



On the Relationship between Synoptic Wintertime Atmospheric Variability and Path Shifts in the Gulf Stream and the Kuroshio Extension

TERRENCE M. JOYCE, YOUNG-OH KWON, AND LISAN YU

WHOI, Woods Hole, Massachusetts

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ABSTRACT

Coherent, large-scale shifts in the paths of the Gulf Stream (GS) and the Kuroshio Extension (KE) occur on interannual to decadal time scales. Attention has usually been drawn to causes for these shifts in the overlying atmosphere, with some built-in delay of up to a few years resulting from propagation of wind-forced variability within the ocean. However, these shifts in the latitudes of separated western boundary currents can cause substantial changes in SST, which may influence the synoptic atmospheric variability with little or no time delay. Various measures of wintertime atmospheric variability in the synoptic band (2–8 days) are examined using a relatively new dataset for air–sea exchange [Objectively Analyzed Air–Sea Fluxes (OAF_{Flux})] and subsurface temperature indices of the Gulf Stream and Kuroshio path that are insulated from direct air–sea exchange, and therefore are preferable to SST. Significant changes are found in the atmospheric variability following changes in the paths of these currents, sometimes in a local fashion such as meridional shifts in measures of local storm tracks, and sometimes in nonlocal, broad regions coincident with and downstream of the oceanic forcing. Differences between the North Pacific (KE) and North Atlantic (GS) may be partly related to the more zonal orientation of the KE and the stronger SST signals of the GS, but could also be due to differences in mean storm-track characteristics over the North Pacific and North Atlantic.

1. Introduction

There must be a dynamical relationship that links atmospheric storminess and sea surface fronts like the Gulf Stream (GS) and Kuroshio Extension (KE), because there is a geographic relationship in the mean between the pattern of upper-tropospheric clouds and subsurface ocean fronts (Fig. 1). A recent study (Minobe et al. 2008) illustrated parts of the link between the mean Gulf Stream position and the mean atmospheric structure from the boundary layer up into the upper troposphere. Some studies (e.g., Hoskins and Hodges 2002) of atmospheric variability associated with synoptic signals have concentrated on the variability itself and not with any underlying oceanic feature that might play a role. Others (e.g., Nakamura et al. 2004) have also addressed the relationship between the mean oceanic SST and the mean wintertime storm track. That there should be some close relationship between ocean fronts like the GS or KE and midlatitude winter storm tracks

does not particularly identify the dynamical cause. Because the system is coupled, there are processes within the atmosphere and the ocean separately that will force an alignment between the two (Hoskins and Valdes 1990, Nakamura et al. 2008): storms can produce vorticity fluxes that enhance the midlatitude zonal jet, which can affect the location of the GS and KE, and then further influence the development of the storms.

Recently Tanimoto et al. (2003) have examined decadal variability in SST and interannual–decadal atmospheric variability for the North Pacific, and Alexander et al. (2006) have looked at large-scale SST and wintertime storm tracks using both model and reanalysis data. These approaches recognize that interannual–decadal time-scale changes in the ocean require one to examine similar time scales for changes in the overlying atmosphere as well. We will follow this line of thinking here, differing in that our focus is interannual–decadal measures of atmospheric synoptic variability and not variations in seasonal means (e.g., Tanimoto et al. 2003, Kelly and Dong 2004). Herein we will not distinguish between storm tracks and baroclinic waveguides (e.g., Wallace, Lim and Blackmon 1988), but we will usually refer to them interchangeably as synoptic variability.

Corresponding author address: Terrence M. Joyce, WHOI, Mail Stop 21, 360 Woods Hole Rd., Woods Hole, MA 02543.
E-mail: tjoyce@whoi.edu

Mean Annual High IR Clouds (ISCCP)

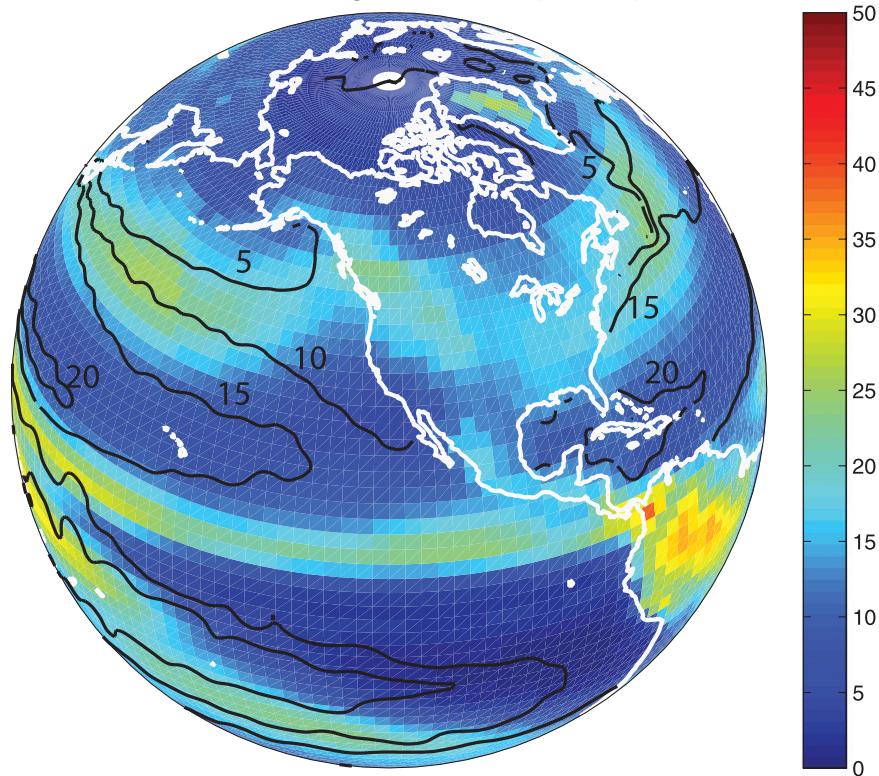


FIG. 1. Climatological (or time mean) distribution of percent cloudiness (colors) in high (IR) clouds from ISCCP cloud climatology (Schiffer and Rossow 1983) from 1983 to 2006. Sub-surface 200-m temperature is contoured every 5°C from 0° to 20°C. Where the 5°–15°C isotherms are tightly grouped demarcates the pathways of the Gulf Stream and the Kuroshio Extension, which underlie the high cloud maxima.

We will not draw any conclusions about the root cause of the oceanic variability. It is clear that the 22-yr record we will be using is insufficient to adequately address the nature of the decadal variability. Rather, we will accept the ocean variability as a given and examine whether synoptic atmospheric variability is responding.

We will make use of two rather disparate datasets not used in previous studies of this type. The first of these is a daily, global ocean–atmospheric flux product: Objectively Analyzed Air–Sea Fluxes (OAFlux; Yu and Weller 2007), which has been augmented (L. Yu, unpublished manuscript) to include surface winds with all variables at sea level (+10 m for winds). These products are a combination of observed satellite-derived fields and 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40) and National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) re-analyses that have been optimally combined using the Coupled Ocean–Atmosphere Response Experiment (COARE) 3.0 bulk flux algorithm. The OAFlux latent

and sensible heat flux estimates are unbiased, and the root-mean-square (rms) difference is less than 8 W m^{-2} when compared with daily flux time series measurements at 107 locations (Yu et al. 2008). The global, ice-free region daily dataset that we will use spans a roughly 22-yr period from 1983 to 2004, although OAFlux monthly datasets are available online from 1958 to 2006. Because we have daily fields, the data are high passed using a filter similar to that used previously (Alexander et al. 2006) to provide variability in a 2–8-day band. We will then examine the rms wintertime variability of different fields from year to year and how they vary in concert with oceanic signals related to, but not identical with, SST front locations.

The second dataset (see also section 3) is the archive of subsurface ocean temperatures that will be used to estimate the spatial mean paths taken by the GS and the KE after they have left their respective coastal boundaries. For the Gulf Stream, the archive of subsurface temperature at 200-m depth is used. The $T(200 \text{ m})$ temperature of 15°C has been routinely used to demark

the north or cold wall of the Gulf Stream for some time (Fuglister 1955), and we have examined smoothed, seasonally resolved locations of this boundary going back to 1955, as discussed in a following section. The leading spatial EOF of this boundary, explaining approximately 50% of the variance, is a north–south shift of the Gulf Stream between 75° and 55°W, a distance of about 1600 km. A similar procedure with annual resolution was used (Joyce et al. 2000) to study the relationship of the Gulf Stream to the North Atlantic Oscillation (NAO) and to subtropical mode water. For the Kuroshio Extension between 142° and 180°E (Qiu et al. 2007), an SST-based index was used to examine coupling between the ocean and atmosphere. They resorted to lagged analyses (SST leading) to extract the influence of the ocean on the atmosphere, because the atmosphere will change SST and affect the SST when the ocean is either in phase with or slightly following the atmosphere (Frankignoul and Sennéchaël 2007). Our Kuroshio Extension index is similar to that derived for the Gulf Stream and is based on the subsurface $T(200\text{ m})$ temperature following the seasonally varying 14°C isotherm, which marks the KE axis (see Kawai 1972). For both the GS and KE, we will choose the index of the wintertime [January–March (JFM)] position.

2. Mean atmospheric fields and variability trends

The mean seasonal fields from the basic OAFlux heat flux data have already been discussed by Yu and Weller (2007) elsewhere; they also included a discussion of trends and monthly variability. The surface wind products are developed by the OAFlux project at Woods Hole Oceanographic Institution (WHOI) under the auspices of the National Aeronautics and Space Administration's (NASA's) Ocean Vector Wind Science Team (OVWST) program. The global wind analysis was developed in parallel with the ocean evaporation, latent and sensible heat fluxes, and radiative fluxes development, and is based on an objective analysis of multiple passive and active microwave sensors [Special Sensor Microwave Imager (SSM/I), NASA Scatterometer (NSCAT), Quick Scatterometer (QuikSCAT), Advanced Microwave Scanning Radiometer for Earth Observing System (EOS; AMSR-E)] and three atmospheric wind reanalyses. The reanalyzed winds provide directional information for passive microwave wind speed retrievals whenever scatterometer measurements are lacking. Validation of the datasets has been made with 100+ in situ buoy time series measurements. The products include the 10-m surface wind (zonal and meridional components and wind speed), wind stress, and wind stress derivative (curl and divergence) fields, and are available on a daily basis from mid-1987 to the present, with two spatial

resolutions (0.25° and 1°). To be consistent with the 1° analysis of latent and sensible heat fluxes, only the 1° resolution winds were used in study. Prior to 1987 daily winds used are based on reanalysis products.

Here we discuss trends in the bandpassed energy, but first show the mean fields for the (relative) vorticity and the divergence, which are two quantities not previously discussed for this dataset. The vorticity (Fig. 2, upper panels) for the North Atlantic and North Pacific regions containing the GS and KE shows a mean structure that is rather different for the two regions. In the North Pacific the strong, positive mean vorticity in the atmospheric boundary layer exists above the subpolar ocean mainly because of decreasing northerly winds (not shown) as one moves eastward into the Aleutian low. This is not as prominent in the North Atlantic. In both regions, the speed of the westerlies increases from north to south, giving rise to a midlatitude vorticity maximum that is generally aligned with the axis of the KE and GS. Small-scale structures in the mean vorticity and divergence (Fig. 3) near the coasts and near strong SST gradients have been discussed elsewhere by various authors (Chelton et al. 2004; Samelson et al. 2006; Song et al. 2006; Sampe and Xie 2007; Spall 2007) and are reflected in our results, which also derive from scatterometer wind measurements, although we have used the OAFlux winds at $1^\circ \times 1^\circ$ resolution.

A clear maximum of mean vorticity develops over the GS downstream of Cape Hatteras and the maximum positive vorticity follows the SST front of the GS much closer than that for the KE. This positive mean vorticity signature has been ascribed as being due to the ocean currents being reflected in the atmosphere boundary layer stress measurements (Chelton et al. 2004), and thus in the neutral 10-m wind product that we are using. We note that the SST front of the GS is more intense and differs from the broad, diffuse meridional frontal zone of the KE, where the Kuroshio flows eastward from Japan near $\sim 35^\circ\text{N}$ and the Oyashio leaves the coast near $\sim 40^\circ\text{N}$. In the North Atlantic, these two frontal zones are compressed into one, where one passes from subpolar to subtropical waters in a meridional distance of 100 km. The GS path also takes an abrupt bend to the north near 50°W as it turns the corner around the Grand Banks. This feature (also clear in Fig. 1) is coincident with atmospheric variables also turning northward, such as latent and sensible heat release (not shown), vorticity, and divergence (Fig. 3), as well as their respective standard deviations in the wintertime synoptic band (lower panels of Figs. 2 and 3). Thus, wintertime variability in the GS is much less zonal than over the KE in the Pacific. The mean divergence is positive to the north of the KE and GS and negative

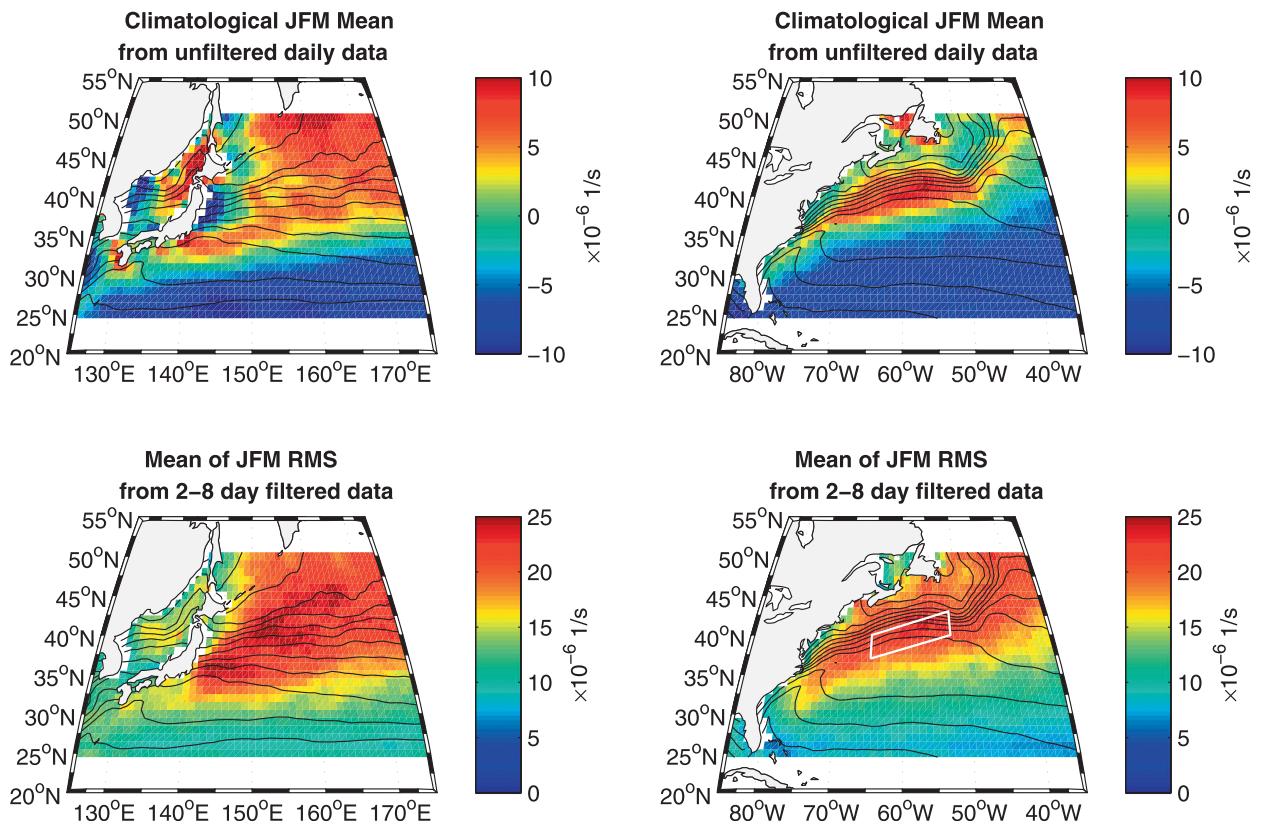


FIG. 2. Time mean (relative) vorticity from the OAFlux surface wind climatology for JFM for regions of the (top left) KE and (top right) GS, with (bottom) the rms synoptic, bandpassed variability for JFM for each region, respectively. The black contours are the climatological JFM mean SST with a 2°C interval. Vorticity units (colors) are 10^{-6} s^{-1} . (bottom right) A box selected for later analysis of variability (Fig. 16) over the storm-track maximum is indicated.

(convergent) to the south. This is consistent with the mean northerly atmospheric flow across the two currents accelerating over the warm water in the cores of these flows (initially resulting from boundary layer thickening) and then decelerating on the southerly sides as the atmosphere warms and adjusts to the warm ocean SST cores (more prevalent in the GS than the KE) disappear. Regions of the large mean sensible heat release to the atmosphere correspond well with those regions of convergent motion, as one would expect if the sensible heat release drove the upward motion of the air in the atmospheric boundary layer (Lindzen and Nigam 1987). Because this air is moist, resulting from the collocation of large latent heat release by the ocean, the convergent flow transports water vapor upward and could, under the right conditions, provide an agent for cloud formation and deeper convection in the atmosphere (Minobe et al. 2008). Overall, the synoptic variability of vorticity and divergence exceeds the mean values, unlike the case for latent and sensible heat release, where synoptic variability is roughly 30%–50% of the mean wintertime values. For these turbulent heat

transport quantities, the variability is clearly strongest where the means are maximum.

One curious finding for the synoptic variability is that there are large trends over the 22-yr record (Fig. 4) that are more significant in the North Pacific than in the North Atlantic. This is consistent with NWP products and Comprehensive Ocean–Atmosphere Data Set (COADS; Chang 2007) and will require that all variables, including the variability measures, need to be detrended. We were initially surprised at the strong relationship between the atmospheric variability in the KE and the oceanic KE index (next section) everywhere in the domain until we detrended the signals. The lower panels of Figs. 2 and 3 are not influenced by the existence of the trends in the JFM mean; the trended and detrended versions of the lower panels of Figs. 2 and 3 are essentially identical. Clearly the reasons for the trends are of interest but not particularly relevant here. These trends and their spatial characteristics will be discussed in a future study. At this point, we will linearly detrend all of the variables to maximize the interannual–decadal signals that are our focus in this study.

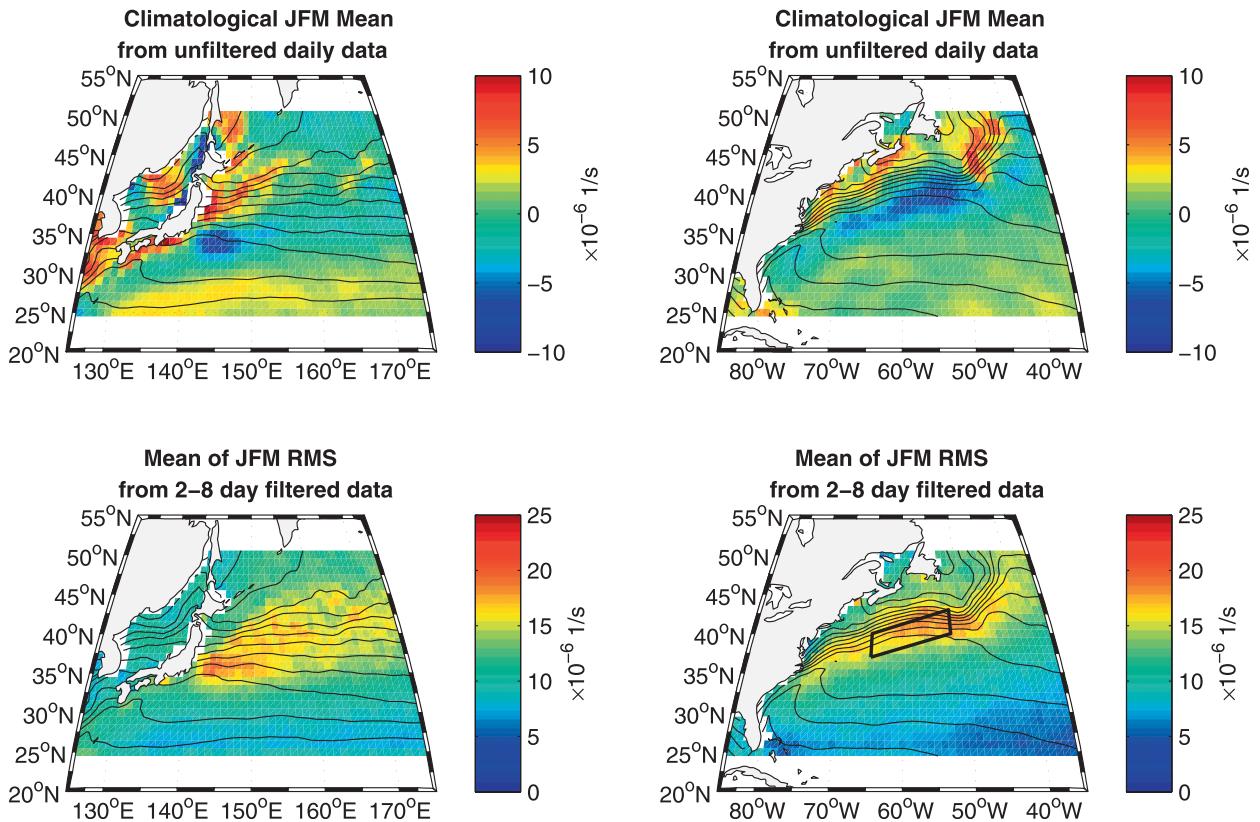


FIG. 3. As in Fig. 2, but for horizontal divergence. Note how the mean fields mirror the vorticity pattern with approximately the same magnitude.

3. Gulf Stream and Kuroshio indices

We have extended the GS path index from earlier work to include seasonally resolved variations. Because the subsurface data for the region are sparse, the data selected [the World Ocean Database 2001 (WOD01), updated using the Global Temperature and Salinity Pilot Program (GTSP) sources] are centered in the midpoint of each season (e.g., JFM for winter) and at each 1° latitude \times 1° longitude, with a time-space range of ± 1 yr, $\pm 1^\circ$ latitude and $\pm 2^\circ$ longitude, with a Gaussian taper having an e^2 -folding time scale of $(1 \text{ yr})^2$, and spatial scales of $(1^\circ \text{ latitude})^2$ and $(2^\circ \text{ longitude})^2$. Thus, the full seasonal signal in the data is substantially smoothed. The gridded temperature data are then used along a path following the mean $T(200 \text{ m}) = 15^\circ\text{C}$, or north wall position, and nine points are selected spanning a longitude range from 75° to 55°W . After an EOF analysis, the leading mode is selected, having temperature increasing or decreasing in time at every point along the mean path. This shift of the GS path is given by the principal component of the leading EOF (Fig. 5), and is normalized to have standard deviation unity. (We have detrended the temperature data before and EOF anal-

ysis and detrended the leading EOF principal component after EOF analysis, with no basic difference in the result.) A similar approach, but using altimeter data along individual Ocean Topography Experiment (TOPEX)/Poseidon or Jason tracks, detecting the largest gradient of the sea surface anomaly (Kelly and Gille, 1990), was recently employed (Peña-Molino and Joyce 2008) to estimate a spatial mean GS path every month over the satellite period starting from late 1992. This index, also normalized, agrees well with the longer, but smoother $T(200 \text{ m})$ estimate. As discussed elsewhere (Frankignoul et al. 2001), these two data sources focus on slightly different portions of the GS: one at the north wall, near the maximum SST gradient, and the second at the center of the strongest surface flow. However, both give reliable information about the changes in the GS path. It is the longer, subsurface GS path index record that will be used subsequently, selecting out the winter period for a more limited 22-yr time window of 1983–2004 corresponding to the OAF flux product. Satellite altimetric data, though providing better temporal resolution and a more uniform spatial sampling, are presently limited to a slightly more than one decade time span, and are too short for the study of decadal

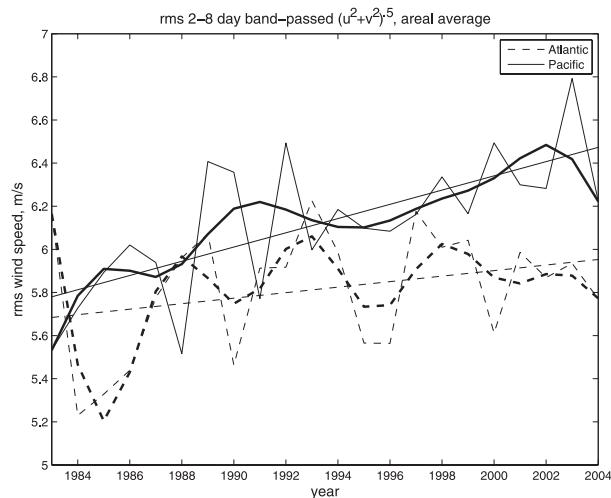


FIG. 4. For the spatial regions shown in Figs. 2 and 3 (24.5° – 50.5° N, 85.5° – 34.5° W or 124.5° – 175.5° E), the winter rms synoptic variability in surface wind speed is plotted for the 22-yr OAF flux record period; KE (solid lines) and GS (dashed lines). Smoothed interannual variability (with a symmetric, low-pass Butterworth filter) is plotted with thicker lines to reduce interannual variability and better show the trends in the North Pacific data. Linear trends using the unfiltered data are also shown.

variability. In fact, the 22 yr for which we have good satellite coverage for the atmospheric variables is arguably too short as well. However, it is the best we can do today.

A similar, but slightly different approach for deriving an oceanic index was taken for the KE. Here, subsur-

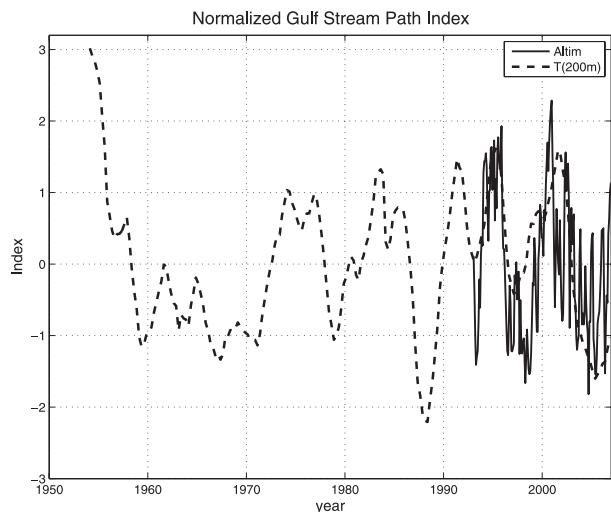


FIG. 5. Normalized time series of the principal component of the GS latitude shift based upon subsurface temperature $T(200\text{ m})$ (dashed line) and satellite altimeter (solid line). Each of these PCs represents the leading EOF mode of seasonal [$T(200\text{ m})$] or monthly (altimeter) variability for the GS path. Only the JFM record for $T(200\text{ m})$ PC was used in our analysis.

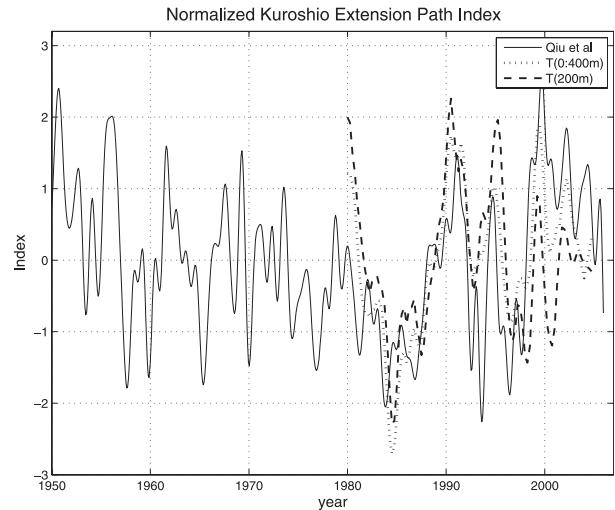


FIG. 6. Normalized time series of KE path changes based upon SST (solid line), $T(200\text{ m})$ (dashed line), and $T(0\text{--}400\text{ m})$ (dotted line) subsurface temperatures, as discussed in text.

face temperature, $T(200\text{ m})$ was again used spanning a longitude range from 142° to 160° E. The KE index is based on the leading PC of the 14°C isotherm at 200 m . The subsurface temperature field was objectively mapped based on the raw profiles from the World Ocean Database 2005 on a seasonal interval and then moving averaged with the 13-month window to mainly retain the annual and longer variability. The data were detrended prior to the calculation of EOFs. The leading EOF of the 14°C isotherm location at 200 m represented a uniform north–south shift of the KE and explained about 70% of the total variance. We have also used the vertically averaged temperature (related to heat content) from the surface to 400-m depth and tracked the maximum meridional gradient in this quantity, again using EOF analysis. Both yield similar time series (Fig. 6) and are closer to one another than to previously published a third index (Qiu et al. 2007), which used only SST data within a box extending from Japan all the way to the date line and covering a latitude range from 32° to 38° N spanning the KE. As we noted earlier, all three were examined, but the SST index did not produce correlations with the atmospheric data that were as robust as the other two. In fact, it was for this reason that we sought some other KE index after initially finding relationships between the synoptic variability measures and the KE index that were less significant than for the GS. This could possibly be due to the wider span of longitudes defining the SST index in which the KE path may occasionally bifurcate (B. Qiu 2008, personal communication). Thus, we have selected the $T(200\text{ m}) = 14^{\circ}\text{C}$ one as being both robust and closer to that previously derived for the GS. The range of path shifts is

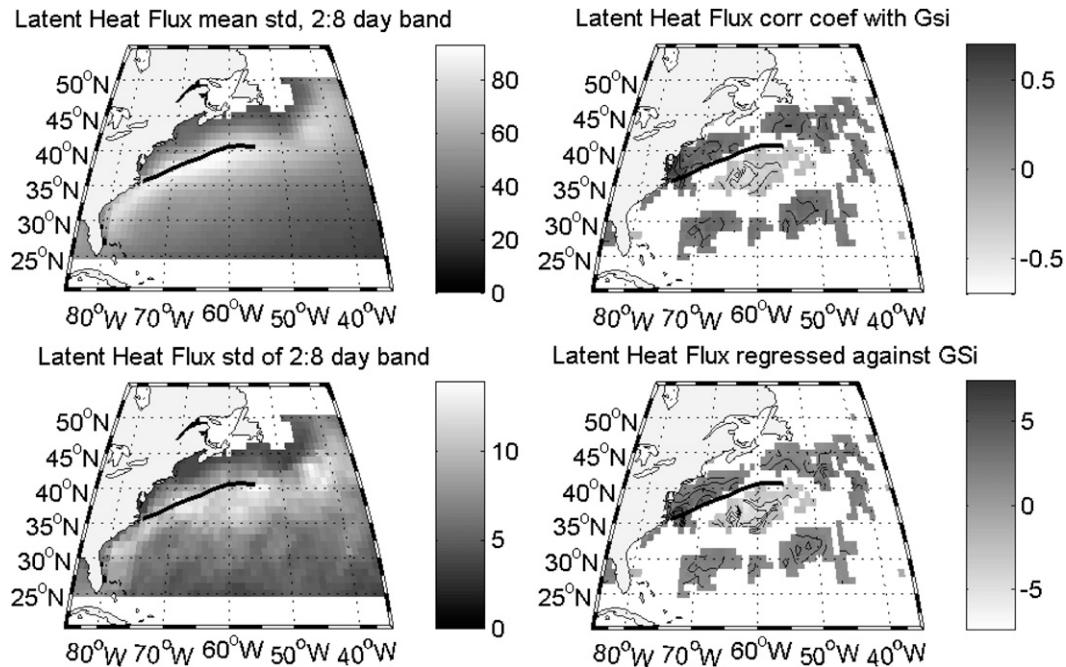


FIG. 7. Latent heat flux for the GS region: (top left) mean rms 2–8-day band for JFM, (bottom left) interannual JFM std of rms 2–8-day band, (top right) correlation coefficient of interannual JFM rms variability with GS index with only significant ($>90\%$) values plotted, and (bottom right) regression against GS index. Darker shades of gray denote regions of positive correlation/regression here and elsewhere, while lighter shades of gray are negative. For all but the correlation coefficient, the units are W m^{-2} . The mean GS path is plotted (solid line). The contour interval for the (top right) correlation coefficient is 0.1 and (bottom right) regression is 0.1 times the grayscale maximum on the plot.

approximately 100 km, which is a small distance meridionally but coherent over an extensive zonal extent.

4. Relation to atmospheric variability

We will present results showing the interannual–decadal relationships between sea level atmospheric variability and the GS and KE indices separately. By stringing together each of the 22 winters, an interannual time series was constructed containing the standard deviation for each winter of the variability of latent and sensible heat release, meridional winds, vorticity, and divergence, as well as winter estimates of the GS and KE indices. These variability time series were then individually compared and regressed against the ocean-based indices. In effect, we are examining the year-to-year winter atmospheric variability and comparing it to the year-to-year index of wintertime ocean signals that are closely tied to SST gradients. We will show all comparisons for the North Atlantic, and then compare and contrast these results to the North Pacific.

a. Gulf Stream

Synoptic variability in the atmosphere can cause widespread outbreaks of cold, dry continental air over

the oceans. During these events there are massive heat losses by the ocean into the atmosphere (CLIMODE Group 2009). Because these turbulent exchanges produce large latent and sensible fluxes across the air–sea interface, they provide energy to cool the oceanic waters that move northward from the subtropics, and after heat release they warm the overlying atmosphere. These processes, which are sensitive to temperature differences between the air and water, are expected to play a role in the air–sea coupling, thus linking atmospheric variability and GS (and KE) path location. There could be other oceanic signals (frontal strength, surface transport) that are important, but here we choose only to examine the path changes, for which a meridional shift would create a large perturbation in the SST field seen by the atmosphere. For latent heat flux variability (Fig. 7), the interannual change in the variability index is roughly 10%–20% of the mean (cf. upper and lower left panels of Fig. 7). The correlation pattern between the variability and GS index indicates a dipole structure on either side of the mean GS, consistent with the atmospheric boundary layer signal shifting northward (or southward) in phase with the GS. We show only correlations that are significant at the 90% significance level, and we also show the regression of the

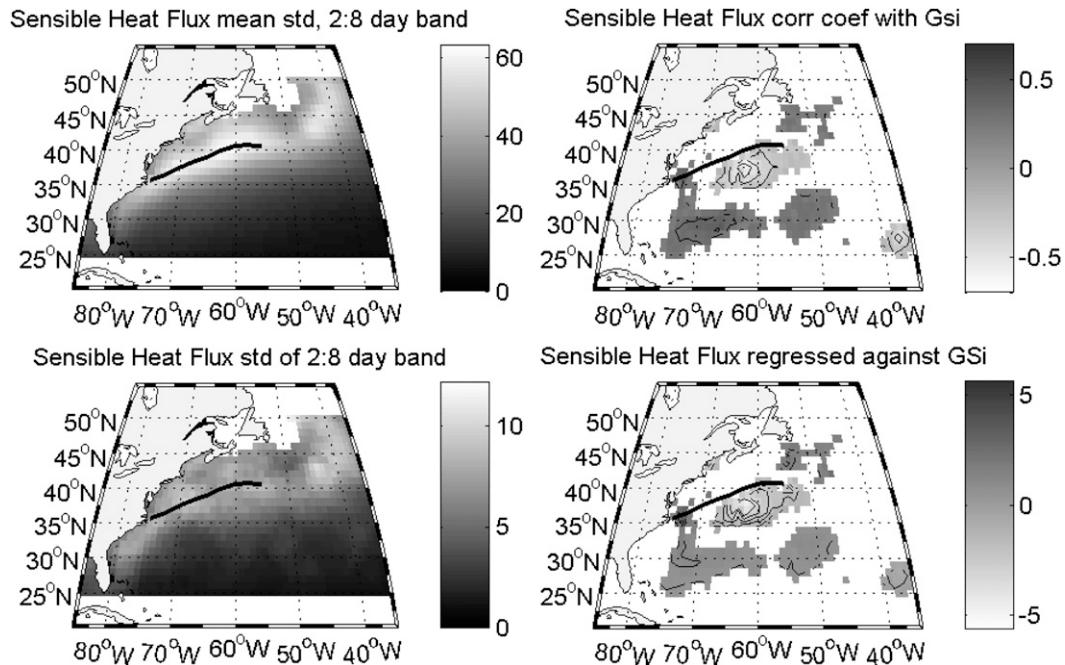


FIG. 8. As in Fig. 7, but for sensible heat flux.

atmospheric signal onto the ocean index. Because the latter has an rms value of 1, the regression shows the amplitude of the atmospheric boundary layer signal that is explained by one standard deviation of the index path shifts. Because the independent variable in our regression is a measure of atmospheric variability, and this is

uncorrelated with the value in the previous or following winter, we have estimated significance based on the 22 independent winter records in the time series (20 degrees of freedom after mean and trend removal). The amplitude of the local latent heat flux regression is $\sim 5 \text{ W m}^{-2}$, which broadens the latent heat loss when the GS is

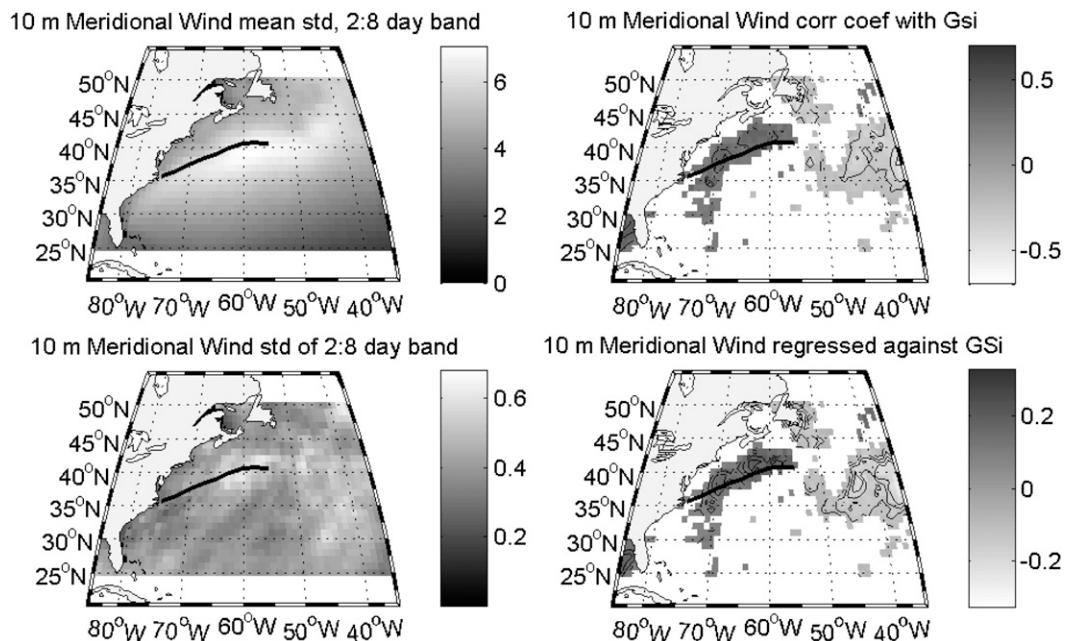


FIG. 9. As in Fig. 7, but for meridional wind. In this case the units are m s^{-1} .

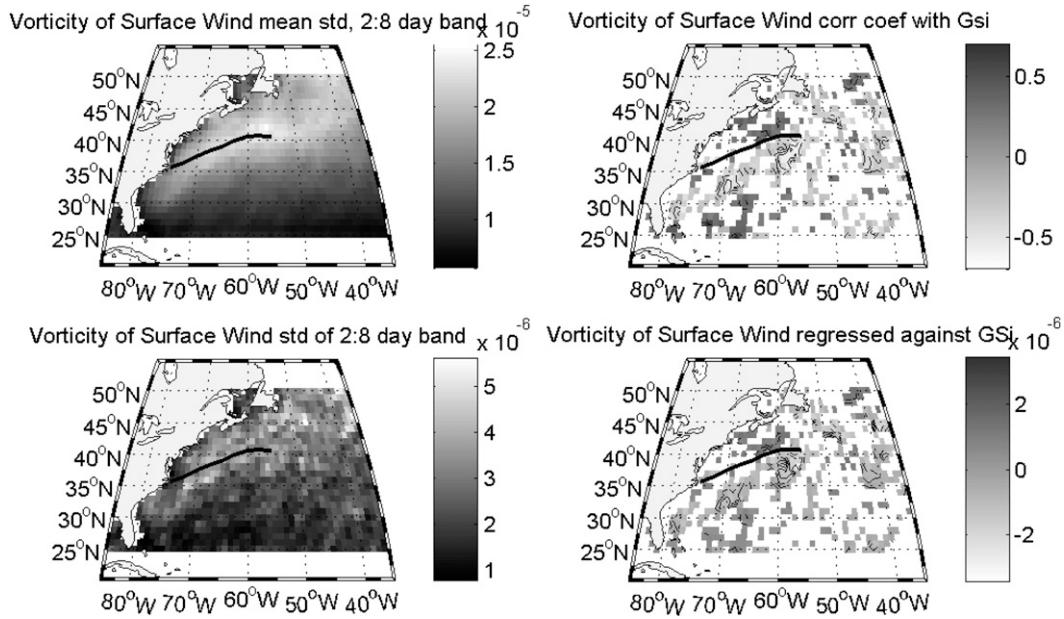


FIG. 10. As in Fig. 7, but for vorticity. The scales are (top left) 10^{-5} s^{-1} , and (bottom left and bottom right) 10^{-6} s^{-1} .

north, and intensifies the local maximum south of the GS when the GS is south. The dipole signal cancels out in a spatial mean to first order and merely represents a change in the location of the latent heat loss depending on the location of the GS. There is also a curious correlation signal outside of the storm track in the subtropics at around 30°N . This is not a local response to path shifts. We have also carried out composite analyses in this and other comparisons (not shown) and arrived

at similar conclusions. Other atmospheric signals (below) will have a similar nonlocal signal in this location.

Similar patterns for the sensible heat exchange (Fig. 8) are also found. However, the mean (not shown) and mean rms (upper left panel) show large values immediately next to the shoreline, and there is not a clear maximum over the GS as for latent heat. This reflects the fact that the evaporation over the colder, nearshore waters is suppressed compared to farther offshore,

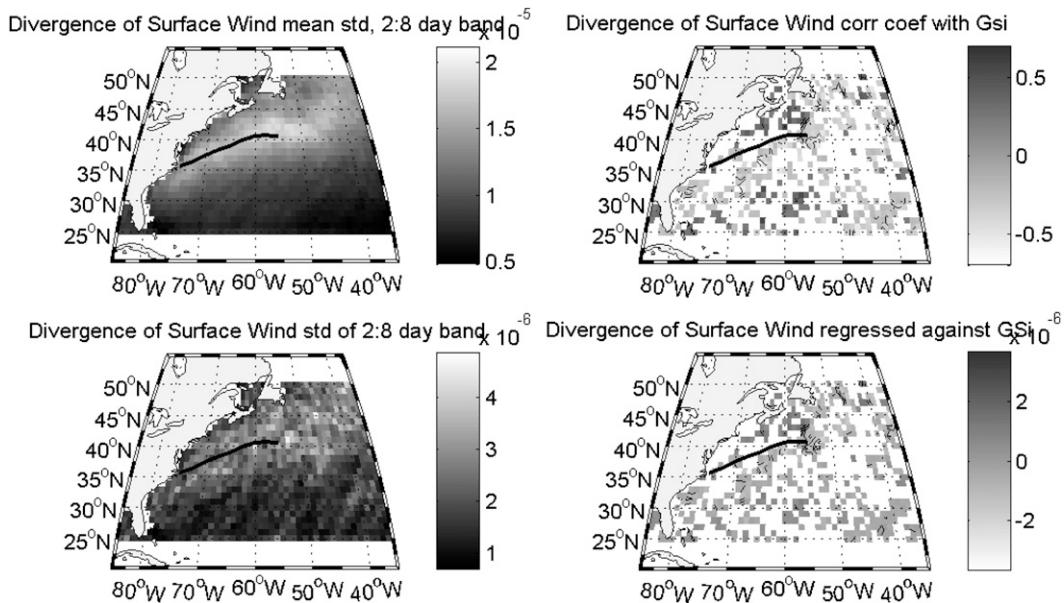


FIG. 11. As in Fig. 10, but for divergence.

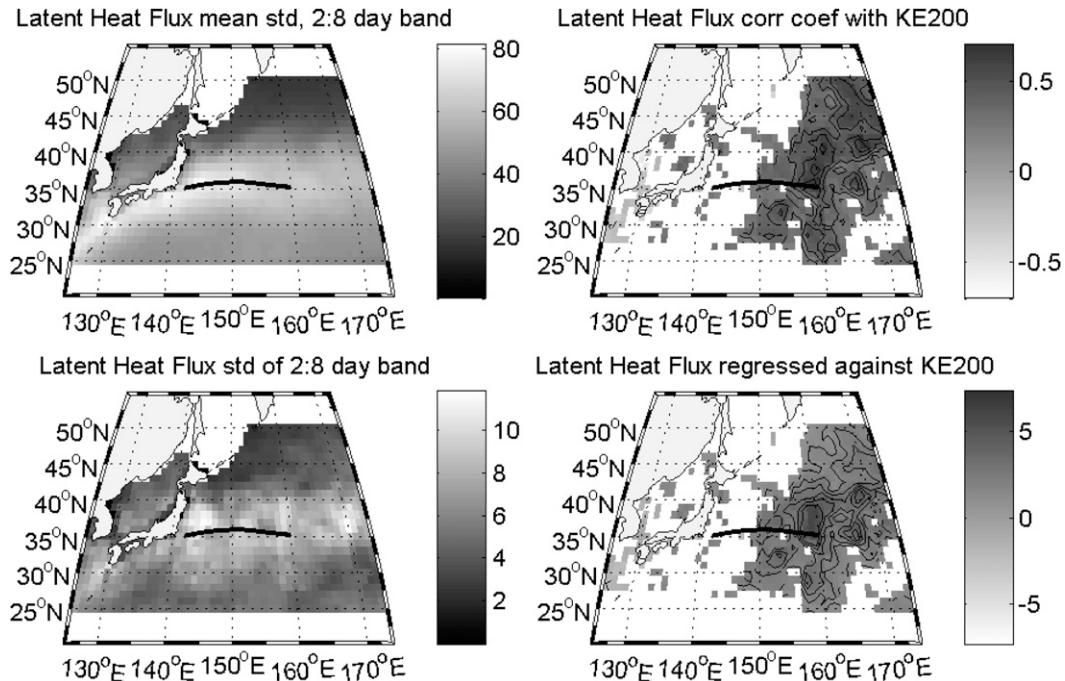


FIG. 12. Latent heat flux for the KE region: (top left) mean rms 2–8-day band for JFM, (bottom left) interannual JFM std of rms 2–8-day band, (top right) correlation coefficient of interannual JFM rms variability with KE index with only significant (>90%) values plotted, and (bottom right) regression against KE index. For all but the correlation coefficient, the units are W m^{-2} . The mean KE path is plotted as the solid line in all panels.

where both the atmospheric boundary layer and the ocean are warmer and the air can hold more water vapor. Because of this difference between the sensible and latent heat patterns, the correlation pattern of the interannual variability does not suggest a simple dipole, but rather an extension of the high sensible heat loss variability to the south when the GS path is in a southerly location. Again, we see a nonlocal response in the atmosphere around 30°N .

Meridional wind variability, often examined in storm-track analyses, has a broad variability maximum located to the south of the mean GS path (Fig. 9). As the GS shifts northward, a shift of this mean pattern to the north would clearly bring stronger variability of meridional winds to this region, while a southward shift of the variability maximum following the GS would not necessarily bring stronger variability to the region immediately south of the mean GS path, because of the broad meridional maximum in this variability measure. We believe this explains the local correlation patterns in Fig. 9. A nonlocal response in the eastern part of the domain is not clearly related to shifts in a variability pattern, but rather to the diversion of more atmospheric variability toward the Azores (38°N , 28°W), just off our map to the east, when the GS is in a southerly location. It is also seen in the zonal wind variability (not shown).

This nonlocal wind response is not collocated with those for the turbulent heat fluxes.

Both vorticity (Fig. 10) and divergence (Fig. 11) are noisier than other variables presented thus far, because smaller-scale atmospheric variability is enhanced when derivatives are taken of the synoptic-scale flow. The mean divergence variability maximum is somewhat more meridionally constrained than for vorticity, but both closely follow the mean GS path. Also, it is apparent that they track northward and southward with path changes, based on the spatial pattern of the correlations in the vicinity of the GS. There is also a hint that in the southeast part of the domain, there is an increase of both of these variability measures because the GS path is southerly. There is also a similarity to the nonlocal response of the turbulent fluxes near 30°N , 60°W .

Because the region to the north of the GS path is relatively narrow, any given synoptic system in the atmosphere that passes over this region will be at least partly over land. Thus, the atmospheric region to the north of the mean GS is substantially different than for the KE in the North Pacific, where the entire region is maritime. And there is no Atlantic counterpart to the Sea of Japan/East Sea in the “upstream” region of the Atlantic storm tracks. We believe these geographic

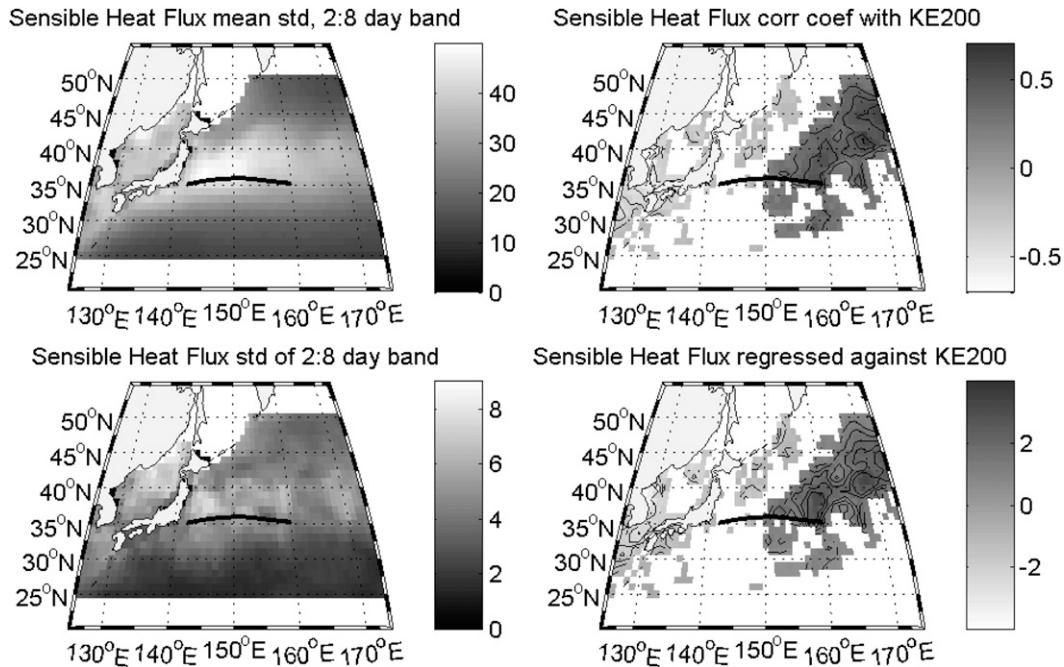


FIG. 13. As in Fig. 12, but for sensible heat flux.

differences are important for the difference in response in the atmospheric variability for the two regions.

b. Kuroshio Extension

Unlike the GS region, the relationship of the latent heat variability (Fig. 12) with the KE index is quite

different: downstream of the region, which defines our KE index, the variability is significantly higher (lower) when the KE is in a northerly (southerly) position. We define our KE index in the region between 142° and 160°E. There is no suggestion of a local response that follows the current (dipole) as the KE shifts in time.

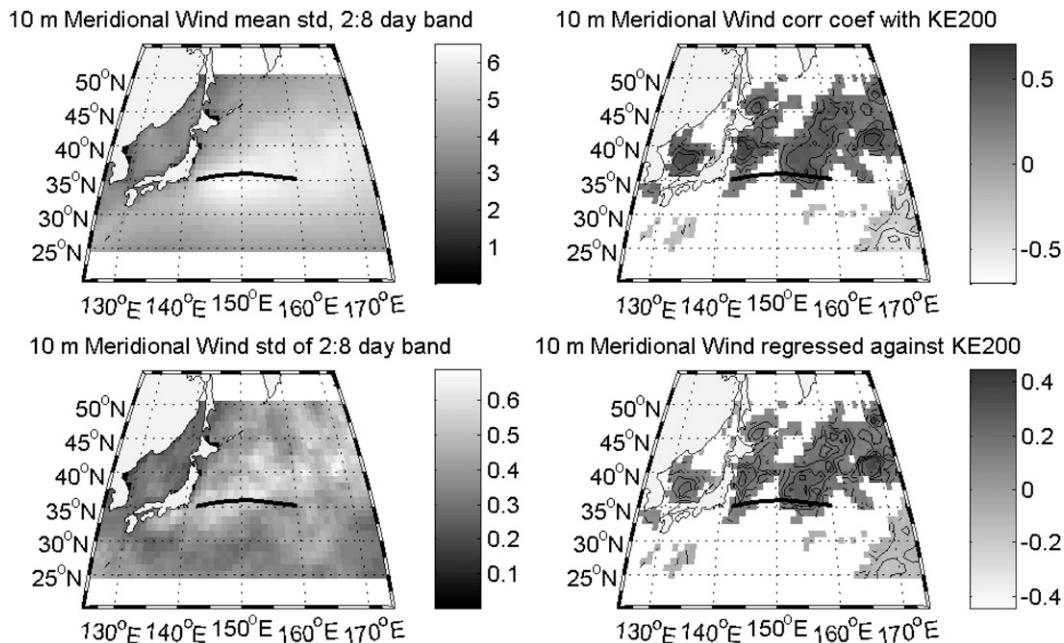


FIG. 14. As in Fig. 12, but for meridional winds. In this case the units are $m s^{-1}$.

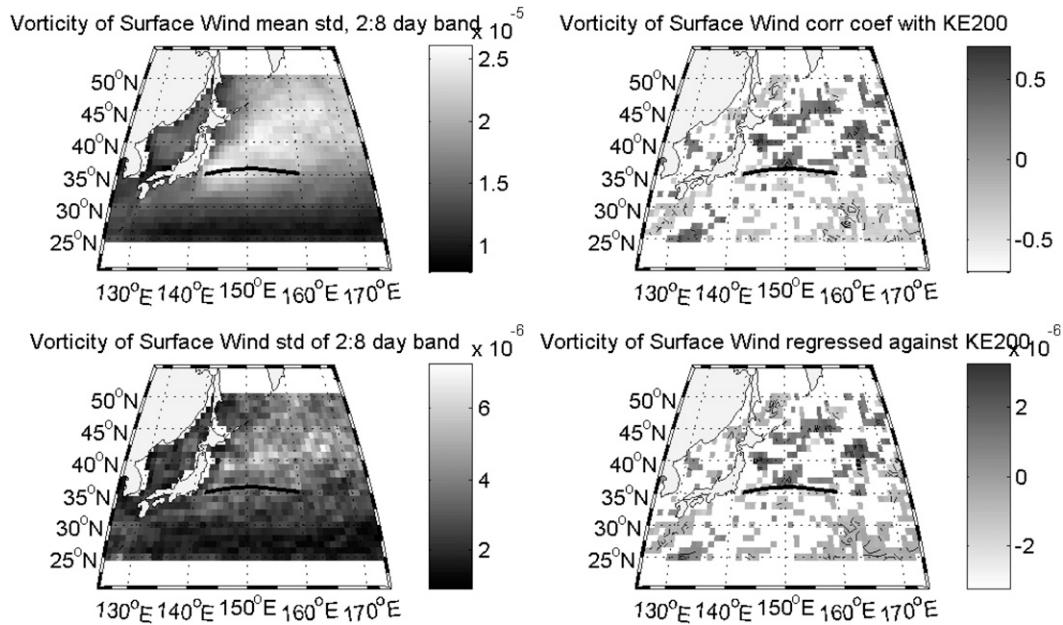


FIG. 15. As in Fig. 10, but for vorticity in the KE region.

This is also true for the sensible heat variability (Fig. 13), although it is not as widespread. Thus, unlike the GS, the spatial mean signal in the KE is one of more turbulent heat loss when the KE is in a northerly position, and vice versa for a southerly position. As well, there seems to be an upstream influence of the KE in the waters surrounding the Korean peninsula and substantial variability throughout the East Sea/Sea of Japan. As for the GS, however, the meridional maximum in sensible heat variability is shifted northward compared to that for latent heat, and the maximum region for variability in meridional winds (Fig. 14) is centered over but south of the mean KE axis. Thus, northward shifts of the winds in concert with the KE will produce a local signal to the north of the KE, but not to the south, because southward shifts of the wind maximum following the KE can produce both signs of correlations depending location. Also, as for the GS, there is an increase of meridional wind variability with southward shifts of the KE in the SE part of the domain, suggesting remote shifts of storm tracks with little or no effects on the thermodynamics of the air–sea exchange. The vorticity map (Fig. 15) is quite interesting and shows the importance of the dual-frontal zone east of Japan. The mean variability is broad and generally shows a high slanting east–northeast from the point where the KE is closest to Japan. However, the part of the signal that varies from year to year (lower left panel of Fig. 15) indicates that this tracks over either the subarctic front or eastward extension from Japan of the

Oyashio. It is possible that this oceanic front may move independently from the KE. We have not examined this here. It is quite apparent that despite the patchiness of the significant correlations between vorticity variability and the KE path, there is a clear sign change in their correlation across the mean KE path. Clearly, the atmospheric boundary layer vorticity variability changes in concert with the KE location, independent of any preexisting alignment between the KE and “mean rms vorticity” signal. Again, we stress that an SST or subsurface temperature index for the subarctic front might show statistical relationships with the atmospheric vorticity variability that are different than for the KE. This is something for future study.

5. Discussion

The relationship between the observed vorticity and divergence can be illuminated by the use of a simple model of the marine atmospheric boundary layer (MABL), where we consider layer velocities (u , v) that obey a balance of

$$\begin{aligned} fu &= -\rho^{-1}p_y - rv \quad \text{and} \\ -fv &= -\rho^{-1}p_x - ru, \end{aligned}$$

where $r = C_d|(u_i, v_i)|/H$,

where f is the (constant) Coriolis parameter and r is a (constant) Rayleigh friction, representing the ocean–atmosphere stress acting over the layer of depth H with

velocity (u, v) . Here we have used a linearized form of the quadratic bottom stress based around a velocity (as yet unspecified) of (u_l, v_l) . The two equations above can be differentiated to obtain the following two related equations:

$$f(v_x - u_y) = \rho^{-1}(p_{xx} + p_{yy}) + r(u_x + v_y),$$

$$f(u_x + v_y) = -r(v_x - u_y).$$

The second illustrates the relationship between the vertical component of vorticity and the horizontal divergence of the boundary layer velocity, namely, that they are linearly related with divergence scaled by a factor of $(-r/f)$ times the vorticity. The first and second can be used to relate the horizontal Laplacian of pressure to the vorticity. In the above, we have assumed that the pressure will not be affected by boundary layer processes as might be expected from heating or cooling (e.g., Lindzen and Nigam 1987). However, the above simple relationship between vorticity and divergence can be used to explain the main features in the observations of the previously presented mean quantities (Figs. 2 and 3), as well as for the bandpassed data. If the mean vorticity and divergence are taken from a selected area over the Gulf Stream (Figs. 2 and 3, boxed region), characterized by its location within the rms vorticity and divergence maximum for the 2–8 day band, we estimate that the ratio of the mean divergence and vorticity is ~ -0.48 . Thus, friction is an $O(1)$ quantity in the MABL. For the variability in the 22 winter records, we have constructed a joint probability distribution function (pdf) from the scatterplot of the bandpassed vorticity and divergence in the storm-track maximum region of the Gulf Stream (Fig. 16). We find for this region that the appropriate nondimensional factor relating divergence and vorticity is $\sim -0.78 \pm 0.06$ for the bandpassed data [95% confidence interval following Garrett and Petrie (1981), using our estimated degrees of freedom]. One might expect this ratio to be the same for mean flow and variability if it truly represents the simple MABL model above. However, because the underlying physics of the drag law upon which the frictional parameterization is based is quadratic, the ratio might vary between the mean and synoptic states. We believe, however, that there is a simpler explanation for the difference between the two ratios: neglect of the imprint of the mean flow of the Gulf Stream on the overlying mean atmosphere (Cornillon and Park 2001). With a mean northwesterly airflow giving a downstream flow of wind over water, the GS will produce a negative imprint of its vorticity on the MABL vorticity. Because the mean MABL vorticity is small compared to the fluctuations and comparable to the ocean-induced signal, one sees an increase in the mag-

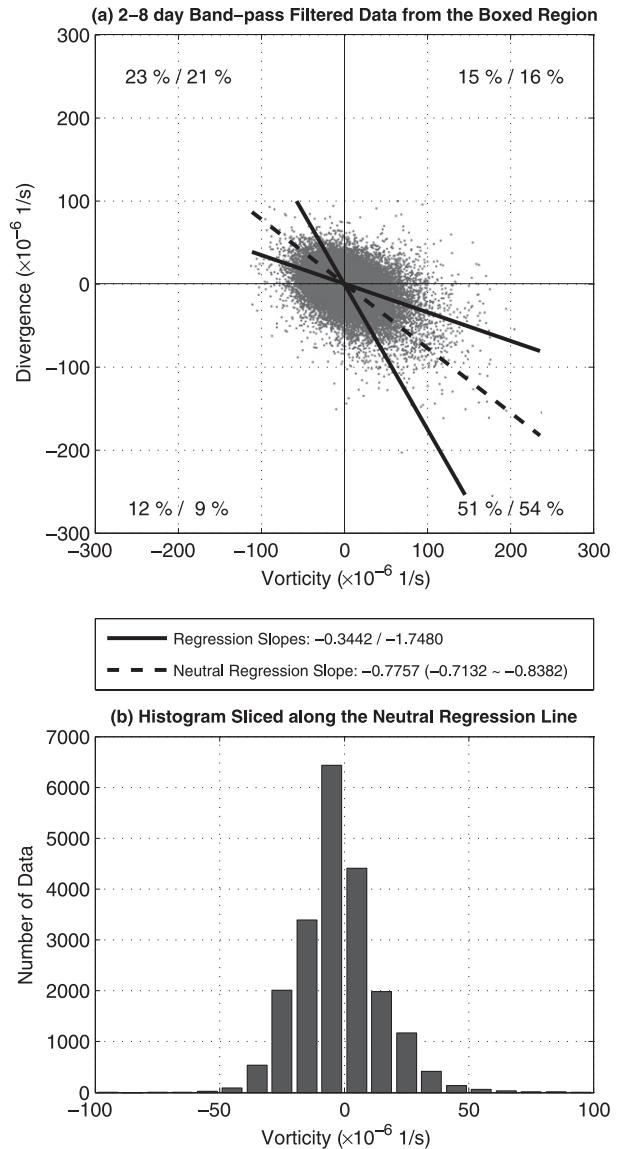


FIG. 16. (top) For the storm-track maximum box over the GS (see Figs. 2 and 3), we show a scatterplot of 2–8-day bandpassed divergence and vorticity, with regression lines and percent variance explained by observations in each of the four quadrants of the plot. Regressions for one variable against the other (solid lines). Along the neutral regression line (dashed line), we show (bottom) the histogram of vorticity indicating a skewed distribution. Here we have filtered the data and removed the means, so the mean vorticity and divergence is zero. The means of these quantities are displayed in Figs. 2 and 3, respectively.

nitude of the positive mean vorticity south of the GS and over the region where we have examined this ratio for the mean state and for the fluctuations.

Beyond these issues, note that the scatter of the synoptic vorticity and divergence data is quite skewed, with an enhanced tail associated with positive vorticities and convergent flows, which is clearly what one would

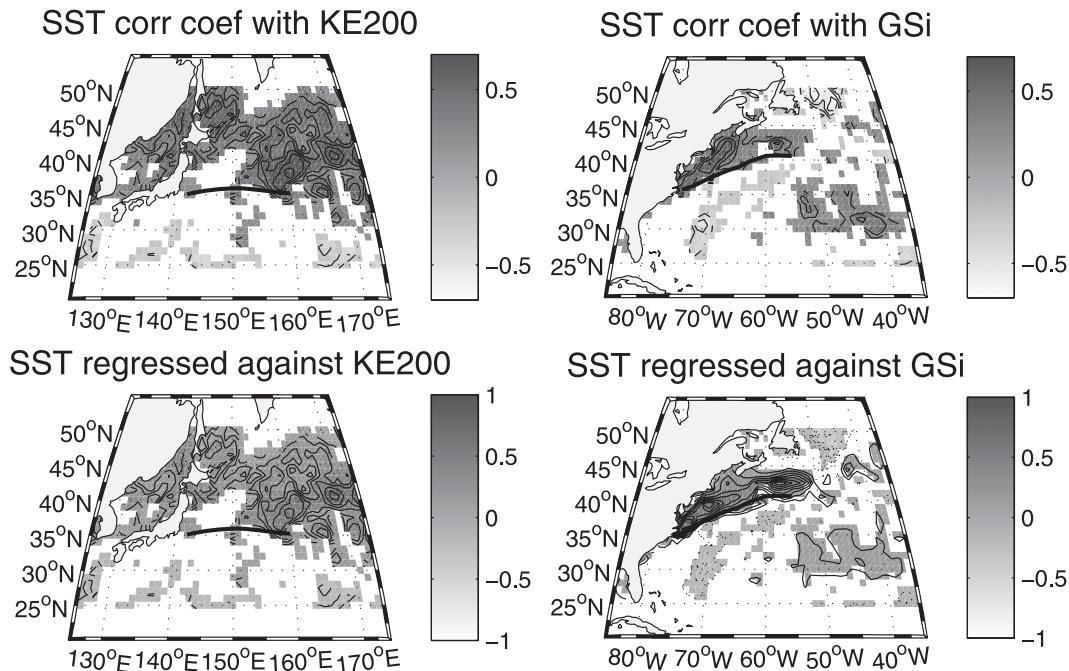


FIG. 17. Wintertime SST signals are correlated with the KE and GS indices to show the effect of these frontal shifts on SST. (top) The correlation coefficient is plotted for significant ($>90\%$ confidence) correlations and (bottom) the regression ($^{\circ}\text{C}$) is shown for the (left) KE region and (right) GS. The contour interval for all panels is 0.1 in the respective units.

expect from intense cyclones with convergent Ekman layers that aid in their development. For variability in this quadrant (southeast of the scatterplot in Fig. 16) of the bivariate distribution, the percent of the total variance explained would be 25% for a symmetric pdf with two independent variables. However, we find 51% (55%) of the total variance of vorticity (divergence) is contained in this quadrant because of a higher probability of extreme events. The northeast and southwest quadrants together explain only about 20% of the variance. To this extent, therefore, our statistical relationships deriving from the covariability between storm signals and GS/KE paths tend to be dominated by convergent wintertime cyclones with weaker, divergent anticyclones, at least over the region where the mean variability signals are largest.

Our analysis of atmospheric variability has been limited by our data to the sea surface, within the MABL. It remains a question how connected the MABL variability is with the rest of the troposphere. That there is an active connection based on the mean fields seems to be established, at least for the GS region (Minobe et al. 2008). Oceanic heating of the MABL in synoptic storms is a vigorous signal and makes up a significant portion of the mean wintertime signal. Our results indicate that rms storm-related sensible and latent fluxes of 60 and 90 W m^{-2} , respectively, are found over the GS (Figs. 8

and 7), and that these are appreciable in terms of mean values, approaching 30%–50% of the mean turbulent fluxes. Similar results apply to the KE region. Thus, winter storms must contribute significantly to the mean signal of diabatic heating throughout the troposphere over the GS and KE and to the energetics that create the storm tracks in the first place (Hoskins and Valdes 1990). However, the vorticity dynamics of the variability also leads to a reinforcement of the westerly wind jet, which, at sea level, can influence the mean latitudes of the oceanic western boundary currents once they leave the coasts. This conundrum of what causes what is partially removed by looking at the variability of the ocean fronts. Because they exhibit path changes on interannual to decadal time scales, observing the covariability of the MABL variability and the GS and KE path locations at zero seasonal lag should enhance the oceanic signals that force the atmosphere. At time lags from one to several years, the oceanic variability caused by the atmosphere will dominate, as has already been established for the KE and GS. That there is a significant change in atmospheric variability in concert with the changes in the GS and KE paths with no appreciable seasonal lag suggests that the ocean is forcing the atmospheric storm track and that there might be skill in predicting future changes of atmospheric storm tracks based on oceanic conditions looking ahead several months to

a year. By itself, the rms variability of the atmospheric variables has no persistence from one season to the next.

The effect of the changing GS and KE path on the overlying atmospheric variability is not the same for both regions, perhaps because of differences in the mean storm tracks, rather than the mean ocean. However, the magnitude of the mean SST gradient in the KE and GS is different (Figs. 2 and 3) as is the response of the SST to changes in the GS and KE paths (Fig. 17). The spatial structure of the SST signal associated with changes in these indices is different for the two regions. For the GS, the associated SST signal looks like more north–south shifts of the GS itself, but with a larger amplitude to the north of the GS than to the south, due to the fact that the SST gradient is larger on the northern flank of the warm core than to the south. For the KE, the SST signal is broad, downstream, and reflects a warmer KE region when the KE itself is north and colder when it is south. These dominant SST signals for the GS and KE are damped by the atmospheric response rather than generated by changes in the storm track; they are of the same sign as changes in storm-related latent and sensible heat flux, with increased (decreased) heat loss over warmer (colder) waters.

For both regions defining the Northern Hemisphere storm tracks, the local, frontal signatures of oceanic forcing of synoptic atmospheric variability are limited to within a few degrees of latitude on both sides of the mean SST fronts. However, we have also found nonlocal signals in the atmosphere, which suggest that changes in the GS and KE path location can influence broad regions of the atmospheric storm climate downstream of the regions of coherent path changes in these currents. The atmospheric boundary layer signals are not always consistent among the variables measured in relation to the oceanic forcing. For example, latent heat release behaves differently in the far field than meridional wind or vorticity variability. The reasons for this and the nonlocal nature of the response are not entirely clear to us at this time. They may in part be due to SST-related forcing by shifts of the ocean fronts, but also seasonal–decadal changes in the low-frequency atmospheric circulation arising from its response to oceanic forcing and not a direct link between path changes and large-scale, nonlocal synoptic variability. Some of these issues can be examined using existing datasets, but one would hope that further dynamical connections can be made using models that both resolve the strong SST gradients of these oceanic flows and allow them to shift in a coherent fashion over substantial zonal distances.

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