ABSTRACT

An integration of a high-resolution coupled general circulation model whose ocean component is eddy permitting and thus able to reproduce a sharp gradient in sea surface temperature (SST) is analyzed to investigate air–sea heat exchanges characteristic of the midlatitude oceanic frontal zone. The focus of this paper is placed on a prominent SST front in the south Indian Ocean, which is collocated with the core of the Southern Hemisphere storm track. Time-mean distribution of sensible heat flux is characterized by a distinct cross-frontal contrast. It is upward and downward on the warmer and cooler flanks, respectively, of the SST front, acting to maintain the sharp gradient of surface air temperature (SAT) that is important for preconditioning the environment for the recurrent development of storms and thereby anchoring the storm track. Induced by cross-frontal advection of cold (warm) air associated with migratory atmospheric disturbances, the surface flux is highly variable with intermittent enhancement of the upward (downward) flux predominantly on the warmer (cooler) flank of the front. Indeed, several intermittent events of cold (warm) air advection, whose total duration accounts for only 21% (19%) of the entire analysis period, contribute to as much as 60% (44%) of the total amount of sensible heat flux during the analysis period on the warmer (cooler) flank. This antisymmetric behavior yields the sharp cross-frontal gradient in the time-mean flux. Since the flux intensity is strongly influenced by local magnitude of the SST–SAT difference that tends to increase with the SST gradient, the concentration of the flux variance to the frontal zone and cross-frontal contrasts in the mean and skewness of the flux all become stronger during the spinup of the SST front. Synoptically, the enhanced sensible heat flux near the SST front can restore SAT toward the underlying SST effectively with a time scale of a day, to maintain a frontal SAT gradient against the relaxing effect of atmospheric disturbances. The restoration effect of the differential surface heating at the SST front, augmented by the surface latent heating concentrated on the warm side of the front, represents a key process through which the atmospheric baroclinicity and ultimately the storm track are linked to the underlying ocean.
1. Introduction

Impacts of midlatitude air–sea interaction on the climate system and its long-term variability are not fully understood. Some of the numerical experiments (e.g., Latif and Barnett 1994) and observational and/or statistical data analysis (Schneider and Cornuelle 2005; Qiu et al. 2007) have suggested that midlatitude air–sea interaction can amplify decadal climate variability over the North Pacific. Nevertheless, how significantly the midlatitude ocean can impact the overlying atmosphere is still under debate. Summarizing the results of numerical experiments carried out previously with atmospheric general circulation models (AGCMs), Kushnir et al. (2002) concluded that no coherent large-scale atmospheric response had been obtained to prescribed midlatitude sea surface temperature (SST) anomalies. Rather, midlatitude SST anomalies in general have been shown to be forced primarily by atmospheric anomalies (Frankignoul 1985) through changes in surface heat fluxes (Cayan 1992; Alexander 1992), Ekman heat transport (Yasuda and Hanawa 1997), and entrainment at the bottom of the oceanic mixed layer (Miller et al. 1994). Thus the midlatitude oceans are believed to be passive to atmospheric anomalies, including those generated under the remote influence of tropical variability (Lau 1997; Alexander et al. 2002; Deser et al. 2004), except that heat exchanges with the adjusted underlying ocean can reduce thermal dumping of the atmospheric anomalies (Alexander 1992; Barsugli and Battisti 1998).

In recent years, however, impacts of midlatitude oceanic frontal zones on the atmosphere have been gaining increasing attention, as their fine structures and variability have become apparent by recent satellite observations and high-resolution numerical models (Xie 2004; Chelton et al. 2004; Nakamura and Kasim 2003; Nonaka et al. 2006; Minobe et al. 2008; Nakamura et al. 2008). Satellite observations have shown that SST and surface wind speed are correlated positively in such midlatitude oceanic frontal zones as those along the Kuroshio Extension (Nonaka and Xie 2003), the Gulf Stream (Sweet et al. 1981; Park and Cornillon 2002; Chelton et al. 2004; Xie 2004; Minobe et al. 2008), the Antarctic Circumpolar Current (ACC; O’Neill et al. 2003; Liu et al. 2007), and the Brazil Current (Tokinaga et al. 2005). The same SST–wind relationship is also found in historical data and in situ observations (Tokinaga et al. 2005, 2006). Unlike the situation where SST anomalies are generated in response to basin-scale wind anomalies, the positive correlation is an indication of oceanic thermal forcing on the planetary boundary layer (PBL), especially in regions of strong ocean currents including midlatitude oceanic frontal zones. This is also supported by observed positive correlation between surface heat flux and SST anomalies (i.e., enhanced upward heat flux over positive SST anomalies) in the western North Pacific subpolar frontal zone (Tanimoto et al. 2003; Yasuda and Kitamura 2003; Nonaka et al. 2006, 2008).

More recent studies have further suggested that the oceanic influence is not limited to PBL but extend into the free troposphere (Liu et al. 2007; Minobe et al. 2008). Atmospheric GCM (AGCM) experiments by Inatsu et al. (2003) and Inatsu and Hoskins (2004) showed that a midlatitude oceanic frontal zone, such as the one over the south Indian and Atlantic Oceans, can anchor the storm track and polar-front jet (PFJ; or subpolar jet) that accompanies it. An idealized AGCM experiment by Brashaw et al. (2008) with zonally symmetric SST distribution showed high sensitivity of the position and intensity of a midlatitude storm track to the meridional SST profile and the strength of a subtropical jet (STJ). Another idealized AGCM experiment by Nakamura et al. (2008) with zonally symmetric SST showed that a midlatitude frontal SST gradient anchors a storm track by energizing migratory cyclones and anticyclones. The resultant increase in eddy transport of westerly momentum from the subtropics strengthens a PFJ, and the enhanced poleward heat transport strengthens surface westerlies along the frontal zone. In a manner consistent with the numerical experiments, collocation is observed over the south Indian Ocean among the cores of the midlatitude SST front, storm track, and PFJ (Nakamura and Shimo 2004; Nakamura et al. 2004). It has been further argued (Hoskins and Valdes 1990; Nakamura et al. 2004, 2008) that the enhanced storm-track activity may in turn act to exert positive feedback on the midlatitude ocean through strengthening the surface westerlies to drive the currents that accompany the oceanic frontal zone.

A midlatitude SST frontal zone can enhance local storm-track activity through moisture supply from the ocean on its warmer flank and through tight cross-frontal gradient of surface air temperature (SAT). The latter dry process is important from a viewpoint of baroclinic storm development via the coupling of upper-level eddies with a surface baroclinic zone (Hoskins et al. 1985), but developing storms act to relax the gradient by transporting heat poleward. Without effective restoration, the SAT gradient would be lessened after the passage of a cyclone, and the recurrent development of atmospheric disturbances would therefore be reduced. Nakamura et al. (2004, 2008) argued that such a relaxed SAT gradient could be restored effectively through associated enhancement of upward and downward sensible heat fluxes at the surface on the warmer and cooler sides of the SST front, respectively.

In this present study, we investigate synoptic and statistical characteristics of the restoration processes that
act to maintain a frontal SAT gradient across a mid-latitude SST front, paying attention to surface heat flux distributions. Through our analysis of output of a high-resolution coupled GCM (CGCM) that reproduces SST fronts, we show that sensible heat flux is enhanced intermittently on either side of a frontal zone in association with migratory atmospheric disturbances, acting effectively to adjust SAT toward the underlying SST and thereby restore the SAT gradient at the front. Our focus is placed on a prominent SST front that forms along the warmer flank of the ACC in the south Indian Ocean sufficiently away from landmass.

This paper is organized as follows. Section 2 outlines the CGCM and an atmospheric reanalysis dataset used in this study, and section 3 describes the impacts of the SST front on distributions of surface heat flux and SAT. These results are discussed in sections 4 and 5 provides a summary and conclusion.

2. Model and reanalysis data

**CFES**

In this study we analyze the first five years of an integration of the CGCM for the Earth Simulator (CFES; Komori et al. 2008a,b), which consists of the Atmospheric GCM for the Earth Simulator version 2 (AFES2; Enomoto et al. 2008) and the Coupled Ocean–Sea Ice model for the Earth Simulator (OIFES; Komori et al. 2005). AFES2 is an improved version of AFES (Ohfuchi et al. 2004), which is based on the Center for Climate System Research (CCSR; University of Tokyo)/National Institute for Environmental Studies (NIES) AGCM version 5.4.02 (Numaguti et al. 1997) but has been modified substantially to achieve the best computational efficiency on the vector-parallel architecture of the Earth Simulator (ES). A land surface model called the Minimal Advanced Treatments of Surface Interaction and Runoff (MATSIRO; Takata et al. 2003) is incorporated in CFES. In OIFES, a sea ice model is coupled with the Ocean model for the Earth Simulator (OFES; Masumoto et al. 2004; Sasaki et al. 2008), which is based on the Modular Ocean Model version 3 (MOM3; Pacanowski and Griffies 2000) developed at the Geophysical Fluid Dynamics Laboratory (GFDL) with substantial modifications added. The sea ice model is based on a model developed at the International Arctic Research Center (IARC) of the University of Alaska (Zhang and Zhang 2001). It has been improved by introducing a scheme for elastic–viscous–plastic rheology (Hunke and Dukowicz 2002). In CFES, surface sensible and latent heat fluxes are calculated by using the bulk formulas whose coefficients are evaluated basically with a method by Louis (1979) in taking the difference in the roughness length between heat and momentum fluxes into account (Uno et al. 1995).

For the particular integration used in this study, the horizontal resolution of AFES2 is set T239 (~50 km) with 48 $\sigma$ levels (including 12 layers below $\sigma = 0.8$). The top is placed at the $\sigma = 0.003$ (~0.3 hPa) level. The horizontal resolution for OIFES is set 0.25° with 54 levels whose intervals are 5 m immediately below the surface. The model maximum depth is 6065 m. The ocean GCM is thus eddy permitting and coupled with the atmospheric GCM in which meso-$\alpha$-scale phenomena can be fully resolved. The initial conditions for AFES2 were taken from the atmospheric fields for 1 November 1982 based on the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; Uppala et al. 2005). For the initial condition for the ocean model, the climatological temperature and salinity fields for January were taken from the World Ocean Atlas 1998 (Antonov et al. 1998a,b;c; Boyer et al. 1998a,b,c) while no motion was assumed. Initially, sea ice was assigned with full concentration where the initial SST was below the freezing temperature. After integrated over two months from the initial state, AFES2 was coupled with OIFES for its spinup.

For comparison with the model output, we use 6-hourly fields of the ERA-40 data provided by the ECMWF data server. We also use monthly-mean SST fields based on measurements by the Advanced Microwave Scanning Radiometer for Earth Observing System (AMSR-E) on the Aqua satellite. The dataset has been produced by the Remote Sensing Systems with horizontal resolution of 0.25°.

3. Results

a. Snapshots

As apparent in snapshots of surface fields in Fig. 1, CFES can well represent atmospheric disturbances and oceanic frontal signatures owing to its high horizontal resolution. On 30 July of year 1 (the first year) of the model integration, an atmospheric low pressure system is migrating eastward around [42°S, 65°E] with a cold southerly wind to the west of the trailing atmospheric cold front (Fig. 1a). The advection of the cold air onto the warmer side of the SST front results in a large SST–SAT difference and therefore strong upward fluxes of sensible and latent heat from the ocean. As obvious in Fig. 1b, the distribution of the SST–SAT difference exhibits pronounced local maxima on the warmer flank of the SST front between 45° and 60°E. The associated maxima of sensible heat flux (SHF), which is proportional...
to the SST–SAT difference, are a manifestation of the influence exerted by the frontal SST structure that is governed dominantly by oceanic processes.

Figure 2 shows latitude–time sections of atmospheric and oceanic variables for the 55°E meridian, the longitude that corresponds to the core of the strong SST front (Fig. 1b). Frequent equatorward spills of cold air masses associated with recurrent development of atmospheric disturbances (Figs. 2a,b) yield intermittent augmentation of the SST–SAT difference and the corresponding enhancement of SHF and latent heat flux (LHF), which occur predominantly on the warmer flank of the SST front (Figs. 2b–d) as in Fig. 1. On the cooler flank of the SST front, in contrast, the poleward advection of warm air ahead of individual cyclones yields strong downward SHF intermittently. Thus the time-mean SHF is maximized on the warmer flank of the SST front (Fig. 2c), where the time-mean SST–SAT difference is also largest. The time-mean SHF deceases rapidly poleward across the SST front, and the negative SHF is strongest on the cooler flank of the front. A similar cross-frontal decline is evident in the time-mean LHF (Fig. 2d), although it is weakly positive even on the cooler side of the SST front.

To confirm the importance of transient atmospheric disturbances for the time-mean SHF, a scatterplot between the surface meridional velocity and SHF is shown in Fig. 3a at the warmer flank of the SST front at 40°S, 55°E (Fig. 3b). It is apparent that SHF is greatly enhanced in association with strong cross-frontal equatorward winds accompanied by transient atmospheric disturbances, consistent with Fig. 2. Furthermore, a contribution from each 5 m s⁻¹ bin of the surface meridional wind velocity to the total SHF observed during the entire bimonthly period of Fig. 2 is examined (Fig. 3b). Only about 30% of total SHF during this bimonthly period (black curve) is accounted by events of weak meridional wind (whose intensity is less than 5 m s⁻¹) whose total duration accounts for about 70% of the entire bimonthly period (black bars), and about 70% of the total SHF is...
released during the less frequent (about 30%) strong meridional wind. More specifically, as much as 60% of the total SHF is released during several intermittent events in which the meridional wind velocity ($v_{10}$) is equatorward ($v_{10} < 0$) and stronger at least by a unit standard deviation (8.3 m s$^{-1}$) than the time-mean value (1.5 m s$^{-1}$ poleward), whose total duration accounts for only 21% of the entire period. Similarly, at the cooler flank of the SST front (45°S, 55°E), intermittent (18.5% frequency) events of strong poleward winds whose intensities exceed the time-mean value by more than a unit standard deviation account for 44% of the total negative SHF (92.4% of the net total SHF, as about half of the negative SHF is canceled as suggested in Fig. 2c) for the entire bimonthly period. These results confirm the primary importance of transient atmospheric disturbance for forming the time-mean SHF field.

Associated with the passage of transient atmospheric disturbances, both the SHF and LHF are highly variable in time, especially within the oceanic frontal zone, where their standard deviations exhibit well-defined maxima (Figs. 2c,d). The SHF variance is distributed rather symmetrically about the frontal axis (Fig. 2c), since fluctuations in the cross-frontal wind component can enhance the flux on both sides of the axis. In contrast, the distribution of the LHF variance is markedly asymmetric about the frontal axis (Fig. 2d). This is because the wind fluctuations can induce large variations in LHF only on the warmer side of the frontal zone, where the Bowen ratio is particularly low because of the nonlinearity in the Clausius–Clapeyron equation. This asymmetric distribution indicates that moisture supply to individual storms occurs mostly from the warmer side of the oceanic frontal zone.

It is noteworthy that bimonthly SAT preserves its frontal gradient across the oceanic frontal zone despite a number of atmospheric disturbances traveling through this region have acted to relax the SAT gradient by

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**Fig. 3.** (a) Scatterplot between the northward (equatorward) surface (10 m) wind velocity $v_{10}$ (horizontal axis) and SHF (vertical axis). (b) Histograms for $v_{10}$ frequency (black bars; left axis) and the total SHF during the bimonthly period (W m$^{-2}$; white bars; right axis) for each bin of $v_{10}$ with 5 m s$^{-1}$ intervals. The accumulative fraction (%) to the total SHF is superimposed (black curve; left axis). All plots are based on 6-hourly CFES output from July through August in year 1 at (40°S, 55°E) on the warm flank of the SST front.

**Fig. 4.** Six-hourly time series for (a) SAT (°C; solid curve) and SST (°C; dashed curve), (b) SHF (W m$^{-2}$; solid curve; left axis) and surface (10 m) wind speed (m s$^{-1}$; dashed curve; right axis), and (c) SHF (W m$^{-2}$; solid curve; left axis) and SST − SAT (°C; dashed curve; right axis). (d) Pressure (hPa)–time section of the atmospheric ML depth (black curve). For reference, thin gray lines are superimposed for altitudes of 200, 1000, and 2000 (m). The top of ML is defined instantaneously as the level at which virtual potential temperature is higher than at the surface by 1°C. All variables are simulated in CFES from July through August in year 1 at 40°S, 55°E on the warm flank of the SST front. In (b) and (c) correlation coefficient between the plotted time series is indicated at the lower-left corner.
transporting heat poleward. This implies a restoration possibly through air–sea heat exchanges in the oceanic frontal zone. Nakamura et al. (2004, 2008) argued that SAT gradient relaxed by atmospheric eddy heat transport augments the SST–SAT difference in magnitude on both sides of the oceanic frontal zone and the resultant enhancement of upward and downward SHF on the warmer and cooler sides of the frontal zone, respectively, can effectively restore the SAT gradient toward the underlying SST gradient. Nakamura et al. (2008) called this restoration “oceanic baroclinic adjustment.” Figure 2c suggests that essentially the same restoration process may be operating in association with individual atmospheric eddies, since upward and downward SHF intermittently enhances predominantly on the warmer and cooler sides of the frontal zone, respectively. The restoration of the SAT gradient should be important for preconditioning the environment for recurrent development of the disturbances and thereby the formation of the storm track.

b. SAT restoration

In the following, we attempt to assess quantitatively how effectively SAT gradient relaxed by atmospheric processes can be restored toward the underlying SST gradient through the oceanic baroclinic adjustment. The restoration time scale can be estimated based on the approximate heat balance in the atmospheric mixed layer (ML), in which we assume temperature to be vertically uniform and thus equal to SAT. Although horizontal advection is critically important as a forcing of SAT fluctuations as discussed in the next subsection, it may be instructive to assess the efficiency of the restoration that SHF can potentially exert on SAT by considering a hypothetical case where a given SAT anomaly is going to be modified primarily via SHF at the ocean surface; that is,

\[
\frac{\partial \text{SAT}}{\partial t} \approx \frac{\text{SHF}}{\rho C_p H}, \tag{1}
\]

where \( \rho \) is air density, \( C_p \) is the specific heat of air at constant pressure, and \( H \) is the ML depth. The bulk formula for SHF may be expressed as

\[
\text{SHF} = \rho C_H C_p W(SST - \text{SAT}), \tag{2}
\]

where \( W \) denotes the surface wind speed. In (2) we assume both \( \rho \) and the exchange coefficient \( C_H \) to be constant for simplicity. We consider a situation where the passage of atmospheric disturbances changes wind direction and SAT without changing wind speed substantially within the ML. In this “stormy” situation SHF can be approximated to be proportional to \( \text{SST} - \text{SAT} \), and then (1) may be written as

\[
\frac{\partial \text{SAT}}{\partial t} \approx \frac{\alpha (\text{SST} - \text{SAT})}{\rho C_p H}, \tag{3}
\]

where \( \alpha \) is a constant of proportionality. In a midlatitude oceanic frontal zone, including the one in the south Indian Ocean, temporal variations in SST tend to be much smaller than those in SAT owing to the deep ocean ML (with its bottom well below 100-m depth especially in winter) and advective effects by strong ocean currents. Then (3) can be rewritten as

\[
\frac{\partial (\text{SAT} - \text{SST})}{\partial t} \approx -\frac{\alpha (\text{SST} - \text{SAT})}{\rho C_p H}, \tag{4}
\]

and the restoration time scale of SAT toward SST due only to SHF can thus be given by \( \tau_R \approx \rho C_p H/\alpha \).

In Fig. 4, time series of SST, SAT, \( \text{SST} - \text{SAT} \), and SHF on the warmer flank of the SST front (40°S) are depicted to validate the above assumptions. Compared to \( \partial \text{SAT}/\partial t \), \( \partial \text{SST}/\partial t \) is indeed negligibly small (Fig. 4a) because of the deep oceanic ML, despite the heat loss through LHF and SHF that spontaneously exceeds 600 W m\(^{-2}\) in association with the passage of atmospheric disturbances (Figs. 2c,d). This steadiness of SST validates the assumption made for (4). The influence of the heat fluxes on SST can be estimated with a heat budget equation similar to (1) but for the oceanic ML with \( C_p = 4000 \text{ J kg}^{-1} \text{ K}^{-1} \) and \( \rho = 10^3 \text{ kg m}^{-3} \) for seawater. For a mixed layer of 100-m depth, heat loss of 600 W m\(^{-2}\) can induce cooling as weak as 0.13°C day\(^{-1}\). In reality, this cooling is compensated by the thermal advection by mean currents, and SST is thus almost constant on the time scale of months. Therefore the SST–SAT difference varies primarily with fluctuating SAT. On time scales of years to decades, however, SST can vary noticeably and may become important in determining the SST–SAT difference.

A comparison between the time series of \( W \) and SHF (Fig. 4b) indicates that \( W \) remains above 5 m s\(^{-1}\) most of the time, and SHF fluctuations are not much dependent on \( W \). In fact, their correlation coefficient is \( r = 0.12 \) for the bi-monthly period. In contrast, the SHF fluctuations are highly correlated with \( \text{SST} - \text{SAT} \) with \( r = 0.91 \) (Fig. 4c). Therefore SHF is not sensitive to \( W \) but nearly proportional to \( \text{SST} - \text{SAT} \) with their proportionality constant estimated as \( \alpha \approx 25 \text{ W m}^{-2} \text{ K}^{-1} \). We regard the top of the atmospheric ML as the level at which virtual potential temperature becomes 1 K higher than at the surface. In the periods of cold-air advection, the ML top thus defined is estimated as \( H \approx 1500 \text{ m} \) based on Fig. 4d. From these estimations with \( C_p = 1004 \text{ J kg}^{-1} \text{ K}^{-1} \) and \( \rho = 1.2 \text{ kg m}^{-3} \), the restoration time scale \( \tau_R \) is estimated
as about $7.2 \times 10^4$ s, or slightly shorter than 1 day. The
time scale could become even shorter if part of the
moisture supplied from the ocean is condensed within the
ML, as discussed in the next subsection.\(^1\) On the cooler
flank of the SST front (Fig. 2c), downward SHF is en-
hanced when a warm air mass is advected by northerly
winds, and its variability leads to an estimation of $\alpha \approx
25 \text{ W m}^{-2} \text{ K}^{-1}$. Under the enhanced stratification over
the cool ocean, the shallowness of the ML with $H$ as small
as $\sim 200$ m (Fig. 4d) renders the restoration even more
effective, with $\tau_R$ well shorter than a day on the cooler
side of the SST front.

This effective adjustment of SAT toward the un-
derlying SST is crucial for the effective restoration of
the SAT gradient via the oceanic baroclinic adjustment
against the relaxing effect of atmospheric disturbances,
which is necessary for their recurrent development to
anchor the storm track (Nakamura et al. 2008). In our
crude estimation presented above, however, we pur-
posefully neglected the contribution from thermal advec-
tion by the cross-frontal wind component. As discussed
in the next subsection, the thermal advection acts against
the SHF contribution to retard the restoration effect
acting on the SAT gradient.

c. Composite analysis

The snapshots shown in Fig. 1 present a typical ex-
ample of how the SST–SAT difference is modified in the
vicinity of the prominent SST front by thermal advec-
tion associated with atmospheric disturbances. To sub-
stantiate this finding based on the snapshots, we have
conducted a composite analysis for strong disturbances
that travel through the frontal zone within the bimonthly
period as depicted in Fig. 2. As a reference index for the
compositing, we used the meridional wind velocity sim-
ulated at the surface (more precisely, 10-m wind $v_{10}$) at
$42^\circ S, 55^\circ E$, to which 8-day high-pass filtering had been
applied to extract the signal of migratory atmospheric
disturbances. The compositing was based on the events
of maxima of the high-pass-filtered $v_{10}$ (i.e., southerlies)
that are stronger than 1.5 times its standard deviation
during the bimonthly period, as shown in Fig. 5a.

Lag composite maps in Fig. 5 show that a region of
large SST–SAT difference and strong upward SHF moves eastward and extends toward lower latitudes,
following a migratory low pressure system and a trailing
cold front. The simultaneous composite maps in Figs. 5f–i
indicate that southerly winds behind the cyclone trans-
port a cold air mass equatorward across the SST front,
augmenting the SST–SAT difference and upward SHF
particularly on its warmer flank, as suggested in the
snapshots (Fig. 1). The cold-air advection relaxes the
SAT gradient in the oceanic frontal zone between $40^\circ$
and $45^\circ S$ (Figs. 5f,j), but the gradient is restored within
$24-36$ h (Figs. 5n,r). The relaxation and the subsequent
restoration of the SAT gradient are also evident in the
meridional profile of SAT for $55^\circ E$ plotted in the right-
most column of Fig. 5. The restoration time scale is
somewhat longer than our crude estimation presented
above, because of the counteracting advective effect by
the cross-frontal wind component. In the longitudinal
sector of $50^\circ-55^\circ E$, for example, a cold front passes
through the SST frontal zone ($40^\circ-45^\circ S$) $\sim 12$ h before
the peak time of the southerly wind. For the next 24 h the
relaxation of the SAT gradient occurs in the frontal zone,
despite the strong upward SHF acting to restore the
SAT–SAT difference. The recovering of the SAT grad-
ient appears to be completed as the cold advection ceases.

At $42^\circ S, 55^\circ E$ on the warmer flank of the front, the sign
reversal of SHF (thin solid line in Fig. 5a) tends to im-
mEDIATELY follow that of the meridional wind $v_{10}$. South-
"erly winds bring cool SAT anomalies (dashed line in
Fig. 5a), augmenting upward SHF that damps the anom-
aliies to zero just after winds become northerly. Indeed, the
heat balance in the atmospheric ML around the SST
frontal zone (Fig. 6) qualitatively\(^2\) indicates that strong
cooling (solid curve in Fig. 6a) is caused by the cool
advection (dashed curve in Fig. 6b with the reversed sign),
followed immediately by warming by SHF (solid curve in
Fig. 6b). This is confirmed by asymmetric lag correlations
between the advection and SHF: $r = -0.43$, $-0.66$, and
$-0.70$ when SHF leads the advection term by 6 h and 0 h
and lags by 6 h, respectively. To understand their phase

\(^1\) Considering the crucial importance of meridional advection as
a forcing of SAT anomalies (Figs. 2 and 4), it may be possible to
estimate a meridional length scale of the atmospheric response by
replacing the time derivative term in (1) with a meridional advec-
tion term. If the ratio of the meridional wind velocity $V$ to the total
wind speed $W$ is nearly constant, the dependency of SHF on $W$ can
be neglected, and the length scale that is insensitive to $W$ can be
estimated. From the aforementioned estimations with $C_H = 1.3 \times
10^{-3}$, the length scale is estimated as about $1500$ km in a Lagrang-
ian sense if $V/W = 1$. In reality, however, the ratio $V/W$ does vary in
time (not shown), and so should the length scale estimated. Since $V$
and cold-air advection across the SST frontal zone both tend to in-
tensify behind cyclones, the ratio $V/W$ and the length scale both tend
to be maximized, acting to reduce the meridional SAT gradient.

\(^2\) Unfortunately, diabatic heating within the ML is not available
and its accurate estimation is quite difficult. Also, from the avail-
able outputs, it is not possible to precisely estimate the entrainment
rate at the top of the ML, which can modify the ML height and
temperature in the ML. Then, the heat balance analysis here can-
not be quantitative. The purpose of our heat balance analysis is
limited to confirming the relation between the temperature ad-
vection and SHF. The investigation of the accurate heat balance in
the ML is beyond the scope of the present study.

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FIG. 5. (a) Eight-day high-pass-filtered time series of surface meridional wind velocity \(v_{10}^1\) standardized by its standard deviation (thick solid curve, left axis), SAT (dashed curve, right axis in °C), and SHF (thin solid curve, right most axis in W m\(^{-2}\)). All sampled at 42°S, 55°E, just around the axis of the SST front (indicated with dots in the left column). Composite maps for (b),(f),(j),(n),(r) SAT; (c),(g),(k),(o),(s) SST and SST − SAT (°C; shaded as indicated at the bottom); (d),(h),(l),(p),(t) SST and SHF (W m\(^{-2}\); shaded as indicated at the bottom); and (e),(i),(m),(q),(u) meridional profiles of SAT (°C; solid curve) and SST (°C; dashed curve) at 55°E. Vectors indicate surface wind. Lags of −12 [(b)–(e)], 0 [simultaneous, (f)–(i)], 12 [(j)–(m)], 24 [(n)–(q)], and 36 [(r)–(u)] hours have been imposed for the compositing on the basis of wind maxima \(v_{10}^1 > 1.5\) as marked with open circles in (a). Contours intervals for SST and SAT are 1°C. For shaded variables, white (black) contours are added for highlighting their positive (negative) values.
relationship, we consider a simple model for the heat balance in the atmospheric ML with the restoration and the advective effects in the appendix. From the simple model, the phase lag between $v_{00}$ and SAT or SHF is estimated as 0.4 day, reflecting the strong restoration effect of SHF. Though under a crude approximation, the phase lag is roughly consistent with the phase relationships observed in Fig. 4a. It should be noted that the advection is not effective for the actual warming, since the advective warming tends to lag behind the SHF warming (Fig. 6b).

The sum of SHF and the advection (dashed curve in Fig. 6a) tends to underestimate the actual warming
tendency (the ratio of variances of the former to the latter in this bimonthly period is 0.34, although they have a high correlation of $r = 0.76$), which is likely because no contributions from diabatic heating and entrainment at the top of the ML to the heat budget in the ML are included in our evaluation. Although the contribution of diabatic heating cannot be estimated quantitatively, time series of cloud cover and outgoing longwave radiation (OLR; Fig. 6c) imply the formation of low-level clouds immediately after some events of the cold-air outbreaks. For such an event around 15–20 July, for example, cloud cover increases after cold advection with relatively high OLR, suggesting the formation of low-level clouds. Although cool advections are not always accompanied by low-level cloud formation, when they are formed the associated diabatic heating can strengthen the restoration of SAT by SHF.

4. Discussion

a. High variability in air–sea heat exchanges within the oceanic frontal zone

As indicated by its meridional profile for 55°E shown in Fig. 2c, temporal variance of SHF exhibits a pronounced maximum within the oceanic frontal zone between 40° and 45°S. As discussed in the preceding section, a sharp cross-frontal contrast in SHF is important for the effective restoration of SAT toward SST. The pronounced variability of SHF is a manifestation of the SAT gradient restoration against the relaxing effect of atmospheric disturbances. The restoration enables the recurrent development of eddies, which is thus critical for the formation of the storm track. To understand why the high variability in SHF is confined to the oceanic frontal zone, meridional profiles of the SHF variance and its components are plotted in Fig. 7.

In bulk formula for SHF (2), we assume both $\rho$ and the exchange coefficient $C_H$ to be constant for simplicity. Then, the variance of SHF can be approximated as

$$\overline{SHF^2} \approx \rho^2 C_H^2 C_P^2 \left[ W^2 \Delta T^2 + W^2 \Delta T'^2 + 2 W \Delta T W' \Delta T' \right].$$

(5)

where $\Delta T = SST - SAT$ and the primes denote deviations from time averages (i.e., anomalies). It is evident in Fig. 7a that the sharp peak of the SHF variance within the oceanic frontal zone is mainly due to the
second term in (5). The term is proportional to the variance of $\Delta T$, which, unlike $W$, indeed peaks within the oceanic frontal zone (Fig. 7b). Since SST is nearly constant in time (Figs. 2 and 4), the variance of $\Delta T$ mostly represents the SAT variability that arises mainly from thermal advection associated with fluctuating meridional wind ($v'$). The storm track that forms along the oceanic frontal zone causes particularly large fluctuations in $v'$ along the tight mean SAT gradient, which contributes to the confinement of the SHF variance to the frontal zone.

As shown in Fig. 7c, the time-mean surface $W$ is stronger in midlatitudes than in the subtropics because of downward transfer of westerly momentum by baroclinically growing disturbances along the storm track (Nakamura et al. 2004). Interestingly, within its broad midlatitude peak, the time-mean $W$ also exhibits a weak but well-defined local minimum over the SST frontal zone and slightly to the south. This is a manifestation of the corresponding local maximum in thermal wind shear associated with tight SAT gradient maintained across the SST frontal zone.

b. Sensitivity to the SST gradient intensity

The results shown above suggest that a strong SST front can localize the heat release from the ocean. This implies that if strong SST frontal structures are not well resolved in AGCMs and/or CGCMs because of the limited resolution of the models or SST data assigned to the AGCMs, the heat release from the ocean would not be well represented. To confirm the suggested influence by strong SST gradient, we examine the SHF distribution during the first few months of the CFES integration. At that time, the ocean component of CFES has not been fully spun up from its initial state at rest, and neither the ocean currents nor the attendant SST front in the south Indian Ocean have been matured (Fig. 8). This gives us a good opportunity to confirm the importance of SST gradient for the SHF distribution without adding any artificial modifications to the model output or performing any additional model experiments. For this purpose, we compare the surface fields on the same calendar day (1 February) between years 1, 3, and 5 of the integration plotted in Figs. 8 and 9. We focus on the
month of February, because frontogenesis is fairly fast in the CFES ocean surface layer and the frontal zone with tight SST gradient has been matured by July of year 1 (Fig. 2). In February of year 1, however, the SST front is still premature (Figs. 8b and 9a) with the SST gradient not as tight (less than 3 K per latitude) as in year 3 (Figs. 8d and 9a; more than 4 K per latitude across the frontal zone). Correspondingly, localized enhancement of the SST–SAT difference around the SST frontal zone in year 1 is less pronounced than in year 3 (Figs. 8b,d), and so is the time-mean SHF contrast between the warmer and cooler flanks of the frontal zone (Fig. 9b). In addition, Fig. 9c indicates that the SHF variability at 55°E in year 1 is much less confined to the frontal zone than in years 3 and 5, which arises from the weaker and less confined variability in the SST–SAT difference in February of year 1 under the weaker SST gradient across the oceanic frontal zone around 43°S. These results suggest that frontal SST gradient is important for the localized enhancement of SHF and thereby for the restoration of SAT gradient. In fact, SAT gradient in February across the SST frontal zone between 20° and 70°E in year 3 (Fig. 8c) is apparently stronger than in year 1 (Fig. 8a).

Figure 9d shows the corresponding meridional profiles of the SHF skewness along the 55°E meridian. In a general meteorological circumstance, skewness is significant for a variable whose distribution is characterized by a zone of its tight meridional gradient (Nakamura and Wallace 1991), and SAT in the vicinity of the SST front is such a variable. Indeed, large skewness is simulated in CFES on both sides of the SST front (Fig. 9d), and the skewness changes its sign across the front from positive on its warmer side to negative on its cooler side. As found in Fig. 2c, this cross-frontal sign reversal is a statistical manifestation of the tendency for the cold (warm) air advection associated with atmospheric disturbances to induce strong upward (downward) SHF, in an intermittent rather than continuous manner, mostly on the equatorward (poleward) flank of the SST front. In February of year 1, in contrast, the weaker SST gradient associated with the premature oceanic front gives rise to more gradual reversal in the sign of SHF skewness than in the later years.

c. ERA-40 data

As shown in section 3b, SST fluctuates much less than SAT on time scales of months, and the SST–SAT difference, with which surface turbulent heat fluxes are correlated positively (Figs. 2 and 4), varies primarily with fluctuating SAT. The steadiness of SST in oceanic frontal zones means that local air–sea heat exchanges may be reproduced qualitatively in atmospheric models or reanalysis data where frontal signatures are resolved to a certain degree in the SST field prescribed. As an example, snapshots on 18 July 2002 based on the ERA-40 data are presented in Fig. 10. In association with a low pressure system (Fig. 10a), the SST–SAT difference is strongly positive on the warmer side of the SST front (Fig. 10b). Compared with the CFES simulation, however, its meridional enhancement is less pronounced because of weaker SST gradient given to the ERA-40 reanalysis (Fig. 10b; meridional SST gradient is generally less than 2 K per latitude). In fact, the corresponding AMSR-E data measured in July 2002 (Fig. 10d) indicate that SST gradient in the frontal zone is as tight (with maximum exceeding 5 K per latitude) as in the CFES simulation and much stronger than in the ERA-40 data.

One may expect that the SST–SAT difference would be stronger if the SST field used for the reanalysis were higher in spatial resolution. Indeed, as indicated with shading in Fig. 10d, the corresponding field of the SST–SAT difference obtained hypothetically by replacing the
SST field used in the ERA-40 reanalysis with the corresponding AMSR-E data can recover the localized enhancement of the SST–SAT difference on both flanks of the frontal zone, especially along the 55°E meridian (Fig. 10c). As shown in Fig. 11, latitude–time sections of SHF and LHF along the meridian for the period from July to August 2002 based on the ERA-40 data can also recover the characteristic features of SHF and LHF found in the CFES integration with respect to their bi-monthly means and standard deviations (Fig. 2), but again their local enhancement and meridional contrast across the SST frontal zone are less pronounced than in the CFES simulation. These results demonstrate that the impact of the SST front on distributions of surface heat fluxes and their variances, which are crucial for maintaining meridional SAT gradient via the restoration process, is found both in CFES and in the atmospheric reanalysis fields of ERA-40. In fact, time-mean meridional SAT profile in the ERA-40 data exhibits a frontal structure over the SST front (Fig. 11b).

5. Summary and conclusions

In the present study, an integration of a high-resolution CGCM is analyzed to examine an impact of a midlatitude SST front on the spatial distribution and temporal variability of turbulent surface heat fluxes (i.e., SHF and LHF) and SAT. For this purpose, the CGCM must be able to resolve not only atmospheric synoptic-scale disturbances but also frontal SST gradients realistically. This requirement is met by CFES, whose atmospheric and oceanic components have (spectral) horizontal resolution of T239 and 0.25°, respectively. The south Indian Ocean is highlighted, where a prominent SST front forms away from landmass. This situation is ideal for extracting the impact of an oceanic frontal zone without any noticeable
FIG. 11. As in Fig. 2, but for July through August in 2002 on the basis of ERA-40.
influence of land–sea thermal contrasts that may mask the impact.

We have shown in snapshots and latitude–time sections based on a CFES integration that cold-air advection behind a cyclone across the SST frontal zone augments the SST–SAT difference mainly on the warmer flank of the front to induce local enhancement of upward SHF and LHF at the sea surface. Likewise, warm air advected ahead of a cyclone suppresses surface evaporation and yields downward SHF on the cooler flank of the frontal zone. Indeed, several intermittent events of cold (warm) air advection, whose total duration accounts for only 21% (19%) of the entire analysis period, contribute to as much as 60% (44%) of the total amount of SHF during the analysis period on the warmer (cooler) flank. We argue that the concentration of the time-mean and temporal variance of SHF and LHF onto the frontal zone is attributable to the tight SST gradient that acts to maintain the SAT gradient. Induced by atmospheric disturbances, both SHF and LHF are highly variable, and the antisymmetric behavior of SHF about the SST front is manifested statistically as its positive and negative skewness on the warmer and cooler flanks of the front, respectively. With this contrasting fluctuations, SHF after averaged in time is still upward and downward on the respective flanks of the SST front, acting to maintain the mean SAT gradient and thereby anchor the surface baroclinic zone along the front and the associated storm track.

We have shown that strong upward (downward) SHF induced on the warmer (cooler) flank of the SST front by cold (warm) air advection due to atmospheric disturbances can restore SAT toward the underlying SST within 1 day. Although impacts of the oceanic frontal zone were not included, Swanson and Pierrehumbert (1997) first suggested the importance of restoration of air temperature to SST, and their rough estimation of it strength is consistent with the result of the present study. With this effective restoration, 925-hPa temperature averaged over July through August of year 1 is characterized by a baroclinic zone with strong meridional gradient across the oceanic frontal zone over the south Indian Ocean (Fig. 12b). As actually observed (Nakamura and Shimpo 2004), the near-surface baroclinic zone that forms along the oceanic frontal zone acts to anchor a low-level storm track. In fact, high-pass-filtered fluctuations in surface meridional wind velocity (Fig. 12a) and 850-hPa poleward heat flux both associated with subweekly disturbances (Fig. 12c) simulated in CFES are strongest along the oceanic frontal zone. Moisture supply from the ocean that occurs predominantly on the warmer flank of the frontal zone also contributes to the development of cyclones along it. In the model upper troposphere (250-hPa level), the storm track marked as variance maxima of high-pass-filtered meridional wind velocity forms along the oceanic frontal zone with its core region shifted slightly downstream of the core of the SST front, as actually observed (Nakamura and Shimpo 2004). A well-defined PFJ is simulated slightly poleward of the oceanic frontal zone, again consistent with the observations. As argued by Nakamura et al. (2004), however, the actual position of a storm track and its activity depend not only on “oceanic baroclinic adjustment” and moisture supply from the ocean but also on atmospheric processes including the intensity of a subtropical jet.

The present study suggests that maintenance of SAT frontal structure by the effective restoration via SHF can be a key process through which a storm track is anchored along an oceanic frontal zone whose structure and variations are determined primarily by oceanic processes (Nonaka et al. 2006, 2008; Taguchi et al. 2007; Qiu 2000; Xie et al. 2000; Schneider and Miller 2001; Seager et al. 2001; Tomita et al. 2002; Kelly and Dong 2004). The present study highlights the specific restoration process, what may be called “oceanic baroclinic adjustment” (Nakamura et al. 2008; Taguchi et al. 2009). The importance of the tight SST gradient in the frontal zone for the tight SHF gradient and thereby for anchoring a storm track implies that resolving SST frontal zones is necessary to simulate storm tracks realistically in AGCMs or CGCMs. The particular importance has been addressed in a suite of AGCM experiments by Nakamura et al. (2008). It may be interesting to examine how differently the storm tracks and associated air–sea heat exchanges are simulated between such a high-resolution CGCM as CFES and its counterpart where the resolution of its ocean component is purposely lowered.

It has been established that extratropical atmospheric variability arises mainly from internal dynamics. Although the particular CFES simulation used in the present study is too short to investigate climate variability, our results imply that oceanic variations might also influence the large-scale atmospheric circulation through changes in SST fronts. The atmospheric circulation anomalies forced by the oceanic influence could drive such oceanic variations as to induce changes in the SST fronts, forming an extratropical air–sea coupling system. It is interesting to investigate whether interannual and longer-term variations in oceanic frontal zones can exert any significant modulations in the nearby

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3 The high variance in surface winds can be partly due to enhanced vertical mixing over warm SST with reduced vertical stability.
storm tracks and air–sea interactions. For this purpose, a longer integration of CFES is now under way, which will be utilized in our future study.

Although the present study has focused on the south Indian Ocean, there are other major oceanic frontal zones in the western portions of the northern Pacific and Atlantic basins, along which major storm tracks are collocated (Nakamura et al. 2004). It is interesting to investigate whether similar relationship can be found between the SST front and the surface heat flux distribution in each of those regions where air–sea interaction has been believed to exert important influence on (inter-) decadal variability over the basin. The corresponding relationship in the frontal zone over the western North Pacific as revealed in regional atmospheric model simulations is discussed by Taguchi et al. (2009). Since 2007, surface buoy monitoring of SST and surface meteorological variables has been conducted to the north and south of the Kuroshio Extension front jointly by the National Ocean and Atmosphere Administration, Pacific Marine Environmental Laboratory (Cronin et al. 2008; Bond and Cronin 2008), and the Japan Agency for Marine-Earth Science and Technology (JAMSTEC). This monitoring and future extension of the buoy array system in that region, if realized, will contribute to our deeper understanding of midlatitude air–sea interactions in which highly variable surface heat fluxes are involved.

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FIG. 12. Horizontal distributions (shaded with conventions on the right of the individual panels) based on the CFES integration for July through August in year 1. (a) Standard deviation of 8-day high-pass-filtered surface meridional wind velocity (m s$^{-1}$), (b) time-mean meridional gradient of 925-hPa potential temperature [0.1°C (100 km)$^{-1}$]. (c) 850-hPa poleward flux of potential temperature (m s$^{-1}$ K) associated with 8-day high-pass-filtered fluctuations, and (d) standard deviation of 8-day high-pass-filtered 250-hPa meridional wind velocity (m s$^{-1}$). In each panel, time-mean SST (°C) distribution is superimposed with contour lines (every 2°C; heavy lines for every 10°C). In (d) time-mean 250-hPa zonal wind velocity is also superimposed with smoothed contour lines (every 5 m s$^{-1}$; heavy lines for every 10 m s$^{-1}$). In (b), shading is omitted where the 925-hPa surface is placed within the surface ML.
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APPENDIX

A simple model for the heat balance in the atmospheric ML

To understand the phase relationship between the advective forcing and SHF, we consider a simple model for the heat balance in the atmospheric ML with the restoration and the advective effects. With the advective effect added to (4), the heat balance can be expressed in a more realistic manner:

$$\frac{\partial (\text{SAT} - \text{SST})}{\partial t} \approx -\tau_R^{-1}(\text{SAT} - \text{SST}) + F,$$

(A1)

where $F$ signifies the advective effect that may be approximated as $-v_{10} \frac{\partial \text{SAT}}{\partial y}$, where the overbar denotes time averaging. In recognition of the thermal advection is associated with migratory atmospheric eddies, we assume $F$ to be periodic with frequency of $\omega$. Then, (A1) may be rewritten as

$$\frac{\partial (\text{SAT} - \text{SST})}{\partial t} \approx -\tau_R^{-1}(\text{SAT} - \text{SST}) + \Delta F \sin \omega t,$$

(A2)

where $\Delta F$ represents the amplitude of $F$. Then, as temporal variations in SST tend to be much smaller than those in SAT (see section 3b), (A2) can be solved for SAT anomaly ($\text{SAT}'$):

$$\text{SAT}' \approx \frac{\Delta F}{\omega} A \sin(\omega t - \theta),$$

(A3)

with $A = [1 + (\omega \tau_R)^{-2}]^{-1/2}$ and $\theta = \tan^{-1}(\omega \tau_R)$. In the lack of the restoration effect by SHF (i.e., $\tau_R \to \infty$), SAT anomaly is merely forced by the periodic advection. Thus SAT also fluctuates periodically with the amplitude of $\Delta F/\omega (A \to 1)$ and the phase lag of $90^\circ$. By contrast, under the extremely effective restoration ($\omega \tau_R \ll 1$), SAT anomaly is approximated as

$$\text{SAT}' \approx \tau_R \Delta F \sin[\omega(t - \tau_R)].$$

(A4)

Thus, the anomaly is strongly damped and fluctuating with small phase lag behind the advective forcing.

In our case, SHF exerts strong restoration effect on SAT, corresponding to $\tau_R \approx 0.8$ day as discussed above. For the bimonthly period from July to August, an autocorrelation of $v_{10}$ indicates the period of the advection $(2\pi/\omega)$ is about 2.5 days. From (A3) the phase lag $\theta$ can then be estimated as $62.0^\circ$ or 0.4 day, reflecting the strong restoration effect of SHF. Though under a crude approximation, the phase lag thus estimated is roughly consistent with the phase relationship observed in Fig. 5a among $v_{10}$, SHF, and SAT.

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