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ATMOSPHERIC RESPONSE TO TROPICAL ATLANTIC SST ANOMALIES: ROLE OF PLANETARY BOUNDARY LAYER ADJUSTMENT

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1. INTRODUCTION

Decadal rainfall variability in Sahel and northeast Brazil is well correlated with a sea surface temperature (SST) pattern called tropical Atlantic dipole, which displays large cross-equatorial SST gradient (Folland et al., 1986; Carton et al., 1996; Mehta, 1998). While simple/intermediate coupled models suggest this dipole as arising from a positive ocean-atmosphere feedback among wind speed, evaporation at the surface and SST (WES; Chang et al., 1997; Xie and Tanimoto, 1998), more sophisticated coupled general circulation models (GCMs) give mixed results (GrÖtzner et al., 1998; Watanabe et al. 1999).

This disagreement among coupled GCMs may be attributed to the different response characteristics of their atmospheric component to the SST dipole. Several atmospheric GCM studies attempt to address how the atmosphere responds to a change in the cross-equatorial SST gradient (Chang et al., 2000; Sutton et al., 2000; Dommenget and Latif, 2000; Okumura et al., 2001). However, they disagree particularly on the magnitude and meridional extent of their trade wind response off the ITCZ, response characteristics key to the WES feedback.

The discrepancy of atmospheric GCM responses might be partially due to the difficulty in separating the "response" from the noisy atmosphere, but it may also arise from the different model physics, especially the representation of the planetary boundary layer (PBL). In response to a dipole pattern of SST anomalies as shown in Fig. 1, the pressure anomalies originating from the PBL act to produce westerly (easterly) components over the warm (cold) SST anomalies and reduce (enhance) sea surface evaporation, in support of the WES feedback. In contrast to the response in baroclinic models (e.g., Lindzen and Nigam 1987), the centers of sea level pressure (SLP) anomalies are, however, shifted poleward of SST anomaly extrema, as part of equivalent barotropic response in the mid-latitudes excited by the anomalous meridional displacement of the ITCZ (Okumura et al. 2001). As a result, the wind anomalies have the same sign over the whole lobe of SST anomalies (Fig. 1a).

In addition to the SLP driving for surface winds that is emphasized in large-scale ocean-atmosphere interaction, Wallace et al. (1989) showed the importance of the vertical mixing of momentum in a sheared PBL in their study of seasonal/interannual variability in the eastern equatorial Pacific. This vertical mixing adjustment is indeed dominant mechanism for surface wind responses in monthly ocean-atmosphere interaction induced by tropical instability waves in the equatorial front (Hashizume et al., 2002). By this mechanism, SST-induced stability changes act to accelerate (decelerate) the surface winds over the warm (cold) SST anomalies, and produce easterly (westerly) components in the presence of easterly shear (Fig. 1b). Thus, the vertical mixing and the resultant adjustment tend to weaken the SLP-driven wind anomalies over the tropical Atlantic that support WES feedback. If the vertical mixing adjustment is too strong in a given model, the wind response may be much reduced at the surface. Here we use an atmospheric GCM to investigate how the PBL physics affects the model adjustment to SST changes and its implications for air-sea interaction over the tropical Atlantic.

2. GCM EXPERIMENTS

The atmospheric model used in our study is CCSR/NIES AGCM 5.6, which is widely used as a Japanese community model (Numaguti, 1999). We use a T42 version with 23 sigma levels in the vertical. For a better representation of the PBL, about 6 levels are placed in the lowest 1km. The vertical diffusion scheme is based on Mellor and Yamada (1974, 1982)'s turbulence closure, level-2 parameterization. In this scheme that is similar to the "K theory", the vertical diffusion coefficient K_m is expressed as a function of the local Richardson number as well as the vertical wind shear. Thus, buoyancy-induced turbulence implicitly alters the vertical

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momentum flux through the dependency of K_m on the local stability.

First, we carry out a set of three simulations with the vertical diffusion coefficient dependent on the stability (interactive K_m runs). In the control run, monthly climatological SSTs are prescribed globally. Then, a dipole pattern of SST anomalies is added in the tropical Atlantic, with northward or southward gradient. As in Okumura et al. (2001), SST anomalies are zonally uniform and anti-symmetric about the equator with extrema of +/-1.0K at 15N/S. Positive and negative dipole runs are made, with positive SST pole in the northern and southern hemisphere, respectively.

In the second set of experiments, we globally prescribe climatological diffusion coefficients (constant K_m runs) that are calculated based on the control run of the first set. Since the stability in the lower troposphere changes more strongly following a diurnal cycle than a seasonal cycle, monthly mean 4-hourly diurnal cycle data has been used and linearly interpolated. Again, we carry out three simulations: control, positive/negative dipole runs. In both interactive and constant K_m experiments, each run is integrated for 10 years (60 years in total).

3. RESULTS

3.1 Surface wind response

The Atlantic Intertropical Convergence Zone (ITCZ) is anomalously displaced toward the anomalously warm hemisphere in response to a change in meridional SST gradient (not shown). The surface wind response is characterized by weakened (enhanced) trade winds over the warm (cold) lobe of SST anomalies (Figs. 2abc). A northward SST gradient induces cross-equatorial southerly wind. The Coriolis force acting upon these northerlies induces westerly (easterly) component north (south) of the equator. These westerly (easterly) winds, superimposed on the mean easterly trades, reduce (enhance) evaporation at the sea surface and thereby reinforce the initial SST dipole, consistent with the WES feedback. The model captures this anomalous wind pattern reasonably well (Figs. 2bc) although the WES-favoring anomalous winds have smaller meridional extent in the northern hemisphere compared to observations (Fig. 2a). These observed wind anomalies poleward of the SST maximum cannot be explained by linear baroclinic models (Lindzen and Nigam, 1987) but are instead part of an equivalent barotropic response to the SST dipole (Okumura et al. 2001). The reduction of the subtropical wind response in the model is presumably due to the short duration of integration (10 years) and the seasonality of the barotropic response.

Figs. 2def show the vertical difference of the magnitude of zonal wind anomalies between the 2 lowest pressure levels. In general, zonal wind anomalies decrease with height due to surface drag. In the composite map based on NCEP2 reanalysis data (Fig. 2d), the reduction occurs where anomalous winds have larger amplitudes. On the other hand, the result from interactive K_m runs (Fig. 2e) displays large localized wind-speed reductions in the eastern subtropics, away from the centers of zonal wind anomalies but roughly coinciding with large SST anomalies imposed. This indicates that these reductions result from the vertical wind shear changes in response to the anomalous stability at the surface (Fig. 3). Over the warm lobe of the SST dipole, weakened stability near the surface enhances the vertical mixing, brings faster-moving air from aloft and thereby accelerates the easterly trades at the surface. The resultant easterly anomalies act to reduce the SLP-driven westerly anomalies at the surface.

In the interactive K_m runs, a change in air-sea temperature difference at the surface is as small as +/-0.1K, about 10% of SST anomalies (Fig. 3b). Thus, vertical mixing adjustment seems a little too strong in the CCSR/NIES AGCM. The eastern subtropical Atlantic, where the vertical mixing adjustment is strong in the model, corresponds to the region with climatologically strong stability (shear) over the cold SST, implying the nonlinearity of the vertical mixing process in the model. By holding K_m independent of SST anomalies, wind anomalies indeed become larger at the surface by a factor of 1.5-2.0 (Figs. 2cf). Fig. 4 shows the vertical profiles of zonal winds averaged over the area of large vertical shear changes. In interactive K_m runs, the maxima of wind anomalies exist at 950hPa and decrease with height by small changes of mean vertical shear (Fig. 4a), whereas they have almost the same magnitudes below 925hPa in constant K_m runs (Fig. 4b). It should also be noted that in Figs. 2ad, the wind anomalies decrease with height more rapidly if NCEP reanalysis data is used instead of NCEP2. With the improvement of model physics, NCEP2 generally provides better representations of the surface wind fields than NCEP (R. Wu, personal communication). These results suggest that the surface wind response may be sensitive to the treatment of the PBL and the vertical shear adjustment in a given model.

3.2 Vertical mixing effect on climatology

Not only the wind response to SST anomalies but also the climatology change by prescribing the vertical diffusion coefficient (Fig. 4). In constant K_m runs, the vertical wind shear is much weaker and the surface wind is stronger by 1.5m/s than in interactive K_m runs. Also, easterly trades reach their maxima at the different heights.

The change in vertical shear may be explained as a result of considerable momentum flux by the vertical mixing adjustments. The mean zonal momentum equation in the PBL can be written approximately as:

$$-fv = -\frac{1}{\rho_0}\frac{\partial P}{\partial x} + \frac{\partial}{\partial z}\left(K_m \frac{\partial u}{\partial z}\right) \tag{1}$$

where all the symbols have their conventional meanings and we assume the density is constant. Separating each variable into mean and fluctuation parts (denoted by overbars and primes, respectively) and averaging over a long period (~year), we can rewrite the mean momentum equations for interactive K_m runs as:

$$-f\overline{v} = -\frac{1}{\rho_0}\frac{\partial \overline{P}}{\partial x} + \frac{\partial}{\partial z}\left(\overline{K}_m \frac{\partial \overline{u}}{\partial z} + \overline{K'_m \frac{\partial u'}{\partial z}}\right) \tag{2}$$

and for constant K_m runs as:

$$-f\overline{v}_* = -\frac{1}{\rho_0}\frac{\partial \overline{P}_*}{\partial x} + \frac{\partial}{\partial z}\left(\overline{K}_m \frac{\partial \overline{u}_*}{\partial z}\right) \tag{3}$$

By subtracting the equation (2) from (3) and integrating it with respect to z, we obtain

$$\frac{\partial \overline{u}_*}{\partial z} - \frac{\partial \overline{u}}{\partial z} = \frac{1}{\overline{K}_m}\left[\begin{array}{l} \overline{K'_m \frac{\partial u'}{\partial z}} \\ + \int \left\{ f\left(\overline{v} - \overline{v}_*\right) - \frac{1}{\rho_0}\frac{\partial}{\partial x}\left(\overline{P} - \overline{P}_*\right)\right\} dz \end{array}\right] \tag{4}$$

The budget analysis of the equation (4) indicates that the integral term on the right hand side is nearly negligible and that a change in the mean vertical shear can be attributed mostly to the covariance of vertical shear and diffusion coefficient in interactive K_m runs. This is a reasonable approximation especially in the area of large mean vertical shear. If the atmosphere becomes anomalously unstable, the enhanced vertical mixing (K_m'>0) reduces the vertical shear of easterly trades ($\partial u'/\partial z$ >0), yielding a positive correlation between them. The decreased vertical shear in turn tends to reduce the vertical mixing (K_m'<0). The former thermal effect seems to be more dominant in the model and the covariance term in the equation (4) tends to be positive (Fig. 5), thereby weakens the easterly shear in the constant K_m runs.

4. DISCUSSION

In summary, our GCM experiments suggest that the disagreement of atmospheric GCM responses to a SST dipole may result from the differences in their treatment of the PBL and vertical shear adjustment. In a given model, the relative importance of the SLP and vertical mixing mechanisms determine the sign and magnitude of surface wind response to SST changes.

In nature, however, their relative importance would depend on the local structure of the PBL and the scale of SST anomalies. When SST fluctuations have a small horizontal scale, the atmosphere adjusts to them with a larger-scale horizontal structure. This results in large changes in air-sea temperature difference and hence the intensity of vertical mixing. A change in SST may affect the full depth of the PBL and cause the capping inversion to move anomalously in the vertical. For example, a surface warming may raise the inversion and leave a strong cooling near the mean inversion height, forming a temperature dipole in the vertical. The hydrostatic effects of this temperature dipole act to reduce the pressure signal at the sea surface (Hashizume et al., 2002).

In our GCM, the vertical mixing adjustment seems to be strong and weaken the surface wind anomalies in the eastern subtropical Atlantic. It calls for revalidating parameters in PBL schemes, which are generally derived from laboratory experiments and not necessarily suitable for large-scale adjustment processes. Equally importantly, we need to improve the representation of PBL physics in GCMs, including direct interactions between the clouds, radiation and turbulence within the stratocumulus-topped PBL (Randall, 2000). In view of its crucial roles in the climate system, the parameterization of PBL processes still remains a weakness of the existing GCMs and requires further improvements.

References

Carton, J. A., X. Cao, B. S. Giese and A. M. da Silva, 1996: Decadal and interannual SST variability in the tropical Atlantic Ocean. *J. Phys. Oceanogr.*, **26**, 1165-1175.

Chang, P., L. Ji and H. Li, 1997: A decadal climate variation in the tropical Atlantic ocean from thermodynamic air-sea interactions. *Nature*, **385,** 516-518.

Chang, P., R. Saravanan, L. Ji and G. C. Hegerl, 2000: The effects of local sea surface temperatures on atmospheric circulation over the tropical Atlantic sector. *J. Climate*, **13**, 2195-2216.

Dommenget, D. and M. Latif, 2000: Interannual to decadal variability in the tropical Atlantic. *J. Climate*, **13**, 777-792.

Folland, C., T. Palmer and T. Parker, 1986: Sahel rainfall and worldwide sea surface temperatures: 1901-85, *Nature*, **320**, 602-606.

GrÖtzner, A., M. Latif and T. P. Barnett, 1998: A decadal climate cycle in the North Atlantic Ocean as simulated by the ECHO coupled GCM. *J. Climate*, **11**, 831-847.

Hashizume, H., S.-P. Xie, M. Fujiwara, M. Shiotani, T. Watanabe, Y. Tanimoto, W. T. Liu and K. Takeuchi, 2002: Direct observations of inversion-capped boundary layer response to slow variations in sea surface temperature on the Pacific equatorial front. *J. Climate.*, in press.

Lindzen, R. S. and S. Nigam, 1987: On the role of sea surface temperature gradients in forcing low level winds and convergence in the tropics. *J. Atmos. Sci.*, **44**, 2418-2436.

Mehta, V. M., 1998: Variability of the tropical ocean surface temperature at decadal-multidecadal time scales. Part I: The Atlantic Ocean. *J. Climate*, **11**, 2351-2375.

Mellor, G. L. and T. Yamada, 1974: A hierarchy of turbulence closure models for planetary boundary layers. *J. Atmos. Sci.*, **31**, 1791-1806.

Mellor, G. L. and T. Yamada, 1982: Development of a turbulence closure model for geophysical fluid problems. *Rev. Geophys. Atmos. Phys.*, **20**, 851-875.

Numaguti, A., 1999: Origin and recycling processes of precipitating water over the Eurasian continent: Experiments using an atmospheric general circulation model. *J. Geophys. Res.*, **104**, 1957-1972.

Okumura, Y., S.-P. Xie, A. Numaguti and Y. Tanimoto, 2001: Tropical Atlantic air-sea interaction and its influence on the NAO. *Geophys. Res. Lett.*, **28**, 1507-1510.

Randall, D. A., 2000: *General Circulation Model Development*. Academic Press, 807pp.

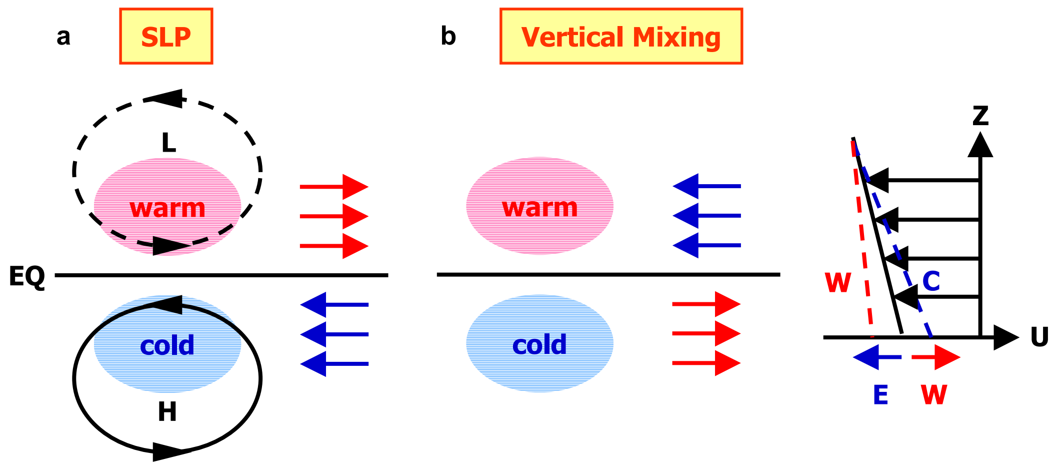
Sutton, R. T., S. P. Jewson and D. P. Rowell, 2000: The elements of climate variability in the tropical Atlantic region. *J. Climate*, **13**, 3261-3284.

Wallace, J. M., T. P. Mitchell and C. Deser, 1989: The influence of sea surface temperature on surface wind in the eastern equatorial Pacific: Seasonal and interannual variability. *J. Climate*, **2**, 1492-1499.

Watanabe, M., M. Kimoto and T. Nitta, 1999: A comparison of decadal climate oscillations in the North Atlantic detected in observations and a coupled GCM. *J. Climate*, **12**, 2920-2940.

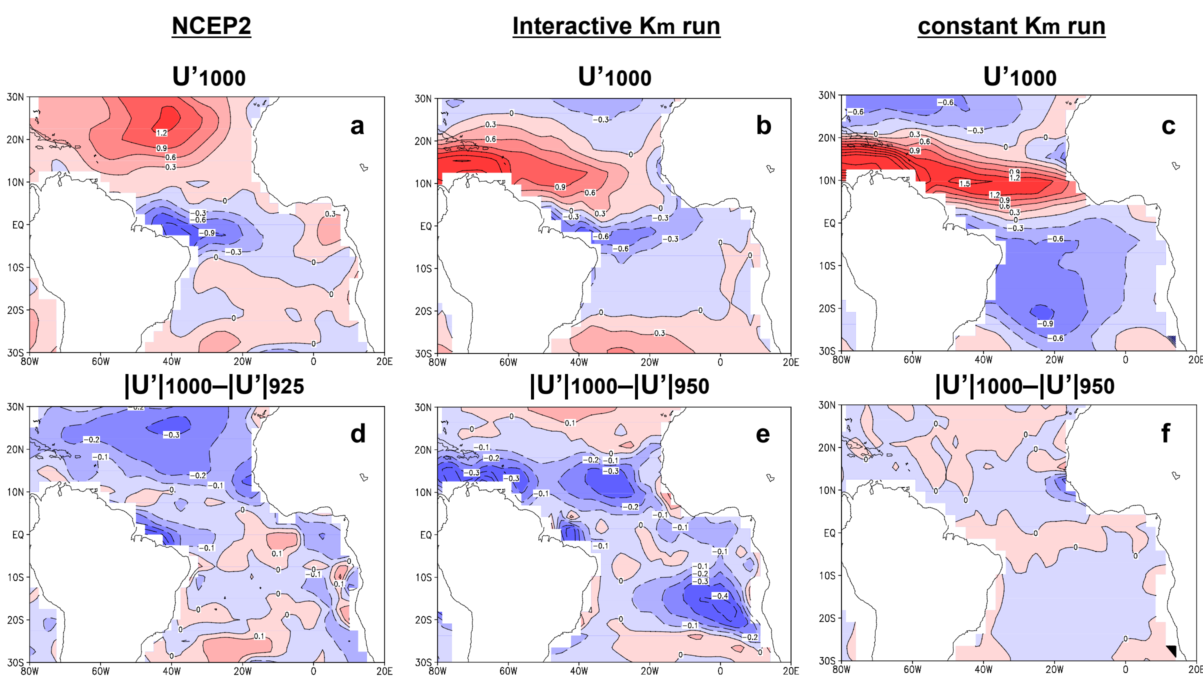
Xie, S.-P. and Y. Tanimoto, 1998: A pan-Atlantic decadal climate oscillation. *Geophys. Res. Lett.*, **25**, 2185-2188.

Fig. 1



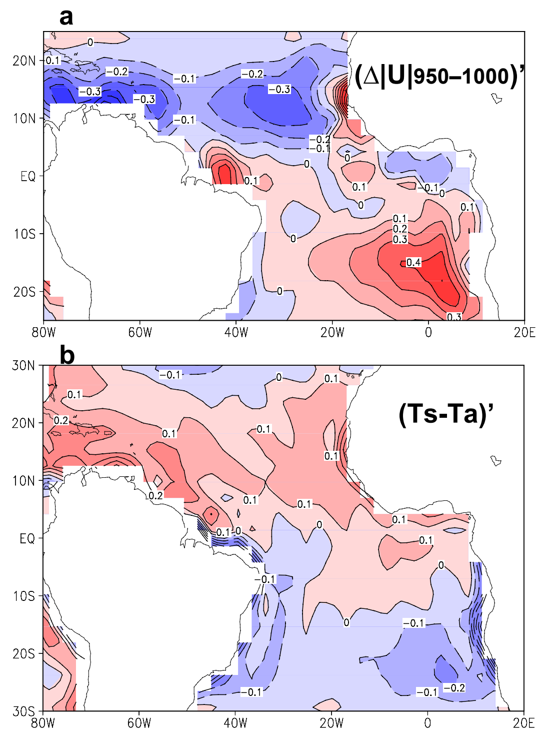
Schematic of surface wind response to SST dipole by (a) SLP forcing and (b) vertical mixing adjustment. Shading represents the SST, black contours SLP and arrows zonal wind anomalies.

Fig. 2



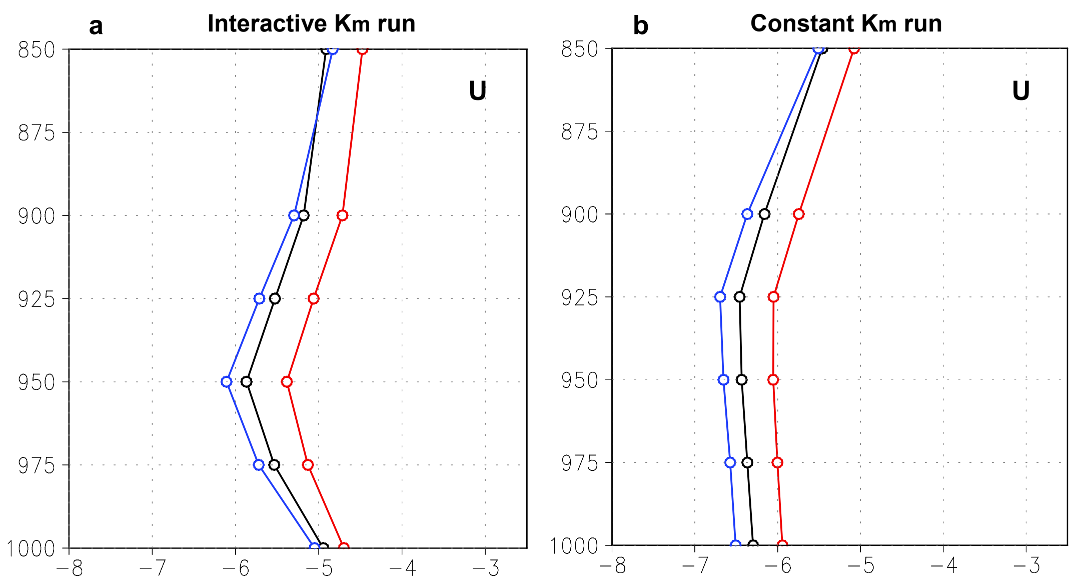
Zonal wind response (ms^{-1}) to a positive SST dipole (the upper panels) and vertical difference of their magnitudes (ms^{-1}) between the 2 lowest pressure levels (the lower panels; 1000-925hPa in (d) and 1000-950hPa in (e)(f)) of the observations (NCEP2) (a)(d), interactive (b)(e) and constant (c)(f) K_m runs. The observed anomalies are defined as the difference between six positive years (1979, 80, 81, 83, 92, 97) and six negative years (1984, 85, 86, 88, 89, 94).

Fig. 3



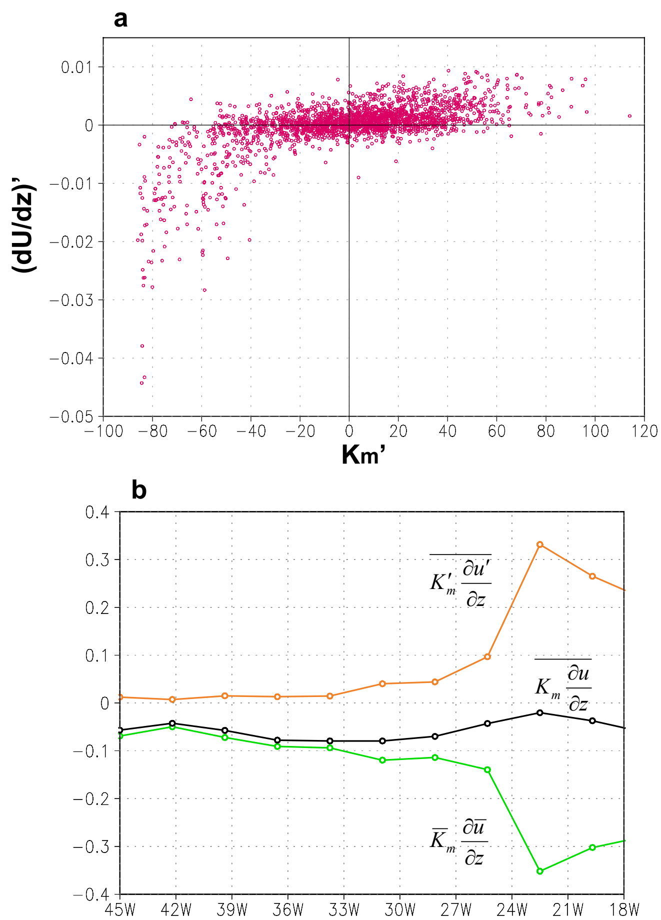
Response to a positive SST dipole in the interactive K_m runs: (a) vertical difference of zonal wind speeds (ms^{-1}; 950-1000hPa) and (b) ocean-atmosphere temperature difference (K).

Fig. 4



Vertical profiles of zonal wind (ms^{-1}) averaged over 10-20N, 20-40W in interactive (a) and constant (b) K_m runs. Black lines correspond to profiles for control runs and red (blue) lines for positive (negative) dipole runs.

Fig. 4



(a) Scatter plots of vertical diffusion coefficient (m^2s^{-1}) and vertical wind shear (s^{-1}) at 20N, 25W, sigma level 0.99. Seasonal cycle has been removed from 4-hourly data and plotted for 1 year. (Therefore, the fluctuations are mainly due to the diurnal cycle and synoptic disturbances.) (b) Annual-mean vertical flux of easterly momentum (upward positive) along 20N (18-45W) (black line) and its components by the mean field (green line) and perturbations (orange line).