Global Ocean Wind Power Sensitivity to Surface Layer Stability

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Global ocean wind power has recently been assessed [Liu et al., 2008] using scatterometry-based 10 m winds. We characterize, for the first time, wind power at 80 m (typical wind turbine hub height) above the global ocean surface, and account for the effects of surface layer stability. Accounting for realistic turbine height and atmospheric stability increases mean global ocean wind power by +58% and −4%, respectively. Our best estimate of mean global ocean wind power is 841 W m$^{-2}$, about 55% greater than the 545 W m$^{-2}$ based on previous methods. 80 m wind power is 1.2–1.5 times 10 m power equatorward of 30$^\circ$ latitude, between 1.4–1.7 times 10 m power in wintertime storm track regions and ∼6 times 10 m power in stable regimes east of continents. These results are relatively insensitive to methodology as wind power calculated using a fitted Weibull probability density function is within 10% of power calculated from discrete wind speed measurements over most of the global oceans.

1. Introduction

Surface wind speed distributions are important for many applications including surface energy and constituent flux, shipping hazards, and wind power assessment. Previous wind power studies range from region [Pimenta et al., 2008] and site-specific [Shata and Hanitsch, 2008] to global land-based [Archer and Jacobson, 2005] and ocean [Liu et al., 2008] surveys. Liu et al. [2008] assessed global ocean wind power using scatterometry-based 10 m winds that assume a neutral wind profile. As shown below, this significantly underestimates wind power at a typical wind turbine hub height (80 m) for two reasons: height and stability. Our purpose is to characterize, for the first time, the global ocean 80 m wind power and to account for and isolate the effects of height and surface layer stability that together increase wind power by a factor of ∼1.5 at 80 m relative to 10 m. Therefore this study, in conjunction with Archer and Jacobson [2005], comprises a global 80 m wind power assessment.

Monin-Obukhov similarity theory (MOST) can be applied within the marine atmospheric surface layer [Edson and Fairall, 1998] to estimate deviations to the logarithmic wind profile due to thermal stratification [Lange and Focken, 2005]. Although MOST applies to conditions which rarely exist, i.e., a horizontally homogeneous surface layer [Arya, 2001], similarity functions have been developed empirically which are generally suitable for practical applications. Thermodynamic, wind speed data (Section 2) and methods (Section 3) used to compute 80 m winds are followed by 80 m wind speed and power global (Section 4.1) and regional analyses (Section 4.2).

2. Data

Without collocated atmospheric sounding observations, vertical wind speed profile estimation given 10 m neutral-stability wind speeds requires surface layer thermodynamic measurements including surface sensible ($H_0$) and latent heat flux ($L_0$), 2 m air temperature ($T_a$) and 2 m specific humidity ($q_a$).

2.1. SeaWinds on QuikSCAT

We use the 7 year (Jan./2000–Dec./2006) Level 3 reprocessed 0.25° × 0.25° QuikSCAT 10 m wind speed dataset available from the Physical Oceanography Distributed Active Archive Center. QuikSCAT uses an empirical algorithm to relate backscatter generated by capillary waves to surface stress. 10 m surface winds (approx. 0600 and 1800 local time) are inferred from these stress measurements by assuming a neutrally-stable atmosphere [Liu, 2002; Liu et al., 2008]. This assumption [Hoffman and Leidner, 2005] introduces a bias during non-neutral conditions. Mears et al. [2001] and Chelton and Freilich [2005] found that 10 m anemometer winds are typically 0.2 m s$^{-1}$ slower than in-situ 10 m neutral-stability winds. Wind vector cells containing the possibility of rain or less than 50% of the timeseries were not included in this study.

2.2. Objectively Analyzed Air-Sea Fluxes

Surface layer thermodynamic data is provided by the Woods Hole Oceanographic Institution third version of global ocean-surface heat flux products released by the Objectively Analyzed air-sea Heat Fluxes (OAFUX) project [Yu et al., 2008]. Bulk aerodynamic formula physical variables originate from an optimal blend of reanalysis data and satellite measurements. These variables are improved through the use of a variational objective analysis technique. Errors for each variable are estimated using in-situ measurements including moored buoys and ship observations. OAFUX surface energy fluxes are computed using the TOGA COARE bulk flux algorithm 3.0 [Fairall et al., 2003]. We bi-linearly interpolate daily OAFUX $H_0$, $L_0$, $T_a$ and $q_a$ from 1.6° × 1.0° to match QuikSCAT spatial resolution.

2.3. NCEP–DOE AMIP-II Reanalysis

NCEP–DOE AMIP-II reanalysis data were provided by the NOAA/OAR/ESRL PSD, Boulder, Colorado, USA, (http://www.cdc.noaa.gov/) [Kanamitsu et al., 2002]. NCEP 2.5° × 2.5° daily mean sea level pressure (MSLP) used to calculate air density is regressed to QuikSCAT spatial resolution. NCEP $H_0$, $L_0$, $T_a$ and $q_a$ are regressed from T62 to QuikSCAT spatial resolution and are substituted where OAFUX data are missing.

2.4. Global Ocean Bathymetry

The one arc-minute global ocean bathymetry and relief dataset from NOAA’s National Geophysical Data Center [Amante and Eakins, 2008] contours coastal regions where depth is less than 700 m.
3. Methods

3.1. Extrapolation to 80 m

Wind speeds at each QuikSCAT overflight time are extrapolated to 80 m using the equation

\[ u_{80} = \frac{u_0}{k} \left[ \ln \left( \frac{z}{z_{0m}} \right) - \psi_m \left( \frac{z}{L} \right) \right] \] (1)

where \( z = 80 \) m, \( u_{80} \) is 80 m wind speed, \( k \) is the Von Karman constant, \( \psi_m \) is a dimensionless stability function. The momentum roughness length \( (z_{0m}) \) is calculated following Charnock [1955] with a Charnock parameter of 1.1 \( \times 10^{-2} \) [Smith, 1980]. Friction velocity \( (u_0) \) is a function of QuikSCAT 10 m wind speed and a neutral drag coefficient [Large et al., 1994]. The Monin-Obukhov length \( (L) \) over the ocean is

\[ L = \frac{u_0^2 \theta_v}{g \bar{w}_v \theta_{v0}} \] (2)

where \( \theta_v \) is the virtual potential temperature and \( \bar{w}_v \) is the surface buoyancy flux

\[ \bar{w}_v \theta_{v0} = \frac{H_0}{\rho_u} + 0.61 T_0 \frac{L_0}{\rho_u} \] (3)

where \( c_p \) is the specific heat at constant pressure and \( T_0 \) is the latent heat of vaporization of water (e.g., Arya [2001]). Air density \( (\rho) \) is calculated using NCEP II reanalysis daily MSLP and virtual temperature derived from OAFLUX data. Following Lange and Focken [2005], stability functions used for unstable and slightly stable cases are

\[ \psi_m \left( \frac{z}{L} \right) = 2 \ln \left( \frac{1 + x}{2} \right) + \ln \left( \frac{1 + x^2}{2} \right) - 2 \arctan(x) \] (4)

+ \frac{\pi}{2}, \text{ for } \frac{z}{L} < 0

\[ \psi_m \left( \frac{z}{L} \right) = -5 \frac{z}{L}, \text{ for } 0 < \frac{z}{L} < 0.5 \] (5)

where \( x = \left[ 1 - 16 \left( \frac{z}{L} \right) \right]^{1/4} \). For very stable conditions the Holtslag and de Bruin [1988] stability function is used

\[ \psi_m \left( \frac{z}{L} \right) = -0.7 \frac{z}{L} - \left[ 0.75 \frac{z}{L} - 10.72 \exp(-0.35 \frac{z}{L}) \right] \]

\[ -10.72, \text{ for } \frac{z}{L} > 0.5 \] (6)

Figure ?? shows wind speed profiles calculated using the methods described above. For faster surface layer wind speeds, mechanical mixing exceeds buoyant mixing. Thus, 80 – 10 m wind speed differences become relatively small across all static stabilities compared to slower speeds.

3.2. 80 m Wind Power Density

The power per unit area using discrete QuikSCAT measurements is proportional to the wind speed cubed and air density

\[ E/A = \frac{1}{2n} \sum_{i=1}^{n} \rho_i u_{80i}^3 \] (7)

where \( A \) is the area swept by the rotors and \( n \) is the number of wind speed observations. Air density \( (\rho) \) is calculated daily and is extrapolated to 80 m using the U.S. standard atmosphere profile.

The cubic relationship between wind power and speed warrants a continuous wind speed distribution for accurately determining wind power. Following Barthelmie and Pryor [2003]; Liu et al. [2008], wind power per unit area is proportional to the third moment of the Weibull PDF and air density, denoted hereafter as \( E_{out}/A \). Scale and shape parameters are calculated from the 80 m height wind speed timeseries [Monahan, 2006; Capps and Zender, 2008].

We estimate wind power using both discrete QuikSCAT measurements (7) and the third moment of a fitted Weibull PDF. Although the two-parameter Weibull PDF closely approximates the observed surface wind speed distribution [Justus et al., 1979; Pavia and O’Brien, 1986], surface winds have non-Weibull characteristics, particularly in the global distribution of skewness [Monahan, 2006]. A Weibull variable has less (more) negative skewness within the tropics (storm tracks) than QuikSCAT data. Thus, \( E_{out}/A \) is 2–6% larger (2–10% smaller) than \( E/A \) within the tropics (northern hemisphere (NH) storm tracks).

4. Results

4.1. Global Oceans

Due to enhanced vertical mixing of momentum, the DJF wind profile is sub-logarithmic (80 m most winds are between 0.5–1.5 m s⁻¹ slower than log profile (80 m log) winds,
Figure 2) and 80–10 m wind speed differences are minimized (0.8–1.5 m s$^{-1}$, not shown) over the northwestern Pacific and Atlantic oceans. Large spatial wind power density (hereafter referring to $E_{pdf}/A$, following Liu et al. [2008]) maxima in excess of 3000 W m$^{-2}$ (1.4–1.6 times 10 m wind power densities) are collocated with these sub-logarithmic wind profiles (Figure 3). DJF super-logarithmic profiles with collocated 80 – 10 m speed differences of 1.6 to 2.1 m s$^{-1}$ are found in the northeastern Atlantic and Pacific oceans where relatively warmer air is advected northward over cold eastern boundary currents east of developing cyclones. 80 m wind power (1500 to 2200 W m$^{-2}$) is 1.6–1.7 times 10 m wind power in these regions.

As expected, wind power densities within the southern hemisphere storm track region vary less interseasonally compared to the NH. Summertime wind speed profiles in both hemispheres between 40$^\circ$ and 70$^\circ$ latitude are super-logarithmic with large 80 – 10 m wind speed differences (> 3.6 m s$^{-1}$) and relatively low wind power densities. Summertime 80 m wind power is 1.6–3.0 times 10 m power over most of these latitudes with cold current regions exceeding 6 times 10 m power (not shown). Seasonally persistent positive 80 – 10 m (not shown) and 80m$_{MOST}$ – 80m$_{log}$ (Figure 2) wind speed differences reside near the Agulhas Return Current region. These coincide with isolated regions of year-round negative sensible heat fluxes due to the deflection of the circumpolar current around the Kerguelen Plateau and other nearby bathymetric features [O’Neill et al., 2005]. Additionally, seasonally persistent reduced 80 – 10 m and 80m$_{MOST}$ – 80m$_{log}$ wind speed differences occupy the Agulhas retroreflection south of Cape Town, South Africa.

A striking feature during the NH summer within the Indian ocean is the Somali Jet. JJA sea surface temperatures (SSTs) in the western Arabian Sea are lowered by evaporative cooling and upwelling in response to the onset of the Somali Jet [Halpern and Woiceshyn, 1999]. Consistent with large interseasonal variability of wind speed characteristics [Capps and Zender, 2008], interseasonal swings in stability regimes, 80 – 10 m and 80m$_{MOST}$ – 80m$_{log}$ wind speed differences and $E_{pdf}/A$ rival those found near the Oyashio and Labrador currents.

With the exception of the eastern Pacific and Atlantic cold tongue regions, the sensible heat flux in the tropics is predominantly just above zero with a muted seasonal cycle (Yu and Weller [2007]). The equatorial flanks of the subtropical highs are characterized with seasonally persistent slightly unstable surface layers and sub-logarithmic wind speed profiles (Figure 2). Swaths of higher power densities (500 – 700 W m$^{-2}$) trace the trade wind regions bounded to the north and south by the lower wind power regions of the horse latitudes and Intertropical Convergence Zone (Figure 3), respectively. 80 m wind power is 1.2 – 1.5 times 10 m wind power equatorward of 30$^\circ$S and 30$^\circ$N latitude (not shown).

Uncertainties in our estimates arise from systematic errors and uncertainties in the surface flux and wind speed data. OAFLUX data has a negligible mean bias with mean absolute uncertainties of 6% and 12% for latent and sensible heat fluxes, respectively, while QuikSCAT has a mean bias of 0.1 m s$^{-1}$ with uncertainties of 1 m s$^{-1}$. The uncertainties in OAFLUX (QuikSCAT) data could translate into wind power uncertainties for 2000–2006 of < 10% (about 10%, for fast wind regions).

4.2. Eastern North America

Increased wintertime vertical momentum transfer over the Gulf Stream and North Atlantic Current results in sub-logarithmic wind profiles, reduced 80 – 10 m wind speed differences and smaller 80 m and 10 m power density differences (Figure 4). Consistent with 10 m winds [Liu et al.,
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5. Conclusions

We estimate the surface layer wind profile over the global oceans using both Monin-Obukhov similarity theory (MOST) and the logarithmic assumption. Year-round slightly sub-logarithmic profile regions between 40°N and 40°S latitude are offset by moderate and extreme super-logarithmic summertime profiles poleward of 40° and east of continents, respectively. Further, most of the southern hemisphere circumpolar region is characterized with year-round logarithmic to super-logarithmic wind profiles. Thus, 2000–2006 global ocean mean 80 m wind power from both methods are nearly equal (863 W m\(^{-2}\) log vs. 841 W m\(^{-2}\) MOST).

Wind power is computed using two methods: discrete twice daily QuikSCAT measurements, and a fitted Weibull PDF. Differences between the two methods are less than 10% for most regions. Spatial patterns of these discrepancies match patterns of QuikSCAT non-Weibull characteristics.

2000–2006 mean global ocean wind power is 841 W m\(^{-2}\), about 55% greater than QuikSCAT 10 m power (545 W m\(^{-2}\)). Regions of high 80 m wind power coincide with high 10 m power regions within the wintertime storm tracks. However, 80 m wind power for these regions is between 1.4–1.7 times 10 m power. Storm track region summer wind speed profiles in both hemispheres are super-logarithmic with 80 m wind power between 1.6–3.0 times 10 m power. Stable surface layers such as those found over the waters east of continents during summer have much lower 10 m power estimates compared to 80 m power estimates. Further, 80 m wind speeds extrapolated using MOST for these stable regions may be underestimated [Lange et al., 2004b]. Regions such as these with insufficient 10 m power may contain enough power to warrant 80 m hub height turbine placement.

Acknowledgments. Level 3 QuikSCAT data were obtained from NASA (http://podaac.jpl.nasa.gov). WHOI third version of global ocean surface heat flux product was obtained from http://oaflux.whoi.edu/. Supported by NSF IIS-0431203, ARC-0714088, and NASA NNX07AR23G. We thank an anonymous reviewer for helpful suggestions.
Figure 4. 2000–2006 DJF (left) and JJA (right) 80 – 10 m wind speed difference (m s$^{-1}$, top), MOST 80 m wind speed minus logarithmic 80 m wind speed (m s$^{-1}$, 2nd row), 80 m wind power density from full Weibull distribution (W m$^{-2}$, 3rd row) and 80 m multiple of 10 m wind power (bottom). Bathymetric contours at $-700$ and $-500$ in magenta. Ta minus SST contours in black (positive contours solid/negative dashed). Regions where > 50% of timeseries is missing are omitted.

References


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