Sea surface cooling in the Northern South China Sea observed using Chinese sea-wing underwater glider measurements

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ABSTRACT

Based on 26 days of Chinese Sea-wing underwater glider measurements and satellite microwave data, we documented cooling of the upper mixed layer of the ocean in response to changes in the wind in the Northern South China Sea (NSCS) from September 19, 2014, to October 15, 2014. The Sea-wing underwater glider measured 177 profiles of temperature, salinity, and pressure within a 55 km x 55 km area, and reached a depth of 1000 m at a temporal resolution of ~4 h. The study area experienced two cooling events, Cooling I and Cooling II, according to their timing. During Cooling I, water temperature at 1-m depth (T1) decreased by ~1.0 °C, and the corresponding satellite-derived surface winds increased locally by 4.2 m/s. During Cooling II, T1 decreased sharply by 1.7 °C within a period of 4 days; sea surface winds increased by 7 m/s and covered the entire NSCS. The corresponding mixed layer depth (MLD) deepened sharply from 30 m to 60 m during Cooling II, and remained steady during Cooling I. We estimated temperature tendencies using a ML model. High resolution Sea-wing underwater glider measurements provided an estimation of MLD migration, allowing us to obtain the temporal entrainment rate of cool sub-thermocline water. Quantitative analysis confirmed that the entrainment rate and latent heat flux were the two major components that regulated cooling of the ML, and that the Ekman advection and sensible heat flux were small.

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1. Introduction

The South China Sea (SCS), with an area of 3,500,000 km², is the largest marginal sea connecting the western Pacific, and is dominated by the East Asian monsoon. In the winter, northeasterly winds overlie the SCS, whereas in the summer, southeasterly winds prevail across the region (e.g., Liu and Xie, 1999; Xie et al., 2003). The upper ocean in the SCS responds to the monsoon, with a deep (~200 m) mixed layer (ML) in the winter and a shallow (~30 m) ML in the summer (e.g., Qu et al., 2007).

The large-scale wind-driven variation in the upper layer of the ocean has been studied by previous researchers. In the northern SCS, surface currents display complicated dynamics, including the Kuroshio Current, cyclonic current, and warm SCS current. All of these large-scale currents are generally associated with wind or wind stress curl (e.g., Shaw and Chao, 1994; Fang et al., 2002; Qu et al., 2007). In the present study, we investigated the oceanic responses to synoptic-scale changes in the wind, which is termed as several hundred kilometers and several days.

Characterization of oceanic response to synoptic-scale weather, i.e. wind events, is vital to understand the exchanges of air–sea heat fluxes through wind stress, heat and fresh water (Ruiz et al., 2012). The synoptic-scale variations also influence the seasonal ocean circulation of SCS (Wang et al., 2009), and even the climate (Qiu et al., 2004). After the passing of the typical synoptic scale weather tropical cyclone, SST decreased, chlorophyll-a increased, near-inertial oscillation was triggered in the thermocline, and momentum fluxes were enhanced (e.g., Large and Pond, 1981; Chu et al., 2000a, 2000b; Tang et al., 2004). Given the strong vertical shear of the near-inertial currents, intense turbulent mixing via shear instability often follows a tropical cyclone (TC) and entrains colder upper thermocline water into the ML, resulting in significant ML cooling, in addition to the cooling by Ekman pumped upwelling and upward heat flux at the air–sea interface (Price, 1981). Another synoptic scale weather was cold surge, which covers the northern SCS every winter to spring. During the cold
surge, the ocean releases more heat to atmosphere than usual (Li et al., 2006), and induces ML cooling. They investigated cold air outbreak-induced SST cooling in the northern SCS (NSCS) using ship-cruise measurements. They estimated the variation in net heat flux and momentum fluxes, but were unable to calculate their contributions to SST due to the low temporal/spatial resolution of the ship route.

There have been few studies on migrations of the upper, ML of the ocean in response to synoptic scale changes in the wind in the open sea of the SCS. This is mainly due to the fact that in situ and satellite observations retain various disadvantages. For example, ship observations provide a vertical structure of the water body; however, they are unable to provide ocean time-series information (e.g., Li et al., 2006). Mooring data can supply continuum measurements, but the vertical resolution is too low to investigate migration of the ML and the process of mixing (Brainerd and Gregg, 1995). Retrieval of the ML depth (MLD) has been attempted with satellite-derived images, but this has only been successful in a few cases (e.g., Pan and Sun, 2012). A slab model developed by Large and Pond (1981) indicates that the ML cooling is suggested to be influenced by air–sea interface heat fluxes, shear instability entraining thermocline water, Ekman pumped upwelling, and Ekman/geostrophic advections. The model has been successfully used to examine the mechanisms of seasonal variations of SST or SST front (Guan et al., 2014; Qu et al., 2007; Qiu et al., 2014). We use the ML model to examine the mechanism of oceanic response to synoptic scale weather.

Currently, gliders carrying sensors deep into the ocean that then swim up to the ocean surface provide us with both the vertical structure of the ocean and continuum data, and have been successfully used to investigate oceanic conditions in several parts of the world. For example, in the Mediterranean Sea, Ruiz et al. (2005) observed the vertical motion of the ocean, and Ruiz et al. (2012) studied mixing of the upper ocean. In the Antarctic Ocean, five glider experiments measured the seasonal cycle of the ML (Swart et al., 2014). In the central Middle Atlantic Bight, the seasonal evolution of hydrographic fields was documented by Castle-lao et al. (2008) using glider observations. In the present study, we observed the upper ocean using glider measurements in the NSCS, and try to examine its mechanisms using ML model. A Chinese sea-wing glider was employed to investigate ML cooling events in the SCS from September, 2014 to October, 2014.

2. Data and methods

2.1. Sea-wing underwater glider measured temperature/salinity

Temperature/salinity data were collected by the sea-wing underwater glider, which developed by Shenyang Institute of Automation, Chinese Academy Sciences (Yu et al., 2013). The communication and navigation subsystem contains iridium satellite communication devices, wireless communication devices, a precision navigation attitude sensor, a Global Positioning System (GPS) device, a pressure meter, and obstacle avoidance sonar. A conductivity–temperature–depth (CTD) sensor with ~6 s sampling resolutions has been installed on the current Sea-wing underwater glider version. In total, 177 profiles (CTD) between surface and 1000 m depth were obtained.

The stations are shown in Fig. 1. Data collection took place from September 19, 2014 to October 15, 2014 in the channel between the Hainan and Xisha Islands. The water depth was > 1000 m, and the along-track Sea-wing underwater glider data resolution was ~5 km. The coverage of the Sea-wing underwater glider track was a square within an area of 55 x 55 km. The climatological mean first-baroclinic Rossby radius of deformation is larger than 100 km

for the deep basin of the SCS (Chen et al., 2011), and the wavelength of the typical diurnal tide is 130 km around Xisha Island (Johnston et al., 2013). The ship started from cross-section 1, and circled the square anticlockwise four times. Before investigating oceanic phenomena, we did data quality check following Chu and Fan (2010): (1) excluded the temperature larger than 35; (2) interpolated all the temperature/salinity profiles into 1-m interval; (3) calculated temperature difference between each profile and mean profile from World Ocean Atlas (WOA) data, and then excluded the difference larger than 3.

To investigate the mixed and mixing process of upper layer, we compared MLDs from different methods, which include temperature difference, density difference (de Boyer Montégut et al., 2004), peak Brunt-Väisälä Frequency (Manucharyan, 2010), and optimal linear fitting method (Chu and Fan, 2010). In the temperature and density difference methods, we define MLDs as depth h where (Qu et al., 2007), and \( \sigma_0 - \sigma_h = 0.125 \) kg/m\(^2\). \( T_f/\sigma_10 \) is the temperature/potential density at 1 m/10 m depth, and \( T_f/\sigma_h \) is the temperature/potential density at depth h. The first three definitions of MLD are subjective, because they were given critical values, which might depend on the study area. The last method is an objective method developed by Chu and Fan (2010), following three steps: (1) fitting the profile data from the first point near the surface to a depth using a linear polynomial, (2) computing the error ratio of absolute bias of few data points below that depth

\[ \text{error ratio of absolute bias} = \frac{\text{bias of depth below}}{\text{bias of depth over}} \]

\[ \text{bias of depth below} = \frac{\text{MLD} - \text{MLD}_1}{\text{MLD}} \]

\[ \text{bias of depth over} = \frac{\text{MLD} - \text{MLD}_2}{\text{MLD}} \]

\[ \text{MLD}_1 \text{ and } \text{MLD}_2 \text{ are the mean MLDs of profile data from } h \text{ and } h+1 \text{ m depth.} \]

\[ \text{error ratio of absolute bias} = \frac{\text{MLD}_1}{\text{MLD}_2} \]

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versus the root-mean-square error of data points from the surface to that depth between observed and fitted data, and (3) finding the depth (i.e., the mixed layer depth) with maximum error ratio. This approach is objective, but might induce some error in the real condition.

Quality index proposed by Lorbacher et al. (2006) could be used to verify the qualities of MLDs. MLD calculation is based on the assumption that the near-surface mixed layer exits quasi-homogeneous properties. The standard deviation of the property about its vertical mean is close to zero. Below the MLD, variance of property should increase rapidly about its vertical mean. Therefore quality index is defined as the ratio of the standard deviation of the observed property in the depth range from the surface to MLD, $h$, to the standard deviation from the surface to the depth of $1.5 \times h$. It is shown below:

$$QI = 1 - \frac{\sigma(T_h - \langle T \rangle)(h_h, h)}{\sigma(T_k - \langle T \rangle)(h_k, 1.5 \times h)},$$

where $QI$ is the quality index, $T_h$ is the observed temperature at depth $h$, $\sigma(\cdot)$ is the standard deviation from the vertical mean value $\langle \cdot \rangle$ in the depth range from the first layer near the surface $h_k$ to the depth of MLD $h$ or $1.5 \times h$, respectively. Quality index $QI = 1$ represents “high-quality” computation of $h$ and progressively lower values imply that either larger volumes of stratified water present above the level of $h$. Note that $h$ is well defined if $QI > 0.8$. $h$ can be determined with uncertainty for $QI$ in the range of 0.5–0.8, and $h$ cannot be identified for $QI < 0.5$ (Lorbacher et al., 2006).

Fig. 2 shows the quality index of MLD from four methods. In the range of $QI > 0.8$, temperature difference method was superior to other methods. The objective optimal linear fitting method was suitable in Chu and Fan (2010)’s study, but not the best method in our case study. Thus, we used the temperature difference method as the criteria of MLD in the present study.

2.1.1. Satellite data

WindSat operates on the Coriolis satellite, which is in a near-polar orbit. WindSat collects daily wind speed variables, and its descending and ascending times in the SCS were around 6 am and 10 am, respectively. We combined the two swaths into one image per day, and the spatial resolution was 25 km × 25 km. Daily Tropical Rainfall Measuring Mission (TRMM) Microwave Imager (TMI) data at a spatial resolution of 25 km were provided by the Remote Sensing System (RSS), and included rainfall rate, wind speed, and SST. Because TMI SST retrievals are influenced by rainfall, we excluded the SST pixels with rainfall flags (Qiu et al., 2009). From September 19, 2014 to October 15, 2014, TMI SST version 4.0 was used to show large-scale variation in the NSCS.

The merged sea surface height (SSH) altimeter data were obtained from Archiving, Validation and Interpretation of the Satellite Oceanographic (AVISO) with a 3.5-day temporal resolution and 0.25° × 0.25° spatial resolution (Ducet et al., 2000). The SSH maps were added to a mean dynamic ocean topography derived from Maximenko and Niiler (2005).

2.2. National Centers for Environmental Prediction/National Center for Atmospheric Research surface turbulent fluxes

Six-hour surface heat fluxes, including net shortwave radiation, net longwave radiation, latent heat fluxes and sensible heat fluxes, were obtained from the National Centers for Environmental Prediction/National Center for Atmospheric Research (NCEP/NCAR) reanalysis project (Kalnay et al., 1996). The surface heat flux was gridded onto a T62 Gaussian grid, which had spaces of around 1.9° in our study area. The heat flux product can capture the timing and relative forcing strength of the synoptic-scale net heat flux (Qiu et al., 2004).

3. Results and discussion

3.1. Cooling event within the ML observed with the sea-wing underwater glider

The sea-wing underwater glider provided temporal variation in sea surface and mixed layer cooling processes (Fig. 3a). $T_1$ was defined as the temperature at 1-m depth, and mean temperature within mixed layer $T_{\text{mean}}$ was $\langle T \rangle$. The black/grey line was the 6-hour averaged $T_1/1\text{m mean}$. We separated observation time into three stages: (1) in Cooling I period (Sep.19–23), $T_1$ decreased by $-1^\circ$C, while $T_{\text{mean}}$ decreased by $-0.5^\circ$C; (2) in Warm period (Sep.24–Oct.3), both $T_1$ and $T_{\text{mean}}$ increased. During this period, the differences between $T_1$ and became large, which indicated the stratification was strong. $T_1$ had significant diurnal variations; (3) during Cooling II period (Oct.4–Oct.15), both $T_1$ and $T_{\text{mean}}$ decreased sharply from 30 to 28.5. After Oct.5, $T_1$ almost equaled $T_{\text{mean}}$, which implied a well mixed upper ocean condition. The two cooling cases are further investigated in the following.

The vertical temperature structure and Brunt–Väisälä frequency ($N^2$) are shown in Fig. 3b and c; MLD (purple line) and maximum $N^2$ (black line) are added in Fig. 3c. As mentioned in Section 2.1, we defined MLD as the depth where the temperature is $0.6^\circ$C below the $T_1$. For Cooling I, surface cooling extended to a maximum depth of 40 m and lasted 3 days. The mean MLD was 35 m, which was in close agreement with the MLD climatology data for September (Qu et al., 2007). From September 21 to September 23, $T_1$ decreased, but the MLD did not increase. This confirms the fact that the MLD does not always have a negative relationship with SST (Chu et al., 2000a, 2000b). Therefore, retrieving the MLD from surface parameters not only requires SST, but also other parameters, such as sea surface height, wind stress, and dynamic height (Ali, 1993; Ali and Sharma, 1994; Swain et al., 2006).

Cooling II extended to a depth of 60 m, and lasted for more than 1 week. In this case, the MLD (60 m) was much larger than the monthly mean MLD (40 m) (Qu et al., 2007). This demonstrates the ability of the Sea-wing underwater glider to detect temporal migration of the MLD and to observe the consequences of heat flux and mixing.

The stability of water column has been illustrated in Fig. 3c, which confirms that MLD (purple line) and maximum $N^2$ (black line) were well-matched during the Sea-wing underwater glider observation period, with the exception of October 4–9. We
of density within ML may be induced by diurnal and semi-diurnal tides revealed in the study area (e.g. Yanagi and Takao, 1998; Zhao et al., 2004). The contributions of entrainment or detrainment will be displayed in the next section.

3.2. Satellite observations of the ocean’s surface cooling

During the period of the Sea-wing underwater glider data collection, the wind direction and magnitude varied with time. Corresponding to the three periods described in Section 2, wind speed and TMI SST had response signals (Fig. 4). During Cooling I (September 19–23), southwesterly wind dominated the NSCS and TMI SSTs were smaller than 27 °C; during warm period (September 24–October 3), weak northeasterly wind speed blew across the NSCS, and NSCS water warmed with SSTs of ~29.5 °C; during Cooling II (October 4 to October 15), strong north winds and low TMI SSTs covered the entire NSCS.

We calculated the zonal means of the wind speed and SST (Fig. 5) within 25 km × 25 km from the Sea-wing underwater glider station (see Fig. 1). Data gaps were filled using linear interpolation. The wind speed magnitude displayed two peaks (Fig. 5a): the first was on September 24 (9 m/s), and the second was on October 6 (11 m/s). The latter lasted for more than 1 week. At the first peak, the zonal mean TMI SSTs decreased from 30.8 °C to 29 °C, and during the second peak, TMI SSTs decreased from 29 °C to 28 °C. We derived the time-serials of TMI SSTs at positions A, B, C, and D, which located at the centers of 1, 2, 3, and 4 sections. Results are shown in Fig. 5c. All of the 4 TMI SSTs had the same trend, and captured the two cooling cases. It indicated that the study area is generally homogenous during study period.

We were unable to clearly delineate the cooling processes due to uncertainties in the TMI SSTs from coastal contamination, swath gaps, heavy rain, or wind speed. In addition, the TMI SSTs had coarse resolutions and only reflected surface cooling phenomena. TMI SSTs overestimated T1 measured by the Sea-wing underwater glider (Fig. 5b), which is consistent with the results of Qiu et al. (2009). Sea-wing underwater glider T1 disappeared at ~1 °C (Fig. 3a), while TMI SSTs increased from September 19 to September 21, and then decreased by 1.8 °C from September 21 to September 24, which indicates that microwave SSTs were insufficient for investigating local variation at the synoptic scale.

Compared to the wind vectors in Figs. 4 and 5, Cooling I was corresponding to a transient increase of wind speed, while Cooling II was related to a continuous strong wind speed. The continuous strong winds observed in our study area (Fig. 5a) might explain the deepening of the MLD through turbulence during Cooling II. Most of wind-induced energy, which has a direct ratio with wind speed and TMI SST had response signals (Fig. 4). During Cooling I, the differences between T1 and mean temperatures within the ML (Tmean) (Fig. 3a), and found that T1 were higher than mean temperatures (T1 – Tmean > 0) during warm period. During Cooling I, the differences between T1 and decreased and were small. During Cooling II, achieved the same value as T1, which implies the occurrence of strong convection. We suggested that during Cooling II, the temperature was independent of depth within the ML. It is worth noting that the diurnal variation in Sea-wing underwater glider T1 (Fig. 3a) was significant under calm conditions (wind speed < 5 m/s, see Fig. 5a) during warm period.

Fig. 3. Glider observed (a) T1 (red line) and mean temperature within mixed layer (blue line), (b) temperature profiles, (c) Brunt–Väisälä frequency profiles, and (d) densities at mixed layer (black) and thermocline (red) depths. The black (gray) line in (a) stands for 6-h mean of (mixed layer mean temperature). The purple/black line in (c) indicates the mixed layer depth defined from temperature difference maximum Brunt–Väisälä frequency. (For interpretation of the references to color in this figure legend, the reader is referred to the web version of this article.)

checked the quality index during these periods, finding that the percentage of quality index > 0.8 achieved 85%, which suggested the MLD definition was suitable. MLD and thermocline depth (the depth where the maximum vertical gradient of temperature stay) were plotted on the isopycnal space in Fig. 3d. The density of MLD was around 1021.5 kg/m³ before October 5, and sharply increased to 1022.5 kg/m³ after October 11. The density of the observed thermocline still fluctuated around 1022.5 kg/m³. It indicates that the denser water within the thermocline layer was entrained into mixed layer. Process of detrainment or entrainment at the base of the MLD, which might be induced by waves, eddies, or advection, controls much of the shoaling/deepening of MLD (Thrope, 2005; Fang and You, 1987). The denser water lasted several days during Cooling II, therefore, the large increase of density within ML from October 5 to 11 might result from wind induced entrainment rather than internal tide. However, the significant diurnal variations
The night cooling have been found to be driven by the buoyancy fluxes, which induce ML deepening, and oceanic heat loss (e.g., Brainerd and Gregg, 1995; Callaghan et al., 2014). This indicates that the Sea-wing underwater glider was able to measure diurnal variation in $T_1$. However, this is beyond the scope of the present study, and serves as a topic for future research studies.

### 3.3. Heat balance within mixed layer

Changes can be estimated quantitatively using the heat budget within the ML (e.g., de Ruijter, 1983; Qiu, 2000; Qiu et al., 2014). The heat budget includes net heat flux, advection (Ekman and geostrophic), entrainment, and residual terms of diffusion, which is shown as follows:

$$\frac{\partial T_{\text{mean}}}{\partial t} = \frac{Q}{\rho_0 c_p h} - \nabla \cdot \mathbf{u} T_{\text{mean}} - w_e (T_{\text{mean}} - T_d).$$

where is the mean temperature within the ML. $Q$ is the net heat flux including short wave radiation, $Q_{\text{SW}}$, long wave radiation, $Q_{\text{LW}}$, sensible heat flux, $Q_{\text{SHF}}$, and latent heat flux, $Q_{\text{LHF}}$. The quantities $\rho_0$ and $c_p$ are water density (here $\rho_0 = 1.027 \times 10^3$ kg/m$^3$), and the specific heat capacity of seawater at a given constant pressure (here $c_p = 4300$ J/kg $\cdot$ °C), respectively. The variable $h$ is the ML depth. $\mathbf{u}$ is the horizontal velocity including Ekman velocity ($u_{\text{Ek}}$) and geostrophic velocity $u_G$, and $w_e$ is the entrainment velocity. $T_d$ is the water temperature 1-m far from the base of the ML.

To calculate the temperature decreases induced by air–sea turbulence terms, we averaged the observed MLD into 6-hour MLD values, which was the same time interval with NCEP/NCAR products. The glider observed and modeled temperature trends $T_{\text{mean}}$ were shown in Fig. 6a. The main trends of the modeled was similar as the observed one, and captured the three stages of Cooling I, warm, and Cooling II, although its magnitude was a little larger than the real temperature during the two Cooling cases. The mean bias between modeled and observed $\partial T_{\text{mean}}/\partial t$ was 0.1 °C/6 h.

### 3.3.1. Heat flux

Temperature trend induced by short wave radiation and long wave radiation are displayed in Fig. 6b. It indicates that the temperature trend induced by the short wave radiation was in the range of 0–0.3 °C/6 h, and the mean value was 0.12 °C/6 h. It had no significant change when Cooling I and Cooling II occurred. The temperature trend induced by long wave radiation was in the range of −0.05–0.02 °C/6 h.

The contributions of latent heat flux and sensible heat flux are shown in Fig. 6c. Prior to Cooling II, the temperature decrease induced by the latent heat flux maintained a steady value ($\Delta T_l = -0.01$ °C/6 h), with the exception of some diurnal variation. From October 4 onwards, the temperature decreased sharply, and reached a minimum value ($\Delta T_l = -0.06$ °C/6 h) on October 10. We
integrated the temperature decreases $\int (\Delta T_2 - \Delta T_1) \, dt$ from October 4 to October 10. The temperature change calculated from the latent heat flux was $0.7^\circ C$, whereas the actual temperature change measured by the Sea-wing underwater glider was $1.2^\circ C$.

If the model error is neglected, this difference of $0.5^\circ C$ might come from the first two terms on the right hand side of Eq. (2). The contribution of sensible heat flux was quite small compared with latent heat flux.

3.3.2. Horizontal advection

The Ekman advection term, $-u_{\text{Ek}} \nabla T_{\text{mean}}$, contains the Ekman velocity $u_{\text{Ek}}$ and temperature gradient $\nabla T_{\text{mean}}$. The Ekman velocity is defined as $u_{\text{Ek}} = \tau \times f/k(\rho \, \Omega)$, where $\tau$ is the wind stress vector. Its components $(\tau_x, \tau_y)$ can be derived from the bulk formula, $\tau = \rho_a C_W u_{\text{a}} \text{w}^2$, where $\rho_a$ is the air density ($1.2 \text{ kg/m}^3$), $u_{\text{a}}$ the wind vector and wind speed, respectively, at 10 m above the sea surface, and $C_W$ is the drag coefficient, given by Large and Pond (1981). We obtained the temperature gradient $\nabla T_{\text{mean}}$ using the TMI SST. Fig. 6d shows the temperature trend induced by Ekman advection. The peak value of $0.037^\circ C/6 \text{ h}$ occurred on October 10, which was after the stable Cooling II period. The magnitude of geostrophic advection term was similar with that of Ekman advection term.

3.3.3. Vertical mixing and Ekman pumping

The entrainment rate is a function of the rate of MLD deepening, $h$, the vertical velocity of the water parcel at the base of the mixed layer, $w_{\text{mb}}$, and the horizontal advection of water below the ML, $u \cdot \text{V} h$ (e.g., Cushman, 1987):

$$w_e = \frac{dh}{dt} + w_{\text{mb}} + u_{\text{Ek}} \cdot \text{V} h,$$

(3)

When $w_e > 0$ cold water parcels below the ML are entrained into the ML, and when $w_e < 0$ warm water parcels within the ML are detrained into the depths below the ML. Qu et al. (2007) confirmed that the entrainment rate is primarily a result of a counter balance between $\frac{dh}{dt}$ and Ekman pumping $w_{\text{mb}} = \text{curl}_k (\rho_f f)$, in which the wind stress can be derived from the bulk formula $\tau = \rho_a C_W u_{\text{a}}$, where $\rho_a$ is the air density ($1.2 \text{ kg/m}^3$), $u$ is the wind vector at the 10-m height above sea
surface, and \( C_0 \) is the drag coefficient given by Large and Pond (1981).

Fig. 6e shows the temperature tendency induced by \( \frac{\partial \theta}{\partial t} \). The integrated \( \frac{\partial \theta}{\partial t} \) was 1.09, \(-5.30, 22.20\, m \) over Cooling I, warm and Cooling II period, respectively. For the two cooling events, cold water was entrained into the ML, causing a temperature decrease in the range of \(-0.2 \, ^\circ C/6 \, h \) to \(-0.6 \, ^\circ C/6 \, h \), which was the same order as that induced by the net short wave radiation and latent heat flux (Fig. 6b and c). Because the deepening rates of the MLD were estimated from the Sea-wing underwater glider measurements, it supported us high-frequency vibration of MLD. It is worth noting that during October 3–5, \( \omega > 0 \), that is to say, cold water was entrained into the ML, leading to a temperature change of \(-0.02 \, ^\circ C/6 \, h \), which is consistent with the values presented in Fig. 3. Entrainment caused by Ekman pumping was of a similar order to that of the Ekman advection, geostrophic advection, and sensible heat flux terms, which were smaller than the entrainment induced by mixed layer depth. Ekman pumping started to cool the ML during cooling events.

Our results revealed that MLD deepening-induced entrainment term \( \partial \theta/\partial t \) was as large as the latent heat flux, and contributed more than Ekman advection, geostrophic advection and Ekman pumping. It indicates that vertical mixing induced by eddies or internal wave rather than Ekman pumping played more important role in ML cooling, which is consistent with the upper ocean response to Typhoon (e.g., Guan et al., 2014) or storm (e.g., Hopkins et al., 2014). Previous studies have generally ignored the contribution of MLD deepening and have only considered Ekman pumping, due to the difficulty in making the corresponding observations (e.g., Li et al., 2006; Qiu et al., 2014). The high vertical and temporal resolution Sea-wing underwater glider allowed us to capture the actual conditions.

4. Concluding remarks

We investigated cooling of the upper layer of the ocean in response to changes in the wind based on 26 days of observation (September 19, 2014 to October 15, 2014) by the Sea-wing underwater glider, during which temperature/salinity profiles were obtained. Two wind change events were captured during the Sea-wing underwater glider observations, corresponding to two separate upper oceanic cooling events, Cooling I and Cooling II. During Cooling I, surface cooling occurred within the ML (40 m), but the MLD did not increase. During Cooling II, longer-duration (>7-day) heavy winds passed over the Sea-wing underwater glider stations, surface cooling extended to a depth of 60 m, and the MLD increased.

Through quantitative analysis of the heat budget, we demonstrated a significant contribution to the cooling of the ocean’s ML by the entrainment induced by MLD deepening. The Sea-wing underwater glider provided temporal and vertical high-resolution data to investigate migration of the MLD, which was directly related to vertical mixing in the SCS. This approach offers potential for investigating multi-scale oceanic phenomena, such as diurnal variation (Fig. 3a). Besides diurnal variations of radiation flux, tide signals were also significant. Nonlinear interactions of tides and near inertial waves might contribute on the MLD deepening. However, the processes accompanied with diurnal variations were beyond our scope in the present study and left for future work.

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