North Pacific Subtropical Mode Water Volume Decrease in 2006–09 Estimated from Argo Observations: Influence of Surface Formation and Basin-Scale Oceanic Variability

IVANA CEROVEČKI
Scripps Institution of Oceanography, University of California, San Diego, La Jolla, California

DONATA GIGLIO
University of Washington, Seattle, Washington

(Manuscript received 5 March 2015, in final form 25 November 2015)

ABSTRACT

Analysis of Argo temperature and salinity profiles (gridded at 0.5° × 0.5° resolution for 2005–12) shows a strong North Pacific Subtropical Mode Water (NPSTMW) volume and density decrease during 2006–09. In this time period, upper-ocean temperature, stratification, and potential vorticity (PV) all increased within the region in and around the NPSTMW low-PV pool, contributing to the NPSTMW volume decrease in two ways: (i) the volume of water satisfying the low-PV constraint that is part of the “mode water” definition decreased, and (ii) some water that was initially in the NPSTMW density range \( \sigma_b = 25.0–25.5 \) kg m\(^{-3}\) was transformed into lighter water. Both changes in density and in PV in the NPSTMW region were a manifestation of basinwide changes. A positive PV anomaly started to propagate westward from the central Pacific in 2005, followed by a negative density anomaly in 2007, which caused a dramatic NPSTMW volume and density decrease.

A Walin estimate of surface formation in the NPSTMW density range accounted better (although not entirely) for the interannual variability of the volume of water in the NPSTMW density range without imposing the PV constraint, than did the same estimate with the PV constraint imposed. This underlines the importance of the PV constraint in identifying the mode water. The mode water evolution cannot be fully described from a density budget alone; rather, the PV budget must be considered simultaneously.

1. Introduction

Some of the largest ocean-to-atmosphere heat fluxes in the world are found in the midlatitude western boundary current (WBC) systems of the Northern Hemisphere, both in the annual mean fields and interannually (Wallace and Hobbs 2006). On the equatorward side of each major WBC system, strong wintertime ocean heat loss to the atmosphere forms deep winter mixed layers. These mixed layers create large pools of water, called mode waters, with exceptionally uniform properties. After their formation, mode waters are advected laterally and consequently extend over a much larger region than that of their initial formation (Hanawa and Talley 2001).

Because of their large volume, mode waters are important subsurface reservoirs of heat and play a fundamental role in determining the upper-ocean heat content and the interaction of the upper ocean with the atmosphere. Mode waters retain the memory of the atmospheric conditions at the time of formation and hence are key to understanding interannual-to-interdecadal climate variability. They also play an important role in biogeochemical processes, such as the storage of anthropogenic CO\(_2\) (Bates et al. 2002) and nutrient supply to the surface layer (Sukigara et al. 2011; Sabine et al. 2004).

In the western North Pacific, the 16°–18°C thermostad (i.e., a minimum in the vertical gradient of temperature) that occupies the upper permanent thermocline on the equatorward side of the Kuroshio Extension (KE) jet is called North Pacific Subtropical Mode Water (NPSTMW) (Masuzawa 1969). The NPSTMW potential

Denotes Open Access content.

Corresponding author address: Ivana Cerovečki, Scripps Institution of Oceanography, UCSD 0230, 9500 Gilman Dr., La Jolla, CA 92093-0230.
E-mail: icerovecki@ucsd.edu

DOI: 10.1175/JCLI-D-15-0179.1

© 2016 American Meteorological Society
density range is 24.8–25.7 kg m$^{-3}$ and potential vorticity (PV) is less than $2 \times 10^{-10}$ m$^{-1}$ s$^{-1}$ (Masuzawa 1969; Suga et al. 1989). PV is defined as $PV = -(f/\rho_0)(\partial \sigma / \partial z)$, where $f$ is the Coriolis parameter, $\rho_0$ is the density of seawater, and $\sigma$ is potential density. NPSTMW forms when the wintertime ocean buoyancy loss to the atmosphere creates cold deep mixed layers (Hanawa and Talley 2001). This erodes the seasonal pycnocline and ventilates the low-PV pool lying above the permanent pycnocline (Hanawa 1987). After its formation in late winter, NPSTMW is advected eastward by the wind-driven subtropical gyre circulation and can be partially reentrained into the deepening mixed layer during the following fall/winter. Thereby, through reemergence, NPSTMW affects not only the regional, but also the global climate (Alexander and Deser 1995).

The recent accumulation of observations with high temporal and spatial resolution, both from satellites and from Argo profiling floats (Roemmich et al. 2001), as well as results from high-resolution ocean general circulation models, have together provided a more detailed picture of NPSTMW variability. All these display variability at quasi-decadal time scales, with apparently strong (although not yet fully understood) interconnections between atmospheric forcing, NPSTMW volume, upper-ocean heat content, mesoscale processes, PV structure, the subtropical gyre recirculation, and the dynamic state of the KE jet.

Qiu and Chen (2006) found a pronounced decadal variability of NPSTMW volume in the observations, showing that the NPSTMW winter mixed layers were thick in 1993–96 and shallow in 1998–2001. This was followed by a period of gradual thickening between 2001 and 2004 in the recirculation gyre (RG) region (31$^\circ$–36$^\circ$N, 141$^\circ$–150$^\circ$E and east of Japan). The authors related this decadal variability in NPSTMW volume to variability in the Pacific decadal oscillation (PDO); when the PDO index is positive (negative), negative (positive) sea surface height (SSH) anomalies and negative (positive) thermocline depth anomalies are generated in the central Pacific, which propagate westward in the form of a first-mode baroclinic Rossby wave and reach the RG approximately three years later (e.g., Oka and Qiu 2012). At that point, the KE jet becomes weak and unstable (strong and stable), with high (low) regional mesoscale eddy activity and small (large) NPSTMW thickness in the RG region (Oka and Qiu 2012). Analyzing decadal variability in the NPSTMW volume in the RG region Qiu and Chen (2006) further showed that the interannual variability in NPSTMW thickness in the RG was closely related to the variability in the dynamic state of the KE system. They found little correlation between the NPSTMW decadal variability and the year-to-year changes in the cumulative wintertime surface cooling in the RG.

Davis et al. (2011) identified NPSTMW variability at quasi-decadal time scales in a high-resolution Estimating the Circulation and Climate of the Ocean (ECCO2) global ocean data assimilating model simulation (Menemenlis et al. 2008) forced by atmospheric reanalysis products. The volume of NPSTMW in their analysis region, bounded by 20$^\circ$–40$^\circ$N, 130$^\circ$E–160$^\circ$W, reached a minimum at the end of each of the three decades analyzed (in the years 1979, 1988, and 1999). The occurrence of two of the NPSTMW volume minima coincided with a PDO shift from a warm to a cool phase (i.e., the PDO index changed from positive to negative) with a zero time lag (Davis et al. 2011). Davis et al. (2011) thus related quasi-decadal NPSTMW variability in the analysis region to the variability in basin-scale atmospheric forcing that is associated with changes in PDO. Davis et al. (2011) showed that the interannual formation of NPSTMW is significantly correlated with the integrated ocean heat loss over the NPSTMW formation area, when considering their analysis region (20$^\circ$–40$^\circ$N, 130$^\circ$E–160$^\circ$W), which was much larger than the RG region considered by Qiu and Chen (2006). In contrast, when considering the much smaller RG region, Davis et al. (2011) confirmed the result of Qiu and Chen (2006) and similarly found little correlation between the interannual variability of NPSTMW volume and cumulative surface cooling. The difference indicates that the NPSTMW in the RG is governed by different dynamics than is the rest of the NPSTMW because of its proximity to the KE (Davis et al. 2011). This was supported by Rainville et al. (2014), who showed that the decadal variability in the NPSTMW thickness in the whole western Pacific is smaller than the NPSTMW thickness variability in the RG.

We here expand on the Rainville et al. (2014) analysis of the strong NPSTMW volume decrease in 2006–09, the only such event that occurred during the Argo period. The focus of this study, the methods, and the dataset used are, however, different from Rainville et al. (2014). We examine the link between the interannual NPSTMW volume and density variability during the Argo years 2005–12 primarily in an analysis region bounded by 20$^\circ$–40$^\circ$N, 125$^\circ$E–160$^\circ$W and the variability in the local atmospheric air–sea buoyancy fluxes that leads to variability in NPSTMW wintertime surface formation. Following Qiu and Chen (2005) and Davis et al. (2011), we additionally consider the role of changes in stratification, and accordingly in PV, that emanate from the central Pacific in association with changes in PDO. Our results suggest that variability in the NPSTMW wintertime surface formation and changes in stratification are related, as the changes in stratification may
precondition the density range of strongest subsequent NPSTMW surface formation.

During the Argo years (2005–12) analyzed here, both the PDO and the KE jet dynamic state changed (e.g., Qiu et al. 2014), as briefly described below. The PDO index changed sign several times; a positive phase in 2002–05 was followed by a relatively neutral phase that lasted until mid-2007, when the PDO changed to a negative phase and remained negative for nearly two years. In 2009, the PDO reverted to a positive phase, which lasted only 10 months (until June 2010), followed by a negative PDO. The PDO remained negative through autumn 2012. Additionally, the time period analyzed here spans both an unstable (2006–09) and a stable (2010–12 and 2005) KE dynamical state (e.g., Oka and Qiu 2012). The time period of volume decrease thus entirely coincided with an unstable KE state, in agreement with Qiu and Chen (2006). The NPSTMW volume decrease followed a PDO shift from positive to neutral in 2005. The strong NPSTMW volume increase in 2010 coincided with a change in the KE dynamic state from unstable to stable, and it also followed the PDO switch to a positive phase in 2009. Clearly, the short Argo record analyzed here (8 yr long) makes it difficult to fully untangle cause and effect in the observed atmospheric and oceanic variability during the Argo years 2005–12.

In this paper, we analyze the most recent updated gridded Argo product obtained by Roemmich and Gilson (2009) that provides monthly temperature \( T \) and salinity \( S \) data on a 0.5° × 0.5° horizontal grid for the years 2005–12. Additionally, we estimate surface NPSTMW formation using the Walin (1982) analysis, with input fields from a modern atmospheric state reanalysis by the European Centre for Medium-Range Weather Forecasts (ECMWF) interim reanalysis (ERA-Interim) (Simmons et al. 2006). The \( T \) and \( S \) fields from the gridded Argo product are also used to investigate the role of oceanic preconditioning.

This article is structured as follows. In section 2, we briefly outline the Walin (1982) analysis used to estimate surface formation rates of water in the NPSTMW density range by air–sea buoyancy fluxes. In section 3, we describe the quasi-decadal variability of NPSTMW volume and density, as revealed by the gridded Argo product. NPSTMW formation by local air–sea buoyancy fluxes is examined in section 4. In section 5, we describe changes in stratification and PV in the region of the NPSTMW low-PV pool during 2005–12 and relate them to basinwide changes in stratification (and, concomitantly, PV) in the North Pacific Ocean. A summary is presented in section 6.

2. The Walin analysis

It is natural to carry out the analysis of water mass formation and destruction in isopycnal coordinates (e.g., Nishikawa et al. 2013). We thus employ the Walin (1982) formalism, which combines conservation of density (buoyancy) and volume. This framework enables us to determine the diapycnal volume flux (water mass transformation) across an isopycnal surface caused by nonadvective density flux, either due to air–sea buoyancy fluxes or due to diapycnal ocean mixing. “Water mass formation” is obtained as the difference of transformation between adjacent isopycnal surfaces (and is conventionally scaled by the density difference between those surfaces).

The conservation equation for density \( \sigma \) may be written in the form

\[
\frac{\partial \sigma}{\partial t} + \nabla \cdot (u\sigma + N_\sigma) = 0, \tag{1}
\]

where \( u \) is the fluid velocity, \( u\sigma \) is the advective flux of density, and \( N_\sigma \) is the nonadvective flux of density due to air–sea buoyancy (heat and freshwater) flux or diapycnal ocean mixing. Only the nonadvective flux of density can give rise to water mass transformation (e.g., Marshall et al. 1999; Nurser et al. 1999).

The Walin analysis considers a layer of water in the density range \( \sigma - \delta \sigma < \sigma < \sigma + \delta \sigma \) with volume designated as \( V(\sigma) \). The layer outcrops at the sea surface and extends to some deeper terminating fixed Eulerian surface across which we wish to compute the volume flux into the ocean interior. Typically that fixed surface is chosen to be the mixed layer depth. The change of volume \( \partial V(\sigma)/\partial t \) of the layer considered can be expressed as

\[
\frac{\partial V(\sigma)}{\partial t} = A(\sigma - \delta \sigma, t) - A(\sigma + \delta \sigma, t) - M(\sigma, t),
\]

where \( A(\sigma - \delta \sigma) \) and \( A(\sigma + \delta \sigma) \) represent diapycnal volume fluxes across the isopycnal surfaces \( \sigma - \delta \sigma \) and \( \sigma + \delta \sigma \) bounding the layer \( \sigma \), and \( M(\sigma, t) \) is the volume flux out of layer \( \sigma \) into the remainder of the ocean across the fixed Eulerian surface (Fig. 1). The diapycnal volume flux \( A \) depends only on nonadvective fluxes, and, using the notation of Garrett et al. (1995), it can be expressed as

\[
A(\sigma, t) = F(\sigma, t) - \frac{\partial D(\sigma, t)}{\partial \sigma},
\]

where \( F(\sigma, t) \) is the diapycnal volume flux due to air–sea buoyancy (density) flux, and \( \partial D(\sigma, t)/\partial \sigma \) is the diapycnal
s
layer
is the buoyancy flux given by

\[ F(\sigma, t) = -\frac{\partial}{\partial \sigma} \int_{A_{\sigma}} B \, dA. \]  

(2)

Here, \( A_{\sigma} \) is the surface area of the outcrop window of the layer \( \sigma \) in the density range \( \sigma - \delta \sigma < \sigma < \sigma + \delta \sigma \), and \( B \) is the buoyancy flux given by

\[ B = \frac{\alpha Q}{C_p \nabla z} - \frac{\beta \rho_0 S_0}{\nabla z} (E - P - R), \]  

(3)

where \( \alpha \) is the coefficient of thermal expansion, \( Q \) is the net air–sea heat flux (positive for ocean heat loss, in \( \text{W} \text{m}^{-2} \)), \( C_p \) is the specific heat of seawater, \( \nabla z \) is the thickness of the top layer of the ocean, \( \beta \) is the haline contraction coefficient, \( \rho_0 \) is a reference density, \( S_0 \) is a reference salinity, \( E \) is the evaporation, \( P \) is the precipitation, and \( R \) is the runoff (all in \( \text{m} \text{s}^{-1} \)). A positive buoyancy flux \( B \) implies a decrease in ocean density, associated with a temperature increase due to ocean heat gain or to a decrease in \( E - P - R \).

The water mass formation rate is obtained as the finite difference of water mass transformation rates \( A \) with respect to density:

\[ \text{Formation Rate} = \frac{A(\sigma) - A(\sigma + \delta \sigma)}{\delta \sigma}. \]  

(4)

The formation rate in Eq. (4) is conventionally multiplied by \( \delta \sigma \) in order to have the same units as transformation rate \( A \) (\( \text{m}^3 \text{s}^{-1} \)). A more detailed description of the Walin (1982) analysis is provided in, for example, Nishikawa et al. (2013) or Marshall et al. (1999).

3. The NPSTMW water mass and its volume variability

a. The NPSTMW water mass

The NPSTMW is here taken to be water in the analysis region (20°–40°N, 125°E –160°W, east of the islands of Japan; Fig. 2), lying between the isopycnal surfaces \( \sigma_0 = 25.0 \) and \( \sigma_0 = 25.5 \text{ kg m}^{-3} \), and having PV \( < 2 \times 10^{-10} \text{ m}^{-1} \text{s}^{-1} \). This definition of NPSTMW is very close to the definitions used in Rainville et al. (2007) and in Davis et al. (2011).

We hereafter consistently use the term “NPSTMW” to refer to waters in the NPSTMW density range that reside inside the low-PV pool on the equatorward side of the KE. These waters satisfy both the density (\( \sigma_0 = 25.0–25.5 \text{ kg m}^{-3} \)) and the low-PV constraint. In contrast, we refer to water in the same region and isopycnal range, but without the PV constraint, simply as “water in the NPSTMW density range.”

b. Interannual variability of the NPSTMW volume

During the time period 2005–12, the most pronounced feature of the NPSTMW distribution in the analysis region of Fig. 2 as estimated from the monthly mean gridded Argo product of Roemmich and Gilson (2009) is a sudden decrease in volume and in density. The density decrease started in 2005 and lasted until 2010 (Fig. 3a), and the volume decrease started in 2006 and lasted until 2009, when the NPSTMW volume reached its minimum (Table 1). The volume and density decrease in the same analysis region and isopycnal range, but without the PV constraint, is not nearly as dramatic (Fig. 3b), highlighting the importance of the PV constraint in identifying the mode water, as emphasized by Deremble and Dewar (2013). Since mode waters are defined by both

Unauthenticated | Downloaded 06/13/22 05:35 AM UTC
density and PV constraints, their evolution can differ from the evolution of water in the same density range but with no PV constraint imposed. The evolution of mode waters cannot be described from a density budget (or, equivalently, a corresponding volume budget) alone; rather, the PV budget must be considered simultaneously.

The difference between the NPSTMW volume variability and the variability of the volume of water in the NPSTMW density range can also be seen in Fig. 4, FIG. 2. The time-averaged (2005–12) thickness of the NPSTMW (m) in the potential density range $\sigma_\theta = 25.0$–25.5 kg m$^{-3}$ with PV less than $2 \times 10^{-10}$ m$^{-1}$ s$^{-1}$ estimated from the gridded Argo product (Roemmich and Gilson 2009). The box outlined in green indicates the analysis region referenced in the text (unless otherwise specified): 20°–40°N, 125°E–160°W, east of Japan.

FIG. 3. Monthly averaged NPSTMW volume (m$^3$) estimated from the gridded Argo product (Roemmich and Gilson 2009) for the years 2005–12, in the potential density range $\sigma_\theta = 24.6$–25.6 kg m$^{-3}$ (color; right bar) (a) with PV $< 2 \times 10^{-10}$ m$^{-1}$ s$^{-1}$ and (b) as in (a), but without imposing the PV constraint. We refer to the former as “NPSTMW” and to the latter as “water in the NPSTMW density range.” Both estimates are obtained over the analysis region 20°–40°N, 125°E–160°W, east of Japan, of Fig. 2. In (a) and (b), the black line is a cubic fit to monthly volume maxima.
TABLE 1. (i) Volume of NPSTMW newly formed each year and (ii) volume of NPSTMW destroyed each year, both in the potential density range 25.0 ≤ \( \sigma_t \) ≤ 25.5, with PV \( > 3 \times 10^{-10} \text{ m}^2 \text{ s}^{-2} \), and (iii) as in (i), but for water in the NPSTMW density range; (iv) as in (ii), but for water in the NPSTMW density range; (v) annually averaged volume of NPSTMW with PV \( > 3 \times 10^{-10} \text{ m}^2 \text{ s}^{-2} \) and (vi) annually averaged volume of NPSTMW with PV \( > 3 \times 10^{-10} \text{ m}^2 \text{ s}^{-2} \) in the analysis region of Fig. 2.

<table>
<thead>
<tr>
<th>Year</th>
<th>NPSTMW newly formed ( (&gt; 3 \times 10^{14} \text{ m}^3) )</th>
<th>NPSTMW destroyed ( (&gt; 3 \times 10^{14} \text{ m}^3) )</th>
<th>(ii) NPSTMW destroyed ( (&gt; 3 \times 10^{14} \text{ m}^3) )</th>
<th>(iii) NPSTMW newly formed ( (&gt; 3 \times 10^{14} \text{ m}^3) )</th>
<th>PV ( &gt; 3 \times 10^{-10} \text{ m}^2 \text{ s}^{-2} ) volume</th>
<th>PV ( &gt; 3 \times 10^{-10} \text{ m}^2 \text{ s}^{-2} ) volume</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>4.4 (3.4)</td>
<td>3.0</td>
<td>2.5</td>
<td>3.1</td>
<td>3.2 ± 0.05</td>
<td>2.3 ± 0.5</td>
</tr>
<tr>
<td>2006</td>
<td>3.0 (3.5)</td>
<td>2.5</td>
<td>2.2</td>
<td>3.2</td>
<td>3.2 ± 0.05</td>
<td>2.3 ± 0.5</td>
</tr>
<tr>
<td>2007</td>
<td>2.7 (4.2)</td>
<td>2.7</td>
<td>2.0</td>
<td>3.1</td>
<td>3.2 ± 0.05</td>
<td>2.3 ± 0.5</td>
</tr>
<tr>
<td>2008</td>
<td>2.3 (3.1)</td>
<td>2.3</td>
<td>2.0</td>
<td>3.1</td>
<td>3.2 ± 0.05</td>
<td>2.3 ± 0.5</td>
</tr>
<tr>
<td>2009</td>
<td>1.7 (2.4)</td>
<td>1.7</td>
<td>1.4</td>
<td>2.8</td>
<td>3.1 ± 0.10</td>
<td>2.2 ± 0.10</td>
</tr>
<tr>
<td>2010</td>
<td>2.9 (3.4)</td>
<td>2.9</td>
<td>2.6</td>
<td>3.1</td>
<td>3.1 ± 0.10</td>
<td>2.2 ± 0.10</td>
</tr>
<tr>
<td>2011</td>
<td>3.7 (4.7)</td>
<td>3.7</td>
<td>2.9</td>
<td>3.1</td>
<td>3.1 ± 0.10</td>
<td>2.2 ± 0.10</td>
</tr>
<tr>
<td>2012</td>
<td>3.8 (4.5)</td>
<td>3.8</td>
<td>2.9</td>
<td>3.1</td>
<td>3.1 ± 0.10</td>
<td>2.2 ± 0.10</td>
</tr>
<tr>
<td>Average</td>
<td>4.0</td>
<td>3.1</td>
<td>2.5</td>
<td>3.1</td>
<td>3.1 ± 0.10</td>
<td>2.2 ± 0.10</td>
</tr>
</tbody>
</table>

Showing annually averaged north–south density sections across the analysis region along 145°E. The vertical spacing between the pool-defining isopycnals 25.0 and 25.5 kg m\(^{-3}\) is always large relative to that between the other isopycnals. The PV-constrained pool (PV \( < 2 \times 10^{-10} \text{ m}^2 \text{ s}^{-2} \)) occupies a variable, although not always large, part of the total area between the pool-defining isopycnals (Fig. 4e). The density of the low-PV pool in this section decreases between 2005 and 2010, when it reaches its minimum (Figs. 4a–f), in a manner consistent with the variation of the entire NPSTMW PV-constrained pool considered over the entire analysis region (Fig. 3a).

The PV increase in 2005–09, evident as a decrease in the vertical distance between the pool-defining isopycnals, was caused by shoaling of the base of the NPSTMW [which actually started a year earlier than the start of the Argo data analyzed here, in 2004 (Qiu et al. 2007)]. In 2009, when the density of water above the low-PV pool decreased, the top of the NPSTMW layer deepened and the PV attained its highest values (Figs. 4c–f). Cross sections of temperature and salinity along 145°E (not shown), indicate that the strong change in PV in 2009 was caused by warming of water in the depth range that includes the upper half of the low-PV pool and cooling of water in the depth range that includes the lower half of the low-PV pool. In 2009 and 2010 salinity also decreased over the whole depth range of the low-PV pool because of excess rainfall in the 2008 warm season (Sugimoto et al. 2013). The volume and density of the low-PV pool started to increase after 2010 (Figs. 4g–h).

Figure 4 also shows that, during years when the KE is in a stable state and the Kuroshio flow strength is greater (in 2005 and in 2011–12), the slope of the isopycnals bounding the NPSTMW layer at its northern edge is steep (Figs. 4a, 4g, 4h). In 2006–09, when the KE is in an unstable state, the isopycnals at the northern edge of the NPSTMW layer are farther apart, consistent with stronger stirring by energetic eddies, which are present when the KE is in an unstable state (Qiu and Chen 2006; Qiu et al. 2007) (Figs. 4b–e).

To further quantify volume changes in 2005–12, we follow Davis et al. (2011), who estimated the volume of newly formed NPSTMW in year \( i \) as the difference between the NPSTMW volume maximum at the end of the winter–spring formation season in year \( i \) and the minimum volume in the previous year \( i − 1 \). They estimated the destruction of NPSTMW as the difference between maximum and minimum volume during the same year. The volume of newly formed water and the volume of water destroyed, both obtained in this manner for each year, are plotted in Fig. 5. Numerical values are given in Table 1 for both the NPSTMW with the low-PV constraint and for the water in the NPSTMW density range without imposing the low-PV constraint.
If the volume displayed an identical seasonal cycle with no interannual variation, the formation and destruction would be equal. This is much more nearly the case for the water in the NPSTMW region without the PV constraint, rather than for the NPSTMW. Still, both cases show a significant interannual variation. The most pronounced feature of this variation is a steady decrease in the volume of newly formed water from 2006 to 2009 (Fig. 5; Table 1). For both the NPSTMW density range without the PV constraint and the NPSTMW, the volume

**FIG. 4.** Meridional section at 145°E of annually averaged potential density, shown both as colors and as thin black contours, with a contour interval equal to 0.25 kg m$^{-3}$. The thick solid black contours are the isopycnals bounding the NPSTMW density range $\sigma_n = 25.0$–25.5 kg m$^{-3}$. The thick gray contour bounds the annually averaged low-PV pool with $PV < 2 \times 10^{-10}$ m$^{-1}$ s$^{-1}$. 

Unauthenticated | Downloaded 06/13/22 05:35 AM UTC
itself reached a minimum in 2009 (Figs. 5a, b and Table 1).

The corresponding formation and destruction estimates of NPSTMW (16° < T < 20°C and PV < 2 × 10^{-10} m^{-1} s^{-1}) by Rainville et al. (2014), also given in Table 1, show the same pattern, but both formation (3.7 ± 0.2 × 10^{14} m^{3}) and destruction (3.6 ± 0.2 × 10^{14} m^{3}) are somewhat larger than the formation (3.1 ± 0.9 × 10^{14} m^{3}) and destruction (3.2 ± 0.6 × 10^{14} m^{3}) estimated here, consistent with the area considered by Rainville et al. (2014) being larger than the analysis region shown in Fig. 2.

c. Interannual variability of the seasonal cycle of NPSTMW volume

We next consider the time series of volume in 0.05 kg m^{-3} wide isopycnal layers spanning the NPSTMW density range (Fig. 6). A seasonal cycle is pronounced in all the lighter isopycnals, while it is strongly attenuated in the densest isopycnals in the NPSTMW density range (Fig. 6), in agreement with Rainville et al. (2014). Between 2007 and 2009, there is a steady decrease in the wintertime peak volume for the lighter variety of NPSTMW (approximately in the density range \( \sigma_0 = 25.05-25.25 \text{ kg m}^{-3} \)), which is not present at even lighter densities (\( \sigma_0 < 25.05 \text{ kg m}^{-3} \)), suggesting that the volume decrease may be associated with decrease in NPSTMW formation. In 2010, the wintertime peak volume in the same density range started to increase. In the denser isopycnals, \( \sigma_0 = 25.25-25.5 \text{ kg m}^{-3} \), the volume decrease started in 2005, consistent with shoaling of the base of NPSTMW observed by Qiu et al. (2007) and Rainville et al. (2014) and seen in Fig. 4. Preferential destruction of volume of denser varieties of NPSTMW thus contributes to density decrease of the low-PV pool.
Volume started to recover around 2010, but only in the $\sigma_\theta < 25.4 \text{ kg m}^{-3}$ density range.

Since many studies attribute NPSTMW variability to atmospheric causes, such as variability in ocean heat loss and surface winds associated with the East Asian wintertime monsoon (Suga and Hanawa 1990, 1995), we next ask the following questions: Was the observed decrease of newly formed NPSTMW volume in 2006–09 (Fig. 5a) predominantly caused by a decrease in surface formation, as suggested by Rainville et al. (2014)? Was the increase in 2010 of newly formed NPSTMW volume (Fig. 5a), as well as the increase of isopycnal volume in the NPSTMW density range (Figs. 6c–j), caused by stronger surface formation in 2010 (compared to formation in the years 2006–09)? Was the observed NPSTMW density decrease, most pronounced in years 2009–10 (Figs. 4e–f), related in some way to changes in surface formation in the NPSTMW density range? We address these questions in the subsequent section.

4. NPSTMW formation by air–sea buoyancy fluxes from Walin analysis

Surface transformation and formation rates have been obtained from the Walin (1982) analysis introduced in section 2. Inputs to the Walin analysis are daily estimates of the air–sea buoyancy flux $B$ [given by Eq. (3)] and of

Fig. 6. The time series of NPSTMW volume (with $PV < 2 \times 10^{-16} \text{ m}^{-1} \text{s}^{-1}$) in the analysis region of Fig. 2 obtained from the monthly gridded Argo product, binned into potential density layers 24.95–25.50 kg m$^{-3}$ (as indicated in the title of each panel).
surface density. Variable $B$ is computed using daily net heat flux $Q$, $E - P - R$, and using sea surface temperature (SST) from ERA-Interim. Daily surface potential density is obtained by combining daily ERA-Interim SST with monthly averaged Argo near-surface salinity (interpolated to daily values). The annually averaged surface transformation rates are calculated from Eq. (2) as the area integral of air–sea buoyancy flux $B$ over outcrop windows $\delta \sigma = 0.1 \text{kg m}^{-3}$ wide inside the geographical region of interest. The annually averaged formation rates are then computed as the finite difference of transformation rates with respect to density, as given by Eq. (4), and multiplied by the bin width $\delta \sigma$. The net surface formation rate for each year in the entire NPSTMW density range is the sum of surface formation rates over all outcrop windows in the entire NPSTMW potential density range $\sigma_\theta = 25.0-25.5 \text{kg m}^{-3}$.

The low-PV constraint has not been imposed in the Walin calculation because ERA-Interim daily averaged temperature data are provided only for the ocean surface, thus not allowing us to estimate vertical density gradients needed to compute PV. On the other hand, monthly averaged Argo temperature data do not allow us to follow the instantaneous isopycnals, which is necessary for the Walin analysis (e.g., Garrett and Tandon 1997; Cerovečki and Marshall 2008). We will, however, show in section 4b that the results of Walin analysis without the low-PV constraint imposed tend to be biased low. This made it necessary to re-define the region over which the buoyancy flux integration was performed. The manner in which this redefinition was done is explained in the following section.

a. Maps of NPSTMW surface formation rates

Maps of annually averaged surface formation rates per unit area in the NPSTMW density range (Fig. 7) show that the NPSTMW formation occurs north of 28°N.
[in agreement with (Oka 2009)] and east of 140°E, with the exception of the year 2005, a large meander year during which warm NPSTMW was also found farther west of 140°E, south of Japan (Oka 2009; Sugimoto and Hanawa 2014).

Although the region of strongest surface formation tends to coincide with the time-averaged position of the NPSTMW outcrop window in January–March, which is the time period when subduction occurs, both the formation region and the location of the outcrop window show large interannual variability (Fig. 7). This variability is associated with changes in the air–sea heat flux (Fig. 8), as well as with large changes in the position of the KE jet (Suga and Hanawa 1990) that accompany different dynamic states and path variations of the Kuroshio. Equation (2) shows that the water mass transformation will be strong when both the area of the outcrop window and the surface ocean buoyancy loss over the outcrop window are large. The location of the NPSTMW outcrop window during the time of the strongest ocean heat loss, and the location of the strongest ocean heat loss, tend to be better aligned with each other in years when the KE jet is stable (2005, 2011, and 2005; Figs. 8a,g,h) than in years when the KE is unstable (which is especially the case in 2009; Fig. 8e). Additionally, the area of the outcrop window is smaller in years when the KE is unstable (2008–09) than in the years when the KE is stable (2005 and 2011–2012; Figs. 7 and 9b), indicating that the thermodynamic processes in the KE region during the period of wintertime NPSTMW formation depends on the dynamic processes in the formation region and that the two cannot be separated.

Our results are thus consistent with the results of Rainville et al. (2007), who showed that, in years when the KE is unstable, the volume of NPSTMW formed in winter is smaller than in years when the KE is stable, because the area of the NPSTMW outcrop window is smaller. Our results additionally show that, in years when the KE is unstable, the location of the NPSTMW outcrop window during the time of the strongest wintertime ocean heat loss tends not to be aligned with the location of strongest wintertime ocean heat loss as well as it is in years when the KE jet is stable. This can be
partly attributed to the spatially convoluted path of the KE jet in the unstable state, as was the case in 2009 (Fig. 8e).

b. Defining the NPSTMW surface formation region

After the NPSTMW subduction ceases, a seasonal thermocline develops in April, and the top 200 m of the water column restratifies, isolating the NPSTMW low-PV pool from the atmosphere (Rainville et al. 2014). At that time, surface water in the NPSTMW density range, although often still south of 40°N, loses contact with water in the low-PV pool (Fig. 10). The surface buoyancy flux acting in the NPSTMW density range consequently no longer directly changes the volume of water in the low-PV pool and should no longer be taken into account in the Walin calculation. This happens in spring, when surface water in the NPSTMW density range is gaining heat, and is being transformed into lighter water, isolating the NPSTMW low-PV pool from the atmosphere. Formation maps, such as those shown in Fig. 7, consequently show destruction of water in the NPSTMW density range in the northernmost part of the analysis region, causing the NPSTMW formation rate estimates to be biased low.

In our analysis, this was especially the case in year 2005. A cross section of stratification at 145°E (Fig. 10) reveals that, in 2005, surface water in the NPSTMW density range, although no longer in contact with the low-PV pool, was still south of 40°N even in June. In 2005, the NPSTMW outcrop window was large, and the wintertime air–sea heat loss was not anomalously low (Fig. 8a). Yet carrying out the Walin analysis over the entire meridional extent of the analysis region 20°–40°N yields a very small formation rate of 2.4 Sverdrups (Sv; 1 Sv = 10⁶ m³ s⁻¹), much smaller than the 2005–12 time average of 4.5 Sv.

Fig. 9. (a) Monthly averaged area of the NPSTMW outcrop window (10¹² m²) for the potential density range $\sigma_\theta = 24.6–25.6$ kg m⁻³ (color; right bar) from the gridded Argo product over all points within the analysis region of Fig. 2 at which the temperature at 300 m is higher than 12°C, hence only points within the subtropical gyre (Mizuno and White 1983) but without imposing the low-PV constraint (see section 4b). (b) Formation rate (Sv) of water in the NPSTMW density range in the subtropical gyre (green), and the area of the outcrop window [blue; as in (a), but with units of 10¹¹ m²]; (c) formation rate (Sv) of water in the NPSTMW density range in the subtropical gyre (green) as in (b), and area-averaged ocean buoyancy loss over the outcrop window in the NPSTMW density range in the subtropical gyre (blue), obtained as the ratio of the transformation rate [Eq. (2)] and the area of the outcrop window (W m⁻²).
The most obvious solution would be to carry out the Walin analysis only over the surface area where PV, $10^2$ $m^2 s^{-1}$ and $25.5 \text{ kg m}^{-3}$ (defining the edges of the NPSTMW low-PV pool) are indicated as gray lines.

The most obvious solution would be to carry out the Walin analysis only over the surface area where PV $< 2 \times 10^{-10} \text{ m}^{-1} \text{ s}^{-1}$, ensuring that the outcrop window is in contact with the NPSTMW low-PV pool. We cannot do this because we only have monthly mean Argo temperature data below the surface from which to estimate PV. Our solution was to use the daily ERA-interim SST (relying on Argo only for a surface salinity field that is interpolated to daily values) and to confine the area over which the surface buoyancy flux integration should be carried out to be within the subtropical gyre, south of the KE, by using the proxy of Mizuno and White (1983), who define the northern edge of the subtropical gyre using the location of the $12^\circ C$ isotherm as a proxy for the KE path. The resulting NPSTMW formation rate estimates are almost always larger than those obtained by considering the entire meridional extent of the analysis region: $5.7 \pm 1.2 \text{ Sv}$ versus $4.5 \pm 1.2 \text{ Sv}$ (Table 2). Only in 2009 was the formation rate estimate over the standard analysis region ($4.0 \text{ Sv}$) slightly larger than over the subtropical gyre ($3.7 \text{ Sv}$). Since the KE was unstable in 2009, the KE path and the shape of the outcrop window were very convoluted (Fig. 8e), and it is possible that the approximation used here did not accurately trace the edges of the pool.

To simplify the choice of the region over which the buoyancy flux integration should be carried out, we use the result of Qiu and Chen (2010) that the KE axis in the NPSTMW formation region is located south of $37^\circ N$. We have thus carried out a Walin analysis over a region that coincides with the original analysis region, except it extends to $37^\circ N$ instead of $40^\circ N$. The results of this analysis are very similar to the results obtained by considering only the buoyancy flux over the subtropical gyre [Table 2, (ii) and (iii)], yielding an averaged formation of $5.2 \pm 0.9 \text{ Sv}$. Hence, in what follows, we describe results obtained considering only the subtropical gyre region, using the proxy of Mizuno and White (1983). Formation rate estimates obtained by integrating to $37^\circ N$ would yield the same conclusions. In section 4d, we discuss the volume budget in the NPSTMW density range in the region $20^\circ^\circ N$, $125^\circ E \rightarrow 160^\circ W$, east of the islands of Japan, and use formation rates estimates obtained over the same region.

![Fig. 10. Meridional section at 145°E of the (negative) vertical gradient of potential density from the gridded Argo product for June 2005: the negative vertical gradient $-\frac{d\sigma_v}{dz}$ is shown in color ($\text{kg m}^{-4}$), and potential density isolines at $\sigma_v = 25.0 \text{ kg m}^{-3}$ and $25.5 \text{ kg m}^{-3}$ (defining the edges of the NPSTMW low-PV pool) are indicated as gray lines.](image)

Table 2. (i) Annually averaged formation rate from Walin (1982) analysis due to air–sea buoyancy fluxes, of water in the NPSTMW density range $25.0 \leq \sigma_v \leq 25.5$, without imposing the low-PV constraint (see text) in the analysis region of Fig. 2 bounded by $20^\circ–40^\circ N$, $125^\circ E \rightarrow 180^\circ$. Inputs to the Walin (1982) analysis [Eqs. (2)–(4)] were the following: daily averaged net air–sea heat flux, evaporation, precipitation, and sea surface temperature all from ERA-Interim, combined with monthly averaged Argo near-surface salinity (at 2.5-m depth). (ii) As in (i), but additionally requiring that the temperature at 300-m depth is higher than $12^\circ C$ (Mizuno and White 1983). (iii) As in (i), but within an analysis region like that of Fig. 2 but reduced to the narrower latitude range $20^\circ–37^\circ N$ in order to include only the region south of the KE jet axis as defined by Qiu and Chen (2010).

<table>
<thead>
<tr>
<th>Year</th>
<th>(i) Annually averaged formation rate (Sv) due to air–sea buoyancy fluxes within the formation region ($20^\circ–40^\circ N, 125^\circ E \rightarrow 180^\circ$)</th>
<th>(ii) As in (i), but additionally requiring that temperature at 300 m exceed $12^\circ C$ (Mizuno and White 1983)</th>
<th>(iii) As in (i), but within the latitudinally restricted formation region ($20^\circ–37^\circ N, 125^\circ E \rightarrow 180^\circ$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2005</td>
<td>2.4</td>
<td>5.4</td>
<td>5.3</td>
</tr>
<tr>
<td>2006</td>
<td>4.6</td>
<td>4.7</td>
<td>4.3</td>
</tr>
<tr>
<td>2007</td>
<td>4.2</td>
<td>6.4</td>
<td>5.0</td>
</tr>
<tr>
<td>2008</td>
<td>5.4</td>
<td>6.2</td>
<td>5.6</td>
</tr>
<tr>
<td>2009</td>
<td>4.0</td>
<td>3.7</td>
<td>3.9</td>
</tr>
<tr>
<td>2010</td>
<td>4.0</td>
<td>5.5</td>
<td>4.9</td>
</tr>
<tr>
<td>2011</td>
<td>6.2</td>
<td>7.6</td>
<td>6.7</td>
</tr>
<tr>
<td>2012</td>
<td>5.5</td>
<td>6.4</td>
<td>5.7</td>
</tr>
<tr>
<td>Average</td>
<td>4.5 $\pm$ 1.2</td>
<td>5.7 $\pm$ 1.2</td>
<td>5.2 $\pm$ 0.9</td>
</tr>
</tbody>
</table>
c. The NPSTMW transformation and formation rate estimates

In each of the years considered, the annually averaged surface formation rates have a pronounced peak within the NPSTMW $\sigma_\theta$ density range 25.0–25.5 kg m$^{-3}$ (Figs. 11b,d). In 2005, the strongest surface formation is at $\sigma_\theta = 25.2$ kg m$^{-3}$. In 2006 and 2008–10, when the density of the low-PV pool is anomalously low (Figs. 3a), the strongest surface formation is at somewhat lighter densities (at $\sigma_\theta = 25.1$ kg m$^{-3}$). In 2011 and 2012, the strongest formation is again at $\sigma_\theta = 25.2$ kg m$^{-3}$, as at the beginning of the period analyzed here. The density change of the peak in surface formation seems to be a result of oceanic preconditioning. The wintertime sea surface potential density $\sigma_\theta$ anomaly for each year, obtained from Argo $T$ and $S$ as a deviation of the wintertime $\sigma_\theta$ for each year from its 2006–12 time mean, displays a large scale, basinwide pattern of sea surface density variability (Fig. 12). In the NPSTMW formation region, sea surface density decreased during 2007–10 (Figs. 12b–e) so that the strongest wintertime ocean heat loss occurred over progressively lighter outcrops. The wintertime sea surface density in the NPSTMW formation region was lowest in 2008–10 (Figs. 12c–e), suggesting that this density decrease, which was part of basinwide density changes, provided oceanic preconditioning for preferential surface formation of a lighter variety of NPSTMW in these years (Figs. 11b,d). Furthermore, the preferential surface formation of a lighter variety of the NPSTMW in the years 2008–10 replenished only the lighter water in the low-PV pool, further decreasing the

FIG. 11. Annually averaged (a),(c) surface transformation rate $A = F$ [Eq. (2)] and (b),(d) formation rate [Eq. (4)] obtained using the Walin (1982) framework during the years 2005–12 over the subtropical gyre within the analysis region of Fig. 2 without the low-PV constraint (units: Sv). Inputs to the Walin (1982) analysis are as follows: daily averaged ERA-Interim air–sea buoyancy flux $B$ and daily averaged sea surface potential density $\sigma_\theta$, obtained from daily averaged ERA-Interim sea surface temperature combined with monthly averaged Argo near-surface salinity.
The surface formation rate of water in the NPSTMW density range was low in 2006 (4.7 Sv); it increased in 2007 (6.4 Sv) but decreased again in years 2008 (6.2 Sv) and 2009 (3.7 Sv) [Table 2(ii)]. Surface formation across the entire NPSTMW density range was smallest in 2009, which was the result of the combined effect of the small area of the outcrop window (Fig. 9b) and of weak wintertime ocean buoyancy loss over the NPSTMW outcrop window (Fig. 9c). A decrease in surface formation in the years 2007–09 [Table 2(ii)] has thus most likely contributed to the observed decrease of newly formed NPSTMW volume during this time, especially for the strong decrease in 2009 (Fig. 5a). Compared to 2009, the surface formation rate increased in 2010 (5.5 Sv), especially for the denser variety of NPSTMW, with $\sigma_\theta$ between 25.1 and 25.4 kg m$^{-3}$ (Fig. 11d). The increase in surface formation in 2010 was caused by an increase in the area of the outcrop window, while the surface ocean buoyancy loss over the NPSTMW outcrop remained relatively weak (Fig. 9b,c). The increase in the area of the NPSTMW outcrop window in 2010 was most likely a consequence of the KE state switching from unstable to stable in 2010. Thus, increased surface formation likely played an important role in the 2010 increase of newly formed NPSTMW volume (Fig. 5a) as well as the increase of isopycnal volume in the NPSTMW density range (Figs. 6c–j). Both changes in the outcrop area in the NPSTMW density range and variations of air–sea buoyancy flux over the outcrop window play an important role in the interannual variability of NPSTMW formation in the years analyzed (Fig. 9).

d. The NPSTMW volume budget

To more accurately relate interannual variability in surface formation and in volume in the NPSTMW density range, we next consider the volume budget in the
NPSTMW density range in the analysis region, restricted however to the 20°–37°N meridional range (which yielded more accurate surface formation rate estimates than those obtained for the original 20°–40°N meridional range). Because our analysis region has open boundaries on all sides, water formed at the surface could have either subducted into the ocean interior or been advected through the sides of the analysis region. We have therefore computed zonal and meridional geostrophic velocities (referenced to 1975 dbar) using the gridded Argo T and S fields in order to estimate volume transports in and out the sides of the analysis region in the NPSTMW density range. Elements of the volume budget in the NPSTMW density range, in the region 20°–37°N, 125°E–160°W, east of the islands of Japan, are shown in Fig. 13.

The time rate of change of volume in the NPSTMW density range is well correlated with variability in surface formation in the sense that years with larger (smaller) volume tend to coincide with years with larger (smaller) surface formation (Fig. 13b), in agreement with Davis et al. (2011) and Rainville et al. (2014). The variability of the NPSTMW volume does not correlate so well with the variability in surface formation. This should be expected, as its variability is additionally determined by PV variability, demonstrated to have a strong influence on the NPSTMW volume variability (Figs. 3a,b). For the same reason, its variability is higher than the variability of water in the NPSTMW density range.

Both net zonal and net meridional transport divergence (given by the difference between the transport in and out of the region analyzed here) are negative, decreasing the volume of water in the NPSTMW density range, as this water is predominantly advected east and south by the subtropical gyre circulation. Thus, accounting for volume transport through the open boundaries reduces the difference between volume variability and surface formation. This difference is predominantly positive (with a 2005–12 average of 5.6 Sv), suggesting the volume budget as analyzed here is lacking processes that destroy water in the NPSTMW density range. On account of the coarse resolution of observations used in this estimate, both in space and time, the difference includes errors in representation of smaller-scale processes that can cause destruction of water in the NPSTMW density range, such as diapycnal ocean mixing or divergence of diapycnal eddy buoyancy flux (Nishikawa et al. 2013) (Figs. 13d,e).

The time variability of water in the NPSTMW density range can be substantially different from the time variability of NPSTMW (Figs. 3a,b and 13a,b). Therefore, the Walin analysis, although providing a natural isopycnal framework for water mass analysis, is not the most suitable tool to analyze the volume budget of a low-PV pool since it does not account for the role of PV fluxes. A study along the lines of the study of the Eighteen Degree Water by Deremble and Dewar (2013) would provide a more complete and accurate description of NPSTMW variability.

5. The interannual variability of density and PV in the region of the NPSTMW low-PV pool

Decadal variability of the NPSTMW volume, such as that shown in Fig. 3, has been linked to basinwide changes of both atmospheric forcing and oceanic properties associated with the PDO, both in observations (Qiu and Chen 2006) and numerical model results (Davis et al. 2011). It has been shown that PDO-related changes in atmospheric forcing generate anomalies in the SSH and in the permanent thermocline depth (Qiu et al. 2007) that modify the background stratification and PV (Sugimoto and Hanawa 2010). These anomalies all propagate westward from the central Pacific in the form of a first-mode baroclinic Rossby wave (e.g., Oka and Qiu 2012).

Qiu et al. (2007) showed that a change in the PDO index from negative to positive in 2002, and the associated shift in the basin-scale surface wind forcing, produced a negative SSH anomaly and a negative thermocline depth anomaly, both propagating westward and causing a density increase, a PV increase, and an increase in thickness of the NPSTMW low-PV pool in the years 2005–06. This is in agreement with the Argo observations considered here: at the beginning of the analyzed period, in the years 2005–06, sea surface density was anomalously high (Fig. 12) and the PV was anomalously low (Fig. 14) in the region of interest, and the bottom of the NPSTMW layer was deep (Fig. 4).

Analysis of the NPSTMW west of 160°E in the ECCO2 output showed that the NPSTMW volume decrease in years 1979, 1988, and 1999 coincided with a change in the PDO index from positive to negative, with a zero time lag (Davis et al. 2011). The present analysis of the NPSTMW in the region west of 160°E shows that the NPSTMW volume decrease that started in 2006 followed a change in the PDO index from positive to neutral in 2005.

Time–longitude plots of the PV anomaly in the NPSTMW σθ range 25.0–25.5 kg m⁻³, averaged over the 32°–34°N latitude range, show anomalously high PV in the eastern and central Pacific, which started to propagate westward from the central Pacific from approximately 160°–170°W in 2005, increasing the PV in the
FIG. 13. Elements of the volume budget in the NPSTMW density range in the region 20°–37°N, 125°E–160°W, east of Japan: (a) the time rate of change of volume of NPSTMW with the PV constraint (thin gray line) and surface formation rate of water in the NPSTMW density range in the same geographical region (thick dark line); (b) as in (a), but for the time rate of change of volume of NPSTMW without the PV constraint (thin gray line); (c) the time rate of change of volume of NPSTMW with the PV constraint minus the formation rate of water in the NPSTMW density range (thick black line), and the time rate of change of volume of NPSTMW without PV constraint minus the formation rate of water in the NPSTMW density range (thin gray line); (d) the time rate of change of volume in the NPSTMW density range without the PV constraint minus the formation rate of water in the NPSTMW density range [also shown in (c); thin gray line], and divergence of zonal (dotted black line) and meridional (solid black line) volume transport into the analysis region, where positive transport divergence increases volume of water in the NPSTMW density range in the analysis region; and (e) 3-month smoothed time rate of change of volume of NPSTMW without the PV constraint minus the surface formation rate of water in the NPSTMW density range (solid black line), and 3-month smoothed divergence shown in (d) (units: Sv). Note the different vertical scales between (a)–(b) and (c)–(e).
region of the NPSTMW low-PV pool in years 2006–10 (Fig. 15). This was followed by anomalously low-PV, which started to propagate westward from the central Pacific in 2008–09, decreasing the PV in the region of the NPSTMW low-PV pool in years 2009–12 (Fig. 15).

We have chosen to average the PV anomaly over the latitude range $32^\circ–34^\circ$N in order to readily compare our results with time–longitude plots derived from SSH altimetric observations by Qiu et al. (2007, their Fig. 5). Qiu et al. (2007) chose the $32^\circ–34^\circ$N latitude range...
because it is characterized by high wind variability. Our results were not sensitive to the choice of latitude range over which PV was averaged, as long as it included the latitude range of the low-PV pool.

Sugimoto and Hanawa (2010) showed that the thickness variation of the NPSTMW layer is predominantly controlled by the variation of the main thermocline depth (as in Fig. 4), and Qiu et al. (2007) similarly showed that the decrease in NPSTMW layer thickness was predominantly caused by shoaling of the base of the NPSTMW low-PV pool, defined by the $\sigma_\theta = 25.5 \text{ kg m}^{-3}$ isopycnal. The time–longitude depth variation of the $\sigma_\theta = 25.5 \text{ kg m}^{-3}$ isopycnal (Fig. 16) shows that the positive PV anomaly that started to propagate westward from the central Pacific in 2005 (Fig. 15) indeed coincided with the shoaling of the bottom of the NPSTMW layer decreasing the thickness of the low-PV pool in years 2005–09. Similarly, the subsequent negative PV anomaly coincided with the deepening of the bottom of the NPSTMW layer, which propagated westward, deepening the low-PV pool (Fig. 16).

Figure 4e shows that deepening of the top of the NPSTMW low-PV pool, defined by the $\sigma_\theta = 25.0 \text{ kg m}^{-3}$ isopycnal, caused the PV in the low-PV pool to attain its highest values. A time–longitude plot of the near-surface density anomaly (Fig. 17), averaged over the top 30 m and in the latitude range 32°–34°N, shows that, in mid-2007, when the PDO index switched from neutral to negative, a negative density anomaly started to propagate westward from the central Pacific, around 170°W–180°, decreasing density in the NPSTMW low-PV pool in the years 2008–10. This density decrease contributed to the decrease of the volume of the NPSTMW low-PV pool in several ways: first, by transforming water from the NPSTMW density range into lighter water; second, by deepening the top of the NPSTMW layer in 2009 and 2010 (Figs. 18 and 4e,f), which further increased PV of the NPSTMW low-PV pool, thus decreasing the volume of water satisfying the NPSTMW low-PV constraint; and finally, as mentioned in the previous section, the negative density anomaly provided oceanic preconditioning for preferential surface formation of a lighter...
variety of NPSTMW, further decreasing the density of the pool.

A particularly clear result of this analysis is that variations of NPSTMW volume are a local manifestation of basinwide shifts in stratification and PV.

6. Summary and discussion

Our analysis of monthly temperature and salinity data from the gridded Argo product of Roemmich and Gilson (2009) for the period 2005–12 shows a strong North Pacific Subtropical Mode Water (NPSTMW) volume and density decrease in 2006–09, as previously observed by Rainville et al. (2014). Rainville et al. (2014) attributed the volume decrease predominantly to a decrease in surface formation. We show that the time rate of change of volume in the NPSTMW density range (estimated without imposing the low-PV constraint) tends to be correlated with the variability in surface formation, in the sense that years with larger (smaller) volume tend to coincide with years with larger (smaller) surface formation of water in the NPSTMW density range. Correlation between the time rate of change of NPSTMW volume (estimated imposing the low-PV constraint) and surface formation is, instead, significantly lower. This result highlights the importance of the PV constraint in defining mode waters. Since mode waters are defined by both density and PV constraints, their evolution cannot be described from a density budget alone (or equivalently its volume budget); rather, the PV budget must be considered simultaneously.

The large variability in the meridional position of the KE jet accompanying different dynamic states and path variations of the KE results in large variability in the meridional position of the NPSTMW surface formation region. Surface formation rates of water in the NPSTMW density range estimated over the commonly considered area bounded by 20°–40°N, 125°–160°E were significantly lower than those estimated over the meridionally more restricted range 20°–37°N. This 20°–37°N range was chosen to exclude regions of surface formation/destruction of water that are in the NPSTMW density range, but are no longer in direct contact with the NPSTMW pool. Surface formation rate estimates obtained by requiring that all points at which surface formation was computed lie within the subtropical gyre equatorward of the KE jet, or lie south of 37°N, were much better correlated with the time variability of volume in the NPSTMW density range compared to the original formation rate estimates obtained considering the entire 20°–40°N, 125°–160°E region.

At the beginning of the period analyzed, in 2005, the base of the NPSTMW layer was deep, the NPSTMW
volume was large, the density in the pool was high, and the PV was low, in agreement with Qiu et al. (2007) and Rainville et al. (2014). The bottom of the NPSTMW layer started to shoal in 2005, when the PDO index switched from positive to neutral. A negative depth anomaly of the bottom of the NPSTMW layer originated in the central Pacific in 2005 and propagated westward. When it reached the NPSTMW low-PV pool, the pool’s thickness decreased, and its PV increased. This PV increase was particularly high in the years 2009–10, a time when the top of the NPSTMW layer additionally deepened because of warming and freshening of the water above it. The observed warming was related to the arrival of a negative density anomaly that developed in the central North Pacific (at about 170°W–180°) in mid-2007, when the PDO index changed from neutral to negative. This anomaly rapidly moved westward, affecting the NPSTMW low-PV pool most strongly in 2008–09: that is, during the time period when the NPSTMW reached its minimum. A strong surface density decrease culminating in 2009 contributed to the NPSTMW volume decrease in several ways. First, some water from the NPSTMW density range was transformed into lighter water. Second, deepening of the top of the NPSTMW layer increased its PV above the value that is used as the NPSTMW low-PV constraint. Finally, the surface density decrease provided oceanic preconditioning for preferential surface formation of a lighter variety of NPSTMW so that only the lighter variety of NPSTMW was replenished, further decreasing the density of the pool.

Surface formation in the NPSTMW density range was smallest in 2009, when the NPSTMW volume reached its minimum. In 2009, both the NPSTMW outcrop window and the wintertime buoyancy loss over it were anomalously low. Also, in 2009, when the KE was unstable, the location of the NPSTMW outcrop window was not as well aligned with the location of strongest wintertime ocean heat loss as it was in years when the KE jet was stable, thus decreasing water mass transformation in the NPSTMW density range.

This study clearly cannot answer all the relevant questions raised in the extensive literature about various processes that govern the quasi-decadal variability of NPSTMW volume because of the limited length of the Argo record. Also, because of their coarse resolution, Argo observations do not represent some processes that were shown to be important for the variability of the NPSTMW volume, such as eddy processes (e.g., Nishikawa et al. 2013). Until a longer and denser array of Argo observations is available, more insight must be sought from high-resolution, long numerical model simulations.
Acknowledgments. We are grateful to two anonymous reviewers for their insightful comments, which substantially improved the paper. We are also grateful to D. Roemmich, A. Subramanian, W. K. Dewar, A. J. Miller, S.-P. Xie, and J. McClean for motivating discussions. IC was supported by NSF Ocean Sciences Grant OCE-0850463 and DOE Grant DE-SC0001933 to Scripps Institution of Oceanography. DG’s participation in the Argo Program was supported by U.S. Argo through NOAA Grant NA10OAR4320156 (SIO CIMEC). The statements, findings, conclusions, and recommendations herein are those of the authors and do not necessarily reflect the views of the National Oceanic and Atmospheric Administration or the Department of Commerce. The efforts of many international partners in planning and implementing the Argo array are gratefully acknowledged. The objectively mapped Argo data were provided by John Gilson. The ERA-Interim data used in this study are from the European Centre for Medium-Range Weather Forecasts (ECMWF 2015).

REFERENCES


FIG. 18. As in Fig. 15, but (a) is for the depth anomaly (m) of the $\sigma_2 = 25.0 \text{ kg m}^{-3}$ isopycnal surface that defines the top of the NPSTMW layer.