Observed Variations of the Atmospheric Boundary Layer and Stratocumulus over a Warm Eddy in the Kuroshio Extension

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ABSTRACT

A research vessel sailing across a warm eddy in the Kuroshio Extension on 13 April 2016 captured an abrupt development of stratocumulus under synoptic high pressure. Shipboard observations and results from regional atmospheric model simulations indicate that increased surface heat flux over the ocean eddy lowered surface pressure and thereby accelerated southeasterly winds. The southeasterly winds transported moisture toward the low pressure and enhanced the air–sea interface heat flux, which in turn deepened the low pressure and promoted low-level convergence and rising motion over the warm eddy. The lifting condensation level lowered and the top of the marine atmospheric boundary layer (MABL) rose, thereby aiding the development of the stratocumulus. Further experiments showed that 6°C sea surface temperature anomalies associated with the 400-km-diameter warm eddy accounted for 80% of the total ascending motion and 95% of total cloud water mixing ratio in the marine atmospheric boundary layer during the development of stratocumulus. The synthesis of in situ soundings and modeling contributes to understanding of the mechanism by which the MABL and marine stratocumulus respond to ocean eddies.

1. Introduction

The Kuroshio Extension (KE) is the area where the northward-flowing warm Kuroshio separates from the Japan coast near 35°N, 140°E (e.g., [Mizuno and White 1983](#)). The Kuroshio Extension develops instabilities

and sheds mesoscale eddies ([Chelton et al. 2011](#)). These active eddies both play an important role in the transportation of ocean heat toward poles ([Dufour et al. 2015](#)) and affect the atmosphere ([Frenger et al. 2013](#)).

SST anomaly effects on wind dynamics and cloud aloft have been investigated by previous literatures as reviewed by [Small et al. \(2008\)](#). Based on satellite data and re-analysis data, [Xu et al. \(2010\)](#) examined the atmospheric

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response to a cold eddy in the KE and found that surface wind speed, precipitation, and cloud coverage were all reduced above the eddy. Frenger et al. (2013) examined over 600 000 individual eddies in the Southern Ocean, and Ma et al. (2015) examined atmospheric responses to over 35 000 oceanic eddies in the Kuroshio Extension region. They found that, in contrast to cold eddies, warm eddies cause surface winds to accelerate and cloud fraction, rain rate, and marine atmospheric boundary layer (MABL) height to increase. These phenomena are attributed to the pressure adjustment mechanism (Lindzen and Nigam 1987) and the vertical mixing mechanism (Wallace et al. 1989; O'Neill et al. 2003; Chelton et al. 2004; Park et al. 2006). The former mechanism results in sea level pressure (SLP) anomalies that tend to be low over warm sea surface temperature (SST) anomalies, yielding an acceleration toward the low pressure of the warm SST anomaly, which can also be expressed as a linear correspondence between wind divergence in the MABL to the Laplacian of the SST field (Lambaerts et al. 2013). Further studies indicated that imprints of warm eddies on the atmosphere were stronger than imprints from cold eddies, possibly due to the discrepancies in heat flux between the two (Ma et al. 2015, Chen et al. 2017).

These previous studies mainly utilized satellite data and reanalysis data, which have limited ability to describe the physical processes in the MABL that respond to eddies at synoptic time scales. In recent years, more and more observational studies have attempted to observe quick atmospheric response to SST fronts and eddies. Rouault and Lutjeharms (2000), Arhan et al. (2011), Messenger et al. (2012), and Messenger and Swart (2016), for example, used in situ observations to certify the responses of MABL to the eddies in Agulhas current. Recently, Jiang et al. (2018, manuscript submitted to *J. Geophys. Res. Atmos.*) used in situ observations to document abrupt variations taking place in the MABL forced by a cold eddy in KE area. In the present work, we apply a suite of shipboard measurements and simulations by a regional model to investigate the response of the MABL and its clouds to a warm eddy. These will enable us to further understand the mechanisms involved in variations in surface winds and clouds in the MABL influenced by a warm eddy.

The rest of this paper is organized as follows. An introduction of the research vessel (R/V) observational data and processing method are provided in section 2. Section 3 analyzes the influence of warm eddy on the stratocumulus with observations. Section 4 shows the results of simulation and experiment made by WRF Model, and a summary with concluding remarks is given in section 5.

FIG. 1. (a) Track of the R/V *Dongfanghong 2* from 1100 UTC 12 Apr to 0900 UTC 13 Apr 2016 with SST (color shading with black edge) and sounding station numbers overlaid on AVHRR OISST (color shading) and AVISO SSHA (black contour at 0.5 m); (b) SST (colored contours at 2°C intervals) at 0000 UTC 13 Apr 2016 from the CTL simulation. The WRF nested model domains are denoted by D1, D2, and D3.

2. Data and methods

As shown in Fig. 1a, R/V *Dongfanghong 2* of Ocean University of China sailed across a warm eddy from 1200 UTC 12 April to 0600 UTC 13 April 2016 (local time: UTC + 10 h). A total of 10 GPS radiosondes (numbered E64–E73 in Fig. 1a) were launched along a zonal track at approximately 1–2-h intervals within the 18-h cruise (Table 1), mainly in daytime. Note that from E68 to E70, it took more than 5 h for the R/V to move a distance of 0.24° longitude (about 20 km), over which there was less than a 1°C nonmonotonic variation of SST. Thus, the changes in atmospheric conditions reported by the R/V within the eddy center primarily represent variability over time rather than space (Table 1, Fig. 1a). Vertical profiles of temperature, humidity, and wind information were obtained from GPS radiosonde soundings at a vertical resolution of 10 m. A shipboard ceilometer (CL31) sequentially measured the

TABLE 1. Station information.

Stations	Time (UTC)	Location
E64	1241 12 Apr	38.38343°N, 147.13962°E
E65	1422 12 Apr	38.32704°N, 147.35551°E
E66	1527 12 Apr	38.30018°N, 147.48822°E
E67	1712 12 Apr	38.36727°N, 147.65359°E
E68	1850 12 Apr	38.45371°N, 147.88304°E
E69	2136 12 Apr	38.25905°N, 147.98123°E
E70	0003 13 Apr	38.28786°N, 148.12169°E
E71	0122 13 Apr	38.25557°N, 148.39087°E
E72	0246 13 Apr	38.24245°N, 148.59103°E
E73	0608 13 Apr	38.33179°N, 148.99615°E

TABLE 2. Specification of WRF modeling.

	Domain 1	Domain 2	Domain 3
Grid spacing (km)	30	10	3.33
Time steps (s)	180	60	20
IC/BC	CFSR and AVHRR	Domain 1	Domain 2
Convection	KF (Kain and Fritsch 1990, 1993)		
Longwave radiation	RRTMG (Iacono et al. 2000)		
Shortwave radiation	RRTMG (Iacono et al. 2005)		
PBL	YSU (Hong et al. 2004)		
Microphysics	Lin (Lin et al. 1983)		
Soil model	Four-layer soil model		

height of cloud base (CBH) and the depth of the MABL, a temperature logger SBE 48 measured SST, and automatic weather station (ATWS) measured surface air temperature (SAT), relative humidity (RH), SLP, visibility, wind speed, and wind direction. Sampling intervals of the ceilometer, temperature logger, and ATWS were 16 s, 1 min, and 1 min, respectively.

Based on the observational CBH along this track and the definition of the low cloud from the World Meteorological Organization, the 3-h average low cloud fraction (LCF) was calculated as the frequency of time that the CL31 reported CBH between 500 and 2000 m, consistent with the hourly ship-log records in the daytime. The sensible heat flux (SHF) and latent heat flux (LHF) along the cruise were calculated from the in situ observations using the Coupled Ocean–Atmosphere Response Experiment (COARE) version 3.0 flux algorithm (Fairall et al. 2003), based on the bulk formulas (Fairall et al. 1996)

$$H_s = \rho_a c_{pa} C_h S (T_s - \theta), \quad (1)$$

$$H_l = \rho_a L_e C_e S (q_s - q), \quad (2)$$

where H_s is the sensible heat flux and H_l is the latent heat flux, S is the surface wind speed, T_s is the SST, θ is the potential temperature, q_s is the sea surface specific humidity, and q is the air specific humidity. Lifting condensation level (LCL) is calculated with the in situ observations based on an approximate formula (Pruppacher and Klett 2004)

$$\text{LCL} = 123(T - T_d), \quad (3)$$

where the T is the SAT and the T_d is the dewpoint temperature.

Reanalysis data are from National Centers for Environmental Prediction Climate Forecast System Reanalysis (CFSR), which is designed and executed as a global, high-resolution, coupled atmosphere–ocean–land surface–sea ice system to provide the best estimate of the states of

these coupled domains (Saha et al. 2010). The visible and infrared data are provided by the Clouds and Earth’s Radiant Energy System (CERES) Synoptic TOA and surface fluxes and clouds (SYN) dataset edition 4A, available at 1-h intervals on a 1.0° grid (Doelling et al. 2013). The daily optimum interpolation (OI) SST product (Reynolds et al. 2007) is from the Advanced Very High Resolution Radiometer (AVHRR), and the sea surface height anomaly (SSHA) data are obtained from Archiving, Validation and Interpretation of Satellite Oceanographic data (AVISO 2016).

The Weather Research and Forecasting (WRF) Model (e.g., Klemp et al. 2007; Michalakes et al. 2005; Wicker and Skamarock 2002; Klemp and Skamarock 2004; Skamarock et al. 2005) version 3.7 is used to investigate the physical processes involved in the responses of MABL vertical structure and its clouds to the warm eddy. A three-layer nesting method is applied in the control numerical simulation (CTL) and the “no eddy” experiment (EXP), and the SST is smoothed around the eddy edge area in the latter simulation. Section 4 provides a detailed explanation about the EXP simulation. The innermost domain D3 has the horizontal resolution of 3.3 km and is centered near the warm eddy center (Fig. 1b). The entire grid system has 46 vertical layers,¹ with 19 levels below 700 hPa in σ coordinates. The CTL and EXP are both initialized at 0000 UTC 10 April and integrated through the end of observation period at 0000 UTC 14 April 2016. The 6-hourly data from the CFSR and SST from AVHRR are used as the initial and boundary conditions. The main sets and physical options of the model are shown in Table 2.

¹ $\sigma = 1.0000, 0.9985, 0.9970, 0.9950, 0.9925, 0.9900, 0.9850, 0.9775, 0.9700, 0.9540, 0.9340, 0.9090, 0.8800, 0.8506, 0.8212, 0.7918, 0.7625, 0.7084, 0.6573, 0.6090, 0.5634, 0.5204, 0.4798, 0.4415, 0.4055, 0.3716, 0.3397, 0.2815, 0.2551, 0.2303, 0.2071, 0.1854, 0.1651, 0.1461, 0.1284, 0.1118, 0.0965, 0.0822, 0.0689, 0.0566, 0.0452, 0.0346, 0.0249, 0.0159, 0.0076, 0.$

FIG. 2. (a) CFSR 700-hPa geopotential height (black contours at 50-gpm intervals) and horizontal wind (arrows); (b) CFSR sea level pressure (black contours at 3-hPa intervals), specific humidity (blue contours at 1 g kg^{-1} intervals), and 10-m horizontal winds (arrows); (c) CTL sea level pressure (black contours at 3 hPa intervals), specific humidity (blue contours at 1 g kg^{-1} intervals), and 10-m horizontal winds (arrows) for 1800 UTC 12 Apr. (d)–(f) As in (a)–(c), but for 0000 UTC 13 Apr. Red circles mark the location of the warm eddy.

3. Observations

The warm eddy was directly under a high pressure ridge at 700 hPa where westerlies were prevailing at 1800 UTC 12 April (Fig. 2a). The high pressure ridge slowly moved eastward during the survey period (Fig. 2d). Descending motion induced by the high pressure ridge favored a temperature inversion at the MABL top, under which marine stratocumulus may develop (Myers and Norris 2013). Coincident with the high pressure ridge, a surface high pressure center was in the northwestern Pacific at 1800 UTC 12 April (Fig. 2b) that also propagated eastward (Fig. 2e). During that period, the warm eddy was located to the northwest of the high pressure center and under surface winds that shifted from northwesterly to southerly, thus bringing warm moist air from the subtropical ocean to eddy area (Figs. 2b,e).

a. Cloud development

As shown in Fig. 1a, the R/V *Dongfanghong 2* sailed eastward through the warm eddy. Over a distance of 170 km along the track, the SST increased from 17°C west of the eddy to nearly 20°C at the center of the eddy (Fig. 1a). The SAT was about $5^{\circ}\text{--}10^{\circ}\text{C}$ lower than the SST, indicating that the atmosphere was being heated by the ocean (Fig. 3b). Over the warm sea surface, a well-mixed layer developed under the major temperature inversion from 1200 to 1500 m, as determined by the

soundings and the ceilometer (Fig. 3a). A moist layer existed under the temperature inversion with RH greater than 80%, consistent with the report of cloud bases by the CL31 that confirmed the presence of low clouds.

According to the observations on the R/V, substantial changes took place after the R/V arrived at E68 close to the center of the eddy, where cloud bases were detected by the CL31 much more frequently than before (Fig. 3a). The LCF began to increase after E67 and showed a rapid increase from 0.5 to 0.82 between E67 and E71 (Fig. 3c). Also, the CBH dropped dramatically from ~ 1300 to ~ 750 m between E69 and E71, consistent with the thickening of the moist layer reported by the soundings (Fig. 3a). The ship log recorded the presence of stratocumulus above the R/V in this period; the photo (Fig. 4) taken on the R/V confirmed the stratocumulus at E70. The CERES cloud and radiative data indicate that high clouds patches covered the low clouds over the eddy during this period (not shown). The independent ceilometer and radiosonde observations from the R/V confirm the rapid development of the marine stratocumulus above the warm eddy.

b. Variations in SLP and heat flux

The SLP reported by the R/V decreased as the ship moved into the warmer sea surface (Fig. 3e) after E68. The SHF and LHF increased noticeably from 50 to

FIG. 3. Observations for E64–E73 stations from R/V *Dongfanghong 2*, including (a) relative humidity (rh, shaded), potential temperature (red contours at 2-K intervals with 284 K in bold), horizontal winds (arrows), and calculated LCL (green circles) from soundings, and the height of cloud base (CBH, black circles) and depth of MABL (red circles) from the ceilometer; (b) SST (blue line) and the difference between SST and SAT (ΔT , black line); (c) SAT (red line), dewpoint (T_d , blue line, on the same scale as SAT), and low cloud fraction (LCF, black line); (d) SHF (black line) and LHF (blue line) calculated by the COARE 3.0 algorithm; and (e) SLP (red line), surface horizontal wind speed (blue line), and surface horizontal winds (barbs) from automatic meteorological station. The time of some soundings are marked by $\blacktriangle$.

100 W m^{-2} and from 100 to 300 W m^{-2} , respectively (Fig. 3d). The enhancement in heat flux started a few hours earlier than the drop in SLP, suggestive of a delayed response of SLP to the heating, as an intermediate step of the pressure adjustment mechanism (Lindzen and Nigam 1987). The fact that SST–SAT decreased while the heat flux increased suggests a substantial contribution of wind speed to heat flux. Indeed, the southeasterly wind speed accelerated from 3 to 10 m s^{-1} during this period (Fig. 3e). These analyses suggest that a positive feedback among heat flux, low pressure, and wind speed may be operating in this case.

The eastward propagation of the surface high pressure center also led to a decrease of SLP and a shift toward south-southwesterly wind (Figs. 2b,e). The reanalysis does not show south-southeasterly winds above the eddy (Fig. 2e), as reported by the R/V at 0000 UTC 13 April, possibly due to the low resolution. A slight low pressure anomaly may be induced by warmer sea surface

FIG. 4. The photo taken on the upper deck of the R/V at E70 station.

(Ma et al. 2015; Chen et al. 2017), thus leading to an increase in pressure gradient force between the low pressure and high pressure center. This could cause a local southeasterly wind acceleration. The strengthening in wind speed produces an increase in surface heat flux which, in turn, lowers the SLP. We will distinguish the effect of the warm eddy from the effect of the background circulation with the simulation and sensitive experiment using the WRF Model.

c. Variations in the heights of the LCL and the MABL

With the shift to southerly winds (Figs. 2b,e), moisture increased in the MABL through both advection (as the specific humidity is higher in south) and evaporation [Eq. (2)]. Between E69 and E71, the T_d increased by $\sim 5^\circ\text{C}$ (Fig. 3c), the LCL dropped from above 1000 to 500 m (Fig. 3a), and the stratocumulus deepened (Fig. 3a). The height of the MABL above the eddy rose to $\sim 1500 \text{ m}$, which is noticeably higher than the MABL outside of the eddy. This change is possibly due to the deepening of the mixed layer produced by local convergence resulting from the decrease of pressure caused by heating both from the warm sea surface and from the latent heat release from clouds (Fig. 3a).

4. Mechanisms of the MABL response

Based on the discussion above, it is possible that the warm eddy contributed to the reduction of surface low pressure and enhancement of stratocumulus. The pressure perturbations are typically very small $O(0.1) \text{ hPa}$ compared to synoptic perturbations in atmospheric pressure $O(1\text{--}10) \text{ hPa}$ (Cronin et al. 2003). Thus it is very hard to distinguish the eddy influence on SLP in in situ observations. In the following section, we confirm this hypothesis by means of the WRF Model.

a. Effects of the warm eddy

The surface synoptic situation from the CTL simulation (Fig. 2c) is similar to the reanalyzed data (Fig. 2b), and we see that the warm eddy was near the center of the surface high pressure and influenced by southerly winds. The vertical cross section from the CTL (Fig. 5a) is consistent with the observations (Fig. 3a), which indicates that the model can reproduce well the synoptic pressure pattern and main characteristics of the MABL. With smoothed SST, the isolines of SST and SAT in the eddy area become more flat in the EXP simulation compared with those in the CTL simulation (Figs. 6a,b), which affects air–sea interaction (e.g., Seo et al. 2016). The differences between the CTL and the EXP simulations (CTL – EXP) are used to highlight the imprint of the eddy on the MABL.

As shown in Fig. 5b, the well-mixed layer in the EXP simulation, unlike CTL, remained under ~ 1200 m without substantial variations in vertical structure along the track. The moist layer was thinner and exhibited less cloud water above the eddy. The horizontal wind in the MABL was southerly without obvious southeasterly winds, especially after 0000 UTC, which confirms that the warm eddy produces winds with more of a southeasterly direction in the CTL than in the EXP. Moreover, the wind response to the eddy in CTL caused surface convergence (Fig. 6c).

b. Atmospheric response in the MABL and low cloud development

As shown in Fig. 6c, the difference between CTL and EXP in the warm eddy is $\sim 3^{\circ}$ – 6°C for SST and ~ 75 – 160 W m^{-2} for heat flux with the ~ 400 -km diameter of the eddy area. Furthermore, SLP declined up to 0.45 hPa in CTL compared to EXP at 0000 UTC 13 April when the stratocumulus started to develop rapidly. Southeasterly wind anomalies (CTL – EXP) become more substantial near the low pressure center and converged with northeasterly wind anomalies. The center of these anomalous changes is collocated with the eddy. All of these changes are consistent with an atmospheric response to the warm eddy (Ma et al. 2015; Chen et al. 2017).

Between E69 and E70, SLP decreases about 0.5 hPa in CTL compared to a 0.2-hPa decrease in EXP; thus, the warm eddy accounts for 60% of the total variation in SLP, slightly larger than the value given by Zhang et al. (2017) and Wang et al. (2017) for the warm flank of an SST front in KE area. Considering that the cloud water mixing ratio increases to 0.09 g kg^{-1} in CTL compared to only 0.005 g kg^{-1} in EXP, the contribution of the eddy to cloud water mixing ratio accounts for 95% of the total (Figs. 5a,b). The fact that between E69 and E70 the wind

FIG. 5. Time–height cross section of relative humidity (rh, shaded), potential temperature (red contours at 2-K intervals with 284 K in bold), cloud water mixing ratio (white contours at 0.02 g kg^{-1} intervals with the first contour at 0.01 g kg^{-1}), and horizontal wind (barbs) along the R/V track for the (a) CTL and (b) EXP simulations; (c) SLP (thick red line for CTL and thin red line for EXP), surface horizontal wind (black barbs for CTL and blue barbs for EXP), and wind speed (thick black line for CTL and thin black line for EXP) along the R/V track. The time of some soundings are marked by ▲. The bottom red line on the time coordinate axis denotes the position of the eddy according to observed SST in Fig. 3.

speed increases 3.2 m s^{-1} in CTL but only increases 2 m s^{-1} in EXP suggests that forcing by the eddy contributes up to 40% of the southerly wind acceleration (Fig. 5c).

Along the mean surface southeasterly wind anomalies direction across the eddy center (white dotted line in Figs. 6a and 6c), ascending motion appeared above the warm eddy on 0000 UTC 13 April (Figs. 7a,b). Under the control of synoptic high pressure, the ascending motion induced by the warm eddy developed below the main temperature inversion and was substantial in the MABL over the eddy center among the surrounding downward motion. A cold anomaly appeared at the MABL top (Fig. 7b), likely the results from strengthened radiative cooling at the cloud top (Lilly et al. 1968; Mechem and Kogan 2003; Myers and Norris 2013).

FIG. 6. SST (shaded, °C) and air temperature at 2 m (black contours at 1°C intervals) at 0000 UTC 13 Apr for (a) CTL and (b) EXP simulations, dotted frames mark the area of smoothing of SST in EXP simulation; (c) CTL – EXP difference at 0000 UTC 13 Apr for SST (black contours at 2°C intervals), SLP (red dotted contours at 0.1 hPa intervals), heat flux (SHF + LHF, green contours at 75 W m⁻² intervals), horizontal wind (arrows), and divergence (Divg, shaded, $1 \times 10^{-5} \text{ s}^{-1}$). The white dashed lines in (a) and (c) mark the position of the transect in Fig. 7.

The ascending motion was favorable for the development of cloud and lifting of the top of the MABL (Fig. 7a). The observations in Agulhas Current indicate that the deepening of MABL over a warm eddy was due to the enhanced the vertical turbulent mixing (Messenger and Swart 2016). In our study, the low SLP anomaly induced by the warm eddy produced the ascending motion to lift the MABL. Note that the warm eddy contributed 80% of the total magnitude of ascent (the ratio between the maximums of ascent in Figs. 7b and 7a). Latent heat release in the development of low clouds might have a positive feedback on the ascent.

The MABL top drops (Fig. 7a) and the LCL lifts with fewer cloud water mixing ratio in the MABL (not shown) downwind the warm eddy with the descending motion.

Two secondary circulations formed in the MABL along the transect. The double secondary circulations were characterized by stronger ascending branches over the warm eddy center and weaker descending branches upwind and downwind of the eddy (Fig. 7b), possibly due to the latent heat released by low clouds. This double secondary circulation pattern is different from the single one induced by local SST fronts (O'Neill et al. 2010; Tomita et al. 2013; Liu et al. 2014) and mesoscale

FIG. 7. (a) Horizontal winds (red barbs, m s^{-1}), vertical velocity (black contours at the interval of 0.08 m s^{-1} and the dashed contour for -0.04 m s^{-1}), and the height of MABL top (blue line) at 0000 UTC 13 Apr along the transect marked by white dashed line in Fig. 6a; (b) CTL – EXP difference at 0000 UTC 13 Apr along the same transect (marked by white dashed line in Fig. 6c) for air temperature (ΔT , shaded), horizontal wind $\times 10$ (red barbs; and wind barbs less than $2 \times 10^{-1} \text{ m s}^{-1}$ are omitted), vertical velocity [black contours, the unit is same as (a)], and the height of MABL top (blue line). The location of the eddy center area (where SST is over 18°C) is marked by red line segment at the bottom of each figure.

eddies (Xu et al. 2010; Chen et al. 2017) suggested by previous studies. According to Ma et al. (2015), changes in wind speed result in divergence upwind and convergence downwind of warm eddies, driving a single vertical secondary circulation. In our case, the southeasterly wind speed is relatively weaker in a high pressure area, which may favor the formation of a heat-induced low pressure anomaly and the double secondary circulation pattern.

c. Sensitivity tests with SST changes

To explore how robust the influence of the warm eddy on the atmosphere, we run three more sensitivity tests with the eddy amplitude 0.75, 0.5, and 0.25 times the real eddy amplitude (amplitude = max SST anomaly). The results are qualitatively consistent with the existing pair of simulations. The relations between these meteorological variables and the eddy amplitude are nonlinear (Fig. 8), and the nonlinear is more significant for ascending motion and the cloud water mixing ratio (Figs. 8c,d).

5. Summary and discussion

During 12–13 April 2016, the R/V *Dongfanghong 2* documented substantial variations in MABL structure and stratocumulus over a warm oceanic eddy in the KE region. Using in situ measurements and the output from control and experiment WRF simulations, the mechanisms driving the variations were investigated. Based on the modeling experiments

described in this paper, the major conclusions and some discussion follow:

- 1) Heated by the warm eddy, a surface low pressure anomaly is induced in a region of synoptic high pressure. The anomalous pressure gradient forces between the low pressure forced by the warm eddy and the surface synoptic high pressure center drive surface southeasterly wind. This southeasterly wind overlaps in phase with the background winds that veered from northwesterly winds to southerly winds, resulting in a southerly wind acceleration over the warm eddy, which increases moisture in the MABL, leading to a substantial reduction of the LCL. The abrupt development of stratocumulus over the warm eddy is attributed to this warm eddy effect (Fig. 9).
- 2) With the deepening of the low pressure, surface convergence and ascent occur over the warm eddy. Quantitative analyses indicate that the warm eddy with a 400-km diameter and 6°C SST anomaly contributes approximately for more than half of the total ascent and total cloud water mixing ratio while for a less significant part of while the total SLP variations. The latent heat release in the development of low clouds might have the positive feedback on the ascent. The double secondary circulations in the MABL along the prevailing wind direction are composed of a stronger ascending branch over the center the warm eddy with two weaker descending ones over the upwind and downwind sides of the eddy.

FIG. 8. (a) From E69 to E70, the SLP decreasing ratio $[(CTL - EXP)/CTL]$ relationship to the eddy amplitude ratio (trial eddy amplitude/real eddy amplitude); (b) as in (a), but for sea surface horizontal wind velocity (vel) increasing ratio; (c) as in (a), but for maximum ascending motion (W) ratio at (38.3°N, 148°E), 0000 UTC 13 Apr; (d) as in (c), but for maximum cloud water mixing ratio (QCLOUD) ratio.

Previous studies indicated that a cold eddy and SST front could force low cloud development in KE area (Zhang et al. 2017; Wang et al. 2017; Jiang et al. 2018, manuscript submitted to *J. Geophys. Res. Atmos.*). Low clouds developed on the upwind area of the cold eddy due to low-level convergence produced by the increase of the wind speed on the upwind edge. Furthermore, under the control of the northwesterly winds, the greater difference between SST and SAT on the warmer area (downwind) of the SST front is the main cause of the increase in local heat flux, resulting in the low pressure developing. In our observations and modeling experiments, the heat flux–low pressure–wind speed mechanism was the main cause of the rapid development of the clouds above the warm eddy center under the force of surface southerly winds. And in turn, the southeasterly wind acceleration over the warm eddy may enhance the surface heat flux and hence deepen the low pressure, which forms a positive feedback effect, as shown in the observations (Figs. 3d,e).

These conclusions are summarized from a single case study. More observations are needed to further investigate the detailed physical processes in the atmospheric response to oceanic eddies. More attention should be

paid to warm eddies since the heat flux there is much larger than that in cold eddies for small-scale air–sea interactions (Jiang et al. 2018, manuscript submitted to *J. Geophys. Res. Atmos.*).

FIG. 9. Schematic of the main process of the development of stratocumulus (Sc) above the warm eddy center under the force of the background westerly winds and synoptic high pressure (H), displaying the local low pressure (L) forced by the increased surface heat flux (HF) over the warm eddy. This yields low-level convergence and rising motion over the warm eddy. The LCL lowered and the top of the MABL rose, thereby aiding the development of the stratocumulus.

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