NOTES

Statistics on Vertical Wind Shear over Oceans

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ABSTRACT

Statistics on boundary layer vertical wind shear were gathered from rawinsonde soundings taken from three small islands and one weather ship. These soundings show a high correlation between surface and 1829 m altitude wind directions. Wind speeds also were vertically correlated except at one tropical station where the winds were light. The directional results suggest that cloud motions could be used with instruments which measure only scalar surface wind speed (such as radar altimeters or microwave radiometers) to infer vector direction and sea surface stress.

1. Introduction

Cloud motion winds have been considered as proxy data for surface winds over oceans because of the lack of other surface observations. The tropical oceans and Southern Hemisphere suffer greatly from the lack of observations because most of the routes of merchant ships are in the Northern Hemisphere between industrial cities. Cloud motions, on the other hand, are an abundant data source because of the numerous lower atmosphere cumuli over the oceans. Cloud motion observations are routinely made from five geostationary satellites around the world, operated by four different countries, which cover all of the earth from 50°N to 50°S latitude.

Comparisons of satellite cloud motions with aircraft wind measurements have shown that the low level cumuli in the tropics move with the winds at altitudes of the bases, usually 920–900 mb (1–2 km), rather than the winds at their tops as might be expected (Hasler *et al.*, 1979). This may occur because satellite images can only resolve groups of clouds. The small individual cumuli are smaller than the resolutions of satellite cameras (1–8 km) in most cases. The group motions appear to be correlated with boundary layer motions that generate the clouds. This suggests that satellite cloud motion data can be combined with other surface based data to improve wind analyses in data sparse areas.

Most wind information other than clouds over oceans is from surface measurements, merchant ships, or the satellite microwave and radar sensors. The microwave and radar sensors respond chiefly to the surface roughness changes caused by the winds. Thus, boundary layer vertical shear information is needed to combine the cloud motions with other wind data.

There have been several studies[2] of the boundary layer vertical shear or the differences between cloud motions. Radiosonde data from island stations have been extensively compiled by Mendenhall (1967) and Gray (1972) for studying the tropical ocean boundary layer and its structure near convective storms. Regional studies have been made by Enfield (1981) in the eastern Pacific and Schott and Fernandez-Partagas (1981) in the Indian Ocean. Comparisons of cloud motions to buoy wind observations have been made in the eastern Pacific by Halpern (1979) and Halpern and Knox (1982), but the largest studies to date have been the ship–cloud comparisons of Sadler and Kilonsky (1981) and Wylie *et al.* (1983) covering most of the tropical Pacific Ocean and Wylie and Hinton (1981, 1982) in the Indian Ocean. These latter studies provided the basis for making combined wind analyses from ship and cloud data.

While the literature on this subject is growing, it does not cover all geographical regions and meteorological conditions. The previous reports give mostly average wind shear statistics for monthly or seasonal ensembles of data. The daily variability is seldom discussed.

A small study of daily variability of wind shear was made from four rawinsonde stations in the fall of 1978 (Table 1). The object of this study was to assess the magnitude of the boundary layer shear problem under a wide range of weather conditions and latitudes. Radiosonde stations indicative of

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[2] A convenient summary of oceanic boundary layer wind studies can be found in Glackin, D. L., 1982: Wind veering in the planetary boundary layer over the ocean. Jet Propulsion Laboratory, Task Plan Rep. 80-1685, Pasadena, CA 91109, 18 pp.

TABLE 1. Station locations.

Name	Latitude (°N)	Longitude (°W)	ID number
St. Paul Island	57	170	70308
Swan Island	17	84	78501
Station P	50	145	99021
Bermuda	32	65	78016

oceanic conditions were chosen. Because these stations were on isolated islands and one weather ship, accompanying measurements of horizontal pressure and temperature gradients were not possible. This lack of data prevented us from studying the dynamical causes of the vertical wind shear. What is shown from the radiosonde wind data is the magnitude of the daily variance in the wind shears under typical oceanic conditions. This information sheds light on the use of cloud motion, or other wind data aloft, to infer surface winds in the absence of data on temperature gradients or thickness fields.

Our discussion centers on the direction and speed components of the wind, rather than the vector zonal and meridional components. This division was made because some satellite sensors do not measure wind direction. We feel that directions can be statistically analyzed from speeds as long as the range in direction is small; that is, the directions oscillate over a hemisphere and do not spin continuously in one direction. Vector components may be mathematically correct, but they don't reflect the methods of measuring winds. Passive microwave radiometers and radar altimeters provide only scalar speed data. For complete analyses of wind fields, directional data will have to be obtained from other sources such as cloud motions or pressure fields. Some *in situ* sensors also measure directions separately from speeds, such as wind vane and anemometer combinations or the Beaufort estimates used on ships.

The radiosonde data discussed here show that the vertical coupling of cloud layer winds to the surface is different for direction as opposed to speed. This information must be considered in the design of algorithms for making composite wind analyses from a variety of data sources.

2. The data

Rawinsonde stations were chosen from three small islands and one ship (Table 1) that regularly reported over the FAA 604 line. Swan Island, located in the Gulf of Mexico, is a very small island with little geographical relief. It was consistently under the easterly trade winds during the period studied. Other tropical and semi-tropical conditions were found at Bermuda in the central Atlantic. High-latitude weather conditions were sampled by Ocean Weather Station (P), a Canadian observing ship, and St. Paul Island, an Air Force station in the Aleutian Island chain of Alaska.

The twice daily soundings from these stations were archived on a semi-regular schedule during the five-day working week from September through November 1978. The wind shear statistics from the surface to 850 mb or 1829 m were extracted from each sounding as reported without any corrections except for St. Paul Island, where northerly and westerly wind conditions had to be discarded because of geographical obstructions upwind of the rawinsonde station. A few northeasterly wind conditions on Bermuda were also discarded for the same reason.

3. Vertical shear statistics

The winds at Swan Island were predominantly from the east or southeast (see Table 2) with fluctuations in speed that were smaller than the other stations. The zonal and meridional components of the wind exhibited standard deviations of only ±2.4 and ±3.3 m s^{-1}, respectively, at the surface level (see Table 3). In contrast to these, much larger wind component fluctuations from ±6.4 to ±7.4 m s^{-1} were found at OWS-P and St. Paul Island. It should be noted that the average wind at OWS-P (4.8 m s^{-1}) was significantly larger than either Swan Island (3.4 m s^{-1}) or St. Paul Island, which was very light (1.4 m s^{-1}). The winds at Bermuda were also light (1.5 m s^{-1}), with moderate fluctuations (±4.6 to ±5.5 m s^{-1}) as found at St. Paul.

The veering of the wind direction, which we define as the direction above the surface (850 mb or other altitude) minus the surface direction, was nearly the same for all four stations studied. It ranged from 12.9° at Bermuda to 15.9° at OWS-P. These values are similar to those reported by Gray (1972) although they are approximately 4° larger than the zonal averages of Gray (1972) for tropical locations. They are also similar to measurements of surface wind veering from the geostrophic wind found over land (Arya, 1975).

The 850 mb wind direction had high correlations with the surface (0.90–0.97, Table 2). This indicates that 81% of the variance in surface wind direction was predicted by the 850 mb direction, since the fractional explained variance is the square of the correlation coefficient. The standard deviations of the veering appear to be large (±26° to ±35°) in comparison to the mean. But it should be noted that range or standard deviation of the measured wind directions was much larger (±69° to ±107°, or two directional quadrants).

Some of the variance in veering could be a result of changes in the "thermal wind," the vertical geostrophic wind shear caused by horizontal temperature gradients (Wiin-Nielsen, 1974). This would be ex-

TABLE 2. Veering between the surface and the 850 mb level. S.D. denotes standard deviation or root-mean-square deviation from the mean.

Station	Number of samples (n)	Pressure level P (mb)	Reported direction Mean ± S.D. (deg)	Reported direction Correlation* of surface value	Veering [$\alpha = \Theta(Z) - \Theta(\mathrm{sfc})$] Mean ± S.D. (deg)
78016	112	sfc	205 ± 99		
Bermuda	112	850	192 ± 103	0.94	12.9 ± 34
99021	79	sfc	256 ± 69		
OWS-P	79	850	240 ± 78	0.90	15.9 ± 35
70308	100	sfc	182 ± 104		
St. Paul	100	850	170 ± 107	0.97	12.1 ± 26
78501	123	sfc	122 ± 86		
Swan Is.	123	850	107 ± 86	0.92	13.9 ± 35

* The correlation between wind direction measurements at 850 mb with the surface.

pected, especially at the high latitude stations OWS-P and St. Paul Island. However, because of the isolated nature of these stations, which were chosen to approximate overwater conditions, horizontal temperature gradients could not be evaluated.

Changes in vertical atmospheric stability also may have caused some vertical veering changes (Brown and Liu, 1982). The temperature profile data from the radiosondes indicated nearly neutral stability conditions for most of the profiles. These profiles are very coarse in their vertical resolution, and without water temperature data from the island stations, the stability changes were difficult to detect.

Some of the variance in veering and variance in wind measurements in general could be a result of the radiosondes making point measurements in a turbulent wind field. Each single radiosonde ascent does not measure the spatially averaged wind field, but rather selective points in this field. Brown (1983) has estimated the error in a single ship wind direction measurement to be around 20° from the spatially averaged field. A similar error may be present in the radiosonde data.

A more detailed vertical description of the vertical wind veering for St. Paul and Swan Islands is given in Table 4. The correlation with the surface decreases slowly with height for direction to 0.91 at 1829 m. However, for the scalar wind speeds, the correlation decreased more rapidly to 0.79 at St. Paul and 0.22 at Swan Island at 1829 m. It appears that the scalar speeds have more of a random component that is not as vertically related as the directions.

The surface wind speeds and directions also were predicted using a simple linear regression with one independent variable from an altitude above the surface, i.e.,

$$X_{\mathrm{SFC}} = a + bX(H), \qquad (1)$$

TABLE 3. Relation of the zonal and meridional wind components of the surface to the 850 mb level. Dispersion is the root-mean-square deviation from the prediction (based on a regression equation) of surface values by 850 mb values. S.D. denotes standard deviation or root-mean-square-deviation from the mean.

Station	Number of samples (n)	Altitude (mb)	Zonal component (u) Mean ± S.D. (m s^{-1})	Zonal component (u) Correlation	Zonal component (u) Dispersion of surface prediction (m s^{-1})	Meridional component (v) Mean ± S.D. (m s^{-1})	Meridional component (v) Correlation	Meridional component (v) Dispersion of surface prediction (m s^{-1})
78016	112	sfc	1.36 ± 4.6			−0.71 ± 5.5		
	112	850	6.44 ± 6.7	.87	2.8	−0.39 ± 11.2	.84	2.5
99021	79	sfc	4.66 ± 6.4			−1.27 ± 6.5		
	79	850	7.52 ± 7.6	.85	3.5	−0.52 ± 7.5	.81	3.8
70308	100	sfc	−1.33 ± 6.0			0.39 ± 7.4		
	100	850	−0.91 ± 9.1	.90	2.6	−1.55 ± 10.6	.93	2.8
78501	123	sfc	−3.29 ± 2.4			0.71 ± 3.3		
	123	850	−5.08 ± 4.1	.56	2.0	−0.38 ± 3.7	.61	2.6

TABLE 4. Velocity data for St. Paul Island and Swan Island. The columns labelled dispersion indicate the rms deviation of the value predicted by a linear regression equation of the form $a + bz$ from actual data values. Therefore, dispersion is representative of the errors we would make in an estimate of speed or direction. S.D. (standard deviation) refers to the rms deviation from the sample mean. When dispersion is significantly less than the standard deviation, the estimate shows skill. The final column on the right is the veering calculated from the regression equation, using the mean value of the direction, and is intended to represent a typical or characteristic value of veering.

			Speed (V)			Reported direction (Θ)			Veering [$\alpha = \Theta(Z) - \Theta(\mathrm{sfc})$]	
Station	Number of samples (n)	Altitude (m)	Mean ± S.D. ($\mathrm{m\ s^{-1}}$)	Correlation to surface	Dispersion ($\mathrm{m\ s^{-1}}$)	Mean ± S.D. (deg)	Correlation to surface	Dispersion (deg)	Regression equation	From mean directions
70308	43	0	9.4 ± 5.0	1.00	0	159 ± 115	1.00	0	0	0.0
		304	12.1 ± 6.7	0.87	2.5	167 ± 112	0.99	14	$9.58 - 0.0135\Theta(Z)$	7.3
		609	11.9 ± 7.0	0.90	2.2	173 ± 109	0.98	25	$19.06 - 0.0281\Theta(Z)$	14.2
		914	11.7 ± 7.2	0.86	2.6	173 ± 107	0.97	28	$19.93 - 0.0366\Theta(Z)$	13.6
		1219	11.7 ± 7.6	0.84	2.8	171 ± 104	0.97	26	$23.76 - 0.0717\Theta(Z)$	11.5
		1829	11.6 ± 7.8	0.79	3.1	165 ± 119	0.95	35	$-7.54 + 0.0776\Theta(Z)$	5.3
78501	52	0	5.6 ± 2.5	1.00	0	111 ± 90	1.00	0	0	0.0
		304	7.4 ± 2.3	0.62	2.0	114 ± 90	0.98	18	$2.01 - 0.0153\Theta(Z)$	3.8
		609	7.7 ± 2.6	0.49	2.2	119 ± 92	0.97	22	$2.50 - 0.0439\Theta(Z)$	7.7
		914	7.3 ± 2.9	0.38	2.3	126 ± 91	0.97	24	$10.43 - 0.0371\Theta(Z)$	18.1
		1219	7.1 ± 3.0	0.28	2.4	130 ± 87	0.96	27	$17.90 - 0.0089\Theta(Z)$	19.1
		1829	7.0 ± 3.1	0.22	3.1	131 ± 84	0.91	37	$17.19 - 0.0188\Theta(Z)$	19.6

where X represents either speed or direction at height H. The coefficients a and b were evaluated for each altitude (H = 304, 609, 914, 1219, 1829 m, or the 850 mb level) and variable independently. We call the standard deviation of the difference between the predicted surface (X_{SFC}) and the measured ($X(0)$) value, the dispersion;

$$\mathrm{Dispersion} = \left[\frac{1}{N-1}\sum (X_{\mathrm{SFC}} - X(0))^2\right]^{1/2}. \quad (2)$$

Thus, the dispersion is the root-mean-square (rms) error in the prediction of surface speed or direction from 850 mb, 304, 609, 914, 1219, or 1829 m wind data.

The dispersion, in all cases (see Tables 3 and 4), was markedly less than the standard deviation of the variable, indicating skill in predicting surface winds. This was most evident for directions where the dispersion ranged from ±14° to ±37°, while the standard deviation (S.D.) of direction in the data sample ranged from ±90° to ±115°.

The scalar speeds did not have as much success in predicting surface observations. From 1829 m the surface speeds could be predicted to $\pm 3.1\ \mathrm{m\ s^{-1}}$ (Table 4). At St. Paul this was near one-half of the surface S.D. in speed, while at Swan Island it was the same magnitude as the S.D. Correlation coefficients indicate that 62% of the speed variance at St. Paul was predicted from 1829 m, while almost none of the variance was predicted at Swan Island.

Dispersion calculations on the zonal and meridional wind components (Table 3) also show the same trends. The dispersion was small for the stations with light winds at Swan Island and Bermuda (± 2.0 to $\pm 2.8\ \mathrm{m\ s^{-1}}$), but this was a large fraction of the S.D. of the components (± 2.4 to $\pm 5.5\ \mathrm{m\ s^{-1}}$). At OWS-P, where the strongest winds were observed, the dispersion ($\pm 3.8\ \mathrm{m\ s^{-1}}$) was a smaller fraction of the S.D. ($\pm 6.5\ \mathrm{m\ s^{-1}}$).

4. Discussion and conclusions

The vertical shear statistics indicate that cloud motions are better predictors of the surface wind direction than scalar speed in the sense that vertical correlation of direction is stronger than that of speed. Wind direction is more vertically coupled than wind speed in the boundary layer.

A useful technique for extrapolating cloud directions to the surface may be possible even if corrections for shear by vertical and horizontal temperature gradients cannot be made. However, improvements in sensing sea surface and air temperatures by satellites may make useful thermal boundary layer corrections to the wind direction possible in the future.

The statistics also indicate a problem in predicting surface speeds under light wind conditions, as evidenced at Swan Island. A previous study by Hasse and Wagner (1971) indicated that surface wind speeds also have a poor relationship to the geostrophic wind when speeds are light.

These findings have strong implications for future wind analyses where data from several sensors will be combined. The winds from passive microwave radiometers and radar altimeter sensors on satellites do not measure direction, as previously mentioned. Records several years long of GEOS-3 altimeter data and Nimbus-7 SMMR data are available for surface wind speed, and more data will be available from the microwave sensor to be flown on the Defense Meteorological Satellite Program (DMSP) and GEOSAT in the next few years. Our results suggest that with the help of cloud motions, these could be turned into vector winds or stress. The scatterometers planned for future satellites (NROSS and ERS) will have directional information, but with possible ambiguities. Cloud motions can be exploited for providing information where these sensors are deficient.

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REFERENCES

Arya, S. P. S., 1975: Geostrophic drag and heat transfer relations for the atmospheric boundary layer. *Quart. J. Roy. Meteor. Soc.*, **101,** 147–161.

Brown, R. A., 1983: On a satellite scatterometer as an anemometer. *J. Geophys. Res.*, **88,** 1663–1673.

——, and W. T. Liu, 1982: An operational large-scale marine planetary boundary layer model. *J. Appl. Meteor.*, **21,** 261–269.

Enfield, D. B., 1981: Annual and nonseasonal variability of monthly low-level wind fields over the southeastern tropical Pacific. *Mon. Wea. Rev.*, **109,** 2177–2190.

Gray, W. M., 1972: A diagnostic study of the planetary boundary layer over oceans. Atm. Sci. Pap. No. 179, Colorado State University, 95 pp.

Halpern, D., 1979: Surface wind measurements and low-level cloud motion vectors near the Intertropical Convergence Zone in the central Pacific Ocean from November 1977 to March 1978. *Mon. Wea. Rev.*, **107,** 1525–1534.

——, and R. A. Knox, 1982: Coherence between low-level cloud motion vectors and surface wind measurements near 0°, 152°W from April 1979–February 1980, *Atmos. Ocean,* **21,** 82–93.

Hasler, A. F., W. C. Skillman, W. E. Shenk, and J. Steranka, 1979: *In Situ* aircraft verification of the quality of satellite cloud winds over oceanic regions. *J. Appl. Meteor.*, **18,** 1481–1489.

Hasse, L., and V. Wagner, 1971: On the relationship between geostrophic and surface wind at sea. *Mon. Wea. Rev.*, **99,** 255–260.

Mendenhall, B. R., 1967: A statistical study of frictional wind veering in the planetary boundary layer. Atm. Sci. Pap. No. 116, Colorado State University, 57 pp.

Sadler, J. C., and B. J. Kilonsky, 1981: Trade wind monitoring using satellite observations. Dept. of Meteor., University of Hawaii Rep. UMET 81-01, 72 pp.

Schott, F. and J. Fernandez-Partagas, 1981: The onset of the summer monsoon during the FGGE 1979 experiment off the east African coast: A comparison of wind data collected by different means. *J. Geophys. Res.*, **86,** 4173–4180.

Wiin-Nielsen, A., 1974: Vorticity, divergence, and vertical velocity in a baroclinic boundary layer with a linear variation of the geostrophic wind. *Bound.-Layer Meteor.* **6,** 459–476.

Wylie, D. P., and B. B. Hinton, 1981: The feasibility of estimating large-scale surface wind fields for the summer MONEX using cloud motion and ship data. *Bound.-Layer Meteor.*, **21,** 357–367.

——, and ——, 1982: A comparison of cloud motion and ship wind observations over the Indian Ocean for the year of FGGE. *Bound.-Layer Meteor.*, **23,** 197–208.

——, —— and Howland, M., 1983: Surface wind observations from cloud motions during FGGE. *Tro. Ocean-Atmos. Newslett.*, No. 18 (May).