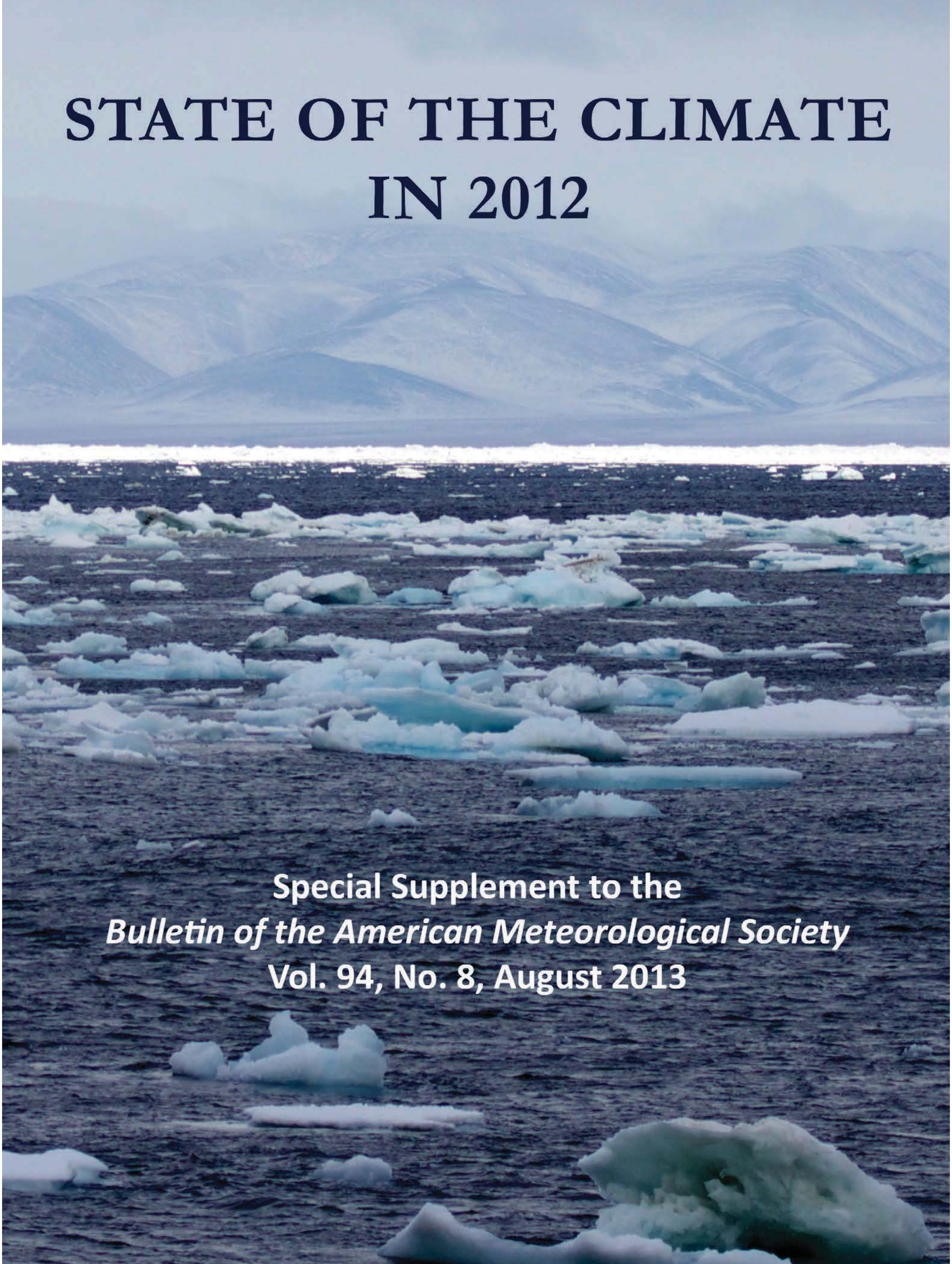

STATE OF THE CLIMATE IN 2012

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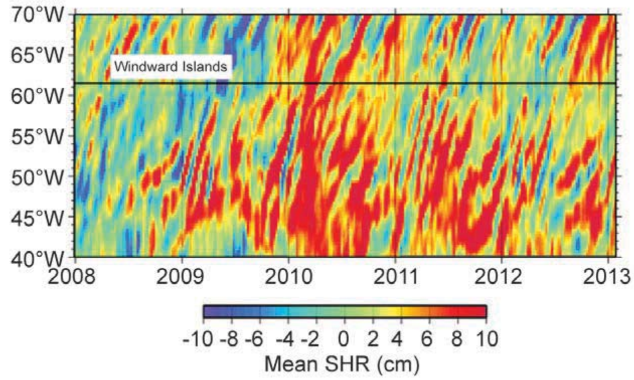

Fig. 3.20. Space-time diagram of deseasoned sea height anomaly (SHR) values along the NBC ring corridor during 2008–12. (Source: http://www.aoml.noaa.gov/phod/altimetry/cvar/nbc.)

h. Meridional overturning circulation and heat transport observations in the Atlantic Ocean—M. O. Baringer, W. E. Johns, G. McCarthy, J. Willis, S. Garzoli, M. Lankhorst, C. S. Meinen, U. Send, W. R. Hobbs, S. A. Cunningham, D. Rayner, D. A. Smeed, T. O. Kanzow, P. Heimbach, E. Frajka-Williams, A. Macdonald, S. Dong, and J. Marotzke

The meridional overturning circulation (MOC) is typically defined as the maximum of the zonally-integrated mass transport stream function. In the Atlantic Ocean the MOC is the large-scale circulation that transports warm near-surface water northward, thereby transferring heat to the atmosphere and returning southward as colder, denser, deeper water. The actual flows are more complicated than this simple description and for further details the reader should see previous *State of the Climate* reports (e.g., Baringer et al. 2011, 2012) or the recent reviews of Srokosz et al. (2012) or Lozier (2012).

The longest time series of a major ocean current contributing to the strength of the MOC is NOAA's Florida Current (FC) data, which began continuous daily measurements in 1982. The full record (Fig. 3.21) shows substantial variability on all measured time scales (Meinen et al. 2010; Baringer and Larsen 2001). The 1982–2012 median transport of daily values is 32.0 ± 0.27 Sv (standard error of the mean based on an integral time scale of about 20 days) with a minimal downward trend of -0.23 ± 0.06 Sv decade^{-1} (90% confidence). In 2012 the annual median was 31.6 ± 1.5 Sv, with the annual mean transport slightly below the average since 2007 (Baringer et al. 2012). However, the 2012 median is within the middle 50% of all annual means. The 2012 daily values of FC transport do show some unusual periods. The daily FC transport values as compared to all previous years (Fig. 3.21, top) indicate that 2012 was unusual in that there were several low transport values (LTP) starting 27 October and ending 24 November. The lowest transport observed occurred on 28 October, reaching only 17.2 Sv. This low value is similar to the lowest transport recorded since 1982 (17.3 Sv on 3 October 1983). The 2012 LTP was preceded by the only high transport event during 2012 that exceeded the 95% confidence limits. This transport exceeded 38 Sv from 3 to 4 October. Low values in the October–November time frame are consistent with the average annual cycle of FC transport (e.g., Meinen et al. 2010). In the last week of October low values are consistent with the passage of Hurricane Sandy northward offshore of the US East Coast. Previous studies have shown that southerly along-shore wind stress can cause a reduction in FC strength, leading to increased sea level along the coast (e.g., Ezer et al. 2013; Sweet et al. 2009). Ezer et al. (2013) quantified this effect using

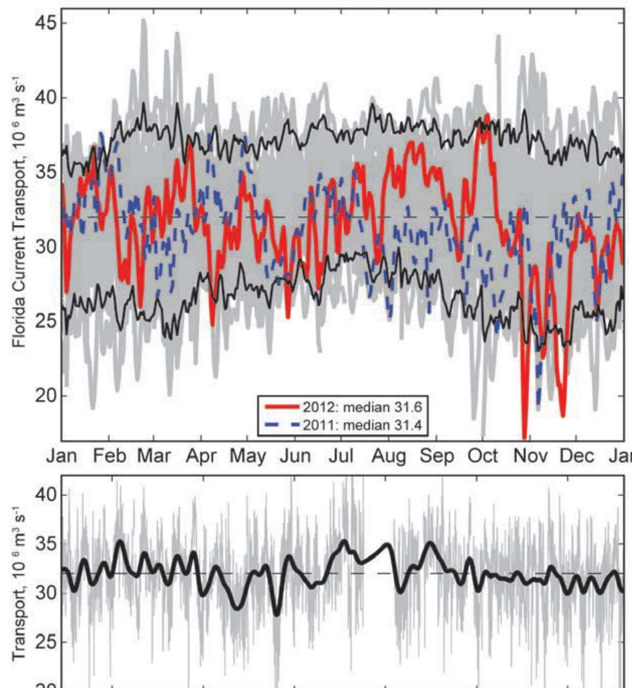

Fig. 3.21. (a) Daily estimates of the transport of the Florida Current during 2012 (red solid line) compared to 2011 (dashed blue line). The daily values of the Florida Current transport for other years since 1982 are shown in light gray and the 95% confidence interval of daily transport values computed from all years is shown in black (solid line); the long-term annual mean is dashed black. The 2012 median transport (31.6 ± 1.5 Sv) lies slightly below the long-term median for the daily values of the Florida Current transport (32.0 Sv). (b) Daily estimates of the Florida Current transport for the full time series record (light gray), and a smoothed version of transport (heavy black line; using a 12-month second-order Butterworth filter) and the mean transport for the full record (dashed black).

a numerical model to show that a 1 Sv decrease in transport corresponds to a 1 cm increase in coastal sea level near these latitudes (see also Boon 2012; Sallenger et al. 2012; Ezer and Corlett 2012).

The only complete daily time series of basin-wide MOC strength at a particular latitude currently available is from the in situ mooring array, RAPID-MOC/MOCHA/WBTS (Fig. 3.22). This mooring array spans the Atlantic Ocean roughly along 26°N (Rayner et al. 2010; Kanzow et al. 2008a). The mean MOC transport centered on this low MOC event (1 April 2009–31 March 2010, referred to as 2009/10) was 12.7 Sv. After this 2009/10 low, the annual mean transport increased to 16.9 Sv in 2010/11, only slightly lower than the 2004–11 mean of 17.5 ± 1.6 Sv (± standard error of annual means). Although the mean transport for 2010/11 represents an increase, there is still a statistically significant (p=0.95) MOC minimum from 13 November–29 December 2010. This MOC minimum is preceded by three minima in low Florida Current transport occurring between 4 October and 10 December 2010 (Fig 3.22). Similar

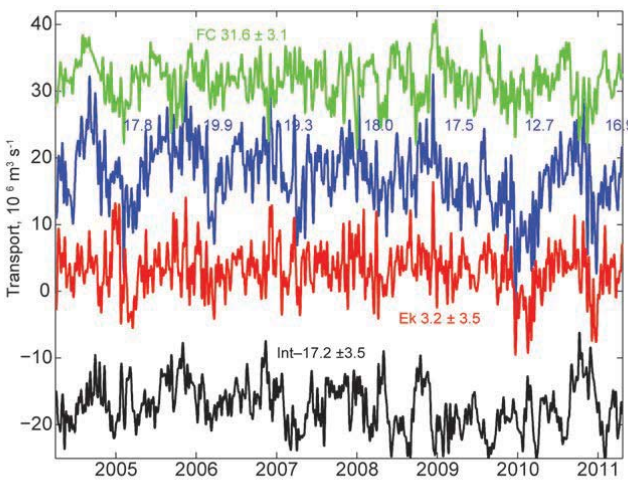

Fig. 3.22. Daily estimates of the strength of the meridional overturning circulation (MOC; blue line) and its components, the Florida Current (FC; green), wind-driven Ekman transport (Ek; red) and the geostrophic interior (Int; black), as measured by the UK National Environmental Research Council Rapid Climate Change Program (RAPID-WATCH), the National Science Foundation's Meridional Overturning and Heat transport Array proposal, and the NOAA Western Boundary Time Series project (WBTS). The interior volume transport estimate (accurate to 1 Sv, Cunningham et al. 2007) is based on the upper ocean transport from April 2004 to April 2011 (see also Rayner et al. 2010; Kanzow et al. 2010), with a 10-day low pass filter applied to the daily transport values. Annual means of the MOC transport starting 1 April of each year are shown in blue text (in Sv).

to the December–March 2009/10 low MOC event, the November 2010–December 2011 low MOC appears most coincident with the unusual and significantly low Ekman transport compared to a mean of 3.2 Sv (2–29 December 2010). In December–March 2009/10, the slowing of the MOC (low FC and Ekman transports) was preceded by a weakening of the upper ocean southward flow east of 77°W (aka the interior transport, Fig. 3.22; 5 October–2 December, mean -11.1 Sv) and followed by a strengthening of the interior southward flow 29 January–6 February 2011. Wunsch and Heimbach (2013) estimate the frequency of occurrence of such monthly extremes to be 14 months in a 1992–2010 monthly transport time series that is assumed to represent a stationary Gaussian process. Overall, however, the interior transport was stronger southward in both 2009 and 2010 (hence contributing to the low MOC transport in those years). In these two years there was interesting phasing between anomalous interior, FC, and Ekman transport. It is the sum of these components that makes up the MOC at this latitude and therefore a clear understanding of this phasing is fundamental to the understanding of the physical mechanisms supporting MOC transport variability (McCarthy et al. 2012).

The North Atlantic MOC is also being monitored by somewhat less direct and complete time series at 41°N and 16°N (Fig. 3.23). Near 41°N, Willis and Fu (2008) developed a technique using Argo profiling floats combined with satellite altimetry and the ECCO2 state estimate (Menemenlis et al. 2005) to estimate the upper ocean circulation with the zonally and vertically integrated upper ocean flow representing the upper layer of the MOC. The 41°N time series shows a similar low MOC event slightly preceding the 26°N winter 2009 low MOC. Near 16°N, a mooring array of inverted echo sounders, current meters, and dynamic height moorings that measures the deep circulation across most of the basin has been in place since 2000 (Kanzow et al. 2008b). Interestingly, the 16°N time series has a high southward flow (hence a large MOC) in the winter of 2009 (13 December 2009–23 January 2010). The three-month low-pass filtering of these time series highlights the seasonal cycles found in all three. There are different phases for each, with 16°N having a maximum MOC in May–July, 26°N having a broad maximum in July–November (Kanzow et al. 2010), and 16°N having a maximum southward flow (and hence stronger MOC) in November–January. Using these time series, various authors have reported MOC trends ranging from

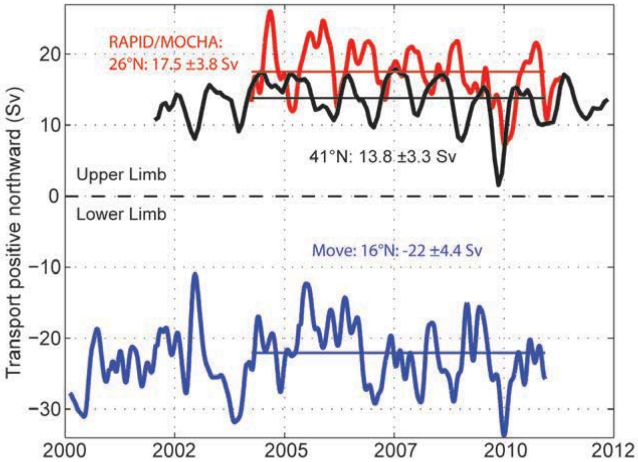

FIG. 3.23. Estimates of the MOC in the Atlantic Ocean from the Argo/Altimetry estimate at 41°N (black; Willis 2010), the RAPID-WATCH/MOCHA/WBTS 26°N array (red; Cunningham et al. 2007), and the German/NOAA MOVE array at 16°N (blue; Send et al. 2011) are shown versus year. All time series have a three-month second-order Butterworth low pass filter applied. Horizontal lines are the mean transport during similar time periods as listed in the corresponding text. For the MOVE data, the net zonal and vertical integral of the deep circulation represents the lower limb of the MOC (with a negative sign for the southward flow).

zero (Willis 2010 at 41°N) to a 10% decrease per decade (Send et al. 2011 at 16°N). Using the overlapping time period of these observations (1 April 2004–10 December 2010), the trend in MOC is -7.1 ± 5.7 Sv decade^{-1} at 41°N and -6.7 ± 5.9 Sv decade^{-1} at 26°N, suggesting a weakening MOC (barely significant at 95% confidence limits). However, at 16°N the deep southward flow is increasing, which is equivalent to an increase in the MOC at -7.7 ± 9.3 Sv decade^{-1} (not significantly different from zero at 95% confidence limits). Using the full time series from either 41°N or 16°N reduces the discrepancy largely by eliminating any significant MOC trend (-1.2 ± 1.7 Sv decade^{-1} at 41°N and 0.5 ± 0.4 Sv decade^{-1} at 16°N). Given the large variability in these short underlying time series and the low transport in the winter of 2009, it is difficult at the present time to determine unambiguously whether the large-scale low-frequency MOC circulation has a trend in the North Atlantic.

In the South Atlantic there is an ongoing estimation of the MOC using upper ocean measurements from expendable bathythermograph sections that measure the upper ocean temperature approximately every three months (Garzoli et al. 2012). The MOC estimate from that data along 35°S since 2002 has suggested no significant trend in the MOC (Dong et al. 2009).

The MOC is related to the meridional transport of heat (MHT) in the oceans, and the variability of MHT can impact heat storage, sea-level rise, and air-sea fluxes, and hence influence local climate on land. MHT has been inferred using the time series data at 41°N (Hobbs and Willis 2012) and 26°N (Johns et al. 2008). Near 41°N, the time series mean MHT is 0.50 ± 0.10 PW (1PW = 10^{15} W) and near 26°N is 1.26 ± 0.40 PW (Fig. 3.24). In the South Atlantic, MHT has been estimated using a combination of expendable bathythermograph (XBT) data and Argo profiling floats (Garzoli et al. 2012; Garzoli and Baringer 2007). The mean MHT near 35°S is 0.55 ± 0.3 PW (±1 standard deviation). Note that the mean MHT from the ECCO-PROD and SODA assimilating models are higher than those computed directly from observations at 41°N, lower near 26°N, and bracket the direct MHT estimates at 35°S. In a detailed intercomparison

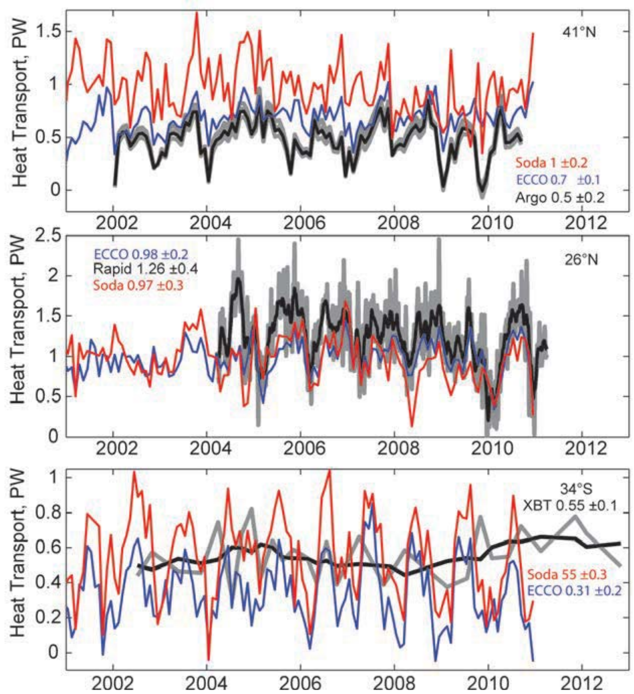

FIG. 3.24. Observed time series of meridional heat transport (MHT) at 41°N (profiling floats), 26°N (mooring/hydrography) and 35°S (XBTs) in the Atlantic compared to the monthly estimates from the ECCO-PROD version 4 (FH12, blue line; Forget et al. 2012, unpublished manuscript) and the Soda version 2.2.4 (red line; Carton and Giese 2008) models. (Top) The black line is the estimate MHT and the gray lines represents the error in the estimate (Hobbs and Willis 2012). (Middle) The black line is the observed data filtered with a 30-day boxcar filter and the gray lines are the underlying 12-hourly data. (Bottom) The gray line is the quarterly estimated MHT from XBTs and the black line is a yearly boxcar filter to those quarterly estimates.

examining the RAPID/MOCHA/WBTS array near 26°N with two climate models, Msadek et al. (2013) found that the low MHT in the models was due to an overly diffuse thermocline rather than a weak MOC. In terms of variability, the low MOC winter of 2009/10 corresponds to a low MHT near both 26°N and 41°N. Both models capture the 26°N event, but not the 41°N event. Near 35°S the assimilation estimates have a much larger MHT seasonal cycle than observations (and, surprisingly, have larger variability). More work is needed to understand the causes of the underlying differences between direct-observation-based and data-assimilating-model estimates of MHT/MOC.

i. Sea level variability and change—M. A. Merrifield, P. Thompson, R. S. Nerem, D. P. Chambers, G. T. Mitchum, M. Menéndez, E. Leuliette, J. J. Marra, W. Sweet, and S. Holgate

Four aspects of sea level variability during 2012 are examined. First updated time series of global mean sea level (GMSL) as determined from satellite altimeter observations are considered. GMSL provides a measure of the temporal change in ocean volume, which is affected by changes in density and the net heat content of the ocean, and by the mass transfer of water between the continents and the oceans via land ice melt and shifting evaporation-precipitation patterns. Second, seasonal anomalies of regional sea level based on satellite altimetry and tide gauge observa-

SIDEBAR 3.2: SLOWDOWN OF THE LOWER, SOUTHERN LIMB OF THE MERIDIONAL OVERTURNING CIRCULATION IN RECENT DECADES—G. C. JOHNSON AND S. G. PURKEY

The Atlantic Meridional Overturning Circulation (AMOC), fed by sinking of North Atlantic Deep Water (NADW), has been continuously monitored for possible changes in the subtropical North Atlantic since 2006 (see section 3h). Annual means of the measured transport of the AMOC range from 13 Sv to 20 Sv between 2006 and 2010 (McCarthy et al. 2012). However, the AMOC is only the upper northern limb of the global meridional overturning circulation (MOC; Lumpkin and Speer 2007). The lower southern limb of the MOC, fed by sinking of Antarctic Bottom Water (AABW), has a transport of similar magnitude (Lumpkin and Speer 2007). In fact, AABW fills more of the world oceans than NADW, including much of the Southern Ocean as well as the deep Indian, Pacific, and western South Atlantic Oceans (Johnson 2008).

There is growing evidence that this lower southern limb is changing over recent decades, including statistically significant warming of deep waters of Antarctic origin nearly globally since circa 1990 (Kouketsu et al. 2011; Purkey and Johnson 2010), as well as freshening of bottom waters near Antarctica in the Indian (Rintoul 2007) and Pacific (Swift and Orsi 2012) sectors of the Southern Ocean. While there is no equivalent of the subtropical North Atlantic AMOC observation system (see section 3h) for the lower southern limb of the meridional overturning circulation, many studies, summarized below, have inferred consistent, worldwide reductions in the strength of this limb by comparing various oceanographic section data from the Global Ship-Based Hydrographic Investigations Program (http://www.go-ship.org/) with data from earlier oceanographic sections occupied during the 1980s and 1990s as part of the World Ocean Circulation Experiment (WOCE) Hydrographic Program.

Changes in the inventories of the coldest, deepest waters around the globe using oceanographic section data from the 1980s to present suggest a strong contraction of the coldest varieties of AABW (Fig. SB3.4; Purkey and Johnson 2012). Their results suggest that deep Southern Ocean waters with potential temperatures below 0°C are disappearing at a rate of over 8 Sv. At 35°S, the northward flow of bottom waters of southern origin is estimated to be slowing down by about 0.7 Sv decade^{-1} from 1968 to 2005 in the Pacific and 0.4 Sv decade^{-1} in the western Atlantic Ocean over the same time interval from a numerical model assimilating these data (Kouketsu et al. 2011). The same assimilation finds no change in the deep Indian Ocean, but oceanographic section data are sparser there since the completion of WOCE. While classic AABW (potential temperatures <0°C) is only found in the Argentine Basin of the western Atlantic at this latitude (Fig. SB3.4), the AABW influence is still strong in the bottom waters of all three oceans (Johnson 2008).

In the Northern Hemisphere, inverse model calculations in the Pacific using data from repeat hydrographic sections along 24°N suggest that the deep northern transport of waters of Southern Ocean origin lessened by 1.5 Sv between 1985 and 2005 (Kouketsu et al. 2009), consistent with results in the South Pacific. In the North Atlantic Ocean at 24°N, one analysis based on geostrophic calculations inferred a remarkable slowdown of northward flow of 4 Sv from four hydrographic sections occupied between 1981 and 2004 (Johnson et al. 2008); although another set of inverse estimates using different reference levels, integration areas, and data extrapolation into u-sampled areas found a slowdown of only 1.3 Sv over those same times, with a resurgence of 0.4 Sv between 2004 and