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SPECIAL SECTION: REVIEW

New developments in mode-water research: Dynamic and climatic effects

Progress of North Pacific mode water research in the past decade

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Abstract This article reviews the progress in research on North Pacific mode waters of the past decade from the physical oceanographic perspective. The accumulation of satellite altimeter sea surface height data, the rapid growth of the Argo profiling float array, and the advancement in eddy-resolving ocean general circulation models have greatly improved the traditional views on the mode waters that were formed prior to the 1990s based on the analyses of historical temperature/salinity data. Areas where significant progress was made include: (1) descriptions of the mode waters' distributions and properties with fine spatial scales, particularly in their formation regions in winter where observational data had been insufficient; (2) clarifications of the mode waters' formation and subduction processes relating to the large-scale mean circulation, as well as to the time-varying mesoscale eddy field; (3) impacts of the mode waters' circulation and dissipation processes on the climate and biogeochemical processes; and (4) dynamic versus thermodynamic causes underlying the mode waters' decadal changes. In addition to the review, future directions for mode water research are also presented.

Keywords Mode waters · North Pacific · Physical oceanography · Seasonal to decadal variability · Eddy modifications · Climate and biogeochemical impacts

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1 Introduction

In the subtropical gyres and part of the subpolar gyres of the world oceans, there are distinct water masses characterized by a nearly vertically homogeneous layer or low potential vorticity (PV) lying just above or within the permanent pycnocline. These water masses, called "mode waters," are formed as a deep mixed layer on the warm side of a strong current or front in late winter because of convective mixing, and are left in the subsurface as a pycnostad after being capped by the seasonal pycnocline in spring. Some part of the formed mode waters is entrained into the mixed layer in the following winter, modulating the sea surface temperature, while the remaining part enters the permanent pycnocline (this process is called subduction) and is then advected to spread over a much wider area than its formation regions, carrying temperature, salinity, and PV anomalies. The formation, circulation, and dissipation of mode waters and their variability are related to ocean-atmosphere interaction and various upper-ocean dynamic and thermodynamic processes on a wide range of time scales. Moreover, they are believed to play an important role in biogeochemical processes such as the oceanic uptake of atmospheric CO_2 (e.g., Bates et al. 2002) and the nutrient cycling in the oligotrophic subtropical gyres (e.g., Palter et al. 2005; Krémeur et al. 2009; Sukigara et al. 2011).

The subtropical gyre in each ocean basin contains three types of mode waters formed in the western, eastern, and poleward parts of the gyre, except that the South Hemisphere gyres share the Subantarctic Mode Water formed just north of their common southern boundary, the subantarctic front (Hanawa and Talley 2001). In the North Pacific, a thermostad of 16–18°C lying in the upper permanent thermocline in the northwestern part of the subtropical gyre was identified and named the subtropical mode water (STMW) by Masuzawa (1969) because it produces a mode in the volume distribution on the temperature-salinity diagram for the subtropical gyre.[1] Since then STMW was little explored for nearly 2 decades, but drew attention again in the late 1980s when the Ocean Mixed Layer Experiment (Toba et al. 1991) was implemented as one of the Japanese activities in the World Climate Research Program. Possible roles of STMW as a heat reservoir memorizing the wintertime ocean-atmosphere interaction have inspired a number of studies based on shipboard observations and data analyses that helped clarify the formation, circulation, and long-term variation of STMW.

STMW is formed as a deep winter mixed layer just south of the Kuroshio and the Kuroshio Extension (KE) between ~132°E and near the dateline, where warm surface water supply from the Kuroshio/KE underlying the strong East Asian winter monsoon results in large oceanic buoyancy loss in winter (Hanawa 1987; Hanawa and Hoshino 1988; Suga and Hanawa 1990; Bingham 1992). Due to the downstream cooling of the Kuroshio/KE, colder and denser STMW tends to be formed progressively in the farther eastern part of the formation region (Suga and Hanawa 1990; Bingham 1992). After spring, the main body of STMW, capped by the seasonal pycnocline and isolated from atmospheric contact, is advected southwestward by the "Kuroshio Countercurrent" and is subducted into the permanent pycnocline (Hanawa 1987; Suga et al. 1989; Suga and Hanawa 1990, 1995a; Bingham 1992).

The volume and properties of STMW formed every winter exhibit significant year-to-year variations (Hanawa 1987; Suga et al. 1989). On interannual time scales, thicker and colder STMW tends to be formed in years with stronger winter monsoons because of larger oceanic heat loss and increased southward Ekman transport (Suga and Hanawa 1995b; Yasuda and Hanawa 1997, 1999; Taneda et al. 2000; Hanawa and Kamada 2001; Hanawa and Yoritaka 2001). On decadal to interdecadal time scales, on the other hand, intensification of the westerlies leads to the formation of warmer STMW through gyre spin-up and increased warm water advection by the Kuroshio (Hanawa and Kamada 2001). The local existence of STMW south of Japan is also influenced by path variations of the Kuroshio (Yoshida 1964; Kawabe 1995). The westward advection of STMW from the region southeast of Japan is blocked when the Kuroshio takes a large meander path (Bingham et al. 1992; Suga and Hanawa 1995b, c).

In the late 1990s, the other two mode waters in the North Pacific were identified in succession, as expected from isopycnal PV maps presented in an earlier study (Talley 1988). Nakamura (1996) and Suga et al. (1997) separately described a thermostad of 8.5–11.5 and 10–13°C lying in the lower permanent pycnocline in the central part of the subtropical gyre using different climatological data and termed it the central mode water (CMW). They inferred that CMW is formed between KE and its northern bifurcation (Suga et al. 1997) and between the northern bifurcation and the subarctic front[2] (Nakamura 1996), which renders CMW much colder, fresher, and denser than STMW. The eastern subtropical mode water (ESTMW) was identified by Hautala and Roemmich (1998) as a weak thermostad of 16–22°C lying in the upper permanent pycnocline in the eastern part of the subtropical gyre. It is formed as a relatively deep winter mixed layer southwest of the subarctic front, which extends in the northwest-southeast direction in this region (Yuan and Talley 1996; Suga et al. 2004). It is often stratified in terms of temperature and salinity in a compensating way within its pycnostad, possibly because of cross-frontal intrusions (Hautala and Roemmich 1998; Sprintall and Roemmich 1999).

The two new mode waters formed and advected anticyclonically in the central to eastern part of the gyre contrast with the classical STMW that is largely confined to the Kuroshio recirculation region in the northwestern part of the gyre. They are interesting from two points of view (Hautala and Roemmich 1998; Hanawa and Talley 2001). First, their low PV signature is a good tracer of the ventilation process in the subtropical gyre. Second, if their volume and properties formed every winter change from year to year, their advection in the permanent pycnocline might transport temperature and salinity anomalies from the mid-latitude ocean surface to the equatorial upwelling region and cause interdecadal climate variability, as hypothesized by Hanawa (1996) and Gu and Philander

[1] The term "mode water," introduced by Masuzawa (1969), was subsequently applied to any thick, broadly distributed, near-surface layer characterized by low PV (Hanawa and Talley 2001). Therefore, mode waters nowadays are not necessarily characterized by uniformity in terms of temperature and salinity. In other words, temperature and salinity might be stratified in a compensating way within a pycnostad of mode waters.

[2] In this review, the term "subarctic front" is used for a density-compensating front between the warmer, saltier water in the subtropics and the colder, fresher water in the subarctics, characterized by the outcrop of the 33.0–33.8 isohalines (Roden 1970, 1972; Zhang and Hanawa 1993; Yuan and Talley 1996), as in the literature of physical oceanography. This front often separates into two or more fronts, particularly in the eastern North Pacific, and is also called the subarctic frontal zone. In the literature of fisheries oceanography (e.g., Favorite et al. 1976; Yasuda 2003), the term "subarctic front" is referred to the front at the southern boundary of subsurface temperature inversions characterizing the subarctics, represented by the 4°C isotherm standing almost vertically below the 100-m depth (Uda 1963; Favorite et al. 1976). This front is called the "polar front" in this review, as in the literature of physical oceanography.

(1997). An initial attempt using historical temperature data (Yasuda and Hanawa 1997) demonstrated that CMW becomes colder during the decade after the 1976/1977 regime shift (e.g., Nitta and Yamada 1989; Trenberth 1990) than the previous decade, probably because of larger oceanic heat loss and increased southward Ekman transport associated with the intensification of the westerlies. On the other hand, Hautala and Roemmich (1998) did not find a substantial change in the ESTMW volume and temperature between 1970–1979 and 1991–1997, although the lack of data during the intervening period might hinder the detection of ESTMW changes associated with the 1976/1977 regime shift.

The aforementioned studies on the North Pacific mode waters prior to around 2000 were comprehensively reviewed by Hanawa and Talley (2001). In the past decade following their review, mode water research has been advanced greatly because of the rapid development of research tools. One of the biggest factors is the accumulation of satellite-based data such as sea surface height and the development of subsurface observing systems represented by Argo (Roemmich et al. 2001). Compared to the historical data until the 1990s that had insufficient coverage away from coasts and were spatially limited to a number of repeat sections, the unprecedented Argo array of 3,000 profiling floats provides us now with temperature and salinity data down to 2,000-dbar depth at 10-day intervals with a horizontal resolution of 3° in latitude and longitude (Roemmich et al. 2009; Freeland et al. 2010). This makes it possible to take snapshots of the entire structure of mode waters at various times of year, providing new insights on their formation, circulation, dissipation, and temporal variations. In addition, the long-term satellite-based data have revealed the variability of currents/fronts and eddy fields that are closely related to mode water variability, particularly in the energetic Kuroshio-Oyashio Extension region (e.g., Qiu 2002) where STMW and CMW are formed (Qu et al. 2002). Another important factor is the introduction and progress of modeling studies of the North Pacific mode waters. As the model resolution improved from non-eddy-permitting to eddy-permitting and further to eddy-resolving (e.g., Hasumi et al. 2010; Masumoto 2010), the structures of the currents/fronts and the associated mode waters became reproduced more realistically, helping to understand the underlying dynamics, particularly on mesoscales that are not fully resolved by observations. It is also worth noting that while most of the studies until the 1990s were performed by a single research group at Tohoku University, Japan, their activities have stimulated many scientists to participate in the mode water research, generating a variety of views and approaches.

This paper reviews the progress of physical oceanographic research on the North Pacific mode waters in the past decade following the review of Hanawa and Talley (2001). Readers may refer to the review of Joyce (2011) in this issue, and Kelly et al. (2010), for the mode water research progress in the North Atlantic and the comparison with the North Pacific. Research progress on the formation and subduction of the North Pacific mode waters is described in Sect. 2, and that on the other processes such as their circulation, modification, and dissipation is summarized in Sect. 3. After reviewing the studies on their decadal variability in Sect. 4, we present future directions for the mode water research in Sect. 5.

2 Formation and subduction

In the classical ventilated thermocline theory (Luyten et al. 1983), water in the subtropical gyre is pushed down (or subducted) from the base of the Ekman layer into the permanent pycnocline because of Ekman pumping and is subsequently transported by the anticyclonic Sverdrup flow along isopycnal surfaces, conserving its PV. Practically, water is subducted not from the base of the Ekman layer but from that of the late winter mixed layer (Stommel 1979), which generally shoals equatorward within the subtropical gyre. Consequently, lateral induction through the sloping mixed layer base increases the subduction rate as compared to the estimation from Ekman pumping alone (Williams 1989, 1991; Marshall et al. 1993; Huang and Qiu 1994; Qiu and Huang 1995).

The PV (Q) of water subducted at the base of the winter mixed layer is expressed as,

$$Q = \frac{f}{\rho_0} \frac{\mathbf{u}_b \cdot \nabla \rho_m}{w_b + \mathbf{u}_b \cdot \nabla h} \quad (1)$$

where f is the planetary vorticity, ρ_0 the reference density, ρ_m the mixed layer density, h the mixed layer depth (MLD), $\mathbf{u}_b$ and w_b the horizontal and vertical velocities at the base of the winter mixed layer, respectively, and ∇ is the horizontal differential operator (Williams 1989, 1991). The distribution of $-w_b$ (called the vertical pumping term) is relatively uniform over the North Pacific subtropical gyre (Huang and Qiu 1994), and the water acquires low Q if $-\mathbf{u}_b \cdot \nabla h$ is large or $-\mathbf{u}_b \cdot \nabla \rho_m$ is small. In other words, mode waters are formed in the case of large lateral induction associated with an MLD front (Kubokawa and Inui 1999; Kubokawa 1999; Nishikawa and Kubokawa 2007) or small density advection associated with a small downstream gradient of the mixed layer density (Fig. 1).

In the early 2000s, the mechanisms of large-scale formation and subduction of the three mode waters in the North Pacific were examined, based mainly on the subduction theory including Eq. 1 and by non-eddy-permitting

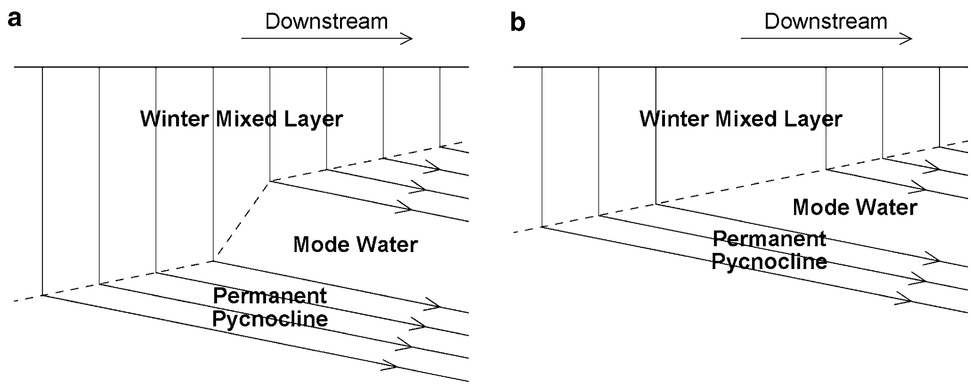

Fig. 1 Schematic illustrating the two mechanisms of mode water subduction: **a** large lateral induction and **b** small density advection. *Solid* and *dashed lines* denote isopycnals and the base of the winter mixed layer, respectively

models with a horizontal resolution of 1° or 2°. The STMW formation region is characterized by the deep winter mixed layer south of the Kuroshio and KE accompanied by a sharp MLD front at its southern end (Suga and Hanawa 1990; Oka and Suga 2003), and STMW is subducted into the permanent pycnocline by crossing this front to the south because of large lateral induction (Kubokawa and Inui 1999; Tsujino and Yasuda 2004). Similarly, CMW enters the permanent pycnocline by crossing an MLD front at the eastern end of the formation region, namely the deep winter mixed layer north of KE (Xie et al. 2000; Qu et al. 2002; Tsujino and Yasuda 2004; Hosoda et al. 2004). In contrast, the ESTMW formation region is characterized by a weak winter MLD maximum centered at 30°N, 140°W that is much shallower than those in the STMW and CMW formation regions. This maximum is formed in spite of relatively uniform, weak wintertime cooling over the eastern North Pacific because stratus clouds (Ladd and Thompson 2000, 2001) and the saline Ekman transport from the south (Toyoda et al. 2004) preclude the development of the seasonal pycnocline in summer and precondition the deepening of winter mixed layer. The ESTMW formation region is also characterized by a small horizontal gradient of the winter mixed layer density because the horizontal changes of the mixed layer temperature and salinity, between warmer, saltier water to the southwest and colder, fresher water to the northeast, compensate each other. Consequently, low PV of ESTMW is generated by the small density advection mechanism (Xie et al. 2000; Hosoda et al. 2001; Ladd and Thompson 2001). These subduction mechanisms proposed by the modeling studies are supported by a winter mixed layer climatology constructed by Suga et al. (2004) for STMW and ESTMW, but not for CMW. In their climatology, the winter MLD gradually decreases eastward in the eastern part of the CMW formation region, which implies that the water is subducted mainly because of small density advection (Suga et al. 2004, 2008). This discrepancy between the models and the observations is probably because the past models tend to produce too deep winter mixed layers in the CMW formation region, particularly in its eastern part (Ladd and Thompson 2001; Tsujino and Yasuda 2004).

As the horizontal resolution of ocean general circulation models improved from non-eddy-permitting to eddy-permitting, they better reproduced the frontal structures in the Kuroshio-Oyashio Extension region that are essential for the formation of STMW and CMW. Qu et al.'s (2002) model with a resolution of 1/4° (longitude) × 1/4° (latitude) reproduced the KE and subarctic fronts, and the formation of STMW and CMW south of the respective fronts. Furthermore, Tsujino and Yasuda's (2004) model with a resolution of 1/4° × 1/6° reproduced the northern bifurcation of KE (Kuroshio bifurcation) as well and captured the formation of two varieties of CMW, the lighter variety (L-CMW) south of the Kuroshio bifurcation front and the denser variety (D-CMW) south of the subarctic front. Although the Kuroshio bifurcation front in the model seems somewhat too strong (Tsujino and Yasuda 2004), the formation of two varieties of CMW is actually observed in a synoptic hydrographic section along 179°E (Mecking and Warner 2001) and a repeat section along 165°E (Oka and Suga 2005).

In the latter 2000s, accumulation of in situ observation data including those from Argo profiling floats enables us to capture more detailed, unsmoothed structures of mode waters at various times of year, particularly at higher latitudes in winter where an insufficient number of historical data had existed because of unfavorable weather conditions for shipboard measurements. An intense hydrographic survey carried out northeast of Japan in July 2002 revealed a new type of pycnostad with $\theta = 5$–7°C, $S = 33.5$–33.9, and $\sigma_\theta = 26.5$–26.6 kg m^{-3} (θ and σ_θ are potential temperature and density; S is salinity) around 43°N, 160°E south of the subarctic front and the polar front[3] (Saito et al. 2007), as suggested by earlier studies[4] (Yasuda 2003; Oka

[3] An intense hydrographic survey conducted in spring 2003 indicates that the subarctic front and the polar front coincide with each other in this longitude range (Eitarou Oka, personal communication 2011).

[4] In Yasuda (2003), TRMW is referred to as "Dense Central Mode Water."

and Suga 2005). This pycnostad is colder, fresher, and a little denser than D-CMW, and was named the transition region mode water (TRMW) by Saito et al. (2007) because its formation region corresponds to the transition region described in the literature of fisheries oceanography (e.g., Favorite et al. 1976; Yasuda 2003). TRMW is often stratified in terms of temperature and salinity within its pycnostad as in the case of ESTMW, probably because of isopycnal intrusion that characterizes the transition region (Saito et al. 2007).

The TRMW distribution region presented by Saito et al. (2007) corresponds to the region where the deepest winter mixed layer in the North Pacific appears in previous climatologies (Huang and Qiu 1994; Ladd and Thompson 2000; Suga et al. 2004), although MLD there might be overestimated because the smoothing process in the climatologies mixes different water types across the subarctic front, causing artificial cabbeling near the sea surface (Ohno et al. 2004; Suga et al. 2004). This MLD maximum is possibly generated for the following reasons. First, it lies in the "stability gap" described by Roden (1970) and Yuan and Talley (1996), a zonal band of lateral minimum in the vertical stability located south of the subarctic front (Ladd and Thompson 2000; Suga et al. 2004; Saito et al. 2007). Second, the quasi-stationary jet associated with the polar front (Isoguchi et al. 2006) brings warmer and saltier water from the south, enhancing the oceanic heat loss and making the surface water denser in winter (Saito et al. 2007). Actually, recently developed high-resolution surface heat flux data and Argo float data revealed locally enhanced oceanic heat loss and deep mixed layer formation just southeast of the polar front in winter (Hiroyuki Tomita, personal communication, 2011).

The entire structure of the CMW and TRMW formation regions and the subduction mechanism of these waters were presented by a recent study using Argo float data during 2003–2008 (Oka et al. 2011a). To the north of KE, two zonally elongated regions of deep winter mixed layer extend along 33°–39°N and 39°–43°N from the east coast of Japan (~142°E) to ~160°W. The southern region corresponds to the formation region of L-CMW, while the northern region to that of D-CMW and TRMW, in the western (eastern) part of which TRMW (D-CMW) is predominantly formed (Fig. 2). Both regions extend beyond the longitude range of the Kuroshio bifurcation front to the west and east, suggesting that the two regions are not separated by the bifurcation front, but rather enhance it, because these two adjacent low PV waters should accompany sharply inclined isopycnals in between. From the eastern part of both regions east of 170°E where the winter mixed layer becomes gradually shallower, warmer, and lighter to the east/downstream, L-CMW with $\theta = 11$–15°C, $S = 34.3$–34.6, and $\sigma_\theta = 25.7$–26.2 kg m^{-3} and D-CMW and TRMW (mostly the former) with $\theta = 7$–11°C, $S = 33.6$–34.2, and $\sigma_\theta = 26.1$–26.4 kg m^{-3} are subducted into the permanent pycnocline because of small density advection as inferred by the previous climatological studies (Suga et al. 2004, 2008). On the other hand, D-CMW and TRMW (mostly the latter) with $\sigma_\theta = 26.5$–26.6 kg m^{-3} formed only in the western part of the formation region

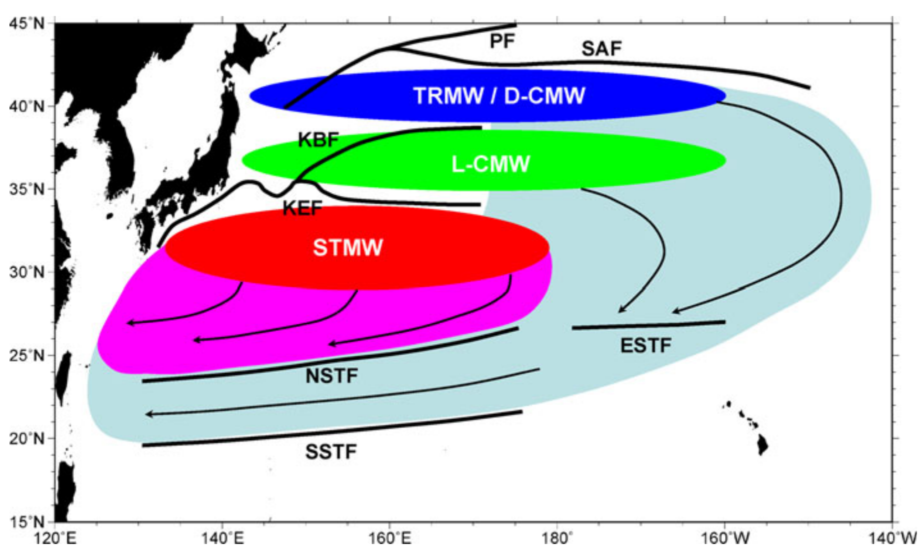

Fig. 2 Schematic illustrating the relationship between the mode waters and frontal structures in the western to central North Pacific. *PF* polar front, *SAF* subarctic front, *KBF* Kuroshio bifurcation front, *KEF* Kuroshio Extension front. NSTF, SSTF, and ESTF are the northern, southern, and eastern subtropical fronts identified in Kobashi et al. (2006), respectively. *Red*, *green*, and *blue ovals* denote the formation regions of STMW, L-CMW, and TRMW/D-CMW, respectively. *Pink* (*light blue*) *shadings* with *arrows* indicate the spreading of STMW (L-CMW and D-CMW) after subduction

where winter MLD gradually increases eastward/downstream are entrained into the mixed layer in the following winter and cannot be subducted into the permanent pycnocline, which likely explains the large discrepancy in the temperature-salinity relations between the winter mixed layer and the permanent pycnocline in this σ_θ range, argued by Suga et al. (2008). The downstream thickening of CMWs in the western part of their formation regions preconditions their permanent subduction in the eastern part, as speculated by earlier studies (Ladd and Thompson 2000; Mecking and Warner 2001).

In addition to the large-scale mode water formation and subduction mentioned above, effects of mesoscale eddies (Marshall 1997; Hazeleger and Drijfhout 2000) and the anticyclonic recirculation gyre, both of which characterize the highly variable and nonlinear KE system (e.g., Qiu 1999, 2002; Ebuchi and Hanawa 2001), on the STMW formation and subduction have been clarified using Argo float data combined with altimetric sea surface height data (Uehara et al. 2003; Pan and Liu 2005; Qiu et al. 2006; Oka 2009; Oka et al. 2011b; Kouketsu et al. 2011) and state-of-the-art, eddy-resolving models with a horizontal resolution of 1°/10° or better (Rainville et al. 2007; Nishikawa et al. 2010). The recirculation gyre consists of several anticyclonic circulations associated with crests of the Kuroshio/KE meander (e.g., Kawai 1972), in each of which thick STMW with a characteristic temperature is formed in winter, possibly because water is effectively cooled and homogenized there (Qiu et al. 2006; Rainville et al. 2007; Oka 2009; Nishikawa et al. 2010). The formation process is also modulated by prevailing mesoscale eddies; thicker (thinner) STMW tends to be formed in anticyclonic (cyclonic) eddies, in which background stratification is weaker (stronger) in association with the deeper (shallower) permanent pycnocline and oceanic heat loss is larger (smaller) because of warmer (colder) sea surface temperature (Uehara et al. 2003; Pan and Liu 2005; Rainville et al. 2007; Nishikawa et al. 2010; Kouketsu et al. 2011).

After spring, STMW continues to show highly variable distributions created by mesoscale activities, being thicker (thinner) in the anticyclonic (cyclonic) part of the flow (Uehara et al. 2003; Rainville et al. 2007; Nishikawa et al. 2010; Oka et al. 2011b). While the recirculation gyre tends to confine STMW in its formation region (Oka 2009), anticyclonic eddies transport part of it southward through the southern boundary of the region across the westward mean flow because of southward migration of eddies that trap STMW, as observed by Takikawa et al. (2005), and to eddy mixing (Rainville et al. 2007; Nishikawa et al. 2010). This also causes STMW to cross the winter MLD front and to be thereby subducted into the permanent pycnocline. Such eddy subduction contributes to approximately half of the total STMW subduction rate (Nishikawa et al. 2010). Thus, the STMW subduction revealed by Argo float data and high-resolution models is quite different from the traditional climatology-based picture that the thick body of STMW is advected by the southwestward mean flow as the season progresses (Sect. 1).

Subduction across the mean flow has also been reported for CMW. In synoptic meridional sections in the formation region, L-CMW (D-CMW) occasionally intrudes southward into the KE (Kuroshio bifurcation) front or exists as isolated patches south of the front (Mecking and Warner 2001; Oka and Suga 2005). As an extreme case, a high-resolution hydrographic survey southeast of Japan found a D-CMW patch trapped in a subsurface mesoscale eddy approximately 1,000 km south of KE in fall, which was estimated to be subducted only a half year ago in association with the instability of the KE front (Oka et al. 2009). Such cross-frontal subduction could modulate the CMW distribution determined by the large-scale subduction by the mean flow, particularly in the density range of $\sigma_\theta = 26.5$–26.6 kg m^{-3}, which is not ventilated through the large-scale CMW subduction (Oka et al. 2011a).

Furthermore, the CMW formation can also be dominated by mesoscale activities, as inferred from intermittent CMW distributions based on Argo float data from a single year (Bingham and Suga 2006) and from eddy-resolving numerical models (Nishikawa et al. 2010). In fact, a recent composite analysis of Argo float data and altimetric sea surface height data revealed that roughly half of winter mixed layers deeper than 150 dbar in the CMW formation region are formed in anticyclonic eddies (Kouketsu et al. 2011). Past shipboard observations also demonstrated that inside anticyclonic eddies (or warm-core rings) pinched off from the KE, thick STMW is modified to CMW or TRMW because of winter cooling and interaction with the ambient water while the eddies are migrating in the region north of the KE for a few years (Tomosada 1986; Yasuda et al. 1992). To further clarify the role of mesoscale activities in the CMW formation and subduction, high-resolution hydrographic surveys north of KE in conjunction with eddy-resolving models are strongly desired.

In addition to the mesoscale variability, shorter time-scale variability has also been demonstrated to affect the mode water formation. The timing of the deepest mixed layers in the North Pacific varies among regions, being February–March (February–April) in the STMW and L-CMW (D-CMW and TRMW) formation regions (Oka et al. 2007; Ohno et al. 2009). The mixed layer deepening prior to these "late winter" months is dominated by episodic events on time scales of several days that are related to the passages of synoptic low-pressure disturbances, as revealed by in situ buoy observations and high-resolution numerical model simulations (Qiu et al. 2004; Rainville et al. 2007; Cronin et al. 2008; Jensen et al. 2011). The importance of synoptic-scale atmospheric forcing is further supported by short-term MLD variations observed by profiling floats (Yoshida and Hoshimoto 2006; Oka et al. 2007), with the use of a relatively small value of temperature and density as a threshold to determine MLDs (e.g., de Boyer Montégut et al. 2004).

3 Circulation and dissipation

The development of observations and models in the past decade enabled us to investigate the circulation, modification, and dissipation of the mode waters that had been inadequately studied until the 1990s. By using data from 20 profiling floats that were concentratively deployed in the anticyclonic recirculation gyre south of the first quasi-stationary meander of the KE under the Kuroshio Extension System Study project (KESS; http://www.uskess.org), Qiu et al. (2006) explored the seasonal evolution of the vertical structure of STMW in its formation region in 2004. As the seasonal pycnocline develops and gradually thickens after April, the underlying STMW is eroded from its top, becoming thinner, whereas its bottom remains largely at a constant depth of $\sim$500 dbar. From a PV budget analysis for this STMW under the assumptions that horizontal advection is negligible (because the KE is in a stable state in 2004, as explained in Sect. 4) and that the vertical eddy diffusivity (K_v) at the lower boundary of STMW is 0.1×10^{-4} m^2 s^{-1} as previously observed in the permanent pycnocline (e.g., Gregg and Sanford 1980; Ledwell et al. 1993), K_v at the upper boundary of STMW is estimated to be 2–5 $\times$ 10^{-4} m^2 s^{-1}, which is an order of magnitude larger than that in the permanent pycnocline. A recent dissolved oxygen budget analysis using data from a profiling float equipped with a fluorometer and an oxygen sensor estimated a similar K_v value of 1.7×10^{-4} m^2 s^{-1} near the upper boundary of STMW (Sukigara et al. 2011). Such large vertical diffusivity possibly stems from the sharp drop in stratification from the seasonal pycnocline to the STMW pycnostad, which acts as a barrier for the downward transmission of internal gravity waves generated in the surface mixed layer and contributes to the enhanced eddy mixing as these waves are reflected at the upper boundary of STMW (Qiu et al. 2006). The large vertical diffusivity has important implications not only for the modulation of sea surface temperature in the STMW formation region (e.g., Hanawa and Sugimoto 2004), but also for the nutrient supply from the subsurface to the oligotrophic surface layer in the subtropical gyre (e.g., Mori et al. 2008; Sukigara et al. 2011).

Meanwhile, shipboard observations using a microstructure profiler indicated that K_v is relatively uniform between the top and bottom of STMW, and is much smaller than the above estimates, being 10^{-6} to 10^{-5} m^2 s^{-1} in summer 2006 and 10^{-7} to 10^{-5} m^2 s^{-1} in winter 2007 (Mori et al. 2008). Why is there such a large discrepancy between the budget-based and observed values of K_v? The most plausible explanation is that like the deepening of the mixed layer in the fall and early winter seasons (Sect. 2), erosion of STMW is dominated by episodic atmospheric disturbances on time scales of several days, during which K_v becomes a few orders of magnitude larger but is missed by shipboard microstructure profiler measurements due to rough sea conditions. In fact, an analysis of data from an acoustic Doppler current profiler moored in the STMW formation region under the KESS project revealed that following each wind storm event during July–December 2004, the upper ocean inertial velocity shear tends to increase in amplitude and to propagate downward to the top layer of STMW (Luc Rainville, personal communication, 2010). Another possible cause for the discrepancy is interannual variation of K_v in STMW resulting from the changing density structure of STMW. Since thinner STMW is formed and more strongly dissipated in 2006 than in 2004 (Sect. 4), STMW is less vertically homogeneous during 2006, which is likely to hinder the intensive breaking of downward-propagating inertial waves at its upper boundary. The third possible cause is the importance of horizontal processes that were assumed to be negligible in Qiu et al. (2006). After spring, STMW patches originating from different regions and having different densities are stirred by mesoscale eddies and circulations and interleave with each other, and this helps destroy the vertically uniform structure of STMW, particularly when the KE is in an unstable state (Oka et al. 2011b; Sect. 4).

As noted in Sect. 2, part of STMW is transported southward from its formation region, being subducted into the permanent pycnocline. Subsequently, it is advected southwestward, carrying temperature and salinity anomalies, and reaches the western boundary one to several years later (Oka 2009). A portion of this “old” STMW is then transported back to the formation region by the Kuroshio/KE and is obducted onto the winter mixed layer, modulating the sea surface temperature there (Qiu and Huang 1995; Qiu 2002; Endoh et al. 2006; Liu and Hu 2007). Besides this anticyclonic circulation of STMW, eastward advection of STMW by the KE from the region southeast of Japan to the east of the dateline over a period of 1 year was suggested from a lag correlation analysis of winter sea surface temperature anomalies (Sugimoto and Hanawa 2005a).

The Kuroshio and KE influence not only STMW, but also CMW and TRMW to the north. Saito et al. (2011) analyzed quasi-Lagrangian observations by an isopycnal Argo float to indicate that TRMW becomes warmer and saltier after its formation in late winter, and is modified to

D-CMW within 1 year. They attributed this change partly to double-diffusive salt-finger convection between TRMW and the warmer, saltier, and equally dense surface water of the Kuroshio origin that is advected from the south. This scenario is consistent with the observed facts that the upper portion of TRMW is characterized by a high Turner angle implying active salt-finger convection and by inhomogeneity in terms of temperature and salinity relative to the lower portion (Toyama and Suga 2010, 2011) and that the proportion of TRMW (D-CMW) decreases (increases) eastward/downstream in their formation region (Oka et al. 2011a; Saito et al. 2011). Thus, TRMW is likely a transient feature in the TRMW/D-CMW formation region and can be regarded as the most upstream portion of D-CMW.

After L-CMW and D-CMW are subducted from the eastern part of the formation regions (Oka et al. 2011a), they are advected anticyclonically in the subtropical gyre toward the western boundary (Suga et al. 2004; Tsujino and Yasuda 2004; Oka et al. 2011a). As they circulate southwestward in the southern part of the gyre, D-CMW takes a more zonal path than L-CMW because of the beta spiral effect (Stommel and Schott 1977; Kubokawa 1999), which makes the two waters overlap each other (Fig. 2; Oka and Suga 2005; Kobashi et al. 2006; Xie et al. 2011). This stack of CMWs in the western to central part of the gyre, as well as STMW in the western part, has been shown to generate the Subtropical Countercurrent (Uda and Hasunuma 1969) and the associated subtropical fronts (Kubokawa 1997, 1999; Kubokawa and Inui 1999; Aoki et al. 2002; Kobashi et al. 2006; Yamanaka et al. 2008; Xie et al. 2011), whose thermal effects on the atmosphere locally enhance precipitation and generate positive wind stress curl anomalies in winter to spring (Kobashi et al. 2008; Xie et al. 2011). This is a new dynamic role of the mode waters in climate, which contrasts with the traditional role as a heat reservoir memorizing the wintertime ocean-atmosphere interaction (Xie et al. 2011; see also the review of Kobashi and Kubokawa 2011 in this issue for details). Another interesting feature is that as L-CMW and D-CMW circulate anticyclonically, their pathways shift to the inner side of the subtropical gyre represented by geostrophic streamlines (Oka and Suga 2005). This is consistent with the modeling study of Nonaka and Xie (2000) that demonstrated that temperature anomalies subducted from the northern part of the gyre in association with the subduction of low PV water circulate more zonally in the southern part of the gyre as high baroclinic mode Rossby waves than a passive tracer subducted from the same location, reaching the western boundary at a latitude higher than that of the North Equatorial Current bifurcation (e.g., Qiu and Chen 2010a), and are therefore unable to flow into the equatorial region.

ESTMW is also advected southwestward from the formation region along the outer path in the subtropical gyre (Hautala and Roemmich 1998), but its low PV signal is rapidly dissipated to the downstream (Ladd and Thompson 2001; Hosoda et al. 2001; Suga et al. 2004; Sugimoto and Hanawa 2005b). This is because (1) ESTMW is much thinner than the other mode waters in the North Pacific and is characterized by a weak PV minimum (Suga et al. 2004; Sugimoto and Hanawa 2007); (2) owing to the subsurface temperature/salinity stratification in the eastern part of the subtropical gyre, salt-finger convection occurs vigorously between ESTMW and the underlying permanent pycnocline water, which erodes ESTMW from its bottom after spring (Sugimoto and Hanawa 2005b, 2007; Shimada et al. 2007; Toyama and Suga 2010, 2011); and (3) owing to stratus clouds over the formation region, ESTMW is not capped by the strong seasonal pycnocline after spring, which allows atmospheric disturbances to continually modify ESTMW from its top (Sugimoto and Hanawa 2007). Nevertheless, subducted temperature/salinity anomalies in association with ESTMW might be transported to the downstream even after the low PV signal of ESTMW is lost. A recent analysis of Argo float data showed that isopycnal θ–S (or spiciness) anomalies propagate in the range of $\sigma_\theta = 25.0$–25.5 kg m^{-3} from the eastern subtropical Pacific to the western tropical Pacific, along a path and with a speed corresponding to the mean geostrophic flows (Sasaki et al. 2010). This propagation might be associated with the subduction and circulation of ESTMW, and its influence on the equatorial region, which has been hypothesized to be important for interdecadal climate variability of the North Pacific (Hanawa 1996; Gu and Philander 1997), should be examined quantitatively in future studies.

4 Decadal variability

The ocean-atmosphere system in the mid-latitude North Pacific exhibits significant decadal to interdecadal variability (e.g., Nitta and Yamada 1989; Tanimoto et al. 1993; Graham 1994; Trenberth and Hurrell 1994). Its state is often described using the Pacific decadal oscillation (PDO) index, which is defined as the leading principal component of North Pacific sea surface temperature variability poleward of 20°N (Hare 1996; Zhang 1996). A positive (negative) PDO index indicates a stronger (weaker) Aleutian low, stronger (weaker) westerlies, and a lower (higher) sea surface temperature except near the North American coast. An abrupt change of sign in the PDO index and the subsequent persistence for a few decades, called a regime shift, occurred in 1925, 1947, and 1977, exerting widespread impacts on the climate and marine ecosystems in the North Pacific (Mantua et al. 1997; Mantua and Hare 2002).

This observed long-term ocean-atmosphere variability is widely believed to be controlled by the ocean, which has much larger heat content and longer time scales than the atmosphere. As an important step for clarifying its mechanism, the response of the ocean interior to the decadal/interdecadal surface forcing has been explored. Pioneering works of Deser et al. (1996) and Schneider et al. (1999) analyzed historical temperature data to demonstrate that following the 1976/1977 regime shift when the PDO index turned positive, cold anomalies were subducted from the sea surface in the central North Pacific and then propagated downward in the permanent thermocline. Since the subduction area corresponded roughly to the central location of decadal sea surface temperature variability (Tanimoto et al. 1993) and also to the CMW subduction region, CMW was expected to play a key role in transmitting the decadal atmospheric changes to the upper ocean in the mid-latitude North Pacific (Suga et al. 1997). Following the data analysis of Yasuda and Hanawa (1997), numerical simulations using various surface forcing were performed in the early 2000s to investigate decadal variability of CMW, with particular attention to the 1976/1977 regime shift (Inui et al. 1999; Xie et al. 2000; Kubokawa and Xie 2002; Ladd and Thompson 2002; Hosoda et al. 2004). Their results, with some differences among models, showed that changes in the thermodynamic forcing (increases of upward surface heat flux and southward Ekman transport) and the dynamic forcing (increase of Ekman pumping) associated with the regime shift resulted in the formation of thicker and colder CMW and the eastward shift of its circulation path in the subtropical gyre, respectively, and their combined effects altered the subsurface thermal structure of the gyre. An analysis of the 180° repeat hydrographic section across the CMW formation region consistently indicated that the CMW temperature abruptly increased by 1°C from 1988 to 1989 when the PDO index transiently turned negative (Suga et al. 2003). Ladd and Thompson (2002) further generalized that when the PDO index is positive, thicker and colder CMW tends to be formed and the gyre circulation transporting it tends to be strengthened, and vice versa. Such a relation between the PDO index and the mode waters with no time lag was also demonstrated by recent modeling studies for CMW (Qu and Chen 2009) and STMW (Davis et al. 2011).

On the other hand, recent observational studies proposed a new mechanism of decadal STMW variability, which involves delayed oceanic response to the PDO-related surface wind forcing. Specifically, Qiu and Chen (2005) identified a decadal oscillation of the KE current system between two dynamic states based on the satellite altimeter sea surface height data during 1993–2004. This oscillation originates from the large-scale wind stress curl forcing in the central North Pacific around 160°W where the forcing has its largest amplitude (Qiu 2003). When the PDO index is positive (negative), negative (positive) sea surface height and permanent thermocline depth anomalies are generated there and propagate westward to the southeast of Japan over a period of $\sim$3 years at the speed of first-mode baroclinic Rossby waves. As a result, the upstream KE jet becomes unstable and weak (stable and strong), and is accompanied by a weak (strong) southern recirculation gyre and high (low) regional eddy activity, which is against our expectation that a stronger mean flow favors baroclinic instability and leads to higher eddy activity (Qiu and Chen 2005, 2010b, 2011; Kelly et al. 2007; Taguchi et al. 2007). In the past 18 years, the KE was in the stable state in 1993–1994, 2002–2005, and 2010 and in the unstable state in 1995–2001 and 2006–2009.

The decadal KE variability is closely related to the STMW formation in the recirculation gyre. An analysis of temperature profiles in the KE region during 1993–2004 revealed that the STMW thickness also exhibited significant decadal variability and was large (small) when the KE was in the stable (unstable) state (Qiu and Chen 2006), which has been confirmed by several studies based on observational data (Pan and Liu 2005; Qiu et al. 2007; Sugimoto and Hanawa 2010; Oka et al. 2011b) and reanalysis product (Miyazawa et al. 2009). This occurs because during an unstable KE period, active southward transport of high PV water from the Mixed Water region across the KE due to the stronger eddy activity (Qiu and Chen 2006; Qiu et al. 2007), as well as the higher background stratification associated with the shallower permanent pycnocline in the weaker recirculation gyre (Sugimoto and Hanawa 2010), is unfavorable for the development of deep winter mixed layer, and the reverse is true during the stable KE period. For the period of 1993–2004, Qiu and Chen (2006) also demonstrated that the STMW thickness was highly correlated with the pre-existing upper-ocean stratification and had little correlation with the wintertime surface heat flux forcing. In other words, the thickness is determined by the dynamic forcing rather than the thermodynamic forcing, which contrasts with our traditional view that the latter forcing is dominant. A subsequent analysis of long-term temperature data during 1971–2007 showed that the contributions from the pre-existing stratification and the wintertime surface heat flux are comparable, with the former (latter) being more important after (before) around 1990 (Iwamaru et al. 2010). It has also been pointed out that the pre-existing stratification is determined not only by the dynamic state of the KE current system, but also by the summer surface heating, whose intensity depends on the activity of tropical cyclones (Kako and Kubota 2007; Tomita et al. 2010).

The decadal KE variability is important not only for the formation of STMW, but also for its subsequent evolution.

Recent analyses of Argo float and high-resolution shipboard observation data (Oka 2009; Oka et al. 2011b) showed that the STMW circulation in 2008 was much more turbulent than that in 2006. In 2006, the STMW formed in several anticyclonic circulations in the recirculation gyre tended to be continually trapped in its respective circulations after spring and to remain in the formation region until late fall (Oka 2009). In 2008, in contrast, it tended to gradually migrate southward. Simultaneously, STMWs with different temperatures formed in different longitude ranges were gradually stirred, then interleaved with each other, and were finally vertically mixed to form a STMW with an intermediate temperature (Oka et al. 2011b). The difference in the STMW circulation between the 2 years is attributable to the dynamic state of the KE current system. The KE was in the unstable state in both 2006 and 2008, but its path was more variable in 2008, which was the peak year of the unstable period of 2006–2009. The associated stronger eddy activity in the STMW formation region in 2008 was hypothesized to enhance the eddy transport of STMW in both the meridional and zonal directions (Oka et al. 2011b).

Although the seasonal evolution of the STMW distributions in the stable KE period has not been investigated in detail (because of the lack of Argo floats in 2002–2005), it is possible to put forth the following hypothesis about the decadal variation in the STMW subduction and dissipation. In the unstable KE period, the volume of STMW formed in winter is smaller than in the stable KE period because the formation region is partially masked by cyclonic eddies (Rainville et al. 2007) and the STMW thickness is smaller (as mentioned above), but the volume of STMW subducted into the south of the formation region is larger because of the enhanced eddy transport. Also, STMW in the formation region is dissipated after spring, mainly by vertical processes in the stable KE period and by horizontal processes in the unstable period. This is because the thicker and more uniform STMW in the stable period is more susceptible to the intensive breaking of downward-propagating inertial waves at its upper boundary (Sect. 3), while the stronger eddy activity in the unstable period helps destroy the vertically uniform structure of STMW (Oka et al. 2011b). Such hypothesized STMW variability would possibly affect the climate and primary production through the impact on the Subtropical Countercurrent (Yamanaka et al. 2008; Xie et al. 2011) and the nutrient supply from the subsurface to the surface layer (Sukigara et al. 2011), respectively.

The decadal variability of STMW temperature has also been explored. Modeling studies have demonstrated that on decadal time scales, stronger westerlies in the central North Pacific lead to the formation of warmer STMW several years later through an increase of the Kuroshio heat transport (Ladd and Thompson 2002; Yasuda and Kitamura 2003; Lee 2009), as previously shown by observational studies (Yasuda and Hanawa 1997; Hanawa and Kamada 2001). In addition, the Argo float data analysis (Oka 2009) suggested that the zonal temperature change in the STMW formed east of the Izu-Ogasawara Ridge ($\sim$140°E) is small in the stable KE period and large in the unstable KE period. This might be explained as follows. In the stable period, the KE takes a relatively straight path and is accompanied by a single, contiguous southern recirculation gyre, within which STMW having a zonally uniform temperature is formed. In the unstable period, on the other hand, the southern recirculation gyre is broken into several anticyclonic circulations, within each of which STMW with a different temperature is formed. In this latter case, the longer KE path results in a greater oceanic heat loss along the path, leading to a gradual decrease of the STMW temperature to the downstream. The float data analysis also revealed a significant interannual change in the STMW salinity (Oka 2009), which may be related to the variability of the North Pacific Tropical Water (Cannon 1966; Suga et al. 2000). With the accumulation of additional Argo float data in the coming years, we can expect the decadal variability in the STMW properties and its mechanisms to be clarified further.

The decadal variability of ESTMW has been least explored among the three mode waters. A hindcast run for 1965–1993 (Ladd and Thompson 2002) showed that the ESTMW density fluctuated on decadal time scales due primarily to the buoyancy flux change, while a recent data assimilation experiment for the 1990s (Toyoda et al. 2011) demonstrated that the ESTMW formation rate varied interannually owing to changes not only in the surface cooling in winter and the pre-existing stratification controlled by the low-level cloud coverage in summer, but also in the Ekman convergence of salt. These studies, however, pertain to the pre-Argo period during which we had inadequate salinity data. Given the important roles played by salinity in the ESTMW formation and dissipation (Sects. 2 and 3), future analyses of Argo float data and the reanalysis product assimilating them are expected to reveal new features and mechanisms of the ESTMW decadal variability.

5 Future directions

We have reviewed the progress in the observational and modeling studies of the North Pacific mode waters from the past decade. It is instructive to end this review by offering some directions for the mode water studies of the next decade. The first target is the role of mesoscale eddies, which prevail in the regions surrounding the KE jet (Mizuno and White 1983; Itoh and Yasuda 2010a, b; Itoh et al. 2011), in the formation and subduction of L-CMW, D-CMW, and TRMW. "Eddy-resolving" synoptic shipboard observations, such as the one conducted by Oka et al. (2011b), in conjunction with eddy-resolving simulations will contribute greatly to the refined depiction of the large-scale CMW formation and subduction revealed by Argo float data (Oka et al. 2011a).

Nearly 5 years have passed since the Argo float array in the North Pacific was completed, and the collected data have improved the climatology-based, classical views on STMW and CMW. However, the formation, subduction, and circulation processes of ESTMW have not been fully investigated using the float data. Future studies are encouraged to verify the various processes of ESTMW put forth by the modeling studies in the early 2000s.

During the past decade, accumulation of satellite altimeter sea surface height data has significantly enhanced our understanding of the decadal variability of the KE current system (Qiu and Chen 2005, 2010b, 2011). In a similar manner, accumulation of Argo float data will undoubtedly unravel the decadal variability of the mode waters in the coming decade. In Sect. 4, we have presented our hypothesis on the decadal variability of STMW associated with that of the KE current system. CMW is also expected to exhibit decadal variability owing to both the thermodynamic (surface heat flux and meridional Ekman transport) and dynamic forcings. We anticipate two types of dynamic forcing on CMW that originate from the PDO-related surface wind forcing in the central North Pacific. One is the decadal variability of eddy activity north of the KE, which would control the number and distribution of anticyclonic eddies that are the probable formation sites of thick CMW (Kouketsu et al. 2011). A recent study by Qiu and Chen (2011) found that the decadal mesoscale eddy variability had an opposite phase between the upstream (140°–152°E) and downstream (152°–165°E) KE regions. These downstream decadally varying eddy signals can potentially induce the long-term CMW variability. The other is the observed (Qiu and Chen 2005) and simulated (Nonaka et al. 2006; Taguchi et al. 2007) decadal shifts of the thermohaline fronts bounding the CMW formation regions. If these shifts are accompanied by changes in the winter mixed layer properties in the interfrontal regions, the CMW density can fluctuate without a significant change in the thermodynamic forcing (Suga et al. 2011). It is of interest to examine the degree to which the large-scale CMW subduction scheme for 2003–2008 (Oka et al. 2011a) applies to other periods. The maximum σ_θ of CMW that is subducted into the permanent pycnocline (26.4 kg m^{-3} for 2003–2008) can fluctuate on decadal time scales, and this may cause the decadal apparent oxygen utilization changes observed in the subsurface centered at $\sigma_\theta = 26.6$ kg m^{-3} in the subtropical and subarctic North Pacific (e.g., Andreev and Kusakabe 2001; Ono et al. 2001; Watanabe et al. 2001; Emerson et al. 2004; Mecking et al. 2008). Investigation into the decadal mode water variability and its mechanisms will also be facilitated by further development of numerical models in the coming decade. Models that are able to reproduce the recently identified structures and variability of currents, eddies, and mode waters are called for.

The decadal variability of STMW and CMW can influence the sea surface temperature in their formation regions through re-emergence (Hanawa and Sugimoto 2004; Sugimoto and Hanawa 2005a) or obduction after circulating in the subtropical gyre (Qiu and Huang 1995; Qiu 2002; Endoh et al. 2006). It can also potentially affect the sea surface temperature in the southern region through its impact on the Subtropical Countercurrent (Yamanaka et al. 2008; Xie et al. 2011; Kobashi and Xie 2011; Nonaka et al. 2011) and the Hawaiian Lee Countercurrent (Sasaki et al. 2011). On the other hand, the decadal variability of ESTMW can possibly modulate the sea surface temperature in the equatorial upwelling region. Quantification of these influences and clarification of the underlying mechanisms will lead to a better understanding of the ocean's role in the decadal climate variability.

Multidecadal variability or longer-term changes of the mode waters have been examined only recently. The STMW temperature observed in the winter formation region at 141°–150°E and in summer at 137°E exhibited an increasing trend of 3–5 $\times$ 10^{-2}°C/year^{-1} for the period of 1971–2007 (Iwamaru et al. 2010). Coupled climate model simulations predicted that in response to global warming, the densities of the mode waters tended to become lower (Lee 2009; Luo et al. 2009), although their subduction rates increased in Lee's model and decreased in Luo et al.'s model. It has also been predicted that banded structures of the sea surface temperature warming that were slanted in the northeast-southwest direction appear in response to global warming because the shallower winter mixed layer in the Kuroshio-Oyashio Extension region results in weaker mode water subduction, weaker Subtropical Countercurrent, and smaller thermal advection from the southwest (Xie et al. 2010, 2011; Xu et al. 2011). Such long-term mode water changes and their consequences are expected to become increasingly important not only from the physical, but also the biogeochemical, perspectives in the coming decades.

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