

The Probability Distribution of Land Surface Wind Speeds

ADAM H. MONAHAN AND YANPING HE

School of Earth and Ocean Sciences, University of Victoria, Victoria, British Columbia, Canada

NORMAN MCFARLANE

Canadian Centre for Climate Modelling and Analysis, University of Victoria, Victoria, British Columbia, Canada

AIGUO DAI

National Center for Atmospheric Research, Boulder, Colorado*

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ABSTRACT

The probability density function (pdf) of land surface wind speeds is characterized using a global network of observations. Daytime surface wind speeds are shown to be broadly consistent with the Weibull distribution, while nighttime surface wind speeds are generally more positively skewed than the corresponding Weibull distribution (particularly in summer). In the midlatitudes, these strongly positive skewnesses are shown to be generally associated with conditions of strong surface stability and weak lower-tropospheric wind shear. Long-term tower observations from Cabauw, the Netherlands, and Los Alamos, New Mexico, demonstrate that lower-tropospheric wind speeds become more positively skewed than the corresponding Weibull distribution only in the shallow (~ 50 m) nocturnal boundary layer. This skewness is associated with two populations of nighttime winds: (i) strongly stably stratified with strong wind shear and (ii) weakly stably or unstably stratified with weak wind shear. Using an idealized two-layer model of the boundary layer momentum budget, it is shown that the observed variability of the daytime and nighttime surface wind speeds can be accounted for through a stochastic representation of intermittent turbulent mixing at the nocturnal boundary layer inversion.

1. Introduction

The probability distribution of surface wind speeds is a mathematical function describing the range and relative frequency of wind speeds at a particular location. It is important for calculating surface fluxes (e.g., Cakmur et al. 2004), wind extremes (e.g., Gastineau and Soden 2009), and wind power climatologies (e.g., Troen and Petersen 1989; Petersen et al. 1998a,b; Burton et al. 2001). While empirical studies of the wind speed probability distribution have a long history (e.g., Brooks et al.

1946; Putnam 1948; Dinkelacker 1949; Sherlock 1951; Anapol'skaya and Gandin 1958; Essenwanger 1959; Narovlyanskii 1970; Baynes 1974; Hennessey 1977), relatively little work has been done to explore the physical mechanisms that determine the character of these distributions. By better understanding the physical controls of the probability distribution of surface wind speeds, we can improve the estimation of surface fluxes in observations and global climate models (GCMs) (e.g., Wanninkhof et al. 2002; Cakmur et al. 2004; Capps and Zender 2008) and the prediction of the wind power resource and extreme surface winds in present and future climates (e.g., Troen and Petersen 1989; Curry et al. 2010, manuscript submitted to *Climate Dyn.*; van der Kamp et al. 2010, manuscript submitted to *Climate Dyn.*; Monahan 2010, manuscript submitted to *J. Climate*).

In previous studies, global-scale characterizations of the probability density function (pdf) of sea surface winds have been interpreted using an idealized model of

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Corresponding author address: Adam Monahan, School of Earth and Ocean Sciences, University of Victoria, Victoria BC V8W 3P6, Canada.
E-mail: monahana@uvic.ca

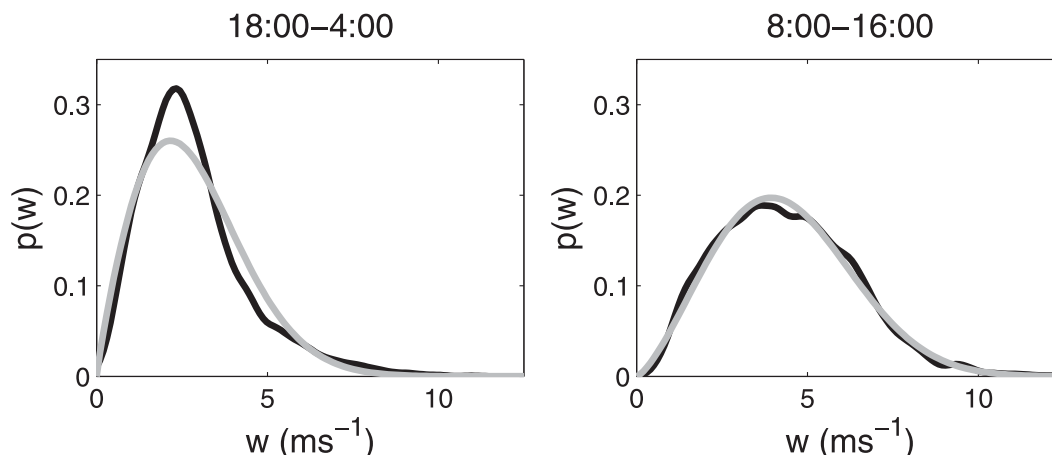


FIG. 1. Black curves: kernel density estimates of the observed JAS 10-m wind speed pdf at Cabauw, the Netherlands, for (left) nighttime (1800–0400) and (right) daytime (0800–1600); all times are local time. Gray curves: best-fit Weibull pdfs to these data.

the boundary layer momentum budget (Monahan 2004, 2006a, 2010). In particular, these analyses have emphasized the manner in which turbulent momentum exchanges (at the surface and with the free atmosphere) influence the shape of the pdf of sea surface winds. Recently, He et al. (2010) analyzed the first three statistical moments of land surface wind speeds—mean, standard deviation (STD), and skewness—at a large number of stations across North America, demonstrating a strong dependence of the shape of the wind speed pdf on surface type, season, and time of day. He et al. (2010) noted that, away from open water, the skewness of nighttime winds is larger than that predicted by the Weibull distribution, the standard parametric model of the pdf of wind speeds. In other words, the Weibull distribution generally underestimates wind speed extremes at night. A number of regional climate models (RCMs) considered in this earlier study were not able to reproduce this diurnal cycle in the surface wind speed pdf.

That the Weibull distribution is a good approximation to the observed wind speed pdf during the day, but underestimates extremes at night, is illustrated in Fig. 1 in which observed wind speed pdfs (black curves) are contrasted with best-fit Weibull pdfs (gray curves) for summertime 10-m wind speeds at Cabauw, the Netherlands. At night, the central part of the observed wind speed pdf is narrower than that of the corresponding Weibull distribution, with a more pronounced tail toward higher wind speeds. In contrast, the daytime wind pdf closely matches the Weibull distribution. Note that the largest wind speeds at night are no larger than those during the day; in this sense, the nighttime winds are no more “extreme” than daytime winds. Rather, the extreme wind speeds *relative to the mean wind speed* are

greater at night than during the day and—for nighttime winds—larger than would be expected of a Weibull-distributed quantity.

The present study extends He et al. (2010) to a global domain, characterizing the land surface wind speed pdf and the observed physical controls on its shape (section 3). A complementary observational analysis of the relationship between the surface wind speed pdf and other boundary layer processes is then carried out using long-term observations from tall meteorological towers (section 4). It is shown that, while the Weibull distribution is a good approximation of the pdf of surface wind speeds under conditions of strong winds or unstable stratification when winds are weak and stratification is stable, the Weibull distribution underestimates wind speed skewness (and therefore the extremes) in the extratropics. The Weibull distribution is used as a point of comparison, not because of any fundamental physical importance of this parametric distribution, but because it is the standard distribution used in studies of surface wind speed variability. The results of these observational analyses are then interpreted using an idealized model of the boundary layer momentum budget incorporating a representation of the diurnal cycle in the stratification of the lower atmosphere (section 5). The predictions of this model demonstrate that, at least in the extratropics, the nocturnal extreme wind speeds (greater than those for a Weibull variable) can be understood as resulting from intermittent turbulence at the inversion capping the stably stratified nocturnal boundary layer. A discussion and conclusions follow in section 6. The appendix provides a review of the empirical wind speed probability distributions that will be considered in this study. Throughout this study, wind speeds will be denoted

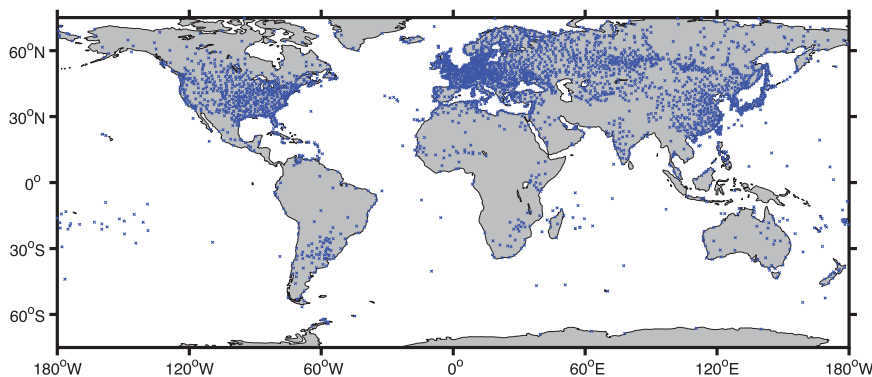


FIG. 2. The network of 3734 surface meteorological stations with at least 12 years of 3-hourly surface wind speed observations considered in this study.

by w and their pdf by $p_w(w)$. Finally, we note that the focus of this study is on the pdf of the “eddy averaged” winds, not the winds on the time scales of turbulence. In the present study, these faster time scales are parameterized rather than investigated explicitly.

2. Data

Three-hourly synoptic weather reports transmitted through the Global Telecommunication System (GTS) and archived at the National Center for Atmospheric Research (NCAR) are used here to characterize the surface wind speed pdf on a global scale. This dataset was described in Dai and Deser (1999). It contains 3-hourly, 10-min-averaged data of surface wind speed usually (but not always) measured at around 10-m height at weather stations and reported in units of knots ($1 \text{ kt} = 0.514 \text{ m s}^{-1}$). Here we used wind data from 1 January 1979 to 31 December 1999 over a global domain (70°S – 80°N , 180°W – 180°E). When observations were available at higher than 3-hourly resolution, mostly for the later years, only the reports near the 3-hourly report times were selected (to ensure consistency of records within and between stations). More than 5000 stations from across the globe are available; for consistency with the analysis in He et al. (2010), we selected from these the 3734 stations that had at least 12 years with a total of 1460 (i.e., 365×4) or more records each year (Fig. 2).

The surface wind speed pdfs will be described by their first three statistical moments (mean, standard deviation, and skewness). The first two of these moments are measures of wind speed magnitude (the average and the variability, respectively); in contrast, skewness is interesting on its own as a measure of the shape of the surface wind speed distribution. These moments are estimated separately for each season [December–February (DJF), March–May (MAM), June–August (JJA), and September–November (SON)] using all years of the 1979–99 data and

for both day (from local sunrise to local sunset) and night (from local sunset to local sunrise). Sunset and sunrise times are calculated using the National Oceanic and Atmospheric Administration sunrise/sunset and solar position calculators, based on Meeus (1999). For each station, there are up to 3832 wind records for each data subset (time of day, season) during the study period. As a result, the leading three statistical moments are expected to be reliably estimated. A preliminary analysis of the observations indicated the presence of a small number of strikingly large wind speed values; other simultaneous measurements were often inconsistent with extreme weather. Because the goal of the present study is to characterize the probability distribution of surface wind speeds in standard conditions, and to exclude erroneous data as much as is possible, the top 0.5% of station wind records was removed before calculating leading moments. More detailed discussions of the processing of these data and of data quality issues are presented in Dai and Deser (1999) and He et al. (2010).

Six-hourly surface ground temperature, 2-m air temperature, and 700-hPa zonal and meridional winds were taken from the National Centers for Environmental Prediction (NCEP)–NCAR reanalysis for the period from 1979 to 1999 (Kalnay et al. 1996). These data are available online (www.esrl.noaa.gov/psd/data/gridded/data.ncep.reanalysis.html).

To investigate the vertical structure of the wind speed moments, data were obtained from the Cabauw Experimental Site for Atmospheric Research (CESAR), which maintains a 213-m tall tower at Cabauw (52°N , 4.9°E) in the western Netherlands (data were downloaded from <http://www.cesar-database.nl/>). As described online (at www.cesar-database.nl/About.do), the tower is located in open, flat pastureland with few nearby wind-breaks, such as houses or trees. Validated observations of 10-min averaged wind (speed and direction) and

temperature were obtained from 2002 through 2009 at heights of 10 m, 20 m, 40 m, 80 m, 140 m, and 200 m. Air temperature observations were also available at a height of 2 m.

Observations of boundary layer wind and temperature were also obtained from a 92-m tower (denoted TA-6) maintained by the Los Alamos National Laboratory. This tower, located at an altitude of 2263 m above sea level near Los Alamos, New Mexico (35°51'N, 106°19'W), is in a very different physical environment than that of Cabauw. Wind and temperature observations with 15-min resolution were obtained from 1 February 1990 through 31 December 2009 at altitudes of 12 m, 23 m, 46 m, and 92 m above the ground, along with temperature and pressure observations at a height of 1.2 m. A detailed description of the tower is available online (<http://environweb.lanl.gov/weathermachine/ta6.asp>) and data is available online at http://environweb.lanl.gov/weathermachine/data_request_green.asp.

The data from Cabauw and Los Alamos were averaged to hourly resolution so as to smooth the associated plots of statistical moments. The results of these analyses are not qualitatively altered if the moments are computed at their original time resolution.

3. Probability distribution of land surface winds: Global observations

The standard empirically based pdf used for studies of surface wind speeds is the Weibull distribution (e.g., Petersen et al. 1998a; Burton et al. 2001). As discussed in the appendix, this distribution is characterized by a particular relationship between statistical moments: the skewness is a decreasing function of the ratio of the mean to the standard deviation (positive when this ratio is small, near zero when the ratio is intermediate in size (~ 3 – 4), and negative when the ratio is large). In broad terms, a similar relationship between moments is also characteristic of sea surface wind speeds (Monahan 2006a,b) and daytime land surface wind speeds over North America (He et al. 2010). That is, the pdf of sea surface wind speeds is broadly consistent with the Weibull distribution, as is that of wind speeds over land in North America in the daytime. However, He et al. (2010) demonstrated that nighttime land surface wind speeds over North America are more strongly skewed than would be the case for a Weibull-distributed variable.

The relationships between observed land surface wind speed moments over North America reported in He et al. (2010) are also evident in observations of surface winds from the global network of stations. Kernel density estimates (cf. Silverman 1986) of this relationship

for all four seasons, and for day and night, are contoured in Fig. 3; also displayed is the curve corresponding to the relationship for a Weibull variable. Plots for wintertime were obtained by combining Northern Hemisphere (NH) observations in DJF with Southern Hemisphere (SH) observations in JJA. Similarly, springtime plots combine NH MAM data with SH SON data; summertime plots combine NH JJA and SH DJF data; and autumn plots combine NH SON data with SH MAM data.

For daytime observations, the relationship between moments broadly follows that of a Weibull-distributed variable; the spread of the observed moments around those predicted for a Weibull distribution reflects both small deviations from Weibull behavior and sampling variability (e.g., Monahan 2006a). However, at night (particularly in spring and summer), the wind speed skewness is consistently larger than that of a Weibull-distributed variable. For the majority of observations, the pdf of nighttime wind speeds has a tail toward large positive values longer than the Weibull distribution would predict for the observed mean and standard deviation of wind speed. This diurnal difference is not a feature of the winds above the boundary layer; for example, at 700 hPa the relationship between wind speed moments for all seasons and at all times of day is characteristic of a Weibull-distributed variable (Fig. 4). The relationship between wind speed moments aloft is broadly consistent with the vector winds being bivariate Gaussian with uncorrelated, isotropic fluctuations (appendix). In fact, the observed free-tropospheric vector wind component fluctuations are close to Gaussian (e.g., Petoukhov et al. 2008).

A revealing illustration of the roles of surface processes and winds aloft in determining the shape of the surface wind speed pdf is obtained by partitioning the global data in terms of the mean surface stratification and the mean wind speed at 700 hPa (as a proxy for the lower-tropospheric wind shear). Surface stratification was computed as the difference between the 6-hourly ground surface temperature and 2-m air temperature from the NCEP–NCAR reanalysis. Statistical moments (mean, standard deviation, and skewness) of wind speed conditioned on mean surface stability and mean wind speed aloft were computed over $0.5 \text{ K} \times 1.5 \text{ m s}^{-1}$ boxes. From the mean and standard deviation the skewness of a Weibull-distributed variable can be computed, and then compared to the observed skewness. Plots of the skewness excess (the observed skewness less that of a Weibull variable) as functions of stratification and 700-hPa wind speed (Fig. 5) demonstrate that the surface wind speed pdfs are close to Weibull where the winds aloft are strong. As mean 700-hPa wind speeds decrease, the surface wind speeds become increasingly positively

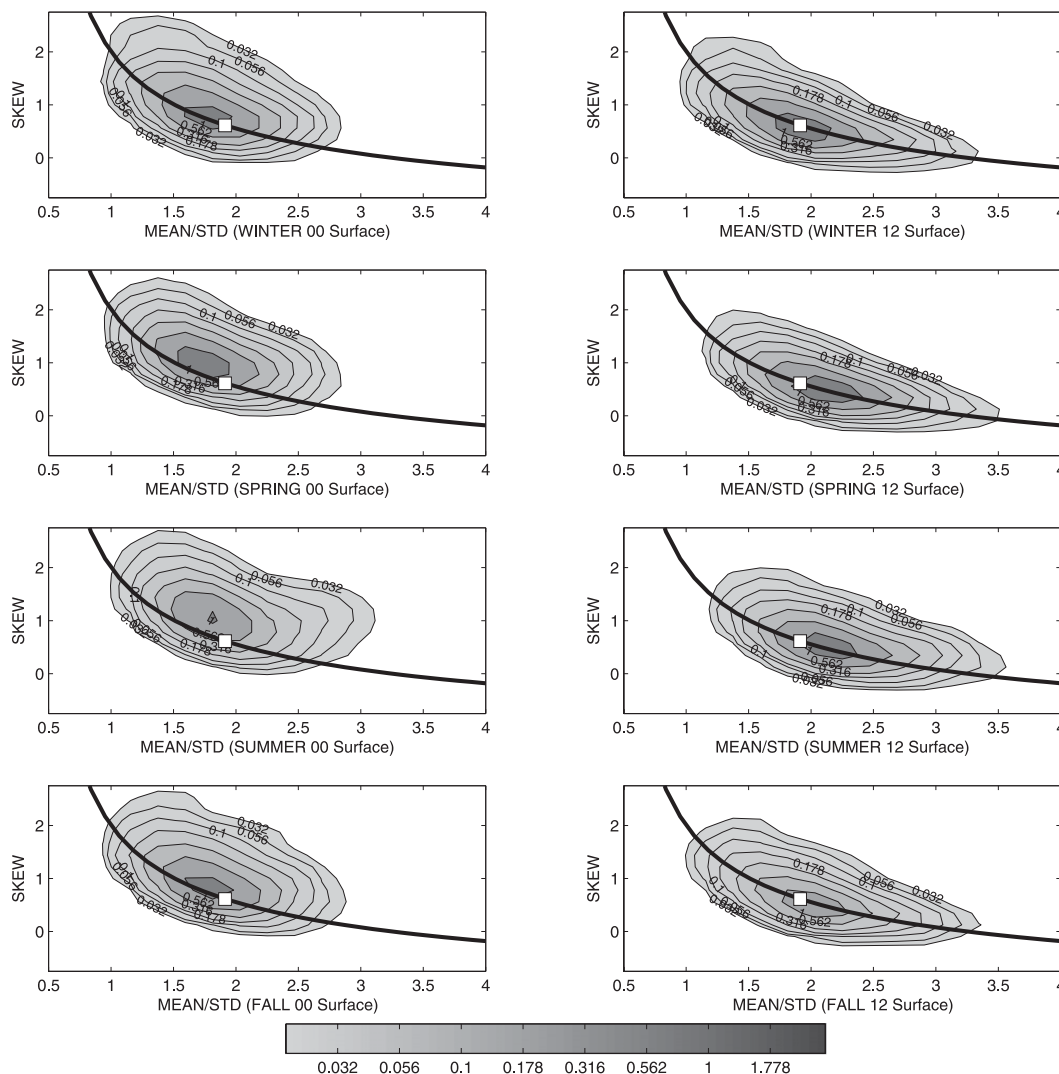


FIG. 3. Estimates of the relationship between surface wind speed skewness and the ratio of the mean to the standard deviation based on surface observations from 1979 to 1999 at the stations shown in Fig. 2. Black curve is the relationship between moments for a Weibull distribution; white square corresponds to the Rayleigh distribution. (left) Nighttime (centered at 2400); (right) daytime (centered at 1200). (first row) Winter data (DJF in NH, JJA in SH). (second row) Spring data (MAM in NH, SON in SH). (third row) Summer data (JJA in NH, DJF in SH). (fourth row) Autumn data (SON in NH, MAM in SH).

skewed relative to a Weibull variable (particularly at night). Furthermore, there is a tendency for the skewness excess to be larger under conditions of stable stratification.

When separated by geographical region (Fig. 6), the tendency for the skewness excess to increase with decreasing mean 700-hPa wind speed and increasingly stable stratification is particularly pronounced in the extratropics (30°S – 90°N). While there is evidence for these relationships in the tropics (30°S – 30°N), the situation in this latitude range is more complicated. Lower-tropospheric wind speeds tend to be lower in the tropics

than the extratropics, and variability more strongly influenced by moist processes. The remainder of this study will focus on the land surface wind speed pdf in the extratropics, where the evidence is clearer of a strong influence of (dry) surface stratification and lower-tropospheric wind shear on the shape of the surface wind speed pdf.

In broad terms, surface wind speed pdfs in the extratropics are close to Weibull when turbulence kinetic energy (TKE) is actively being generated at the surface (by wind shear or unstable stratification), but have larger-than-Weibull extremes when TKE is being consumed. When TKE is actively being generated, turbulent transport

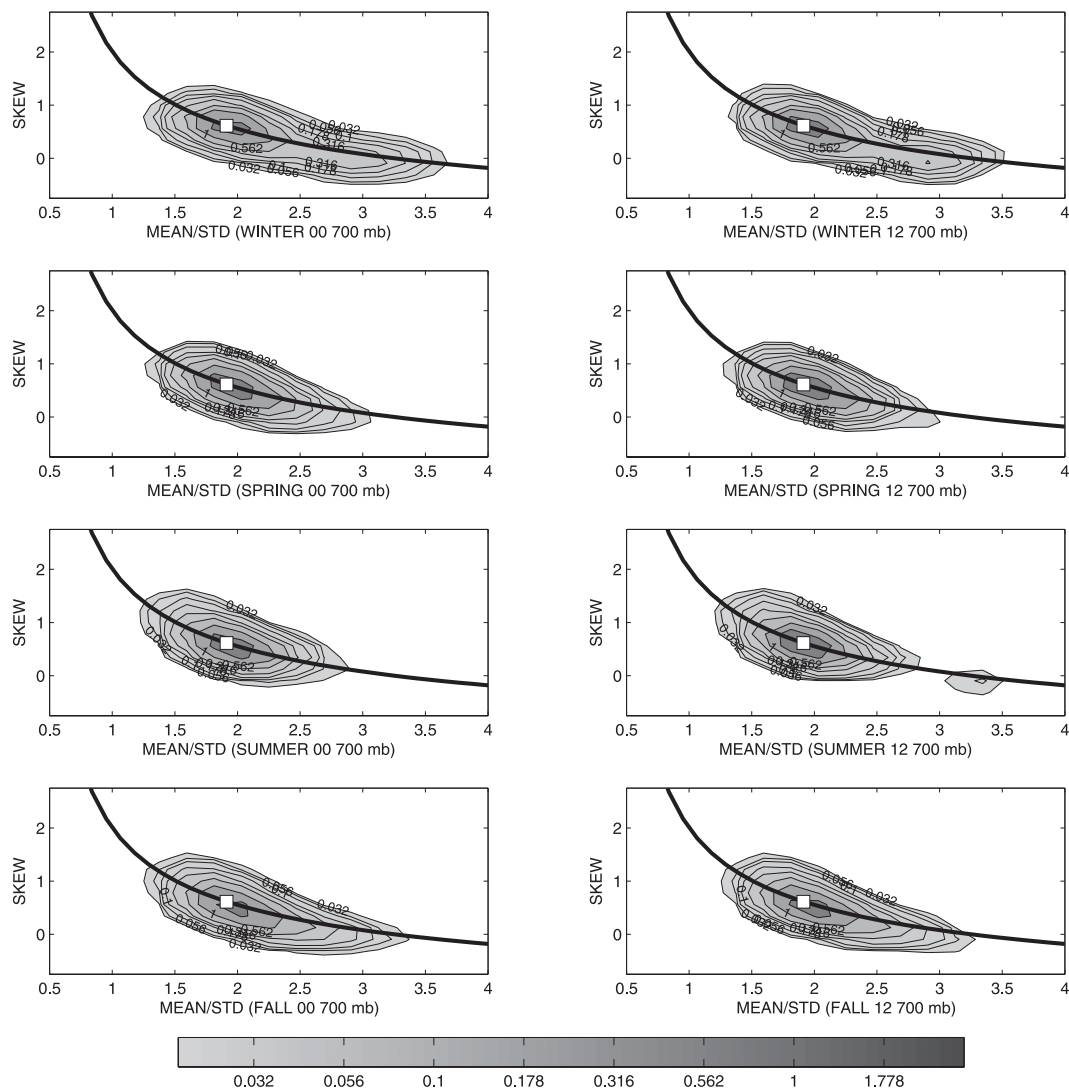


FIG. 4. As in Fig. 3 but for 700-hPa wind speeds from the NCEP–NCAR reanalysis.

of momentum within the boundary layer is enhanced and the surface winds are more strongly coupled with the winds aloft. As the vector winds aloft are close to Weibull, the present results are consistent with the statistical character of the surface winds, reflecting that of the winds aloft when these fields are strongly coupled. In contrast, the weak (on average) turbulence associated with strongly stable stratification and weak wind shear reduces the coupling of surface winds with those aloft and enhances the importance of processes internal to the boundary layer and at the surface. In this interpretation, the relatively large wind speed extremes illustrated by the positive wind speed skewness excess are produced by these local processes. As discussed above, this picture is more complicated in the tropics—possibly because of the complicating influence of moist processes.

4. Probability distribution of land surface winds: Tower observations

The results of section 3 provide insight into the physical controls of variations in the shape of the surface wind speed pdf. However, the surface meteorological observations and reanalysis products provide only a crude characterization of boundary layer processes and their relation to surface winds. A more detailed characterization of these processes is provided by observations with (relatively) fine vertical resolution from tall meteorological towers. We will consider observations from two towers (Cabauw, the Netherlands, and Los Alamos, New Mexico; cf. section 2) that are tall enough (200 and 92 m, respectively) to span the average nocturnal boundary layer and have at least several years of

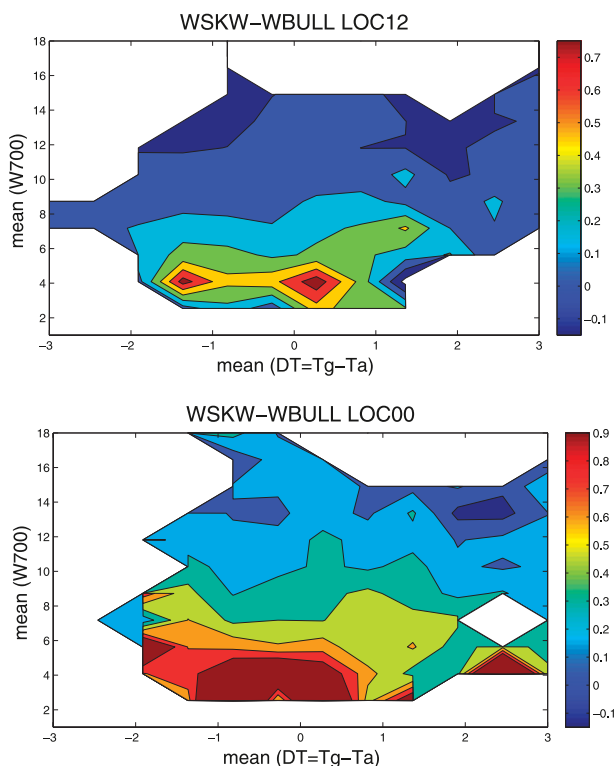


FIG. 5. Observed wind speed skewness excess (observed skewness minus that of a Weibull distribution) as a function of mean surface stability (computed as the difference between the ground temperature T_s and the 2-m air temperature T_a , both from NCEP–NCAR reanalysis) and mean 700-hPa wind speed (also from NCEP–NCAR reanalysis). (top) Observed daytime winds (centered at 1200); (bottom) observed nighttime winds (centered at 2400).

temporally high-resolution data available. Hourly averaged data for late summer [July–September (JAS)] are considered in the following analyses; similar results are obtained for the “raw” data (10 and 15 min for Cabauw and Los Alamos, respectively) and for other seasons.

We first consider variability at the Cabauw tower. The mean summertime diurnal cycle of temperature (uppermost panel of Fig. 7) demonstrates a canonical fair-weather boundary layer evolution (e.g., Stull 1997). Surface warming during the day drives the buoyant generation of turbulence and active mixing in the boundary layer. The daytime active turbulent mixing is particularly evident in the mean diurnal cycle of potential temperature θ (second panel of Fig. 7): from just before noon until late afternoon, the potential temperature is well mixed from the surface to 200 m. At around 1800 (all local times) stable stratification starts to develop in response to surface cooling. In particular, a temperature inversion starts to form at the lowest levels. As evening progresses, the stratification becomes stronger and the inversion layer grows to occupy the lower 50–100 m. Between

0600 and 0700, surface warming generates static instabilities and the boundary layer mixing begins anew.

The diurnal evolution of the mean wind speed (third panel of Fig. 7) also follows the canonical pattern (e.g., Dai and Deser 1999): active turbulence during the day efficiently mixes high momentum fluid down toward the surface and low momentum fluid aloft, so the wind speed gradient is weak. Mean surface wind speeds are highest during the day, while those aloft are weaker than at other times in the diurnal cycle. As the boundary layer becomes stably stratified, the mean surface wind speed decreases while the mean wind speed higher up increases. Weak mean surface wind speeds are maintained between 2000 and 0600, while maximum mean wind speeds aloft occur between midnight and 0200 (likely associated with the timing of the mean nocturnal jet peak). The diurnal cycle of the standard deviation of surface wind speeds (Fig. 7, fourth panel) also displays familiar features: variability in surface winds is greatest during the day when the boundary layer is (on average) actively turbulent, while it is weakest during the period of weak mean wind speeds when the mean surface inversion is well developed. Aloft, the standard deviation has a pronounced semidiurnal cycle: variability is large during the actively turbulent daytime and at night when the surface inversion is strongest. The time at which the nocturnal jet takes its maximum speed is expected to be influenced by the angle between the boundary layer wind and the geostrophic wind when the surface inversion forms (e.g., Thorpe and Guymer 1977); variability in this timing is likely responsible for this nighttime maximum in the standard deviation, $\text{std}(w)$, aloft. Variability in the winds aloft is weakest in midevening and early morning.

The diurnal cycle in observed wind speed skewness is displayed in the fifth panel of Fig. 7, while the bottom panel shows the difference between the observed skewness and that predicted for a Weibull-distributed variable from the mean and standard deviation using Eq. (15) (i.e., the skewness excess). During the day, values of $\text{skew}(w)$ are around 0.5 at all levels, and the skewness excess is close to zero—daytime wind speeds are nearly Weibull at all levels. As the surface inversion develops, the low-level wind speeds become strongly positively skewed, consistent with the global results discussed in section 3: these wind speeds are much more positively skewed than would be a Weibull variable with the same mean and standard deviation. The positive skewness excess persists until the surface inversion breaks down in the early morning. From late in the night until early morning, the winds aloft acquire a slightly negative skewness excess.

During the day, when the boundary layer is (on average) actively turbulent, a clear linear relationship exists

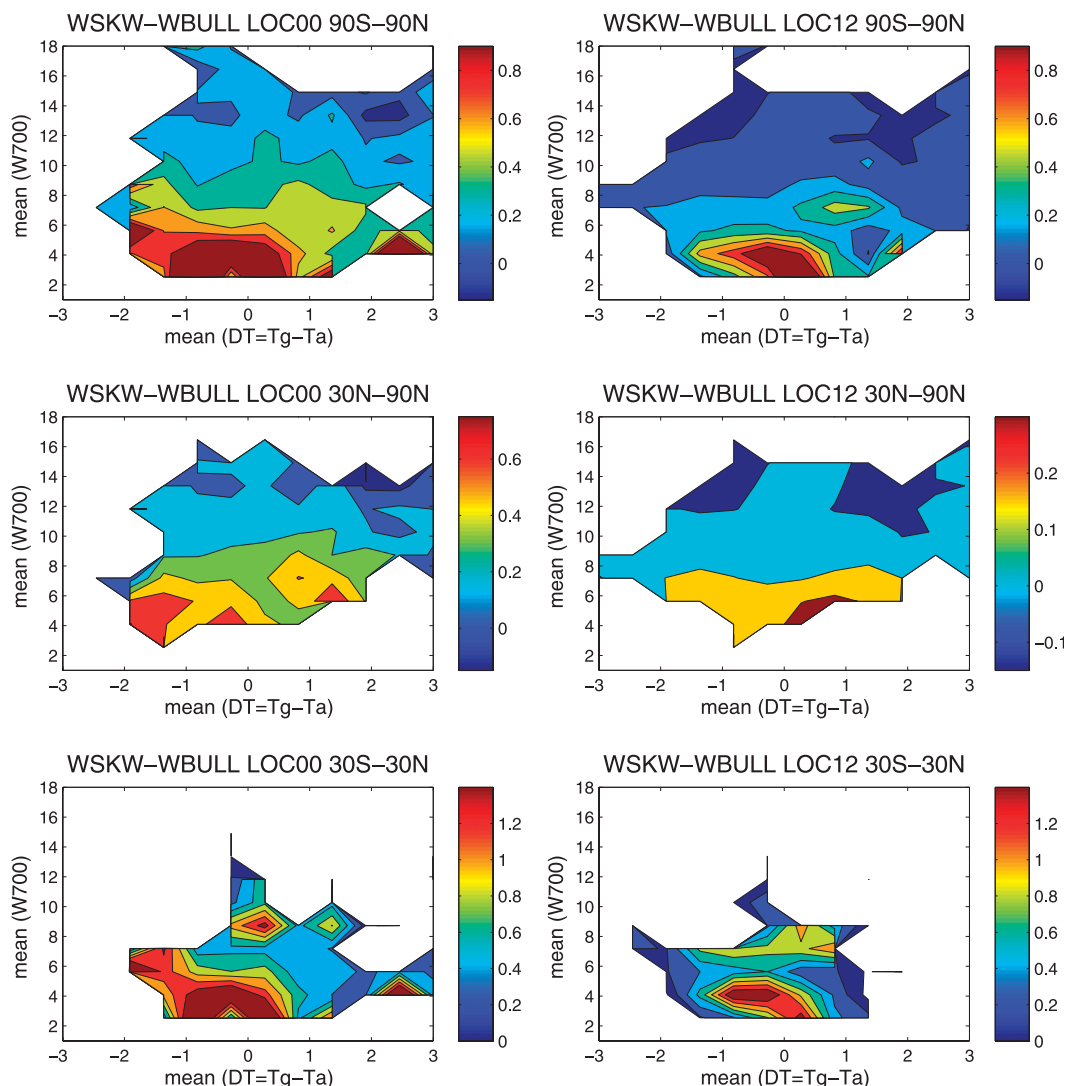


FIG. 6. As in Fig. 5 but for different spatial domains. (first row) All observations. (second row) NH extratropics (30° – 90° N). (third row) Tropics (30° S– 30° N). (first column) Nighttime moments; (second column) daytime moments.

between the wind speeds at 10 and 200 m. This relationship is evident in a scatterplot of the wind speeds at these two levels at 1400 (Fig. 8, second panel). On average, the winds at 200 m are roughly 50% faster than those at 10 m. The relationship between the wind speeds at these two levels is more complicated at night. A scatterplot of wind speeds at these two levels at 0200 (Fig. 8, first panel) suggests the presence of two distinct populations. The more populous of these is characterized by a larger shear than that of daytime winds, such that the 200-m winds are roughly three times faster than those at 10 m. The other, less populous, population has a shallower slope and an average wind speed shear that is much more like that during the day.

These differences in the vertical structure of the wind are features, not just of these particular times, but of all hours in the day and the night. Joint pdfs of wind speeds at 10 and 200 m were estimated for both nighttime hours (1800–0400, first and second panels of Fig. 9) and for daytime hours (0800–1600, third panel of Fig. 9). A clear linear relationship between the wind speeds at 10 and 200 m is evident during daytime hours. Such a clear relationship is not evident in the joint distribution at night (black contours, first and second panels of Fig. 9): in particular, the distribution of w_{10m} broadens as the value of w_{200m} increases. That this spread can be decomposed into two distinct populations is demonstrated by partitioning the data into two sets: those times when the

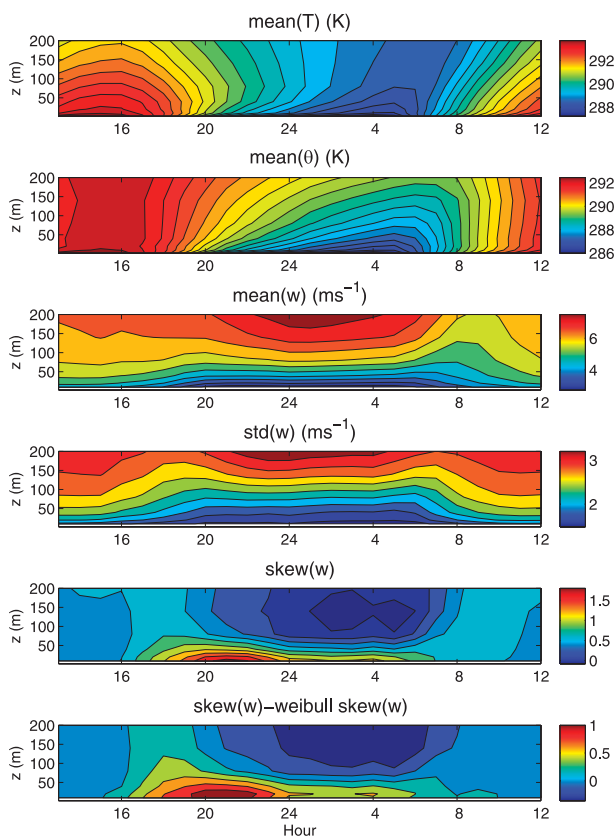


FIG. 7. Diurnal cycles of (first row) mean temperature, (second row) mean potential temperature, (third row) mean wind speed, (fourth row) standard deviation of wind speed, (fifth) skewness of wind speed, and (sixth) difference between observed wind speed skewness and that of a Weibull-distributed quantity computed from hourly averaged data for JAS at Cabauw.

surface stratification is strongly stable ($\theta_{40m} - \theta_{2m} > 1$ K, red contours in the top left panel of Fig. 9) and those times when the stratification is unstable or weakly stratified ($\theta_{40m} - \theta_{2m} < 1$ K, red contours in the top

middle panel of Fig. 9). Under conditions of strongly stable stratification, the wind speeds increase linearly with height so that those at 200 m are between two and three times those at the surface (a much stronger shear than seen during the day). In contrast, the relationship between the wind speeds at the two levels is more similar to that during the day during conditions of weakly stable or unstable stratification. Based on the 1-K threshold, strongly stable stratifications are never (for all intents and purposes) observed during the day (lower right panel, Fig. 9). At night, when the stratification is more likely to be strongly stably stratified, daytimelike conditions of weak stability or instability are also not infrequent (Fig. 9, lower left panel).

The diurnal cycles of hourly averaged July–September wind speed moments observed at tower TA-6 near Los Alamos (Fig. 10) are broadly consistent with those at Cabauw. Strong diurnal cycles in the mean and standard deviation of w at the surface are observed, although the semidiurnal cycle characteristic of winds at Cabauw is not evident. As at Cabauw, the strongly positive wind speed skewness excess is mostly confined to the shallow and stably stratified nocturnal boundary layer. Furthermore, the relationship between surface wind speeds (at 12 m) and those aloft (at 92 m) differs between day and night (Fig. 11). During the daytime, a clear linear relation between wind speeds at the two altitudes is evident (third panel of Fig. 11). At night the slope of the relationship for light surface winds exceeds that of the daytime relationship; for larger wind speeds, the slope of the relationship between surface winds and those aloft is similar to that for the daytime. When the wind data are stratified into those times when the surface stratification (defined as the potential temperature difference between 46 and 1.2 m) exceeds 2 K (“strongly stratified conditions,” Fig. 11) and when the stratification is less than 2 K (“weakly stratified or unstratified conditions,” Fig. 11), evidence of two

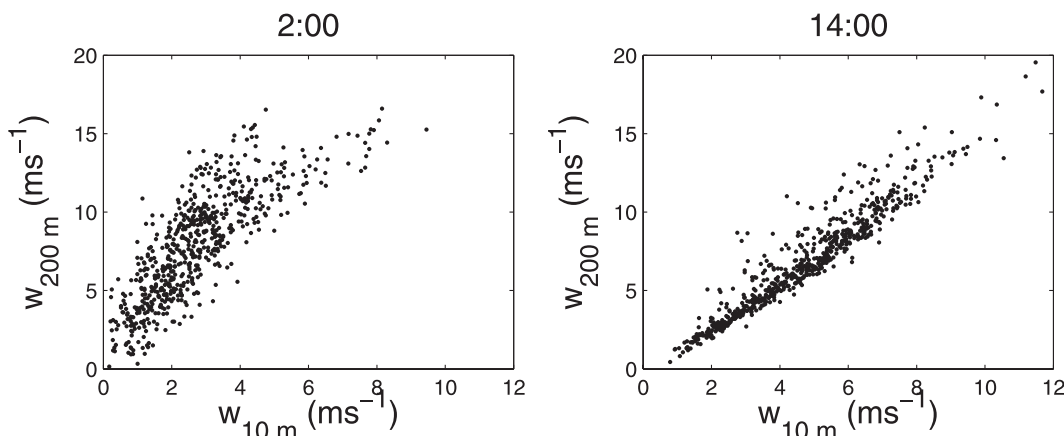


FIG. 8. Scatterplot of 10 and 200 m JAS wind speeds measured at Cabauw at (left) 0200 and (right) 1400.

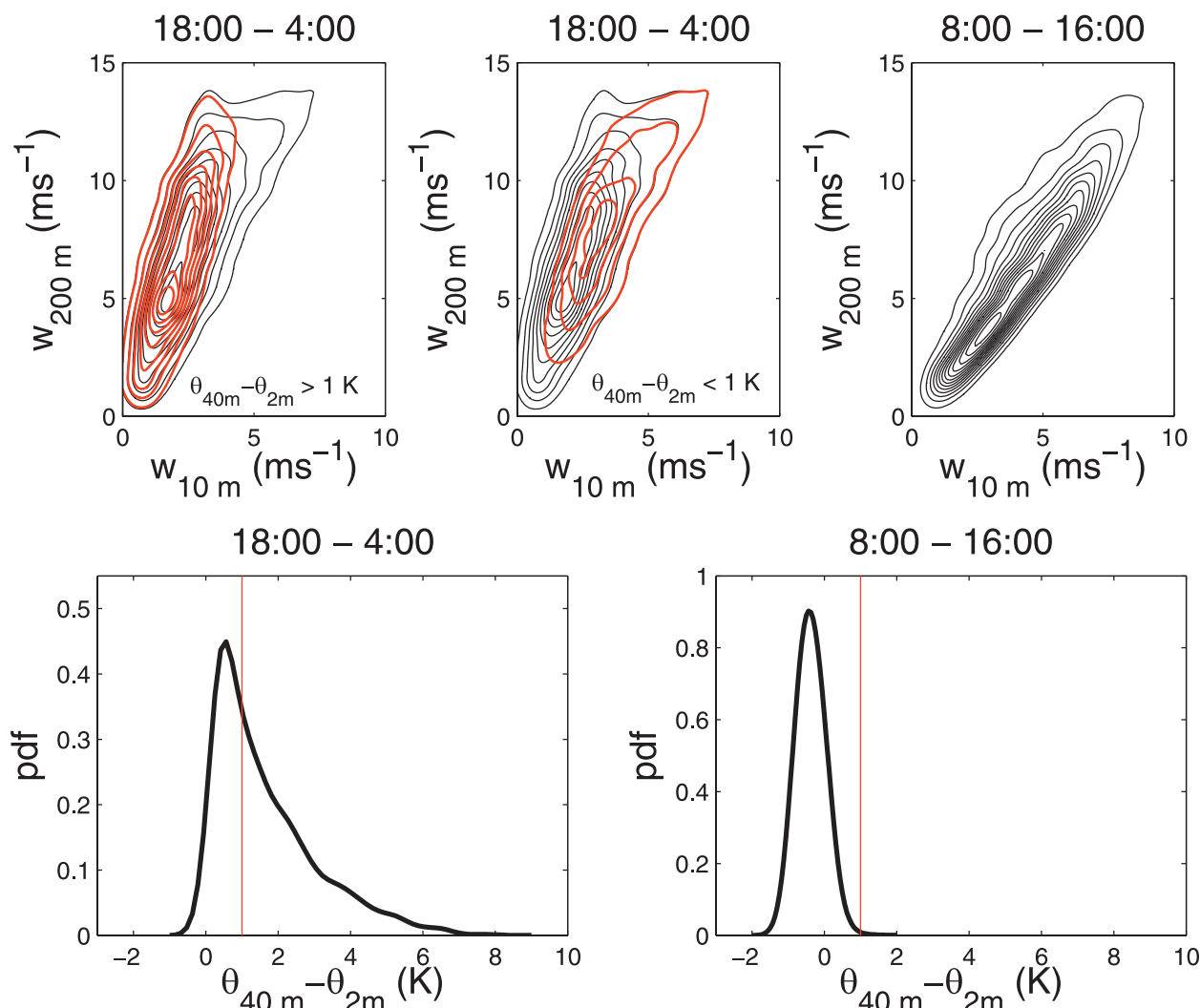


FIG. 9. (top) Kernel density estimates of joint pdfs of wind speed at 10 and 200 m for late summer (JAS) observations at Cabauw. Black contours in the first and second panels illustrate the joint pdf at night (between 1800 and 0400). Red contours in the first (second) panel illustrate the joint pdf for those hours when the lower-layer static stability $\theta_{40m} - \theta_{2m}$ is greater than (less than) 1 K. Black contours in the third panel are the joint pdf of wind speeds at 10 and 200 m for daytime hours (between 0800 and 1600). (bottom) The pdf of $\theta_{40m} - \theta_{2m}$ for nighttime and daytime hours. Red vertical line in the bottom two panels marks the 1-K threshold between (in our usage) strongly stable stratification and from weakly stable to unstable stratification; this threshold is used to separate the populations in the first and second panels.

populations emerges. Strongly stable stratification is characterized by a steep slope in the relationship between w_{12m} and w_{92m} , while the slope for weakly stable or unstable stratification is much like that of the daytime winds. While not as clearly separated as those at Cabauw, these two populations are also suggestive of two “modes of operation” of the nocturnal boundary layer.

As noted, the diurnal cycle in skewness excess in the Cabauw and Los Alamos observations is broadly consistent with that seen throughout the global extratropics (section 3). Furthermore, this diurnal cycle is concentrated in the lower 50 m of the atmosphere. As this is a very

shallow feature, it is not surprising that it is not well captured by global or regional climate models (e.g., He et al. 2010).

5. Idealized model of the diurnally varying boundary layer momentum budget

Data from the global network of surface meteorological stations illustrate that, although extratropical surface wind speeds are nearly Weibull under conditions favorable to the local generation of TKE (unstable surface stratification and/or strong lower-tropospheric wind

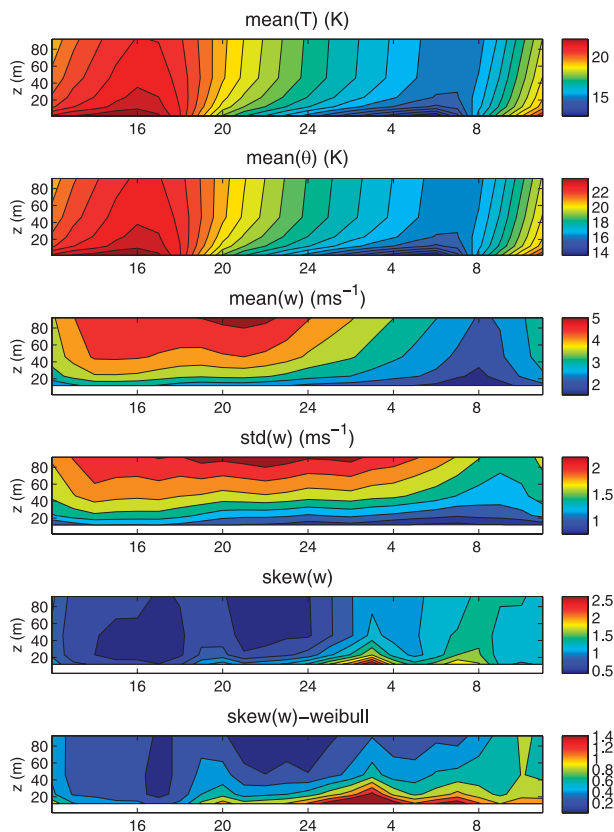


FIG. 10. As in Fig. 7 but for observations from the TA-6 tower near Los Alamos. Hours are in Mountain Standard Time (MST).

shear), under stably stratified conditions with weak wind shear the pdf of w has a tail toward large surface wind speeds longer than that expected of a Weibull-distributed variable. The tower observations considered in the previous section have demonstrated that the skewness excess evident in the diurnal evolution of the boundary layer is concentrated in the shallow nocturnal boundary layer. Furthermore, two populations are evident in the vertical wind structure at night, distinguished by the strength of the surface stratification. Under conditions of strongly stable stratification, the vertical wind speed shear is much greater than that during the daytime. Under weakly stable or unstable stratification, the vertical structure of wind speeds is much closer to that of the daytime. We will investigate the physical mechanisms responsible for these behaviors using an idealized model of the boundary layer momentum budget in which variations of the large-scale geostrophic wind are modeled stochastically (e.g., Monahan 2006a; He et al. 2010).

Following Thorpe and Guymer (1977), we assume that we can represent the daytime boundary layer as a single homogeneous slab of thickness h , while at night

the boundary layer is composed of two well-mixed layers: a stably stratified surface layer of thickness h_1 above which sits a residual layer of thickness $h_2 = h - h_1$. The boundary layer structure is assumed to switch abruptly from the daytime configuration to the nighttime configuration each day at $t = t_n$ (dusk) and back again at $t = t_d$ (dawn). For simplicity, it is assumed that day and night are both 12 h in duration. The assumption that we can represent the nocturnal boundary layer as two clearly distinct, homogeneous layers is a useful but highly idealized model simplification. These layer thicknesses may not be easily diagnosed from observations (e.g., Fernando and Weil 2010).

The momentum vector is assumed to be vertically uniform throughout any single model layer. During the day, the tendency of the horizontal momentum $\mathbf{U}$ (vertically averaged across the boundary layer) is

$$\frac{d\mathbf{U}}{dt} = \underbrace{-f\mathbf{k} \times [\mathbf{U} - \mathbf{U}_g(t)]}_A - \underbrace{\frac{w_e}{h} [\mathbf{U} - \mathbf{U}_g(t)]}_B - \underbrace{\frac{c_d}{h} |\mathbf{U}| \mathbf{U}}_C. \quad (1)$$

Term A represents the ageostrophic residual between the Coriolis and pressure gradient forces, expressed in terms of the geostrophic wind $\mathbf{U}_g$ (with f the Coriolis parameter). Term B represents the turbulent exchange of momentum between the boundary layer and the free atmosphere, represented in terms of a finite-differenced eddy diffusion with the entrainment velocity w_e . It is assumed that the geostrophic wind is a good representation of the lower-tropospheric winds just above the boundary layer. Implicit in the assumption of a constant boundary layer thickness h is the presence of subsidence or restratifying processes offsetting mixed layer growth through entrainment of free tropospheric air. Finally, term C parameterizes the surface turbulent momentum flux in standard Monin–Obukhov fashion in terms of the drag coefficient c_d . For simplicity, the dependence of the drag coefficient on surface wind speed and buoyancy fluxes is neglected. Furthermore, horizontal advection of momentum is neglected in this single-column approximation.

The geostrophic wind is modeled as a red-noise process with mean $\overline{\mathbf{U}}_g$ and isotropic uncorrelated fluctuations with standard deviation, $\text{std}(u) = \text{std}(v) = \sigma$, and autocorrelation time scale τ :

$$\mathbf{U}_g(t) = \overline{\mathbf{U}}_g + \boldsymbol{\eta}, \quad (2)$$

where

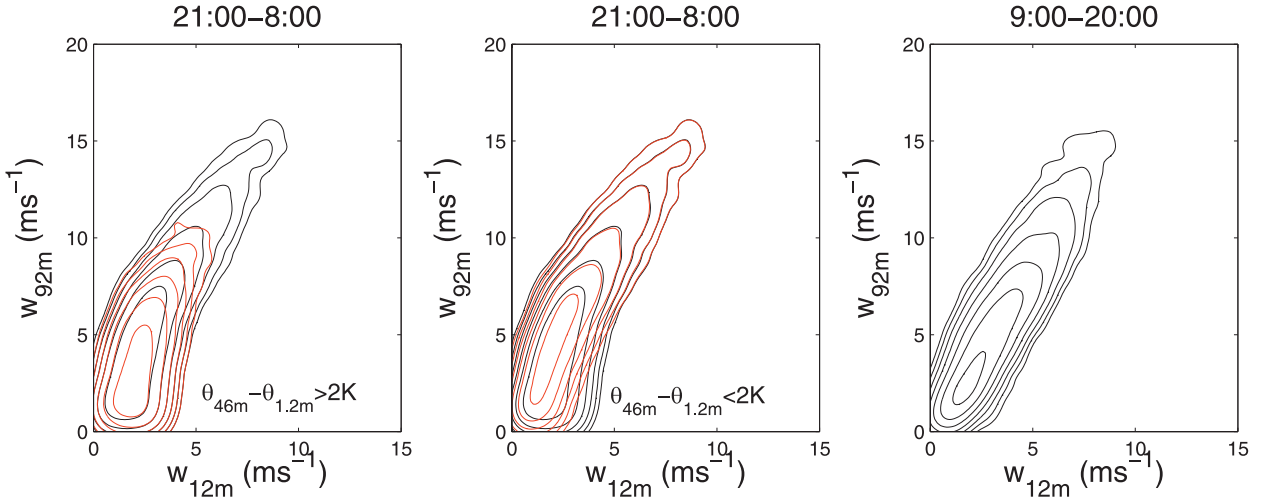


FIG. 11. Kernel density estimates of joint pdfs of wind speed at 12 and 92 m for late summer (JAS) observations at TA-6 near Los Alamos. Black contours in the first and second panels illustrate the joint pdf at night (from 2100 to 0800 MST). Red contours in the first (second) panel illustrate the joint pdf for those hours when the lower-layer static stability $\theta_{46m} - \theta_{1.2m}$ is greater than (less than) 2 K. Contours in the third panel illustrate the joint pdf of wind speeds at 12 and 92 m for daytime hours (from 0900 to 2000 MST).

$$\frac{d\eta}{dt} = -\frac{1}{\tau}\eta + \frac{\sigma}{\sqrt{2\tau}}\dot{\mathbf{N}}, \quad (3)$$

such that $\dot{\mathbf{N}}(t) = (\dot{N}_1(t), \dot{N}_2(t))$ are mutually uncorrelated white-noise processes:

$$\mathbf{E}\{N_i(t_1)N_j(t_2)\} = \delta_{ij}\delta(t_1 - t_2). \quad (4)$$

This representation was chosen for its simplicity and because it produces Gaussian geostrophic vector wind fluctuations (consistent with observed free-tropospheric vector wind variability; e.g., Petoukhov et al. 2008, as discussed in section 3). With this representation, the system of equations becomes a stochastic differential equation (SDE). Introductions to the theory of SDEs are presented in Gardiner (1997) and Penland (2003); Kloeden and Platen (1992) give an introduction to the numerical solution of SDEs. The parameters σ and τ are set at values representative of typical midlatitude synoptic-scale variability (Table 1).

At night, the wind vectors $\mathbf{U}_1$ and $\mathbf{U}_2$ (in the lower and upper layers, respectively) evolve separately, such that

$$\frac{d\mathbf{U}_1}{dt} = -f\mathbf{k} \times (\mathbf{U}_1 - \mathbf{U}_g(t)) - \frac{c_d}{h_1}|\mathbf{U}_1|\mathbf{U}_1 - \frac{\zeta(t)}{h_1}(\mathbf{U}_1 - \mathbf{U}_2) \quad (5)$$

and

$$\frac{d\mathbf{U}_2}{dt} = -f\mathbf{k} \times (\mathbf{U}_2 - \mathbf{U}_g(t)) - \frac{\zeta(t)}{h_2}(\mathbf{U}_2 - \mathbf{U}_1). \quad (6)$$

Although the lower layer is subject to surface drag, the upper layer does not exchange momentum turbulently with the free atmosphere since this layer is assumed to be well mixed but not actively turbulent. Turbulent momentum exchange between the two layers is described by a finite-differenced eddy diffusion with eddy entrainment velocity $\zeta(t)$. The matching conditions at t_n and t_d simply conserve depth-integrated momentum:

$$\mathbf{U}(t_d) = \frac{h_1}{h_1 + h_2}\mathbf{U}_1(t_d) + \frac{h_2}{h_1 + h_2}\mathbf{U}_2(t_d) \quad (7)$$

and

$$\mathbf{U}(t_n) = \mathbf{U}_1(t_n) = \mathbf{U}_2(t_n). \quad (8)$$

It has been observed that the turbulence intensity in strongly stratified boundary layers is intermittent with extended quiescent periods broken by episodes of intense turbulent stirring (e.g., Mahrt 1989, 1999; van de Wiel et al. 2003; Meillier et al. 2008). These periods of enhanced turbulence may result from one or more of a broad range of “submeso” processes (e.g., Mahrt 2010), including internal gravity waves (e.g., Meillier et al. 2008), density currents, and local feedbacks between boundary layer and radiative processes (e.g., van de Wiel et al. 2002). We model the turbulent mixing between the nocturnal surface layer and residual layer as being weak on average but occasionally becoming much stronger. In particular, we model the eddy entrainment velocity ζ as a random process.

TABLE 1. Idealized model parameters and standard values.

Parameter	Quantity	Value
f	Coriolis parameter	10^{-4} s^{-1}
$\overline{U}_g$	Mean zonal geostrophic wind	$1\text{--}10 \text{ m s}^{-1}$
$\overline{V}_g$	Mean meridional geostrophic wind	0 m s^{-1}
σ	Standard deviation of geostrophic wind	$1\text{--}5 \text{ m s}^{-1}$
τ	Geostrophic wind autocorrelation time scale	48 h
h_1	Depth of nocturnal stratified boundary layer	50 m
h_2	Depth of nocturnal residual mixed layer	200 m
c_d	Surface drag coefficient	7.5×10^{-3}
w_e	Daytime boundary layer top entrainment velocity	1 m s^{-1}
p	Probability a night will display intermittent turbulence	0.4
ζ_T	Entrainment velocity on turbulent night	0.015 m s^{-1}
ζ_0	Mean submesoscale entrainment velocity	$5 \times 10^{-3} \text{ m s}^{-1}$
n	Submesoscale intermittency parameter	2
τ_x	Submesoscale autocorrelation time scale	3 h

Two parameterizations of $\zeta(t)$ are considered. In the first, the entrainment velocity across the nocturnal boundary layer inversion is represented as a two-state random variable that (potentially) varies from night to night but is constant over a given night. At each dusk ($t = t_n$), ζ takes the value $\zeta_T > 0$ with probability p and the value 0 m s^{-1} with probability $1 - p$. In the standard simulation, $p = 0.4$ and $\zeta_T = 0.015 \text{ m s}^{-1}$. The second parameterization represents $\zeta(t)$ as the stationary stochastic process:

$$\zeta(t) = \zeta_0 \frac{|x|^n(t)}{\mathbf{E}\{|x|^n\}}, \quad (9)$$

where $x(t)$ is a Gaussian red-noise process with mean zero and unit variance

$$\frac{dx}{dt} = -\frac{1}{\tau_x}x + \sqrt{\frac{2}{\tau_x}}\dot{N}_3. \quad (10)$$

By construction, the stochastic process $\zeta(t)$ is non-negative, of mean ζ_0 , and positively skewed; the degree of skewness is determined by the exponent n . Larger values of n are associated with higher probabilities of weak mixing, such that most of the time $\zeta(t)/\zeta_0 \ll 1$. Since the mean of ζ is independent of n , larger values of n are associated with longer tails toward high values—that is, with burstier stochastic processes. The parameter τ_x is set so that “bursts” of ζ persist for a few hours. In representing the intermittent turbulence as a stochastic process, it is assumed that the mechanism driving intermittency involves physical processes or state variables not explicitly represented in the model, and that the time scale of intermittency is not very short compared to the wind adjustment time scale (in which case deterministic parameterizations would be appropriate;

e.g., Monahan and Culina 2011). These representations of ζ were selected to produce submeso entrainment velocities that are often small and occasionally large; they are not intended as quantitatively accurate, physically based representations of any individual physical process.

Although it is highly idealized, this model is still too complex to admit an analytic solution. Numerical simulations were carried out for values of $\overline{U}_{gx}$ between 0 and 10 m s^{-1} and σ between 0.5 and 5 m s^{-1} (with other parameters set at the values listed in Table 1). For each pair $(\overline{U}_g, \sigma)$, 50-yr-long realizations of the wind speeds w_j ,

$$w_j = \begin{cases} \|\mathbf{U}_j\|, & t_n < t < t_d \quad \text{“night”} \\ \|\mathbf{U}\|, & t_d < t < t_n \quad \text{“day”}, \end{cases} \quad (11)$$

were computed for both levels ($j = 1, 2$) using a forward Euler discretization (Kloeden and Platen 1992). From these, wind speed statistics (mean, standard deviation, and skewness) for simulations with $\zeta(t)$ modeled as a binary variable were computed from each realization at the model time points just before dawn (representing nighttime behavior) and just before dusk (representing daytime behavior) (Fig. 12). Over the range of simulated values of the ratio $\text{mean}(w)/\text{std}(w)$, the daytime skewness (asterisks) lies close to that of a Weibull-distributed variable (solid gray curve). At night (open circles), the simulated skewnesses are considerably larger than those of a Weibull variable. Qualitatively similar results were obtained with the parameterization of intermittent mixing given by Eq. (9). As well, the qualitative features of the results were insensitive to reasonable changes in model parameter values.

The variability of the nighttime eddy entrainment velocity is a necessary component of this model for

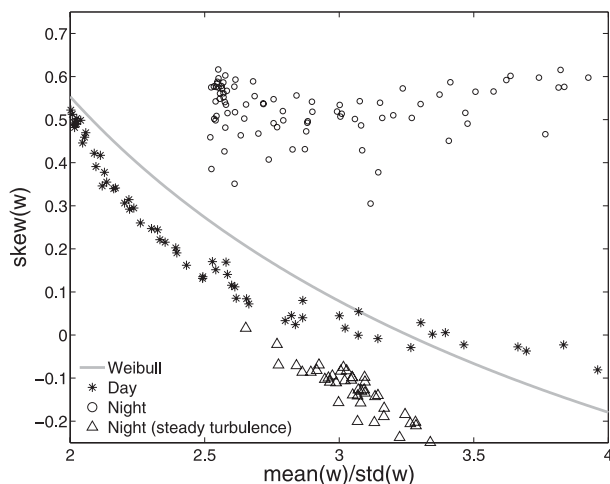


FIG. 12. Relationship between moments from the idealized model of the boundary layer momentum budget with intermittent mixing at the nocturnal boundary layer inversion modeled as a binary variable. Stars denote moments for a particular time during the day (just before dusk), while open circles denote moments for a particular time during the night (just before dawn). Triangles denote the nighttime moments when the intermittent mixing $\zeta(t)$ is set at its constant mean value.

simulating the enhanced nighttime surface wind speed skewnesses. When $\zeta(t)$ is replaced by its constant mean value ζ_0 , the skewnesses of the simulated surface winds are lower than those of a Weibull-distributed variable (Fig. 12, open triangles). In this model, the enhancement of wind speed skewness is a consequence of

the intermittency of the mixing at the nocturnal boundary layer inversion.

A simulation for which the mean and standard deviation of w_1 were similar to those observed at Cabauw was obtained with $\overline{U}_g = 3 \text{ m s}^{-1}$ and $\sigma = 3 \text{ m s}^{-1}$. The joint pdf of w_1 and w_2 in daytime (Fig. 13, third panel) is by model construction a straight line with a slope of 1. At night the joint pdf (Fig. 13, first and second panels) shows two populations: one with a strong wind speed shear between the two levels and the other with a weaker wind speed shear more similar to that observed during the day. When partitioned between those nights without turbulent exchange between the layers and those nights with it, these two populations are cleanly separated (as illustrated by the red contours in the first and second panels of Fig. 13). The separation of the nighttime winds into two populations is also predicted by the alternate parameterization of intermittent turbulence given by Eq. (9). That the qualitative features of the model predictions are independent of which of the two parameterizations of intermittent turbulent mixing are used suggests that the simulation of nocturnal surface wind speeds with strongly positive skewness and two distinct populations is more sensitive to the presence or absence of a representation of intermittency of turbulent mixing than to the precise details of the representation.

While the predictions of this model are in broad agreement with the observations discussed in sections 3 and 4, this agreement is not perfect. The two nighttime

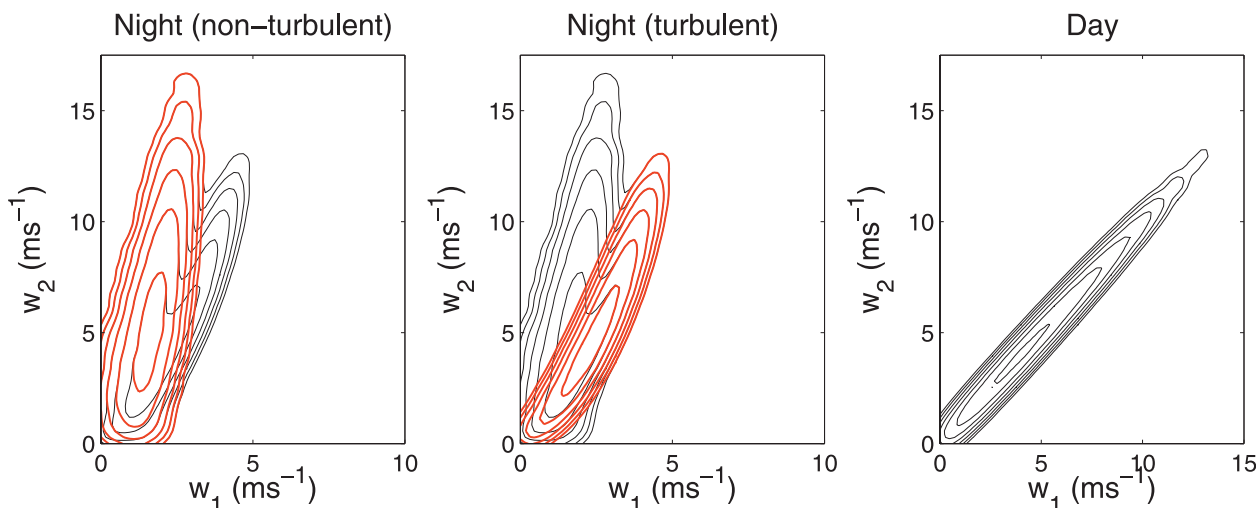


FIG. 13. Kernel density estimates of joints pdfs of wind speeds in lower and upper model layers for the idealized model of the boundary layer momentum budget, with $\overline{U}_g = 3 \text{ m s}^{-1}$, $\sigma = 3 \text{ m s}^{-1}$, and the parameterization of $\zeta(t)$ as a binary variable. Black contour in the first and second panels illustrates the joint pdf for a nighttime point (just before dawn), while that in the third panel illustrates the joint pdf for a daytime point (just before dusk). Red contours in the first and second panels illustrate the joint pdfs on nights that are nonturbulent and turbulent, respectively.

populations are more cleanly separated in the model than in the observations. Furthermore, the difference in shear between turbulently mixing nights and daytime is greater in the model than in observations (the observed daytime winds demonstrate a nonzero vertical wind speed shear that by construction is absent in this model). However, the qualitative agreement between model predictions and observations demonstrates the plausibility of the proposed mechanism for the dependence of the wind speed pdf on boundary layer processes.

6. Discussion and conclusions

The probability distribution of surface wind speeds is a mathematical object of importance for estimating and simulating surface fluxes, for understanding surface extremes, and for characterizing the distribution of wind power in present and future climates. Although there is a long history of empirical studies of this distribution, only recently have there been systematic efforts to understand the physical mechanisms that determine this distribution (e.g., Monahan 2004, 2006a; He et al. 2010). The present study has focused on the pdf of land surface winds. Observational data from a large array of surface meteorological stations were used to characterize the seasonal and diurnal evolution of the surface wind speed pdf and to investigate the dependence of the pdf structure on local surface stratification and the strength of the winds aloft. The vertical structure of the surface wind speed pdf was then investigated using long-term measurements from two tall meteorological towers (one at Cabauw, Netherlands, and the other near Los Alamos, New Mexico). Finally, the results of the observational analysis were interpreted in terms of an idealized model of the boundary layer momentum budget.

This study reached the following conclusions.

- The diurnal cycle in the shape of the pdf of surface wind speeds, first noted over North America in He et al. (2010)—that, while daytime wind speeds follow the Weibull distribution to a good approximation, the nighttime winds are more positively skewed—is a feature of winds across the global network of stations. This diurnal cycle is most pronounced in summer, least pronounced in winter, and absent for free-tropospheric winds at 700 hPa.
- When partitioned based on local lower free tropospheric (700 hPa) wind speed and local surface stability, there is a tendency for the wind speed skewness excess (observed skewness less that of a Weibull variable) to be strongly positive under conditions of weak winds aloft and strongly stable surface stratification. Under conditions of strong winds aloft or unstable stratification,

the surface wind speed pdf is closer to Weibull. This relationship is most pronounced in the extratropics. In the tropics, the dependence of the skewness excess on lower-tropospheric wind shear is similar, but the stability dependence is less pronounced.

- Observations from tall meteorological towers demonstrate that the skewness excess at night is localized to a shallow surface layer (a few tens of meters deep) within the stably stratified nocturnal boundary layer. Above this layer, or during the day, the wind speed skewness excess is generally close to zero. This result suggests that the inability of the RCMs considered in He et al. (2010) to simulate the diurnal cycle in the shape of the wind speed pdf is linked to their difficulties in simulating the shallow nocturnal boundary layer.
- The vertical structure of wind speed at nighttime can be separated into two populations. In one of these, associated with strongly stable stratification, the wind speeds above the inversion capping the nocturnal boundary layer are much larger than those at the surface. In the other, associated with weakly stable or unstable surface stratification, the relationship between wind speeds aloft and those at the surface is more similar to that observed during daytime. While these populations do not correspond to cleanly separated “climate regimes,” they do provide a useful classification scheme for the connection between boundary layer stratification and the vertical structure of wind speeds.
- An idealized model of the boundary layer momentum budget with a crude representation of the diurnal cycle in stratification can account for the nocturnal skewness excess in surface winds and the two populations in the vertical structure of the nocturnal wind speed. A component of this model essential to reproducing these results is the parameterization of intermittent turbulence at the inversion capping the nocturnal boundary layer as a “bursty” stochastic process. The qualitative behavior of the model is not sensitive to different formulations of this parameterization: what is important is not how the intermittent turbulence is modeled but, rather, that it is modeled.

Again, we emphasize that our focus on the Weibull distribution in this study is not because of any basic physical claim it has to describe wind speed fluctuations. Rather, as the standard parametric distribution used in studies of surface wind speed variability (e.g., Petersen et al. 1998a,b; Capps and Zender 2008), it serves as a useful benchmark. Other parametric forms of the wind speed pdf have been suggested (e.g., Li and Li 2005); as the present study is focused on observations of the diurnal cycle of the shape of the wind speed distribution

and consideration of its underlying mechanism, these alternate parametric representations have not been considered.

Because the model of the boundary layer momentum budget considered in this study is highly idealized, the qualitative features of its predictions are more important than the quantitative details. More quantitatively precise analyses of the physical mechanisms driving the diurnal cycle in the shape of the land surface wind speed pdf require more comprehensive models. Using a single-column model (SCM) developed from the Canadian Centre for Climate Modeling and Analysis third-generation atmospheric general circulation model (Scinocca et al. 2008), we find that simulation of this diurnal cycle requires sufficiently fine vertical resolution to resolve the nocturnal boundary layer and a careful parameterization of the Richardson number dependence of mixing processes under stably stratified conditions. Furthermore, inclusion of a stochastic parameterization of intermittent turbulence in stable layers is needed to reproduce the positive wind speed skewness excess at night. This companion study, which will be reported on in a later publication, indicates that the relationship between intermittent turbulent mixing and nocturnal wind speed extremes is not simply an artifact of the idealized model considered here.

As a result of the nature of the data considered in this study, the physical mechanism responsible for producing the wind speed skewness excess in strongly stably stratified conditions was inferred indirectly. In particular, the importance of intermittent turbulence around the nocturnal boundary layer inversion in producing these surface wind speed extremes remains a hypothesis. We have not presented direct evidence of a causal relationship between enhanced turbulence (as measured by, e.g., an increase in temperature variance on turbulent time scales) and surface wind speed extremes. A focused investigation of the relationship between turbulence intermittency and surface wind speed extremes is an important direction of further study. An improved understanding of this relationship would also inform physically based parameterizations of intermittent turbulent mixing by “submeso” processes (e.g., Mahrt 2010) and improve upon the ad hoc formulations considered in this study and in the companion SCM study.

The tower data and idealized model considered in this study provide insight into the physical mechanisms influencing the shape of the wind speed pdf in the extratropics. The surface wind observations indicate that these mechanisms may be somewhat different in the tropics. As tall tower data (of the necessary temporal resolution and duration) are not readily available in the tropics, the degree of similarity between the diurnal evolution of tropical

land boundary layers and those observed at Cabauw and Los Alamos is unknown. Furthermore, while the neglect of moist processes in the idealized model is a reasonable first-order approximation in the extratropics, the accuracy of this approximation cannot be expected to be as good in the tropics. In particular, the difference between temperatures at the ground and at 2 m is not likely an adequate measure of stability in the tropics. Future work will consider the physical controls on the tropical land surface wind speed pdf in more detail, with a particular focus on moist processes.

In the context of large-scale wind power generation, the winds at some tens of meters above the ground are more relevant than those at 10 m (e.g., Archer and Jacobson 2005). Future work will use tower observations and SCM simulations to more thoroughly investigate the pdf of wind speeds at altitudes typical of turbine hub heights and the physical influence of boundary layer processes on the wind resource.

An understanding of the character of surface wind variability is increasingly important because GCMs increasingly focus, not just on surface fluxes of energy, momentum, and moisture, but also of gases and aerosols (e.g., Wanninkhof et al. 2002; Cakmur et al. 2004). Also, as a result of their social and economic costs, there is increasing interest in surface wind extremes (e.g., Gastineau and Soden 2009). Interest in the potential of wind power to offset energy production by combustion of fossil fuels is also increasing: an understanding of the variability of surface winds is central to assessing the wind power resource in present and future climates. This study presents results from—what is to our knowledge—the first global-scale investigation of the physical controls on the probability distribution of surface winds over land, representing a first step in bringing a mechanistic and process-oriented perspective to bear on this quantity of basic geophysical importance.

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APPENDIX

Empirical Wind Speed Probability Distributions

A Weibull-distributed variable w has the pdf

$$p_w(w) = \begin{cases} \frac{b}{a} \left(\frac{w}{a}\right)^{b-1} \exp\left[-\left(\frac{w}{a}\right)^b\right], & w > 0 \\ 0, & w \leq 0, \end{cases} \quad (\text{A1})$$

where the shape parameter b determines the skewness of the distribution and the scale parameter a is the primary factor in determining its mean. For a Weibull-distributed variable,

$$\text{mean}(w^k) = a^k \Gamma\left(1 + \frac{k}{b}\right), \quad (\text{A2})$$

where $\Gamma(y)$ is the gamma function. For a Weibull-distributed variable, the skewness

$$\text{skew}(w) = \text{mean}\left(\left[\frac{w - \text{mean}(w)}{\text{std}(w)}\right]^3\right) \quad (\text{A3})$$

is given to a very good approximation by

$$\text{skew}(w) = S\left(\left[\frac{\text{mean}(w)}{\text{std}(w)}\right]^{1.086}\right) \quad (\text{A4})$$

in which

$$S(x) = \frac{\Gamma\left(1 + \frac{3}{x}\right) - 3\Gamma\left(1 + \frac{1}{x}\right)\Gamma\left(1 + \frac{2}{x}\right) + 2\Gamma^3\left(1 + \frac{1}{x}\right)}{\left[\Gamma\left(1 + \frac{2}{x}\right) - \Gamma^2\left(1 + \frac{1}{x}\right)\right]^{3/2}}. \quad (\text{A5})$$

The Weibull skewness is positive for small values of the ratio $\text{mean}(w)/\text{std}(w)$, near zero for values of the ratio ~ 3 –4, and negative for larger values (Fig. A1). In fact, that the skewness of a Weibull variable is a unique function of the ratio of the mean to the standard deviation can be used as a qualitative measure of the degree to which pdf of observed wind speeds is Weibull. A thorough discussion of the Weibull distribution, its properties, and its sampling variability is presented in Monahan (2006a).

The special case of the Weibull distribution with $b = 2$, the Rayleigh distribution, corresponds to the distribution of wind speeds assuming that the vector wind fluctuations are Gaussian isotropic, uncorrelated, and of mean zero. In the more general case that the mean vector wind has magnitude $\bar{U}$, the joint distribution of the horizontal wind components is

$$p_{uv}(u, v) = \frac{1}{2\pi\sigma^2} \exp\left(-\frac{(u - \bar{U})^2 + v^2}{2\sigma^2}\right) \quad (\text{A6})$$

(with u and v the along- and cross-mean wind components, respectively); the wind speed pdf is

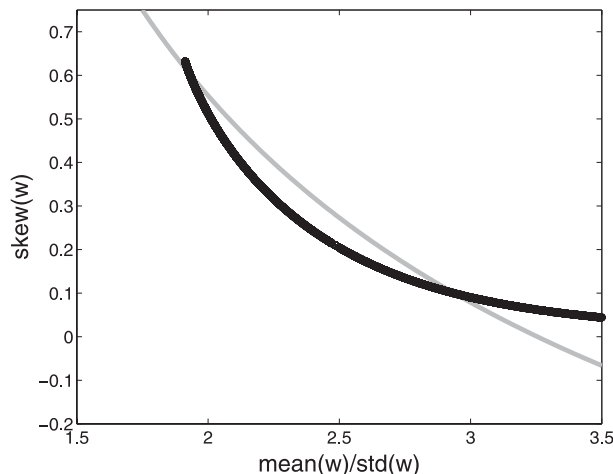


FIG. A1. Relationships between moments for wind speeds modeled as a Weibull variable (gray curve) and for speeds computed assuming the vector winds are bivariate Gaussian with uncorrelated, isotropic fluctuations (black curve). In both cases, the wind speed skewness is a decreasing function of the ratio of the mean to the standard deviation [Eq. (18)]. Black curve was computed for a range of mean vector wind magnitudes $\bar{U}$ between 0 and 10 m s⁻¹ and a range of standard deviations σ between 0.25 and 6 m s⁻¹.

$$p_w(w) = I_0\left(\frac{w\bar{U}}{\sigma^2}\right) \frac{w}{\sigma^2} \exp\left[-\frac{1}{2\sigma^2}(w^2 + \bar{U}^2)\right], \quad (\text{A7})$$

where σ is the standard deviation of the vector wind components; and I_0 is the modified Bessel function of order zero (Narovlyanskii 1970; Cakmur et al. 2004; Monahan 2006a, 2007). As with the Weibull distribution, the skewness of this distribution is a decreasing function of the ratio of the mean to the standard deviation (Fig. A1), although, in contrast to the Weibull distribution, the skewness does not become negative. Over the range of values of $\text{mean}(w)/\text{std}(w)$ between about 2 and 3.5, both the Weibull distribution and that given by Eq. (7) are associated with similarly skewed distributions.

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