate energy, and the fact that it covers about 71% of Earth's surface, mean that the ocean plays a key role in the planetary radiation budget, and SLR is responsive to changes in that budget.

Currently, Earth is in long-term radiative imbalance, gaining energy at the top of the atmosphere at a rate of ~ 0.5 x 10^{15} W on average for the past few decades (Hansen et al., 2005; Murphy et al., 2009). Observational estimates over the last two decades suggest ocean warming accounts for a huge amount of this energy: 0.37 (± 0.06) x 10^{15} W (Lyman et al., 2010; Purkey and Johnson, 2010). This is about 25 times humankind's rate of energy consumption, or the equivalent of each of the 6.9 billion people currently living on the planet running about four-dozen 1200-watt hair dryers continuously. Much of the remainder of the energy imbalance has gone into warming the atmosphere and the land, as well as melting both sea ice and land ice (Levitus et al., 2005a). The ocean's huge thermal inertia and long equilibration time scale mean that it would continue warming and sea level would continue to rise for centuries even if greenhouse gas forcing were stabilized at present levels immediately (Meehl et al., 2005; Church et al., 2011, in this issue).

In recent decades, the observed SLR of about 3 mm yr^{-1} (Leuliette and Willis, 2011, in this issue) has been caused in roughly equal amounts by steric expansion and ocean mass addition (Bindoff et al., 2007; Cazenave et al., 2008). The mean reduction in ocean density arises primarily from warming (Antonov et al., 2005; Domingues et al., 2008), with little contribution from freshening (Munk, 2003). The main process driving the other half of past SLR is mass addition by melting land ice (Pfeffer, 2011, in this issue). There are also geophysical drivers of sea level rise (Tamisiea and Mitrovica, 2011, in this issue).

Here we discuss past regional and global evolution in ocean steric SLR and examine the challenges involved in reconstructing these changes from available observations. We then examine exciting future prospects for tracking ocean steric changes with a truly global ocean observing system.

VIEWING THE PAST THROUGH THE CHANGING LENS OF THE OCEAN OBSERVING SYSTEM

The two major challenges in evaluating past steric sea level changes and their global averages are: (1) changing instruments and platforms used to measure ocean temperatures, salinities, and pressures, and (2) incomplete sampling of global ocean volume in the face of energetic short-term variability. The evolution of ocean observation tools must be taken into account in reconstructing past changes.

Instrument Types

Starting in the early twentieth century, ocean temperatures were measured using reversing thermometers on Nansen bottle casts, with a minimum uncertainty of about ± 0.01°C. Depth could be estimated to an accuracy of about ± 1% by comparing measurements from protected and unprotected thermometers, which read differently because of the effects of pressure on temperature (Emery and Thomson, 1998), while salinities were derived from chemical analysis of the water to an accuracy of ± 0.005 PSS-78. Nansen casts produced the most accurate temperature and salinity observations (typically from research vessels) until the conductivity-temperature-depth (CTD) instrument became widely adopted by the oceanographic community in the 1980s (Warren, 2008). CTD data begin to appear in the archives in about 1961 (Boyer et al., 2009), but only in quantity starting in 1967 (Figure 1). By the 1980s, it was possible to achieve an accuracy of ± 0.002°C in temperature, ± 0.002 in salinity, and a pressure (hence depth) accuracy of ± 0.05% (± 3 dbar) through careful use of a shipboard CTD. CTDs are also deployed on other platforms such as moorings and autonomous profiling instruments (ocean gliders and profiling floats). Profiling floats have made a significant impact on broad-scale ocean sampling via the Argo program (Gould et al., 2004). Because they are only calibrated prior to deployment, CTDs on Argo floats have a target

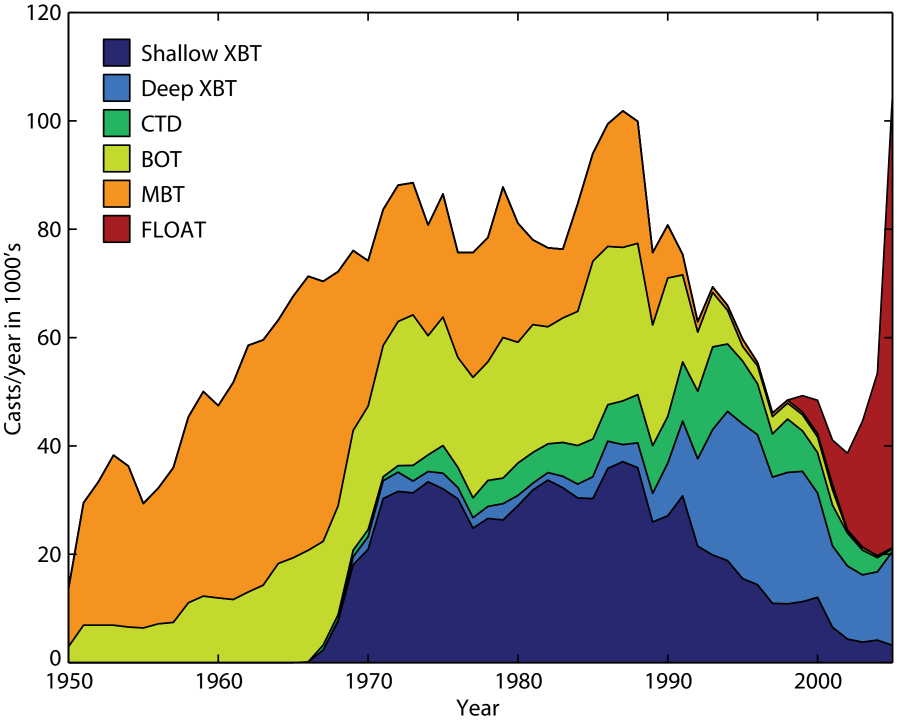

Figure 1. A changing mix of ocean observation platforms. The number of ocean profiles taken per year in historical archives is charted by instrument type: expendable bathythermographs (XBTs), mechanical bathythermographs (MBTs), conductivity-temperature-depth (CTD) instruments deployed from ships, Nansen and Niskin bottle casts (BOT), and CTD-equipped profiling floats (FLOAT).

accuracy of ± 0.01°C in temperature (apparently better from post-mission recalibration of a few recovered units), ± 0.01 in salinity (the typical field accuracy based on comparisons to shipboard CTD data), and a pressure accuracy of ± 0.12% (± 2.4 dbar; a reasonable assessment for most float CTDs based on returned engineering data).

While Nansen and CTD casts, moorings, and autonomous profilers are largely fielded by research agencies, a large amount of thermal data has been collected using ships of opportunity in naval, shipping, and fishing fleets, enabled by instruments deployable from moving vessels. The mechanical bathythermograph (MBT) made it possible for the first time to measure subsurface temperature from a moving ship, but with maximum sampling depths ranging from 140 to at most 295 m. MBT data first appear in the archives in 1941, with numbers peaking in 1966 (Figure 1; Boyer et al., 2009). The MBT has at best a depth accuracy of 1% (Gouretski and Reseghetti, 2010) and a temperature accuracy of ± 0.1°C (Emery and Thomson, 1998). The free-falling expendable bathythermograph (XBT) has largely replaced the MBT. XBT data appear in the archives from 1966, with numbers peaking in 1990. Early XBT models (and the bulk of XBTs dropped) had a maximum depth of 460 m (shallow XBTs; Figure 1) with models capable of 760 m maximum (deep XBTs) available later. The manufacturers' accuracy specifications for the XBT are ± 0.1°C for temperature, and ± 2% for depth.

Historical Sampling Patterns

While ocean temperature data in the archives date back to the eighteenth century, it was only circa 1957, the International Geophysical Year, that ocean measurements began to be taken on a near-global scale. In fact, only a small fraction of even upper-ocean temperature was sampled at all, much less on a yearly basis, until shortly after 1967. The pre-1967 time period included some full-depth, high-quality, coast-to-coast oceanographic sections done using reversing thermometers, but most upper-ocean temperature measurements were taken around the periphery of the North Pacific and North Atlantic Oceans (Figure 2a) at that time, along with many shallow MBT profiles (Figure 1; Boyer et al., 2009).

Starting in 1967, with the introduction of the shallow XBT, sampling along shipping lanes greatly increased available ocean-temperature measurements for the upper 460 m (Figure 2b). This upper-ocean coverage increased from about 40% of the ocean sampled per year in 1967 to near 70% in the 1990s (increasingly extending to 760 m by deep XBTs). Nonetheless, the subtropical, and especially the high-latitude

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Southern Hemisphere oceans all suffer from a notable dearth of observations through about 2004.

Full-depth, coast-to-coast, high-accuracy oceanographic sections using CTDs commenced in the 1980s, ramping up in number and global coverage through the 1990s during the World Ocean Circulation Experiment (King et al., 2001). This program provided a global set of baseline measurements for ocean climate studies. A subset of these sections has been reoccupied since the Climate Variability and Predictability (CLIVAR) program began in 1995, revealing statistically significant warming in the deep Southern Ocean and much of the abyssal global ocean (Fukasawa et al., 2004; Purkey and Johnson, 2010). But even global ship-based sampling programs have a general bias toward summer conditions, especially in high latitudes. For instance, only a few truly dedicated teams attempt to measure the temperature of the Labrador Sea from a ship in February.

The Argo program is an international array of freely drifting profiling floats equipped with CTDs designed to sample the upper 2000 dbar of the global ice-free ocean every 10 days (Gould et al., 2004). While implementation began in 2000, Argo first achieved sparse near-global coverage in 2004, and reached its initial target of 3,000 globally distributed floats in late 2007. It currently provides over 100,000 high-quality temperature (and salinity) profiles distributed around the global ice-free upper ocean with no seasonal bias—a radical shift in our ability to track ocean salinity and temperature fields on a broad scale (Figure 2c).

In summary, broad-scale sampling of the upper ocean really began in

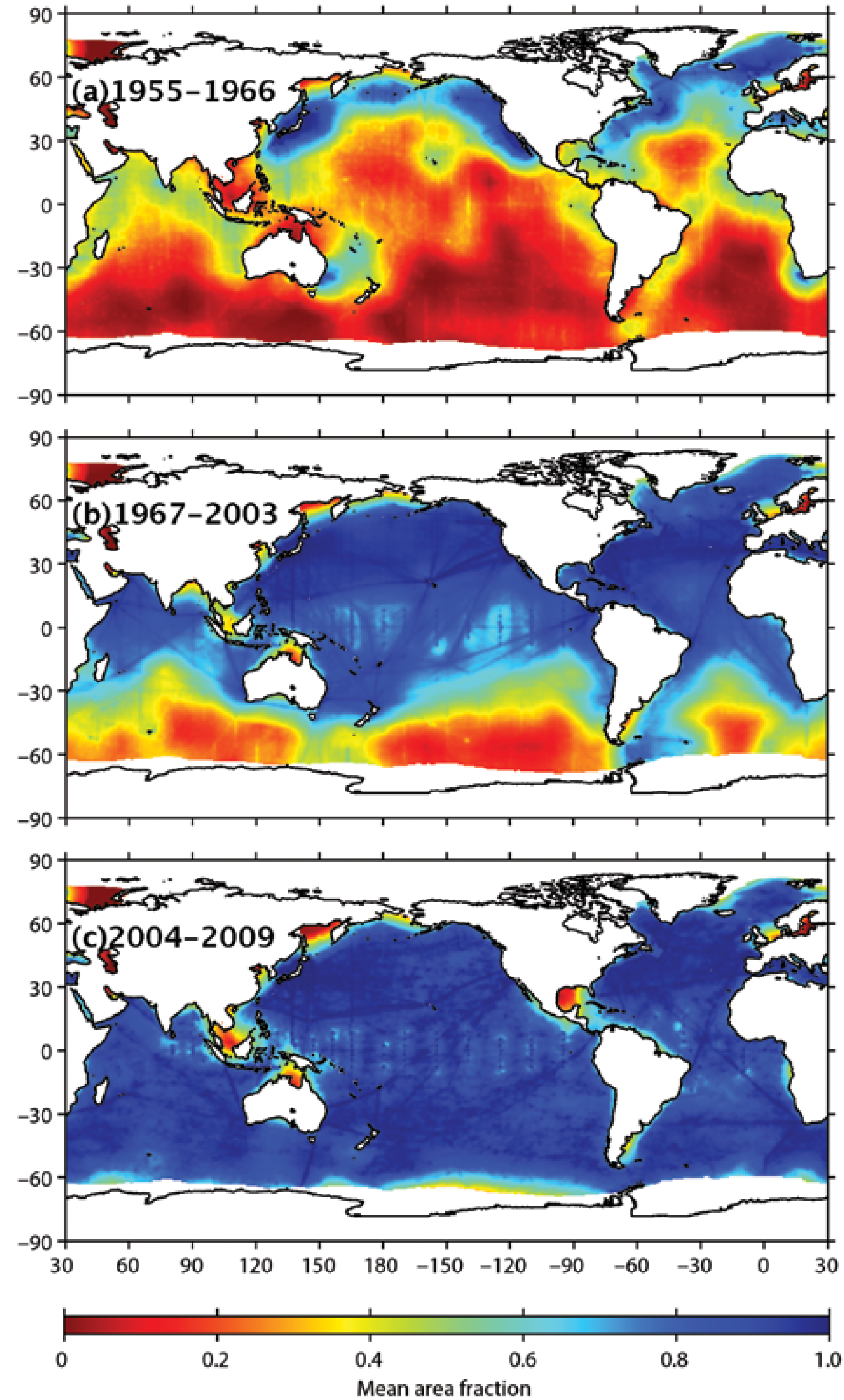


Figure 2. Mean of annual "observed" area coverage for upper-ocean temperature (profiles reaching at least 350-m depth) from (a) 1955–1966, (b) 1967–2003, and (c) 2004–2009, updated from Lyman and Johnson (2008). The figure is based on a simple objective map that contains both small (~ 100 km) and large (~ 1,000 km) scales in its covariance function. Salinity coverage (not shown) is even sparser, especially prior to Argo. *Courtesy of John Lyman*

the 1950s and 1960s, and measurements then were largely shallow. While MBTs and later XBTs provide the most common form of temperature profiles, they are of lower accuracy compared to reversing thermometer or CTD data (more on this topic below). High-quality temperature and salinity data are largely confined to regions and seasons sampled by research fleets until the advent of the Argo program, which now provides nearly global and complete seasonal coverage of broad-scale ocean temperature and salinity fields in the upper 1,000–2,000 dbar. However, the ocean volume that lies below these depths, more than half its total, remains only infrequently sampled along sparse repeat lines established mostly in the 1990s.

> "FINDING THE OPTIMAL MIX OF TECHNOLOGIES AND THE RESOURCES TO EXPAND HIGH-ACCURACY ROUTINE TEMPERATURE AND SALINITY MONITORING INTO ICE-COVERED SEAS AND THE DEEP OCEAN REMAINS AN OUTSTANDING CHALLENGE."

Temporal and Regional Variability

The large gaps in spatial coverage are not the only challenge we face in mapping past, slow, broad-scale ocean changes. The ocean is a dynamic environment. On synoptic or shorter temporal and spatial scales, internal waves, tides, and eddies can have a strong effect on any measured temperature or salinity profile, meaning that sufficient data must be collected to average out these features. A substantial seasonal cycle must also be accounted for when estimating ocean heat content over annual and longer time scales.

The large seasonal buoyancy (temperature and freshwater) fluxes into and out of the ocean (e.g., Moisan and Niiler, 1998) are modulated by interannual variations of winds and surface buoyancy fluxes such as those associated with the El Niño-Southern Oscillation (ENSO), which shifts vast amounts of warm water (hence heat) across the tropical Pacific (e.g., Johnson et al., 2000). Decadal modes of variability that redistribute ocean heat and salt (e.g., Yashayaev, 2007), such as the North Atlantic Oscillation (NAO; Dickson et al., 1996) or the Pacific Decadal Oscillation (PDO; Mantua et al., 1997), have large regional impacts but perhaps less imprint on global averages. These natural large-scale decadal cycles can dominate patterns of regional change over 10–20 year records, masking the pattern of longer-term multidecadal change (Harrison and Carson, 2007).

All of these temporal and spatial variations, together with the sparse and inhomogeneous distribution of historical ocean measurements, make quantifying the temporal evolution of global mean contributions of ocean warming to sea level rise a challenging task. A mean temperature must be defined for a given time, but until Argo began operation, the global upper ocean was never well sampled in any given year, and the ocean volume below the present 1,000–2,000-dbar limit of Argo still suffers from this deficiency. A climatological seasonal cycle must be defined, but again, prior to Argo, it was not feasible in much of the higher-latitude ocean, which had been largely sampled in the summer. Most of the studies of ocean temperature variability published to date suffer from these difficulties to a greater or lesser degree, because Argo is only now approaching a long enough time (at least five years) to allow robust estimation of a seasonal cycle (Roemmich and Gilson, 2009).

After estimating a mean and seasonal climatology, it is necessary to decide how to treat gaps in the historical record, especially when attempting to estimate temporal change in the global mean ocean temperature. A variety of approaches are used, each with trade-offs. Objective analyses (e.g., Ishii and Kimoto, 2009) relax back to the mean in the absence of data, and thus may underestimate trends in data-sparse regions (Gille, 2008). Calculating averages only for the ocean volumes sampled in any given year and applying the mean anomaly for the sampled portions of the ocean to the unsampled portions (Lyman and Johnson, 2008) is a simple but debatable choice. A more complicated method projects data onto large-scale modes of variability that are estimated statistically (Domingues et al., 2008). This method assumes that large-scale modes of variability are stationary, but sometimes they are not (e.g., Bond et al., 2003).

PAST OCEAN WARMING TRENDS

Global Integrals

While individual studies have used different approaches to gap-filling, reference climatologies, and inclusion of various instrument types, they all show that the upper-ocean global heat content (OGHC) over the past 50 years rises by about 100–150 x 10^{21} J of heat in the upper 700 m (Palmer et al., 2010). Although this multidecadal change is robust, there is less agreement on the detailed year-to-year evolution of OGHC (and the associated thermosteric sea level rise) from the 1950s to the present.

Our view of how OGHC has evolved from the 1950s to the present has undergone a major shift due to the discovery and removal of a time-dependent warm bias in XBT data (Gouretski and Koltermann, 2007; Wijffels et al., 2008; Levitus et al., 2009; Ishii and Kimoto, 2009). This bias was a major source of spurious decadal variability in the OGHC trajectory over time, which is now much reduced in most bias-corrected estimates (Figure 3). However, even when the bias is remediated, OGHC does not change smoothly and monotonically upward, but appears to be characterized by periods (years to a decade) of stasis or weak cooling punctuated by periods of rapid warming. Some of the cooling events appear to be related to major Plinian volcanic eruptions that increase Earth's albedo via stratospheric aerosols (Domingues et al., 2008), but not all of the apparent periods of slow warming are yet well understood, including a recent one (Trenberth, 2010). A shift in observing systems from mostly XBT to mostly Argo floats coupled with systematic instrument biases can introduce spurious changes in OGHC (Willis et al., 2009), and even within Argo, there are subtle pressure errors that can affect global averages and regional patterns (Barker et al., in press). Nonetheless, in the last relatively well-sampled 15 years, the spread in published bias corrections for XBTs currently appears to dominate the error budget for OGHC (Lyman et al., 2010).

Upper-Ocean Regional Patterns

As discussed above, understanding the drivers of regional sea level change requires an appreciation of the natural modes of variability as well as external drivers of warming or cooling such as greenhouse gases, aerosols, or solar irradiance changes. Regional decadal change patterns can be dominated by natural modes rather than persistent but weaker (on decadal time scales) anthropogenic forcing. For instance, the pattern of total SLR due to upper-ocean temperature changes (Figure 4a) as well as the total sea level change pattern measured by satellites since the early 1990s (Figure 4b) show the western tropical Pacific Ocean rising and the eastern Pacific falling. However, the pattern of longer-term thermosteric changes since 1955 (Figure 4c) is opposite in the tropical Pacific. This reversal may be due to changes in PDO, which peaked in the 1990s and has been weakening since (Feng et al., 2010), or in ENSO patterns (Timmerman et al., 2010). This reversal, and many other differences between the

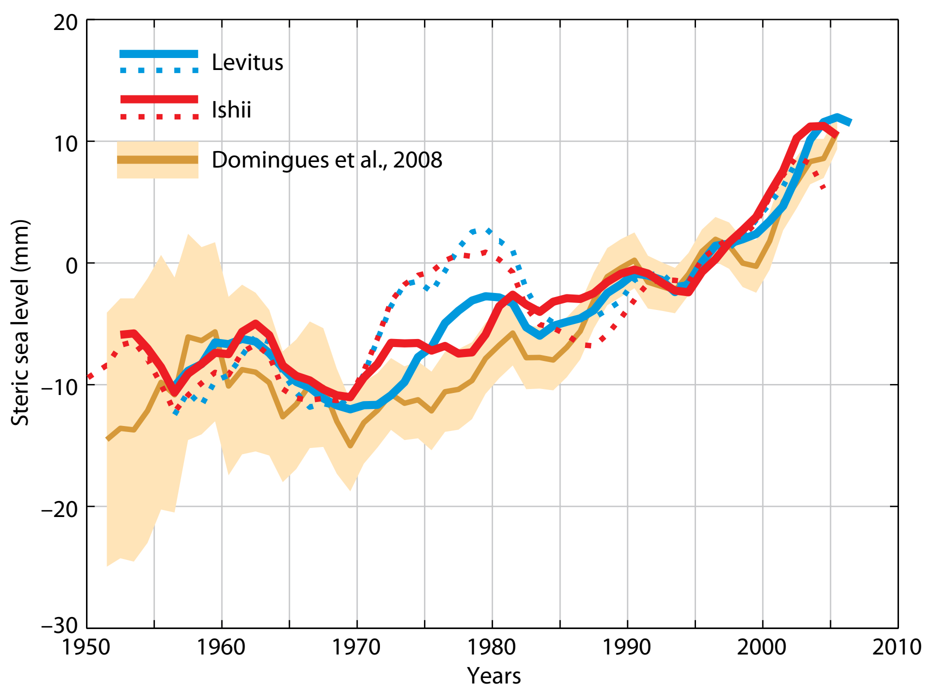
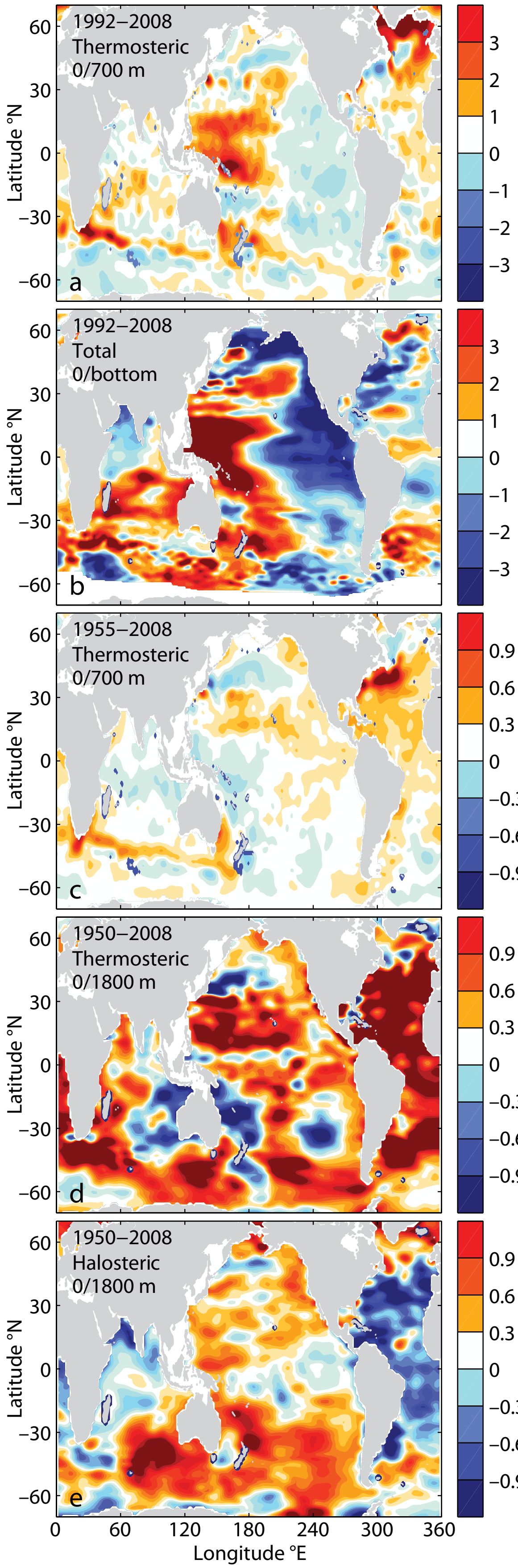


Figure 3. Impacts of instrument bias on observation-based estimates of thermosteric sea level (mm) relative to 700 m based on temperature anomalies from Levitus et al. (2005a, 2009), Ishii et al. (2006), Ishii and Kimoto (2009), and Domingues et al. (2008; with shaded error range). Estimates in dashed lines were affected by now well-documented biases in XBTs (see text), and are distinct from the newer bias-corrected estimates (solid curves). The latter are more similar to each other despite different choices of data types, bias-correction schemes, and mapping methods. *After Church et al., 2010*

Figure 4. Sea level component trend maps (mm yr^{-1}). (a) Thermosteric only, 0–700 m, 1992–2008 based on analyzed fields in Levitus et al. (2009). (b) Total, surface to bottom, 1992–2008, based on satellite altimeter measurements but with the global average adjusted to match (a) in order to allow comparison of spatial patterns. (c) Following (a), but for 1955–2008. (d) Thermosteric, 0–1800 m, 1950–2008, based on the global analysis of Durack and Wijffels (2010). (e) Following (d), but with halosteric only.



patterns of trends over the most recent 15 years compared to 50-year changes, underscores the importance of natural modes in regional patterns at decadal and shorter time scales.

Salinity and Halosteric Effects

Fewer studies have attempted to document global changes in sea level driven by ocean salinity changes, as salinity measurements are sparser than those of temperature. Despite this difficulty, a consistent picture of multidecadal changes is emerging (Levitus et al., 2005b; Helm et al., 2010; Durack and Wijffels, 2010). Overall, salinity changes reflect an enhancement of existing salinity patterns, consonant with an enhancement of Earth's hydrological cycle in global warming scenarios (e.g., Held and Soden, 2006). Net salinity changes associated with the dilution of the ocean due to ice melt are tiny and thus do not drive any significant halosteric changes when averaged globally (Munk, 2003).

However, regional salinity changes can significantly impact regional sea level change patterns. Based on just a thermosteric analysis, sea level in the Atlantic Ocean appears to be rising faster than in other basins over the past 50 years (Figure 4c,d). However, this expansion is greatly compensated by a halosteric contraction (salinification) while the more slowly warming Pacific is undergoing halosteric expansion (freshening; Figure 4e). The result of this difference is that for the full steric change, the Atlantic is more comparable to the other basins (not shown).

On a more local basis, strikingly compensating temperature-salinity changes have been observed over several decades in Labrador Sea Water (LSW; Yashayaev, 2007). Decadal variability in this water mass largely follows changes in the NAO, with cold, fresh LSW formed during positive NAO phases, and warm, salty conditions building into the region during negative NAO phases. These changes are largely density compensating—the colder LSW formed during negative NAO phases is only slightly denser. These temperature-salinity variations are exported to the south, as LSW is a significant component of North Atlantic Deep Water, which spreads southward in the Atlantic limb of the global thermohaline circulation.

Vertical Structure of Ocean Warming Trends

The vertical structure of long-term warming trends illuminates some of the processes involved. Zonal averages of long-term trends versus latitude and depth (Figure 5a) show warming predominant in the upper 700 m, with only a few localized areas of cooling. Warming generally weakens from the surface with depth (Figure 5b), as might be expected for a change driven at the ocean surface.

The nonuniformity of the signal reflects many influences (Figure 5a). For instance, regional variations in the depth to which surface forcing penetrates may be part of the large signal around 60°N. The subsurface cooling on the equator may reflect the influence of large-amplitude, short-time-scale shifts in ocean circulation owing to ENSO or longer-term shifts associated with PDO, as discussed above. Finally, the weak warming signal south of 30°S may result from lack of data in that region, as careful analysis of existing data suggests strong deep warming trends there (Gille, 2002; Alory et al., 2007), likely due to long-term changes in local winds and currents (Böning et al., 2008; Gille, 2008).

Warming trends estimated using data from repeats of coast-to-coast, full-depth, closely spaced CTD sections from 1981 to 2010 suggest that the warming in the upper Southern Ocean extends to the ocean floor (Figure 5c; Purkey and Johnson, 2010), a result that might be anticipated if top-to-bottom fronts of the Antarctic Circumpolar Current shift southward along with the westerly winds in the region (Gille, 2008). In addition, warming in Antarctic Bottom Water appears to be maximum in the global average near 4500-m depth. While the cause of this bottom warming is not yet clear, there is a global pattern of the strongest abyssal warming around Antarctica, weakening into the North Pacific, East Indian, and West Atlantic Oceans (Purkey and Johnson, 2010). The global nature of this change might be expected given the dominance of Antarctic Bottom Water in filling the deep ocean (and indeed the total ocean volume) around much of the globe (Johnson, 2008). These deep changes contribute to both global and regional SLR budgets.

THE FUTURE

While satellite systems have demonstrated that total sea level change and variability can be monitored from space, subsurface ocean measurements are needed to separate out the components—mass change, thermosteric, or halosteric—and to assess their vertical distribution. Sustaining the international

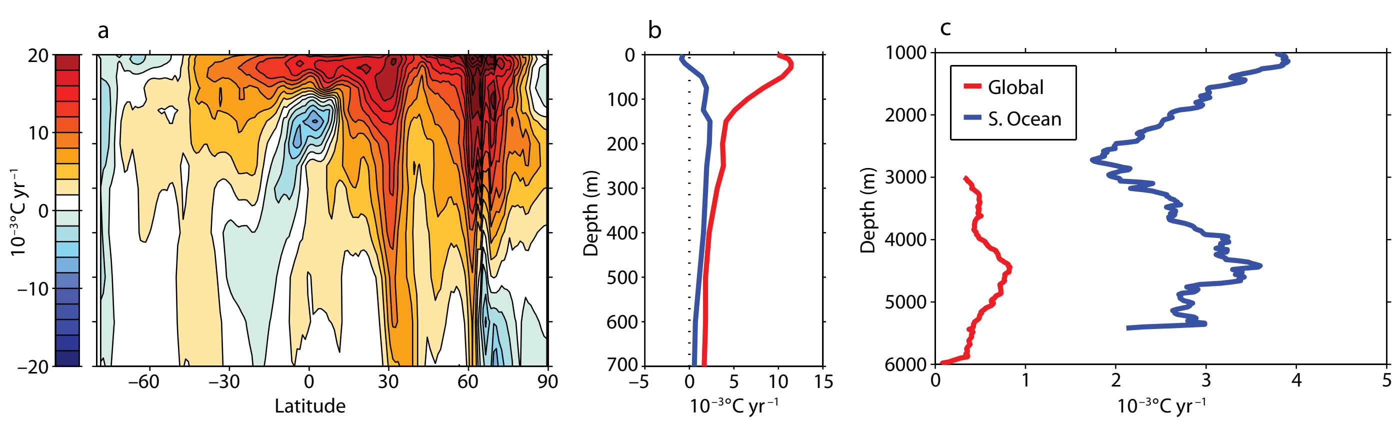


Figure 5. (a) Zonally averaged temperature trend from 1967–2009 based on analyzed fields of Levitus et al. (2009). (b) Area-averaged temperature trend from 1967–2009 for the globe (red) and the Southern Ocean (blue), defined here as the area south of the Sub-Antarctic Front from Orsi et al. (1995). (c) Global (red) and Southern Ocean (blue) depth profiles of deep temperature trends centered about 1992–2005 after Purkey and Johnson (2010).

Argo array of profiling CTD floats over the coming decades will allow more insights into evolving regional and global temperature and salinity patterns, and their role in sea level change and climate in the ice-free regions of the global ocean. However, changes in these properties in higher latitudes are also of great interest for sea level and other climate analyses. Extending Argo into seasonally ice covered regions is now feasible (Klatt et al., 2007), and there are also techniques for collecting CTD profiles under year-round pack ice (Krishfield et al., 2008). In addition, warming (and also freshening) in deep and bottom waters below the 2,000-dbar profiling target for Argo floats also contributes to sea level change, both regionally and globally. Autonomous deep gliders have collected CTD data to nearly 6,000 m (Charles C. Eriksen, University of Washington, *pers. comm*., December 9, 2010), giving hope for efficient broad-scale sampling of the deep half of the ocean volume currently sampled only sparsely in space and time by repeat shipboard CTD sections (e.g., Purkey and Johnson, 2010), mostly by international collaboration at decadal intervals (http://www.go-ship.org). Finding the optimal mix of technologies and the resources to expand high-accuracy routine temperature and salinity monitoring into ice-covered seas and the deep ocean remains an outstanding challenge (Hall et al., 2011).

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