

Estimation of Low-Frequency Wind Stress Fluctuations over the Open Ocean

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ABSTRACT

A simple, approximate formula for mean wind stress is given in terms of the mean and variance of the wind fluctuations over the averaging period. The formula is nonlinear with respect to the mean wind speed.

The formula is tested using 3 h wind observations from eight North Atlantic Ocean Weather Ships. Mean wind stress is calculated 1) by vector averaging the 3 h wind stresses, and 2) by applying the approximate formula. For an averaging period of 4 months the two methods agree to within ± 0.025 Pa, 95% of the time. For an averaging period of 1 month the approximate formula slightly overestimates the stress. This is due to skewness in the probability density function of the observed 3 h wind fluctuations. An expression for the modification of the mean stress due to skewness is given.

A straightforward method is described for the estimation of vector mean wind and variance fields, and thus mean stress fields, over the open ocean. To check the method, the long-term stress field of the North Atlantic, and the seasonal Sverdrup transport across 31°N , are computed and compared with the values given by Willebrand, and Bunker and Leetma. Good agreement is obtained. The zonally integrated Sverdrup transport across 45°N is also calculated and shown to exhibit significant interannual fluctuations.

1. Introduction

Wind stress is of fundamental importance to the study of ocean circulation and the associated fluxes of heat, nutrients, etc. Although the long term mean stress field over the World's oceans has now been relatively well defined (Bunker, 1976), the determination of slow fluctuations about this mean state is hampered by the paucity of wind data over the open ocean. The estimation of such slow stress variations is the subject of this paper.

A method is presented here for the estimation of mean stress fields from mean air pressure maps. The method is mainly based on a recent result obtained by Wright and Thompson (1983) which expresses a time-averaged stress in terms of the mean and variance of the wind fluctuations occurring within the averaging period. The result is of greater generality than the linearized quadratic stress law derived by assuming that the fluctuations dominate the mean.

It is shown that the spatial and temporal distribution of the wind variance over the North Atlantic is dominated by a seasonal cycle with a geographically dependent amplitude. However, this cycle and thus the wind variance for any given month may be readily determined from seasonal wind statistics published,

for example, in the U.S. Navy Marine Climatic Atlas (1974). Thus, the problem of determining the mean stress field for a given period is simplified to the estimation of the mean wind field.

It is well known that air pressure maps can provide good estimates of the large scale features of the wind field by applying an empirically determined contraction and rotation to the geostrophic wind vector. Such an approach has been taken in this paper. In this respect the stress estimation approach is similar to that suggested by Fofonoff (1962). The novel feature of the present method however is that the high-frequency wind fluctuations have been explicitly taken into account thus leading to a time-averaged stress formula which is both seasonally and geographically dependent. It is hoped that this method may prove useful in the estimation of slow stress fluctuations over those periods, or ocean areas, for which there are insufficient wind data to compute the stresses directly.

2. Approximate formula and applicability to wind data

Wind stress averaged over a period in the range 1 h–2 days is most simply determined from a bulk aerodynamic formula of the form

$$\tau = \rho_a c_d (|\mathbf{v}|) |\mathbf{v}|, \quad (1)$$

where τ is the surface wind stress, ρ_a is the density of air, \mathbf{v} is the mean wind (usually recorded at 10 m) and c_d is an empirically determined drag coefficient.

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Recent measurements of the drag coefficient show a significant dependence on wind speed. Large and Pond (1981), for example, suggest the following form for c_d over the open ocean

$$10^3 c_d = \begin{cases} 1.14, & 4 < |v| \leq 10 \text{ m s}^{-1} \\ 0.49 + 0.065|v|, & 10 < |v| < 26 \text{ m s}^{-1}. \end{cases} \quad (2)$$

This is the form used throughout the rest of this paper. It agrees well with the results obtained by Smith (1980) using an independent set of data; the increase of c_d with wind speed is also consistent with the theoretical work of Charnock (1955). The influence of stability has not been considered as its effect on the mean wind stress over the open ocean is expected to be small (Fissel *et al.*, 1977).

It has long been recognized that a stress law of the form (1) will significantly underestimate the stress if the winds are averaged over a period exceeding about 2 days; account must be taken of the fluctuations occurring within the averaging period. Following Wright and Thompson (1983), this underestimation can be quantified by first expressing the wind vector as

$$v = v_0 + v_1,$$

where suffix 0 and 1 denote the mean and fluctuating components respectively. The mean, or expected, stress τ_0 is then given by

$$\tau_0 = \rho_a \iint_{-\infty}^{\infty} c_d(|v|)|v|vP(u_1, v_1)du_1dv_1, \quad (3)$$

where P is the probability density function (pdf) of the fluctuating winds. Wright and Thompson (1983) have shown that if P is assumed to be an isotropic Gaussian pdf with variance σ^2 , i.e.,

$$P = P_g = \frac{1}{2\pi\sigma^2} \exp\left[-\frac{1}{2\sigma^2}|v_1|^2\right], \quad (4)$$

then τ_0 may be approximated by $\hat{\tau}_0$, where

$$\left. \begin{aligned} \hat{\tau}_0 &= \rho_a c_d(a) a v_0 \\ a^2 &= |v_0|^2 + 4\sigma^2 \end{aligned} \right\} \quad (5)$$

If $|v_0|$ and σ are assumed to be in the ranges (0–15) m s^{-1} and (4–11) m s^{-1} , respectively (see next section for justification, then the two estimates, $|\hat{\tau}_0|$ and $|\tau_0|$, agree to within 4%; with the further restriction that $|\tau_0| > 0.1 \text{ Pa}$, the relative error in $|\hat{\tau}_0|$ is less than 2%. Thus the approximation of (3) by (5) is good.

The relationship between $|v_0|$, σ and $|\tau_0|$, according to (2), (3) and (4) is shown in Fig. 1. Clearly the stress will be seriously underestimated if just the mean wind speed is substituted in (1), i.e., $\sigma = 0$ in (5). It is also apparent from Fig. 1 that, for the values of $|v_0|$ and σ of interest here, the time-averaged stress law is significantly nonlinear in $|v_0|$.

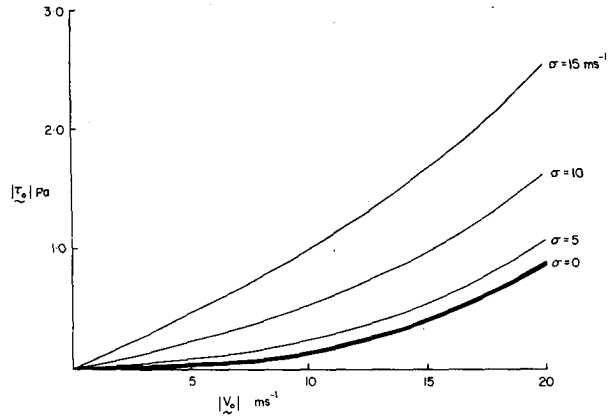


FIG. 1. Mean stress $|\tau_0|$ as a function of $|v_0|$ and σ , according to (2), (3) and (4).

The attraction of (5) lies in the fact that only knowledge of the mean and variance of the wind over a given period is required to determine the mean stress. The applicability of (5) to the determination of mean stress over the open ocean has been tested using 3 h wind observations from eight North Atlantic Ocean Weather Ships (OWS). (See Fig. 2 for OWS positions.) Monthly stresses were first calculated by vector averaging the 3 h stresses determined from (1). This is the most laborious yet most accurate way to determine τ_0 if sufficient wind data are available. Mean stresses were also calculated by substituting monthly estimates of the mean wind and variance in (5). The variance σ^2 was calculated each month from the average of the wind variances in the east and north directions. The two sets of monthly stress estimates, τ_0 and $\hat{\tau}_0$ respectively, were then compared to test the usefulness of (5).

The agreement in the directions of τ_0 and $\hat{\tau}_0$ is good, with over 90% of the direction differences being less than 15° for $|v_0| > 2.5 \text{ m s}^{-1}$ and $|\tau_0| > 0.05 \text{ Pa}$. The magnitudes of τ_0 and $\hat{\tau}_0$ agree to within $\pm 0.05 \text{ Pa}$, 90% of the time (Fig. 3). However, there is a tendency for (5) to slightly overestimate the monthly stresses.

The reason for the stress overestimation is probably related to a skewness of P , in the direction of v_0 . Skewness at OWS C is shown in Fig. 4. It is a function of both $|v_0|$ and the averaging period over which it was determined. To calculate the effect of such an asymmetry on the accuracy of (5), the approach used by Wright and Thompson (1983) has been extended to determine τ_0 for the following skewed pdf:

$$P(u_1, v_1) = P_g(u_1, v_1) \left\{ 1 + \frac{\nu_1}{6} \left[\left(\frac{u_1}{\sigma} \right)^3 - \frac{3u_1}{\sigma} \right] \right\}, \quad (6)$$

where ν_1 is the skewness [i.e. the expected value of $(u_1/\sigma)^3$]. Note that the mean wind has been taken, without loss of generality, along the x -axis. (The

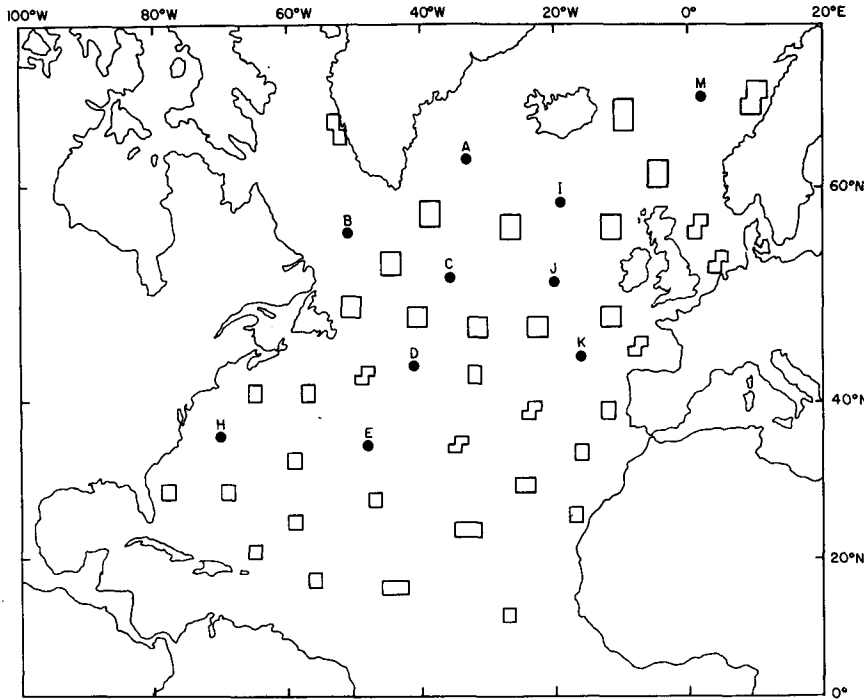


FIG. 2. Positions of the 10 OWS (dots) and 36 additional areas where σ_s was estimated.

above probability density function is a truncated Gram-Charlier expansion). The influence of the skewness on the mean stress can then be incorporated in (5) by adding the term

$$\tau_0^c = \frac{\rho_a \nu_1}{6} \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} c_d(|\mathbf{v}|) |\mathbf{v}| \nu P_g(u_1, v_1) \times \left[\left(\frac{u_1}{\sigma} \right)^3 - \frac{3u_1}{\sigma} \right] du_1 dv_1. \quad (7)$$

Over the range of $|\mathbf{v}_0|$, ν_1 and σ of interest here, it was found that

$$\tau_0^c \approx (0.94 \times 10^{-4} \nu_1 \sigma^3, 0) \text{ Pa}, \quad (8)$$

where σ is in m s^{-1} , with a relative error less than 8%.

During winter, the monthly mean wind speed is typically 7 m s^{-1} (hence $\nu_1 \approx -0.4$ from Fig. 4) and $\sigma \approx 9 \text{ m s}^{-1}$. Thus (5) gives a stress of 0.262 Pa and (8) suggests that this is an overestimate of 9%. During summer, typical values of $|\mathbf{v}_0|$ and σ are 3 and 5 m s^{-1} , respectively, and thus $\nu_1 = -0.1$ from Fig. 4. Eq. 5 gives a stress of 0.045 Pa , and (7) suggests that this is a 3% overestimate. Both results are consistent with the bias at OWS C evident in Fig. 3. Skewness is almost halved as the averaging period is increased from 1 to 4 months (Fig. 4). Thus, according to the above explanation, the overestimation of stress would also be expected to decrease. This is indeed found to be the case as shown by the scattergraph of Fig. 5.

For an averaging period of 4 months, (5) provides an unbiased stress estimate which is accurate to within $\pm 0.025 \text{ Pa}$, 95% of the time.

3. Stress estimation method

It has been shown that, given \mathbf{v}_0 and σ for an averaging period of 4 months or longer, (5) can provide accurate estimates of mean wind stress. Thus, the problem of obtaining stress fields has been reduced to the determination of these two quantities.

Monthly variations of σ and $|\mathbf{v}_0|$ at OWS C are shown in Fig. 6. Apart from a seasonal variation common to both time series, no obvious relationship is evident between them. (Similar results were obtained for the other OWS). Thus to estimate mean stress fields for the North Atlantic, the mean wind velocity and variance fields must be determined separately. A practical method for this is given below, and the resulting stress fields are compared with independently obtained estimates.

a. Estimation of σ

The most striking feature of the temporal fluctuations of σ at the eight OWS is a well defined seasonal variation which has an amplitude of about 1.5 m s^{-1} and reaches a peak in January (Fig. 6, Table 1). By splitting the monthly changes of σ into an OWS-dependent, seasonal component (σ_s , given in Table 1) and an anomaly component (σ_a where $\sigma = \sigma_s + \sigma_a$),

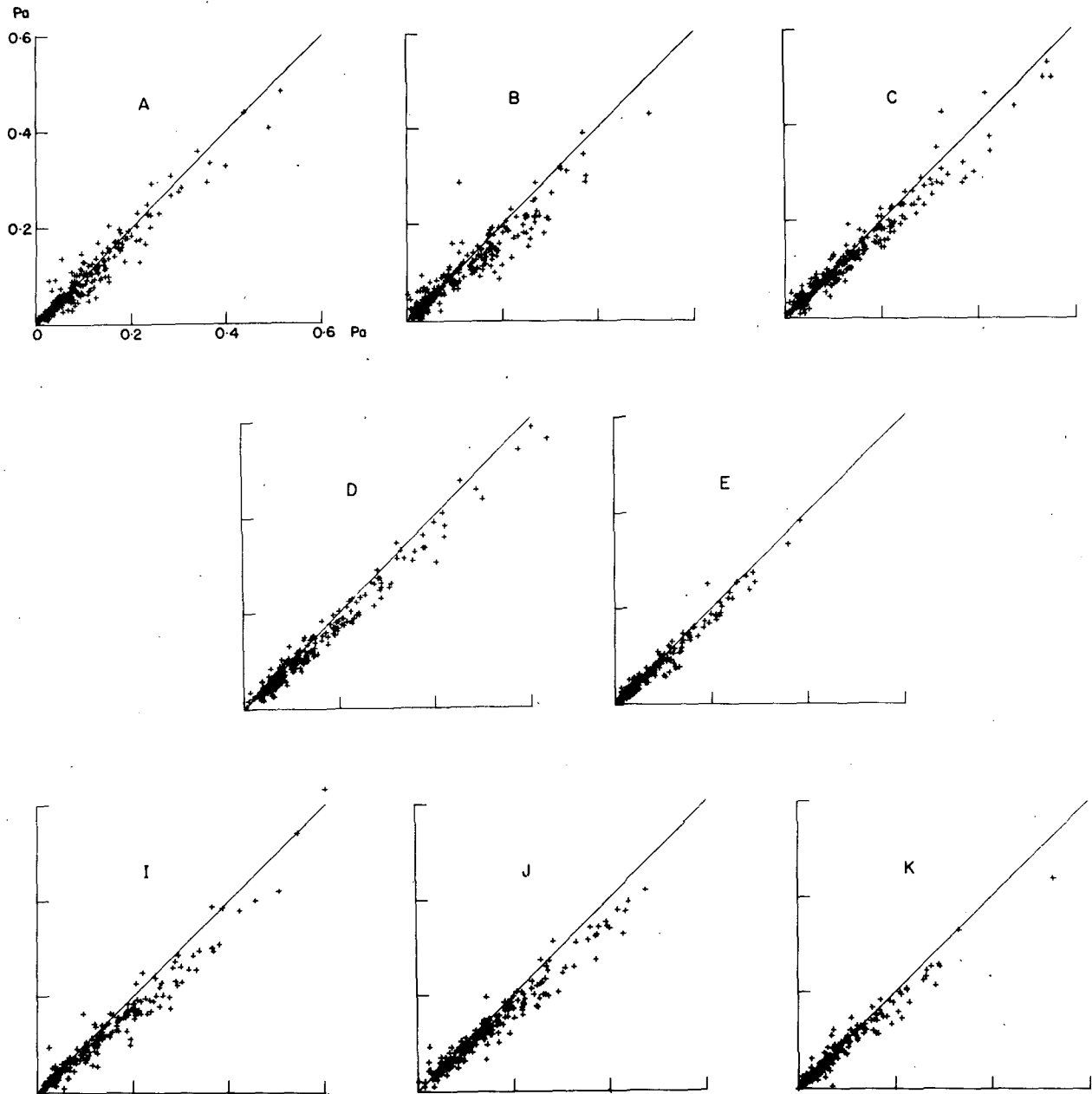


FIG. 3. Scattergraphs of $|f_0|$ (x -axis) against $|r_0|$ (y -axis) for a 1 month averaging period.

it was found that $|\sigma_a/\sigma_s| < 0.25$ for 95% of all σ values. In other words, approximating monthly σ by σ_s will result in a relative σ -error less than 0.25, 95% of the time. For a 4-month averaging period the relative σ -error is below 12%, which implies that the stress magnitude can be estimated to within 20%, given the appropriate v_0 and σ_s .

From Table 1 it can be seen that σ_s at OWS E is systematically lower than the other values throughout the year. This is due to the increased constancy of the winds nearer the equator, i.e., the Trades are steadier

than the Westerlies (Roll, 1965). In order to estimate stress over the whole North Atlantic the spatial variation of σ_s has been quantified.

The U.S. Navy Marine Climatic Atlas (1974) provides mean vector ($|v_0|$) and mean scalar ($|v|_0$) wind speed as a function of season for the 46 areas and OWS shown in Fig. 2. These statistics were calculated from all available wind data recorded over the North Atlantic and are thus the best estimates of the mean scalar and vector wind speeds. If the pdf of the fluctuating winds is assumed to be of the form (4), Fig.

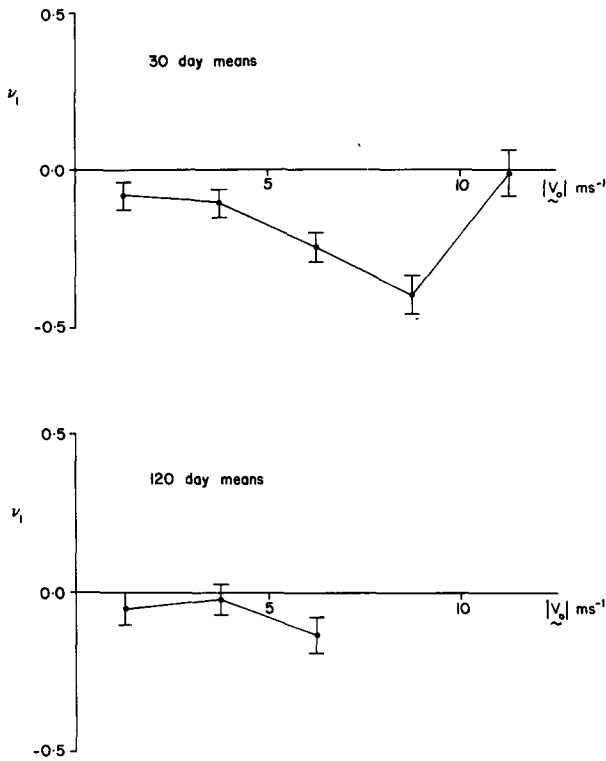


FIG. 4. Skewness (ν_1) in the direction of \mathbf{v}_0 as a function of $|\mathbf{v}_0|$ and averaging period at OWS C. The first skewness estimate is for $|\mathbf{v}_0| \in (0.0, 2.5) \text{ m s}^{-1}$, second for $|\mathbf{v}_0| \in (2.5, 5.0) \text{ m s}^{-1}$, etc. The vertical bars indicate the standard error of each mean.

7 can be used to uniquely define σ_s from $|\mathbf{v}_0|$ and $|\mathbf{v}|_0$. This has been done for each month of the year at the 46 positions shown in Fig. 2. Spatially smoothed, continuous representations of the σ_s fields have then been obtained by fitting bicubic splines to the com-

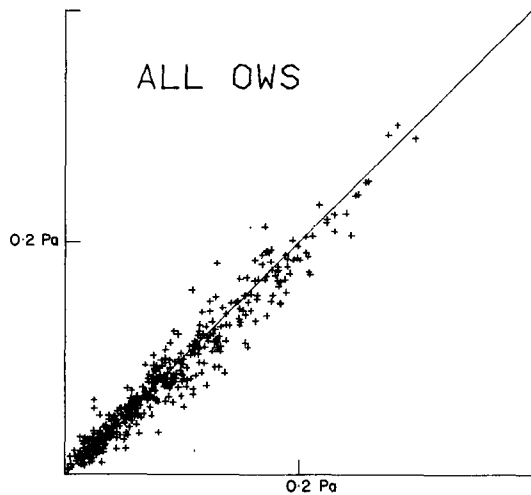


FIG. 5. Scattergraphs of $|\hat{\tau}_0|$ (x-axis) against $|\tau_0|$ (y-axis) for a 4 month averaging period.

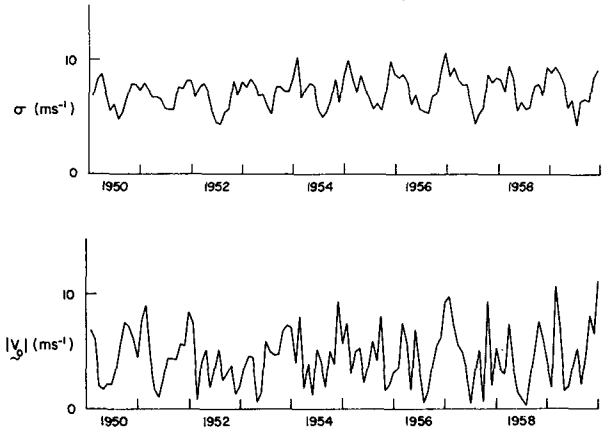


FIG. 6. Monthly variations of $|\mathbf{v}_0|$ and σ at OWS C, 1950–59.

puted σ_s values. Two sample fields are shown in Fig. 8 to illustrate the large spatial and temporal changes of σ_s . (No interior knots were required in the fitting of the splines. The standard deviation of the residuals, over all positions and months, was 0.4 m s^{-1}). It is now clear that the adjustments to the bulk stress formula (1), required to utilize mean winds, are both seasonally and geographically dependent.

b. Estimation of \mathbf{v}_0

A well-tried and successful method of estimating winds over the open ocean is to first calculate the geostrophic wind vector from surface air pressure maps and then apply an empirically determined contraction and rotation to obtain the surface wind vec-

TABLE 1. Seasonal variation of σ , in m s^{-1} , at each OWS (σ_s) for the period 1950–59. A sinusoid with a period of 1 year has been fitted to each column and the resulting amplitude (H), and day of the year when the maximum is attained (D), are also given.

Month	Ocean weather ship							
	A	B	C	D	E	I	J	K
Jan	9.3	9.1	8.5	8.3	7.0	8.5	7.8	7.3
Feb	7.8	8.9	8.2	8.5	7.4	7.8	7.8	7.3
Mar	8.5	8.2	8.1	8.5	6.7	8.5	7.7	6.8
Apr	7.5	7.5	7.5	7.5	5.8	7.8	6.6	6.8
May	6.7	6.7	6.6	6.7	5.0	6.9	6.4	6.0
Jun	5.6	6.2	6.0	6.0	4.1	6.0	5.7	5.2
Jul	5.1	5.7	5.1	4.9	3.5	5.5	5.0	4.4
Aug	5.4	6.2	5.9	5.2	3.7	5.9	5.9	5.1
Sep	7.1	7.4	6.7	6.8	4.8	6.8	6.7	5.5
Oct	8.2	8.1	7.7	7.3	6.2	7.6	7.3	6.5
Nov	8.5	8.1	8.0	8.1	6.4	8.0	7.8	7.6
Dec	8.6	8.3	8.7	8.9	6.7	8.8	7.9	7.8
Mean	7.4	7.5	7.3	7.2	5.6	7.3	6.9	6.4
H	1.8	1.4	1.5	1.7	1.8	1.4	1.3	1.4
D	8	10	13	19	19	17	6	14

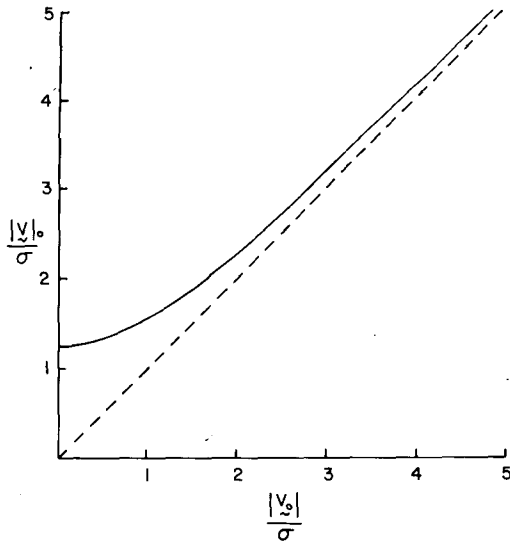


FIG. 7. Mean scalar wind speed ($|v_0|$) as a function of mean vector wind speed ($|v_g|$), both scaled by σ , for a bivariate, isotropic Gaussian pdf of the form (4).

tor (see, e.g., Willebrand, 1978). The main advantages of this method are that, first, air pressure differences give spatially integrated geostrophic winds and thus the large scale features of the wind field tend to be well defined. Second, there is good spatial and temporal coverage of air pressure data over the World's oceans. Lamb and Johnson (1966) for example have calculated reliable monthly air pressure maps for the North Atlantic from about 1880 onwards. Both of these advantages are of considerable importance in the estimation of stress fields for those areas or periods suffering a paucity of reliable wind data. The main disadvantages of the method are that the contraction and rotation corrections to the geostrophic wind vector are not exactly known. Further, the method does not resolve small scale features of the wind field and is not applicable in areas of pronounced isobar curvature or close to the equator. In spite of these shortcomings the method has been used in this paper.

To determine geostrophic winds over the North Atlantic, bicubic splines were fitted to monthly air pressures defined over the grid $15(5)70^\circ\text{N}$, $-10(10)90^\circ\text{W}$ (where available) for the period 1950–75. These air pressure data were obtained from the publications of "Die Grosswetterlagen Europas", Deutscher Wetterdienst, Offenbach. Four equispaced interior knots for each direction were used in the fitting of the bicubic splines. The standard deviation of the residuals, over all grid positions and all months, was 0.4 mb. Monthly geostrophic winds could then be calculated at any position over the North Atlantic from air pressure gradients derived from the spatially continuous spline representation.

A comparison of the monthly estimates of geostrophic (v_g) and surface wind (v_0 , determined from

the vector averaged 3 h wind observations) is shown in Fig. 9 for OWS C. (The results from OWS C are typical). There is considerable scatter which is largely due to the severe spatial smoothing implicit in the determination of the v_g field. However, there is a well defined relationship of the form

$$v_0 = C \begin{bmatrix} \cos\phi & -\sin\phi \\ \sin\phi & \cos\phi \end{bmatrix} v_g, \quad (9)$$

where C describes a contraction and ϕ an anti-clockwise rotation of v_g to get v_0 . Eq. 9 has been fitted to monthly v_0 and v_g at each OWS by the method of constrained least-squares (Thompson, 1979) to simultaneously determine the optimal C and ϕ (Table 2). Clearly $C = 0.7$ is a reasonable approximation for all OWS. However ϕ does vary significantly over the range 5 – 25° . This is probably related to the mean stability of the atmospheric boundary layer over the North Atlantic (Arya, 1975). For the purposes of this paper, the mean values of $C = 0.7$ and $\phi = 15^\circ$ have been used. Such values are in accord with those used by Willebrand (1978) in the estimation of 12 h wind

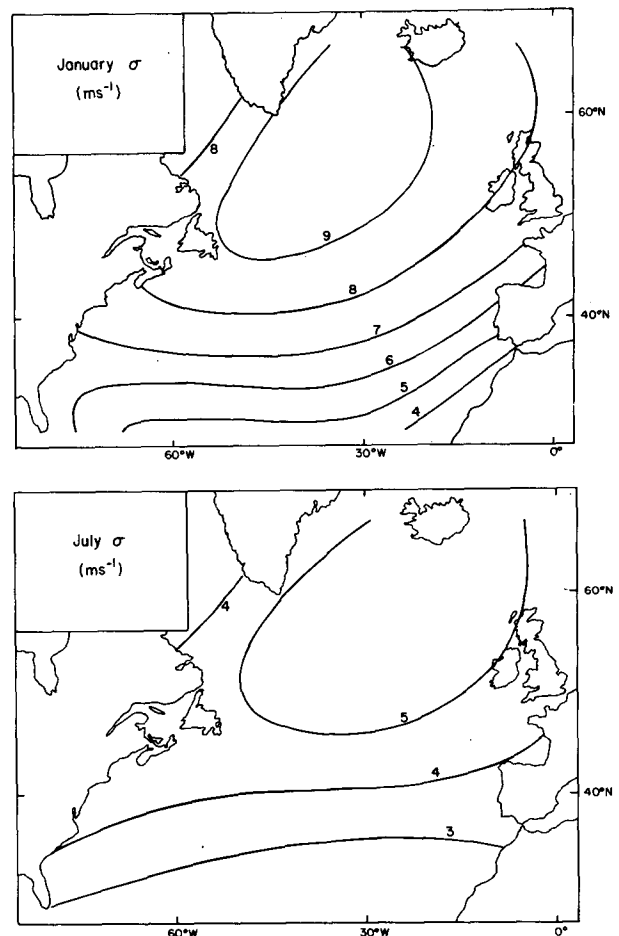


FIG. 8. Seasonal σ fields over the North Atlantic for January and July.

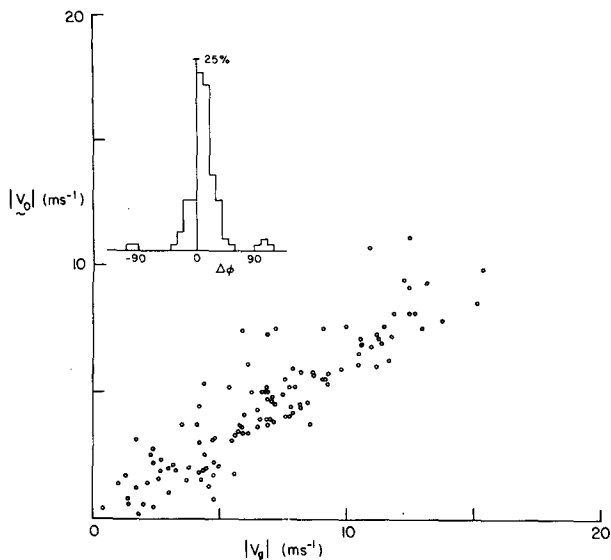


FIG. 9. Scattergraph of $|v_g|$ (x-axis) against $|v_0|$ (y-axis) for 1 month averaging period at OWS C. The histogram of differences in the directions of v_g and v_0 ($\Delta\phi$) is also shown. ($\Delta\phi$ is in deg and increases as v_0 rotates anti-clockwise from v_g).

fields over the same area. (The above fitting procedure was also applied to the winter and summer data separately to determine any seasonal variation in C and ϕ . The differences were within the expected scatter at all OWS except OWS D where the summer and winter (C , ϕ) values were (0.66, 9°) and (0.76, 2°), respectively).

c. Validation of method

Mean stress maps for the 4 seasons of the year have been calculated for the North Atlantic using (5) and the appropriate seasonal v_g (hence v_0) and σ_s distributions. Their average, the long term mean stress

TABLE 2. Results of the regression relating monthly changes of v_g and v_0 according to (9) at each OWS. The 95% confidence intervals for the contraction and angle shifts are shown in parentheses. (This table was completed at an early stage in the study, prior to the use of bicubic splines. The above geostrophic winds were calculated from a 5th-order pressure surface, defined in terms of latitude and longitude, fitted over a grid covering the North Atlantic).

Ocean weather ship	Contraction C	Angle shift ϕ (deg)
A	0.69 (0.06)	20 (5)
B	0.71 (0.04)	25 (3)
C	0.68 (0.03)	9 (3)
D	0.72 (0.04)	5 (3)
E	0.72 (0.04)	10 (3)
I	0.69 (0.05)	17 (4)
J	0.70 (0.04)	16 (3)
K	0.70 (0.05)	5 (4)

field, is presented in Fig. 10. (Note that it was necessary to take such an average, rather than just use the annual mean v_g and σ distributions, because a given mean wind generally gives a larger stress in winter than in summer due to the seasonal variation of σ). As a check on the present method of stress estimation, the mean distribution has been compared with that calculated by Willebrand (1978) for the same area. This author averaged 12 h stress distributions over the period 1973–76. The stresses were calculated from 12 h geostrophic winds using a drag coefficient similar to the one employed here. The maps shown in Fig. 10 are in excellent agreement with the results published by Willebrand (1978).

As a further check on the method, the Sverdrup transport across 31°N has been calculated for the 4 seasons of the year and compared with the values given by Bunker and Leetmaa (1978). The stress fields used by these authors were calculated from approximately 8 million ship observations over the period

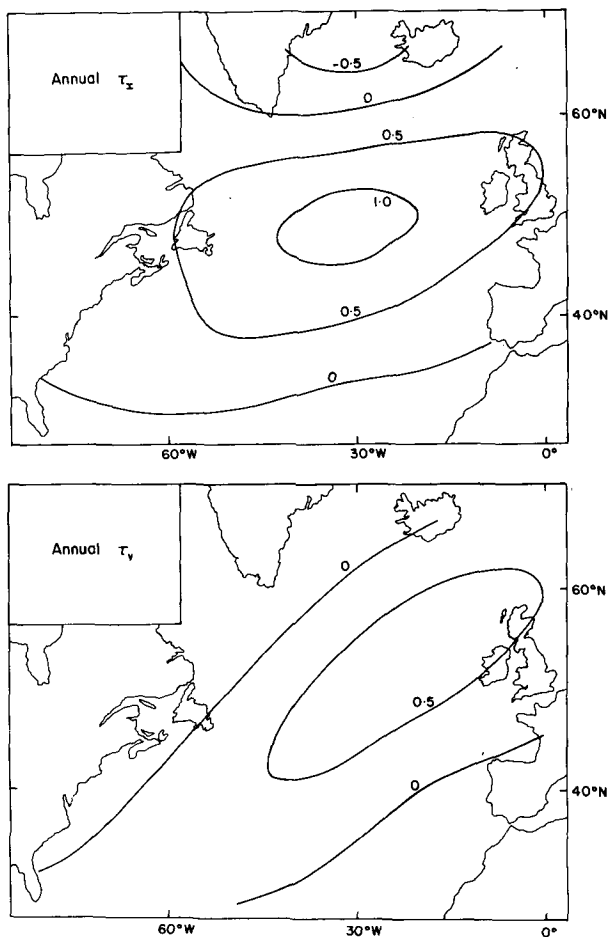


FIG. 10. The long term mean wind stress field based on the average of the 4 seasonal stress fields calculated from the appropriate σ_s and v_g distributions, using (5). The contour interval is 0.05 Pa.

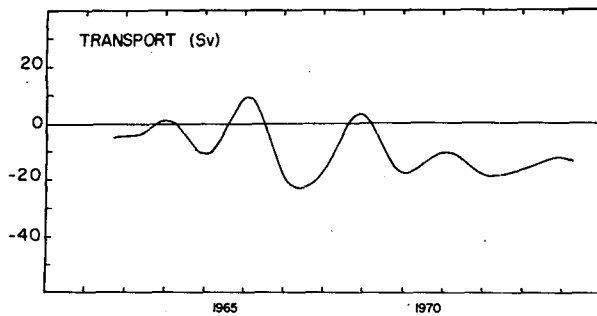


FIG. 11. Low-pass filtered fluctuations of the zonally integrated, meridional Sverdrup transport across 45°N . A positive transport corresponds to a northward flow.

1941–72. They used a drag coefficient which was significantly different from (2) and so, to facilitate comparison, a smoothed version of their neutral stability drag coefficient was used in (5). The two sets of transports were: Winter 41(48), Spring 35(38), Summer 23(26) and Autumn 16(17) Sv, with the Bunker and Leetmaa (1978) values given in parentheses. The present Sverdrup transports are slightly lower than those computed by Bunker and Leetmaa (1978). This is probably due to the greater smoothing of the stress fields by the present method resulting in a reduction in the estimated wind stress curl and thus the transport (Saunders, 1976).

As noted in the Introduction, the purpose of this paper is to present a method which can be used to determine slow fluctuations of the wind stress fields. These can readily be derived from the v_g and σ_s fields. For example, low-pass filtered fluctuations of the meridional Sverdrup transport across 45°N are shown in Fig. 11. (A Cartwright low-pass filter was used with a half-power point at 0.45 cycles per year (cpy) and a cut-off frequency of 1 cpy). It is clear that, across this latitude, there are large annual fluctuations (almost ± 20 Sv) in the wind-induced transport. In fact, as the long term mean transport is only 10.2 Sv, the fluctuations are large enough to temporarily reverse the direction of the total meridional flow.

4. Conclusions

Mean wind stress over a period of about 4 months can be accurately estimated from the mean and variance of the fluctuations over the averaging period using (5). Thus the problem of determining the mean stress distribution over a given area has been reduced to the determination of the v_0 and σ distributions.

The monthly variations of σ over most the North Atlantic are dominated by a seasonal cycle. In fact, σ for any given 4-month period will generally (i.e., 95% of the time) differ from the corresponding long-term seasonal mean by less than 12%. The spatial variations of the seasonal σ -fields over the North Atlantic were readily determined from wind statistics

derived from many years of observations and published in the U.S. Navy Climatic Atlas (1974). Thus, an approximate σ for any season and position over the North Atlantic could be obtained. This same approach could be used to estimate seasonal σ -fields over other ocean areas where similar statistics are available (e.g., South Atlantic Ocean). The mean wind fields were estimated from mean air pressure maps by applying an empirically determined contraction (0.7) and down pressure gradient rotation (15°) to the geostrophic wind vector. The advantages of such an approach are that (1) mean pressure distributions, and thus surface stresses, are available for many years and for large areas of ocean and (2) the large scale features of the wind field are well defined. The main disadvantages are that it does not resolve the small scale features of the wind field and is not applicable in areas close to the equator or of pronounced isobar curvature.

The stress estimation procedure was tested by comparing the long term mean stress and seasonal Sverdrup transports across 31°N with independently determined values. Good agreement was found between the two long term mean stress fields; the present method slightly underestimated the Sverdrup transports because the small scale stress variations, and thus curls, were underestimated.

Slow fluctuations of the meridional Sverdrup transport across 45°N were given to illustrate both the method and meteorological variability that can occur in the North Atlantic on time scales greater than one year. The fluctuations were almost ± 20 Sv about a mean of 10 Sv thus implying that the total transport temporarily changes sign on such a time scale.

In summary, a simple yet effective method has been outlined for the estimation of the large scale features of mean stress fields from air pressure maps. The novel feature of the method is that the magnitude of the high frequency wind fluctuations have been explicitly included in the time averaged form of the stress law. It is believed that the method could be used to advantage for those periods or ocean areas where there are insufficient wind data to determine the stress distributions directly.

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