Provisionally Corrected Surface Wind Data, Worldwide Ocean–Atmosphere Surface Fields, and Sahelian Rainfall Variability

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ABSTRACT

Worldwide ship datasets of sea surface temperature (SST), sea level pressure (SLP), and surface vector wind are analyzed for a July–September composite of five Sahelian wet years (1950, 1952, 1953, 1954, 1958) minus five Sahelian dry years (1972, 1973, 1982, 1983, 1984) (W − D). The results are compared with fields for a number of individual years and for 1988 minus 1987 (88 − 87); Sahelian rainfall in 1988 was near the 1951–80 normal, whereas 1987 was very dry. Before performing the analyses, an extensive study of the geostrophic consistency of trends in pressure gradients and observed wind was undertaken, motivated by the suggestion that changes in observational practice on board ships have introduced an upward trend in reported wind speed during the last 40 years.

The results suggest that, during the period 1949–88, there is a mean increase in reported wind speed of about 16% that cannot be explained by trends in geostrophic winds derived from seasonal mean SLP. Estimates of the wind bias are averaged for 18 ocean regions. The trend in every 2° lat × 2° long wind time series is adjusted according to the mean bias of the region in which the box resides.

A map of correlations between Sahelian rainfall and SLP in all available ocean regions is shown to be field significant. Remote atmospheric associations with Sahelian rainfall are consistent with recent suggestions that SST forcing from the tropical Atlantic and the other ocean basins may contribute to variability in seasonal Sahelian rainfall. It is suggested that wetter years in the Sahel are often accompanied by a stronger surface monsoonal flow over the western Indian Ocean and low SLP in the tropical western Pacific near New Guinea, and there is also a suggestion of increased cyclonicity over the extratropical eastern North Atlantic and northwest Europe. All of these features were present in W − D and 88 − 87 fields. In the tropical Atlantic, W − D shows many of the features identified by previous authors. However, the 88 − 87 fields do not reflect these large-scale tropical Atlantic changes. Instead there is only local strengthening of the pressure gradient and wind flow from Brazil to Senegal. Further individual years are presented (1958, 1972, 1975) to provide specific examples of some of the patterns identified in W − D, while July–September 1983 is shown to have an SLP anomaly pattern over the tropical oceans that is quite unique in the period 1949–88.

1. Introduction

The Sahelian region of Africa (Fig. 1a) receives 80%–90% of its annual rainfall in the months July–September. The large interannual and interdecadal variability in the seasonal rainfall (Fig. 1b) demands explanation, and various mechanisms have been proposed (reviewed by Druyan 1989). Variations at the ocean–atmosphere interface of the tropical Atlantic are clearly associated with the rainfall variability (Lamb 1978a,b; Hastenrath 1984; Lough 1986). Folland et al. (1986, 1991), Palmer (1986), and Rowell et al. (1991) have argued that sea surface temperature (SST) in all the major ocean basins forces Sahelian rainfall variations. This paper identifies some of the surface atmospheric variations over the global oceans that accompany the variations in SST and Sahelian rainfall. A further purpose of the paper is to estimate the extent to which results are influenced by any evolving bias in ship wind speed reports (Ramage 1987; Cardone et al. 1990). This is achieved by forming a new wind dataset in which wind trends in the period 1949–88 are forced to be geophysically consistent with trends in surface pressure gradients.

2. Data

a. Rainfall

The annual standardized Sahelian rainfall anomaly series used in this study (Fig. 1b) is based on Nicholson (1985). For values since 1984, the series has been updated using available CLIMAT reports [details of the updating method are in Colman (1988)]. The standardized series does have limitations as a measure of the relative rainfall season performance over the Sahel as a whole. A varying station network (Fig. 1c) introduces heteroscedasticity to the series (Katz and Glantz 1986), and this could affect the ranking of years. The
station data then interpolated to a grid of 2.5° lat × 3.75° long, using the methods described in Hulme (1992). Note that to qualify for the station network, a station must have data in over 83% of the months during 1949–90, and if a grid box does not contain a station with data, then its monthly value is set to missing. The gridded rainfall values for July, August, and September were summed to give a seasonal dataset (used in Figs. 6 and 10).

b. Near-surface atmosphere

The Comprehensive Ocean–Atmosphere Data Set (COADS) 2° lat × 2° long (2 × 2) trimmed monthly means for the years 1949–88 were used for the analyses (Woodruff et al. 1987). Data were further processed as follows.

(i) For a comparison of geostrophic and observed winds—monthly means of sea level pressure (SLP), zonal wind component (u), meridional wind component (v), and wind speed for 1949–88 were all used. The following details about the monthly means were also used: number of observations, standard deviation of the observations, mean day of the observations in the month, mean location of the observations within the 2 × 2 box. Currently, data for 1980–88 contain only the mean and number of observations for each of the meteorological parameters. To overcome this deficiency, a mean of each of the missing details (except day of the month) was calculated for 1960–79 and these were used for the 1980–88 data. To improve data coverage, the monthly dataset was transformed into a dataset of 3-month seasons, using

\[
V_s = \frac{\sum_{m=1}^{3} V_m N_m}{\sum_{m=1}^{3} N_m}
\]

where \( V_s \) is the value for season \( s \), \( V_m \) is the value of variable \( V \) in month \( m \), \( N_m \) is the number of observations in month \( m \), and summation is over the 3 months that constitute the season.

(ii) For studying relationships with Sahelian rainfall—the monthly COADS data for SLP, \( u \) and \( v \) were combined into a July–September dataset applying the same principle as Eq. (1). Mean values (normals) were calculated for the years 1969–88, the period with best data coverage. To calculate the normals, the data were initially averaged over 2° lat × 4° long (2 × 4) boxes, to improve data coverage. At least eight observations were required for a 2 × 4 box to have a seasonal value in a given year. To calculate a normal, at least 15 out of the 20 seasons during 1969–88 were required to have a seasonal value. The boxes were overlapped by 2° long. Longitudinally adjacent 2 × 4 normals were averaged, giving a set of 2 × 2 normals with effectively a 1:2:1 smoothing in the longitudinal direction.
Anomaly datasets on the $2 \times 2$ scale were created for the July–September seasons 1949–88. Results in this study are presented on the $10^\circ$ lat $\times 10^\circ$ long ($10 \times 10$) scale. The $10 \times 10$ seasonal anomaly dataset was formed:

$$V_{10\times10} = \frac{\sum_{i=1}^{25} V_i N_i}{\sum_{i=1}^{25} N_i}$$

where $V_{10\times10}$ is the $10 \times 10$ anomaly for variable $V$, $V_i$ is the seasonal anomaly of $V$, and $N_i$ is the number of observations in the $i$th constituent $2 \times 2$ box. A maximum value of 100 was set for $N_i$, so that the (rare) $2 \times 2$ boxes that have many more than 100 observations in a season do not dominate the estimate of the $10 \times 10$ seasonal anomaly. For inclusion in the $10 \times 10$ dataset, at least 20 observations were required for the $10 \times 10$ box as a whole. The optimum way of combining the $2 \times 2$ anomalies to achieve the best estimate $10 \times 10$ anomaly will be a weaker function of $N_i$ than in Eq. (2). The weighting in Eq. (2) effectively assumes that the seasonal anomaly being estimated is uniform across the $10 \times 10$ box, that is, the effective number of independent $2 \times 2$ boxes is 1. A common alternative to Eq. (2) is to give each $2 \times 2$ anomaly equal weight. However, the approximation in Eq. (2) is particularly useful in regions where data coverage is dominated by narrow shipping lanes (e.g., tropical South Atlantic) where the number of observations in adjacent $2 \times 2$ boxes for a season may vary from over 100 to less than 5. In such a situation, considerably more weight should be given to the $2 \times 2$ anomaly based on over 100 observations.

c. Sea surface temperature

SST data were taken from the Meteorological Office Historical Sea Surface Temperature dataset version 4 (MOHST4) (Bottomley et al. 1990), which contains monthly SST anomalies, where available, for $5^\circ$ lat $\times 5^\circ$ long ($5 \times 5$) grid boxes extending over the World Ocean. The $5 \times 5$ monthly values for July, August, and September were averaged with equal weight into a $10 \times 10$ seasonal dataset for July–September. As little as one constituent monthly $5 \times 5$ anomaly was allowed when calculating a seasonal $10 \times 10$ anomaly.

3. Corrections for the wind dataset

a. Background

Cardone et al. (1990) suggest that the wind speed reported by ships has increased in recent decades due to changes in observational practices. The main cause that they identify is an increasing fraction of anemometer readings, typically now made at a mean height of about 20 m, relative to Beaufort force estimates, which are converted to 10-m winds (using a biased conversion scale) before insertion into computerized datasets like COADS. Therefore, before studying the relationship between Sahelian rainfall and surface winds, an attempt has been made to verify the presence of the wind trend (Cardone et al. 1990) in all ocean areas, and where necessary to remove the spurious trend from the data. To achieve this, trends in seasonal mean vector winds are compared with trends in seasonal mean pressure gradients, and the results are tested for consistency based on the geostrophic wind relation. A preliminary study by Wright (1988) on an area in the tropical Pacific suggested that trends in reported wind speeds in that area were unlikely to be real because they were not supported by changes in surface pressure gradients. A wind correction method based on this approach assumes that there is no substantial time-varying bias in the estimated pressure gradients (appendix A), and that the difference between the geostrophic wind trend and observed wind trend can be used to isolate time-varying bias in wind observations. The derived geostrophic wind is a function of pressure difference, so it will largely be independent of any time-varying bias that may exist in ship pressure data, though other problems such as changes in bias toward reports during fair weather may influence the geostrophic wind trends in some regions. Trends in boundary-layer stability could also cause a difference between geostrophic wind trends and observed wind trends by modifying the ratio of observed to geostrophic wind. Air–sea temperature differences are a good measure of boundary-layer stability. Using Meteorological Office Night Marine Air Temperature data and SST data, large-scale trends in air–sea temperature differences during 1949–88 were found to amount to at most between $+0.4^\circ$ and $-0.4^\circ$C. Such differences probably imply, for a geostrophic wind of about $7$ m s$^{-1}$, a change in the near-surface wind of typically between $-3\%$ and $+3\%$ (e.g., Bumke and Hasse 1989). Such values are quite small compared to the change in reported wind speed over the last 40 years (see next section), so trends in air–sea temperature differences and boundary-layer stability are not considered further in this paper. The geostrophic wind is not a good approximation to the true wind near the equator. It was found that within $4^\circ$ lat of the equator, the correlation between geostrophic and observed wind during the 40 years analyzed was close to zero, so these series were excluded from the analysis. All series poleward of $4^\circ$ were included, but the contribution of each $2 \times 2$ box to the overall results was weighted according to the correlation between the geostrophic and observed wind, thereby ensuring that little weight was given to those boxes where the derived geostrophic wind seemed to provide a poor approximation of the true wind.

b. Quantifying the discrepancy between wind and pressure gradient trends

Where possible, an estimate of the seasonal mean geostrophic wind in a given direction $\alpha' (G_{\alpha'})$ is made
from the COADS seasonal mean SLP (see appendix A for details). Each estimate of \( G_{\alpha} \) is adjusted for friction and other ageostrophic wind components to yield \( \bar{G}_{\alpha} \), which is compared with the corresponding \( V_{\alpha} \), the observed seasonal vector wind in direction \( \alpha \), where \( \alpha \) differs from \( \alpha' \) by 15° to allow for backing of the wind due to friction. [Fifteen degrees is a generally accepted value (e.g., Riehl 1979), but its choice is not critical, and results reported were largely unchanged by setting the backing angle to 0°.] Below, trends in \( V_{\alpha} \) and \( \bar{G}_{\alpha} \) are compared only in those 2 \( \times \) 2 boxes that had mean observed and derived seasonal vector wind (during 1949–88) \( \bar{V}_{\alpha} \) and \( \bar{G}_{\alpha} \), \(-1\) m s\(^{-1}\) or \( \bar{V}_{\alpha} \) and \( \bar{G}_{\alpha} \), \( +1\) m s\(^{-1}\), so that the vector trends can provide approximations for the percentage increase in the magnitude of the wind speed (see appendix B).

For most 2 \( \times \) 2 boxes, the direction \( \alpha \) was close to the longitudinal direction. This choice of analyzing zonal wind reflects the preference for the 2 \( \times \) 2 boxes to have \( V_{\alpha} \) as different from zero as possible (appendix B). However, in narrow shipping lanes, \( \alpha \) was chosen so that the required pressure gradient to estimate \( \bar{G}_{\alpha} \) was directed approximately along the shipping lane, hence permitting maximum use of the available SLP data (see appendix A).

For every 2 \( \times \) 2 ocean box included in the analysis, and (separately) for each of the four seasons, the following linear model was fitted:

\[
(V_{\alpha} - \bar{G}_{\alpha}) = a_0 + a_1 t + \epsilon
\]

where \((V_{\alpha} - \bar{G}_{\alpha})\) is the difference between the observed and geostrophic seasonal wind in year \( t \), \( \epsilon \) is white noise with mean 0, \( a_0 \) is a constant, and \( a_1 \) represents the trend in the difference between \( V_{\alpha} \) and \( \bar{G}_{\alpha} \). Usually \( a_1 \) is fitted using a least-squares procedure. However, the magnitude of the fitted parameter is then biased low by observational error on the variables (Kendall and Stuart 1961, p. 413). To avoid this, \( a_1 \) was estimated:

\[
a_1 = \frac{(V_{\alpha} - \bar{G}_{\alpha})_2 - (V_{\alpha} - \bar{G}_{\alpha})_1}{t_2 - t_1}
\]

where the overbar indicates quantities measured during period 1 (1949–68), and period 2 (1969–88). Only those years with an estimate of \( V_{\alpha} \) and \( \bar{G}_{\alpha} \) were included, and only 2 \( \times \) 2 boxes with estimates in at least 50% of years in period 1 and period 2 were included. In terms of the process in Eq. (3), \((V_{\alpha} - \bar{G}_{\alpha})\) is an unbiased estimate of \((V_{\alpha} - \bar{G}_{\alpha})\) at \( t = t_1 \). The slope of the line joining the two unbiased estimates of the process at \( t_1 \) and \( t_2 \) [Eq. (4)] gives an unbiased estimate of the trend \( a_1 \) (units are m s\(^{-1}\) yr\(^{-1}\)). The implied spurious percentage increase \( S \) in the observed wind vector over 40 years is

\[
S = \left( \frac{a_1 \times 40}{V_{\alpha}} \right) \times 100.
\]

Two criteria were identified that influence the degree of confidence attributed to each \( a_1 \):

(i) **Number of years with data, \( N \) equals number of years with an estimate of \((V_{\alpha} - \bar{G}_{\alpha})\).**

(ii) **Meteorological reliability.** A simple estimate of this is given by the correlation \( r \) between \( V_{\alpha} \) and \( \bar{G}_{\alpha} \). The correlation may be low due to noise in the data or due to geostrophy providing a poor approximation to the real wind. In either case, confidence in \( S \) is reduced. Note that the magnitude of spurious trend proposed by Cardone et al. (1990) is sufficiently small to have little impact on \( r \).

Combining these two considerations, each spurious trend \( S \) was allotted a weight equal to \( N \times r \). The weighted average of \( S \) over all seasons and 2 \( \times \) 2 boxes is 16.1%, which is the correct order of magnitude to support Cardone et al. About 16.1% is the increase predicted by Cardone et al. for a change in the percentage of anemometer readings from 0% in 1949 to 60% in 1988, in a neutral boundary layer and mean wind speed of 5 m s\(^{-1}\).

As a first attempt to derive geographically varying estimates of \( S \), the values have been averaged over the regions shown in Fig. 2. It was assumed that changes in observational practice, and therefore the required corrections to the data, would be more coherent within shipping lanes than between shipping lanes. So, the divisions in Fig. 2 attempt to include shipping lanes in the same region. Regional divisions were also made where marked and coherent variations in the values of \( S \) were noted (e.g., extratropical North Atlantic). Finer resolution or more objective selection of the regions was not thought justified at this stage. Equatorward of 4° lat, no estimates of \( S \) were made, but equatorial regions are allotted to a region in Fig. 2 for the purpose of correcting the whole wind dataset in the next section. Most equatorial 2 \( \times \) 2 boxes were allotted to the nearest region. The few exceptions (e.g., in the eastern tropical Pacific) were made so as to avoid the allocation of a 2 \( \times \) 2 box to a region that was clearly dominated by a different shipping lane.

Of the 18 regions in Fig. 2, 13 have mean values of \( S \) in the range +10% to +25%. North of 50°N, the spurious wind trend actually appears negative, especially in the Atlantic. This negative trend is present in the different geostrophic analyses performed (not shown) with different length scales for estimating the pressure gradients, different data inclusion criteria, etc. Those areas that Cardone et al. studied (marked with crosses in Fig. 2) all show positive values for \( S \).

c. **Correcting the wind data**

Figure 2 indicates that trends in the estimated pressure gradients and trends in the observed winds are not geostrophically consistent. Cardone et al. (1990) have presented a strong argument for time-varying bias
in reported wind speeds, a bias that is of the right order of magnitude to explain most of the results in Fig. 2. Other possible explanations, such as a genuine change in the ratio of geostrophic to ageostrophic wind, or a time-varying bias in the pressure gradient estimates, are unlikely to be able to explain the size of the values of \( S \) in Fig. 2. Thus, to a first approximation, the values of \( S \) are interpreted as isolating a spurious trend in the wind data. For studying relationships with Sahelian rainfall in the next section, a corrected July–September wind dataset has been calculated using the results in Fig. 2. For each 2 \( \times \) 2 vector wind time series, the corrected data are calculated:

\[
\begin{align*}
\hat{u}_t &= u_t - \left[ u_t(S/100) \times (t - t_b)/40 \right] \\
\hat{v}_t &= v_t - \left[ v_t(S/100) \times (t - t_b)/40 \right]
\end{align*}
\] (6)

where \( \hat{u}_t \) and \( \hat{v}_t \) are the corrected \( u \) and \( v \) seasonal wind vectors for time \( t \), \( u_t \) and \( v_t \) are the observed COADS seasonal wind vectors for time \( t \), and \( S \) is the 1949–88 mean spurious percentage change in wind speed for the region (Fig. 2) in which the 2 \( \times \) 2 box resides; note that the midpoint of the season is used for \( t \) (e.g., July–September 1949, \( t = 1949.71 \)). Here \( t_b \) is an arbitrarily selected time that acts as the base time for the corrections. For example, if \( t_b = 1949.0 \), then when \( t = 1949.0 \), \( \hat{u} = u \), and \( \hat{v} = v \); 40 years later when \( t = 1989.0 \), the wind vectors are reduced in magnitude by \( S\% \), (or increased by \( S\% \) if \( S \) is negative). To correct the dataset for use in the next section on Sahelian rainfall, \( t_b \) was set to the midpoint of the 1969–88 July–September normals period (1979.2). To illustrate the impact of the corrections on the 10 \( \times \) 10 anomaly dataset, Fig. 3 compares July–September anomalies of \( u \) and \( \hat{u} \) for a box in the trade wind regime of the tropical North Atlantic. There is a strong negative trend in \( u \), corresponding to a strengthening of the reported easterly wind component. The trend is statistically significant at the 1% level and amounts to 1.22 m s\(^{-1}\) over the 40 years 1949–88, whereas the trend in the corrected wind time series \( \hat{u} \) is much smaller, amounting to 0.12 m s\(^{-1}\) over the 40 years (not statistically significant at the 10% level). A time series of the difference in SLP anomaly in the boxes to the north and south (crosses in Fig. 3) contains no statistically significant trend, supporting the corrected wind data, though the SLP data do suggest some strengthening of the July–September trade wind in this region (the trend in the SLP differences amounts to 0.28 mb over 40 years).

4. Sahelian rainfall variations
   a. Introduction

Folland et al. (1986) calculated the difference in SST for the five wettest and five driest Sahelian years (W – D) in Fig. 1b during the period 1950–84. Figure 4 compares the SST composite map (here recalculated using the new version of MOHSST, see section 2) with corresponding maps for SLP and surface vector wind. At least 8 out of the 10 years were required to have data for a 10 \( \times \) 10 box to be included. The local statistical significance of the SST and SLP differences is assessed using a \( t \) test (Afifi and Azen 1979). Maps for 1988 minus 1987 are also presented (88 – 87, Fig. 5); 1988 and 1987 represent two recent years with a considerable difference in Sahelian rainfall (1988 much wetter than 1987, Fig. 1b). To check the scale and coherence of rainfall variations in W – D and 88 – 87, rainfall difference maps have been calculated using the gridded rainfall values (Fig. 6). In W – D (Fig. 6a),
rainfall differences are a little larger in the western Sahel than in the eastern Sahel, but in the eastern Sahel they are still generally significant at the 5% level and greater than 100 mm. The significant rainfall differences extend southward to about 10°N. In 1968–1987 (Fig. 6b), rainfall differences in the Sahel are more noisy than in W – D, but generally quite large positive rainfall differences exist across the Sahel and southward to around 10°N.

The fields for W – D (Fig. 4) and 1968–1987 (Fig. 5) are discussed in section 4b. However, it is important to be clear on the value and limitations of Figs. 4 and 5. The years in W – D are extremes taken from within periods of persistent rainfall anomaly. The composite may therefore contain the influence of processes that operate on interdecadal time scales and interannual time scales. One can object to the comparison of W – D and 1968–1987 since the wetter of the two years in 1968–1987 is actually below the long-term normal. However, in terms of high-frequency (time scale less than 10 years) Sahelian rainfall variability, as defined by the residual from the filter applied in Fig. 1b, 1988 is actually the second wettest year during the period 1949–1988, while 1987 is a dry year. So a cautious comparison of W – D and 1968–1987 is useful in the context of searching for common processes operating on interannual time scales. The composite method averages fields over a number of years, so in a statistical sense, the method reduces noise, which in principle is desirable. However, in a meteorological sense, the benefit may be reduced because in different years Sahelian seasonal rainfall anomalies may result from different large-scale processes, which may have different signals in the surface fields. Thus, the W – D fields cannot be used to deduce the form of a unique teleconnection structure with Sahelian rainfall, but they can identify regions where consistent variations occurred in at least a number of the composited years. Study of individual years is then required to illustrate the kinds of patterns that have led to the mean teleconnection relationships.

A further potential problem with W – D is that, using just a few years, the result may be greatly influenced by just one year in which the ocean–atmosphere anomalies are very large. However, it has been confirmed that the variations in W – D that are emphasized in the next sections are not sensitive to the removal of any individual year. One advantage of W – D is that it does enable a clear assessment of the impact of the wind corrections, and comparisons with the SLP and wind patterns of 1968–1987 offer some verification of the wind correction procedure, since 1968–1987 fields are likely to be reliable due to good data coverage and little difference in observational practice from 1987 to 1988. The values in W – D (Fig. 4) are, in a statistical sense, unbiased estimates of the mean difference between the two sets of years, and they are in meteorological units. Physical interpretation of composite fields, and in particular their gradients, is therefore easier than for many statistical measures of association. For example, a map of correlation coefficients is difficult to interpret because correlation is biased low by random error on observations (Kendal and Stuart 1961) and is in nonmeteorological units. Another limitation of Fig. 4 is that, with only a few years included, statistical significance is difficult to both assess and achieve, even though the differences may be physically linked to Sahelian rainfall variations. The shaded areas in Fig. 4 provide useful guidance on which anomalies are most statistically significant, but the level of significance used (10%) is not a severe one, and the
FIG. 5. July–September means, 1988 (Sahelian average) minus 1987 (Sahelian dry) for:
(a) sea surface temperature, (b) sea level pressure, and (c) corrected vector wind (because
there is little difference in time between 1988 and 1987, the corrected and uncorrected wind
maps are almost identical, hence the uncorrected is not shown).
The results are not intended to be interpreted in terms of a purely statistical result.

Some of these deficiencies have been partly overcome in two ways:

(i) In section 4e some individual years are discussed to highlight some years that are similar to that suggested by the composite $W - D$ and other years that are not similar in some or most of the tropics.

(ii) A summary of the statistical relationship between SLP and Sahelian rainfall in all years 1949–88 is provided by a correlation map (Fig. 7) showing the correlation between Sahelian rainfall and SLP in all available ocean regions. The statistical significance of this map has been carefully tested. The statistical significance of each correlation is assessed using a $t$ test, allowing for serial correlation when estimating the degrees of freedom (Bartlett 1935; Folland et al. 1991). Then the field significance of the correlations has been assessed using a Monte Carlo method (Livezey and Chen 1983). For this, 500 random time series were created—each with 40 values. The time series were forced to have the properties of a first-order autoregressive process with model coefficient $a_1$ equivalent to that for a model fitted to the Sahelian rainfall series 1949–88 ($a_1 = 0.72$). These 500 simulated rainfall series were used to create 500 correlation maps like the one in Fig. 7, using the same method to assess local significance as used in Fig. 7. Table 1 shows the fraction of occasions in the Monte Carlo simulation on which
the area covered by locally significant boxes exceeded the significant area in Fig. 7.

In the tropical Atlantic, 28.4% of the sampled ocean area is covered by significant correlations. Such a coverage was achieved in less than 2.5% of the Monte Carlo simulations. Taking the Atlantic as a whole, a smaller percentage of the area is covered with significant correlations (21.4%), but the field significance of the result is actually a little better (1.6%). This is possible because of the larger number of spatial degrees of freedom in the Atlantic compared to the tropical Atlantic alone. The field significance of the Pacific correlations is almost as good as for the Atlantic, but the Indian correlations, despite covering 17.3% of the sampled area, only achieve field significance at about the 7.6% level due to the small number of spatial degrees of freedom. The SLP correlations north of 30°N are field significant at the 2.4% level, suggesting Northern Hemisphere extratropical circulation is probably related to Sahelian rainfall variations. The field significance of the correlations in all available regions is a little better than that for any of the subdivisions shown in Table 1. Folland et al. (1991) performed a similar analysis using SST for 1901–86 and found field significance for the correlations with SST for the globe as a whole, and for each ocean basin and the tropics and extratropics separately. So there is strong statistical evidence for linkages between Sahelian rainfall and remote large-scale atmospheric circulation and sea surface temperature variability.

b. Variations in different ocean regions

1) TROPICAL NORTH ATLANTIC

In W–D (Fig. 4b), close to Africa at 10°–30°N, negative SLP values are bordered by relatively higher values equatorward and near 35°N. The corresponding wind pattern is consistent and fairly insensitive to the wind corrections (Figs. 4c, d), with enhanced convergence near and to the north of the latitude of the Sahel, resulting mainly from southerly flow, but also from some enhanced northeasterly in the corrected data around 25°–35°N. In 88 – 87, widespread low SLP is not found. However, confined to the eastern side of the basin, the meridional pressure gradients are directed into a belt near 20°N, and the wind flow is consistent with this showing enhanced convergence around 20°N

<table>
<thead>
<tr>
<th>Ocean region</th>
<th>(i) Percentage of area</th>
<th>(ii) Probability</th>
</tr>
</thead>
<tbody>
<tr>
<td>All available areas</td>
<td>17.7</td>
<td>0.012</td>
</tr>
<tr>
<td>Tropics (30°S–30°N)</td>
<td>21.5</td>
<td>0.014</td>
</tr>
<tr>
<td>North of 30°N</td>
<td>14.1</td>
<td>0.024</td>
</tr>
<tr>
<td>Tropical Atlantic</td>
<td>28.4</td>
<td>0.024</td>
</tr>
<tr>
<td>Atlantic</td>
<td>21.4</td>
<td>0.016</td>
</tr>
<tr>
<td>Indian</td>
<td>17.3</td>
<td>0.076</td>
</tr>
<tr>
<td>Pacific</td>
<td>14.5</td>
<td>0.020</td>
</tr>
</tbody>
</table>

Notes:
(i) Percentage of ocean region having significant correlations (at the 5% level) with Sahelian rainfall.
(ii) Probability of achieving a larger percentage than that in column (i) (estimated by Monte Carlo test).
resulting from southwesterly flow from the south and northeasterly flow from the north.

In 88 – 87 (Fig. 5a), negative SST values prevail throughout the tropical North Atlantic, opposite to that found for W – D (Fig. 4a and Folland et al. 1986) and in other studies (e.g., Lamb 1978a,b). Furthermore, there is no similarity between the SLP and wind fields in Figs. 4 and 5 away from the extreme eastern side of the basin.

The wind corrections have a large impact at 15°N on the western side of the basin in W – D. The raw data (Fig. 4d) suggest a substantial weakening of the trade winds in wet years, but the corrections leave only a slight weakening (Fig. 4c).

2) TROPICAL SOUTH ATLANTIC

In both W – D and 88 – 87, the region close to Brazil has low SST and high SLP. In W – D the raw wind data (Fig. 4d) suggest a small enhancement of the cross-equatorial flow near Brazil, whereas the corrected data (Fig. 4c) show a stronger southerly flow that is similar to that which occurred in 88 – 87. The similarity in the cross-equatorial pressure gradient near Brazil in W – D (Fig. 4b) and 88 – 87 (Fig. 5b) makes the corrected wind more plausible.

In 88 – 87 (Fig. 5a), positive SST values prevail in much of the rest of the tropical South Atlantic, opposite to that found in W – D (Fig. 4a and Folland et al. 1986). In terms of the cross-equator dipole of tropical Atlantic SST anomalies that Lough (1986) related to Sahelian rainfall, 1988 should have been drier than 1987 in the Sahel, not wetter (Figs. 1b and 6b).

In W – D (Fig. 4b), the high SLP near Brazil extends to a high center near 35°S, 10°W. This produces a northward pressure gradient over much of the tropical South Atlantic west of 0°. In the corrected wind data (Fig. 4c), the flow indicates stronger Southern Hemisphere trade winds over these longitudes, though perhaps a little stronger than that expected from the SLP gradients in Fig. 4b. In 88 – 87, both the SLP and wind suggest that no such large-scale enhancement occurred. So in the tropical South Atlantic, the SST and the surface atmospheric features in W – D are very different from those in 88 – 87, except for close to Brazil. Indeed, the region close to Brazil is the only part of the tropical South Atlantic that has statistically significant linear correlations between SLP and Sahelian rainfall during the years 1949–88 (Fig. 7).

3) EXTRATROPICAL NORTH ATLANTIC

and MEDITERRANEAN

In W – D (Fig. 4a), positive SST values exist throughout the extratropical North Atlantic, except close to the United Kingdom and the Iberian peninsula. The positive values are statistically significant near Greenland and at about 35°N. However, the cooling in the North Atlantic near Greenland, observed over the last two decades (Folland et al. 1990), showed no sign of reversal in 1988; indeed 1988 was actually colder than 1987 in this region (Fig. 5a).

In W – D (Fig. 4b), the main feature of the atmospheric maps is a region of negative SLP near the United Kingdom, with accompanying cyclonic circulation. This association is statistically significant when measured using correlation coefficients for 1949–88 (Fig. 7). Large negative correlations with SLP are also found in the eastern Mediterranean (Fig. 7), and in W – D and 88 – 87 this region has low SLP and high SST.

4) INDIAN OCEAN

In W – D (Fig. 4b) and 88 – 87 (Fig. 5b), SLP values decrease by about 1.0–1.5 mb from the equatorial African coast to the coast of the Indian subcontinent. Enhanced Indian monsoonal flow is expected from such a pressure gradient. Enhanced monsoonal flow is found in the 88 – 87 map (Fig. 5c) and in the corrected data for W – D (Fig. 4c), but not in the raw data for W – D (Fig. 4d). This gives further support to the corrected wind data.

The low SLP north of the equator in W – D is accompanied by high SLP south of the equator, just north of the climatological center of the Mascarene high. The pressure gradient across the equator appears enhanced. An enhanced cross-equatorial flow is therefore expected, but this is not present, even in the corrected data. However, the equator marks a boundary in the regions used for the wind corrections; if the corrections had been smoothed across this boundary, the data just to the south of the equator would have had substantially larger corrections applied (Fig. 2), making them more consistent with the SLP in W – D.

Except in parts of the eastern Indian Ocean and close to the southeastern African coast, the negative SST values of W – D (Fig. 4a) are also found in 88 – 87 (Fig. 5a). However, the stronger Mascarene high and easterly winds of W – D are the direct opposite of the low SLP and stronger westerlies in 88 – 87. Over the whole 40-year period, there is no statistical correlation between July–September SST and surface atmospheric features (SLP or vector wind) in this region. This suggests either that there is little ocean–atmosphere interaction on a seasonal time scale in this region, or that the data are too unreliable to detect the interaction, or that the interaction is more complex. During 1949–88, the linear correlation between SLP near the Mascarene high and Sahelian rainfall is actually quite weak (Fig. 7). It is the negative correlations between Sahelian rainfall and SLP in the north Indian Ocean that are statistically significant.

5) PACIFIC OCEAN

Fields for 88 – 87 (Fig. 5) are dominated by the switch from El Niño (1987) to La Niña (1988) that
occurred in boreal spring 1988. It is known that the overall statistical link between ENSO and Sahelian rainfall is fairly weak (Ropelewski and Halpert 1987 1989). However, this might not be true for certain kinds of ENSO variations. The period from 1987 to 1988 was remarkable for the very great and rapid transition between strong El Niño (1987) and strong La Niña (1988). It is therefore important to study features of the 88 – 87 maps in the Pacific that may have caused this particular ENSO cycle to have a pronounced impact on the difference in Sahelian rainfall between the two years. This possibility is enhanced by the apparently inconsistent changes in SST in the Atlantic between W – D and 88 – 87 discussed previously.

The 88 – 87 SST field (Fig. 5a) has a very strong gradient directed north to south over the longitudes 110°–170°W. This is accompanied by a strong south-to-north SLP gradient and southerly wind flow, leading to strong convergence near 15°N. These features, albeit much weakened, are present in the W – D fields of SST (Fig. 4a), SLP (Fig. 4b), and corrected wind (Fig. 4c). In the western equatorial Pacific the 88 – 87 SST field has another strong gradient, this time principally directed west to east, resulting from large negative SST values in the central equatorial Pacific and large positive values near New Guinea, which form a local maximum in the SST difference values since there are also lower values to the west in the Indian Ocean and to the north. The W – D SST field does also have a relative maximum of SST values near New Guinea. The magnitudes of the gradients are weaker than in 88 – 87 but this is to be expected since W – D is a composite of years, only some of which may have had similar patterns to 88 – 87 (the next section discusses some further individual years). The atmospheric features in W – D and 88 – 87 are also quite similar over the longitudes of the western Pacific. Both have a pair of low-pressure areas centered near New Guinea and southern Japan, where high SLP values are found to the east. Correlations between SLP and Sahelian rainfall during 1949–88 identify this pattern as statistically significant (Fig. 7), and these correlations make the main contribution to the field significance of the correlations over the Pacific Ocean (Table 1).

c. Discussion (including some individual years)

The W – D composite may contain clues to processes that operate on interdecadal and interannual time scales. Other authors have started to separate the processes empirically using various kinds of empirical orthogonal function (EOF) analysis (Lough 1986; Wolter 1989; Folland et al. 1991). All of these studies identified a mode in the tropical Atlantic that was mainly related to short-period (<5 years) variability in Sahelian rainfall. Wolter (1989) also included the tropical Indian Ocean in his analyses and found SST here to be involved in a mode that operated mainly on interdecadal time scales. This mode may be part of an interhemispheric variation of SST, an idea introduced in Folland et al. (1986) and isolated as the third EOF of worldwide SST variations in Folland et al. (1991). Wolter’s EOF included atmospheric data as well, but may have been influenced by the problem in wind speed data discussed in section 3. In particular, implicit in Wolter’s interdecadal time-scale EOF is the suggestion that a stronger Indian monsoon tends to accompany reduced rainfall in the Sahel, whereas the opposite is suggested by the SLP correlation map (Fig. 7) and by the corrected wind in W – D (Fig. 4c).

The three SST EOFs that Folland et al. (1991) discuss in relation to Sahelian rainfall are shown here in Fig. 8. The July–September time coefficients of these EOFs have been standardized for the period 1949–88, and the mean difference in the values has been calculated for W – D and 88 – 87 (Table 2). The W – D composite (Fig. 4a) contains a strong component of interhemispheric variation in SST (Northern Hemisphere warmer than Southern Hemisphere). The very negative W – D coefficient difference for EOF3 (Fig. 8b and Table 2) reflects this. The EOF3 difference for 88 – 87 is also negative (Table 2), but this is unlikely to explain all the rainfall difference (also shown in Table 2), especially since Folland et al. (1991) show that EOF3 is mainly related to low-frequency variations in Sahelian rainfall, not large interannual fluctuations, such as occurred in 88 – 87.

The 88 – 87 fields in Fig. 5 do not show the large-scale tropical Atlantic features found in W – D and in other studies of Sahelian rainfall variability. The SST rotated EOF2 (Fig. 8c) that Folland et al. (1991) use to represent the relationship between Sahelian rainfall and tropical Atlantic SST has for 88 – 87 a slightly positive value, in contrast to the statistically significant negative value for W – D (Table 2). The SST EOF (Fig. 8a) that Folland et al. use to represent El Niño does show a very large difference 88 – 87, consistent with the ENSO variation that occurred in

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<th>Table 2. Differences in Sahelian rainfall and SST EOFs for selected sets of years.</th>
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Notes:
Values are of annual Sahelian rainfall (Fig. 1b) and the July–September time coefficients of the SST EOFs (Fig. 8); units are standard deviations (all four series were standardized to have mean of zero and standard deviation of 1 for 1949–88).


FIG. 8. Covariance eigenvectors (EOFs) of seasonal $10^\circ$ lat $\times 10^\circ$ long sea surface temperature anomalies for 1901–80. (a) Unrotated EOF2, (b) unrotated EOF3, and (c) rotated EOF2 (derived by VARIMAX rotation of EOFs 4–13). (Taken from Folland et al. 1991.)
these years. However, as previously noted, there is only a weak link between measures of ENSO (such as this EOF) and Sahelian rainfall; the difference in the coefficient for $W - D$ is not statistically significant (Table 2).

The particular kind of atmospheric surface pattern in $88 - 87$ (Figs. 5b, c) in the tropical Indian and western Pacific oceans does closely resemble the pattern of variability that Meehl (1987) identified and related to stronger Indian summer monsoons this century. The results in $W - D$ and especially the SLP correlation map (Fig. 7) indicate that the $88 - 87$ tropical pattern of seasonal variation through the longitudes of the Indian and western Pacific oceans may have occurred quite often in the past with a similar association to Sahelian rainfall. GCM integrations forced with the observed SST for 1987 and 1988 have successfully simulated substantially more rainfall in the Sahel in 1988 than in 1987 (Rowell et al. 1991; Palmer et al. 1992). Given the small changes in SST in the tropical Atlantic in $88 - 87$ and the large changes in the tropical Pacific (Fig. 5a), it seems that Pacific SSTs can in some years have a strong influence on Sahelian rainfall. However, it is clearly insufficient to identify whether a year is "El Niño" or "La Niña." This is well illustrated by the fact that two "El Niño" years, 1972 and 1958, appear in the dry and wet composites, respectively. The SST and SLP anomalies for these two years (Fig. 9a-d) have an El Niño signal in the central tropical Pacific. However, in the western tropical Pacific and tropical Indian oceans, the SST and SLP, especially in terms of their gradients, are generally opposite and broadly consistent with the teleconnection relationships identified in Figs. 4, 5, and 7. The tropical Atlantic is also broadly consistent with the teleconnection relationships, though in the south tropical Atlantic, the southern part of the tropical Atlantic SST dipole is close to the Guinean coast in 1958, but weaker and farther south and west in 1972. Janicot (1992b) relates such variations to the existence or absence of a rainfall anomaly dipole across about $10^\circ$N in West Africa. Consistent with Janicot (1992b), the dipole is present in 1958, but not in 1972 (Fig. 10a, b).

Lamb and Peppler (1992) discuss 1975 in the context of it being the wettest year in the period 1970–88 in their west sub-Saharan rainfall index, which covers the latitudes $11^\circ$–$18^\circ$N and longitudes west of $10^\circ$E. The 1975 rainfall anomaly map (Fig. 10c) shows generally near zero anomalies and in some places large positive anomalies (relative to 1951–80) over sub-Saharan Africa north of $7.5^\circ$N, confirming that 1975 was a substantially wetter year within the drought period. In 1975, a La Niña year, the SLP anomaly pattern (Fig. 9f) in the western tropical Pacific and Indian ocean resembles the pattern in $W - D$ and in $88 - 87$ and furthermore, there is a maximum in SST anomalies near New Guinea (Fig. 9e). So a relatively wet year in 1975 is consistent with the teleconnection relationships identified over the tropical Indian and tropical western Pacific in Figs. 4, 5, and 7. Lamb and Peppler (1992) cannot fully account for the return to nearer normal rainfall in 1975 through reference to tropical Atlantic fields alone, and in Fig. 9e, the tropical Atlantic SST anomalies do not seem consistent with a relatively wetter year. Lamb and Peppler (1992) are also unable to account for the widespread drought of 1983 (Fig. 10d) using tropical Atlantic fields alone, which they note have a particularly unusual structure. In addition to the tropical Atlantic, the tropical Indian and tropical Pacific fields are very unusual in this year. There are very large positive SST anomalies in the tropical Pacific (Fig. 9g), but confined much farther east than in typical July–September phases of El Niño (Halpert and Ropelewski 1989). In fact, the SST composite for $W - D$ (Fig. 4a) is quite sensitive to the inclusion of 1983, which may result in too much emphasis on SST changes in the eastern tropical Pacific. The atmospheric pattern in the tropical Pacific and Indian oceans is also unusual in 1983 (Fig. 9h). Correlating the 1983 tropical Indian and Pacific SLP anomalies in Fig. 9h with those of other years, the highest positive correlation (0.40) is found with 1949, which is a moderate La Niña year. In contrast, the SLP anomaly pattern in the tropical Indian and Pacific in 1972 seems more typical; there are five years with a positive correlation $r > 0.45$ (1986: 0.46, 1979: 0.49, 1965: 0.55, 1987: 0.68, and 1982: 0.70) and a number of years when a generally opposite SLP anomaly pattern has prevailed ($r < -0.60$ for 1988: $-0.61$, 1970: $-0.62$, 1955: $-0.63$, 1964: $-0.68$, and 1975: $-0.72$). The conclusion from these results is that the relationship between the 1983 drought in the Sahel and large-scale ocean–atmosphere interactions, which appears to be physically real given the model results of Folland et al. (1991), has a structure in the western tropical Pacific and Indian oceans that is different from that identified in Figs. 4, 5, and 7.

5. Conclusions

Analyses in this paper have used a surface wind dataset that has been provisionally corrected for time-varying bias in wind speed reports from ships. The estimates of bias (Fig. 2) support Cardone et al. (1990), who argue for a careful appraisal of the methods of wind observation on ships. The bias estimates are based on comparisons of derived (mainly geostrophic) winds with reported winds. A more accurate derived wind incorporating ageostrophic components more rigorously, would justify calculation of corrections with better temporal and spatial resolution, improving on the current corrections, which are linear in time and uniform over large areas and all seasons of the year. Nonetheless, the current corrections provide useful quantification of the uncertainty in the ship wind data and of the impact the data problems may have for de-
FIG. 10. July–September rainfall anomalies (millimeters, relative to 1951–80 mean) for (a) 1958, (b) 1972, (c) 1975, and (d) 1983.
tecting decadal climate variability. The impact of the corrections on the physical interpretation of the $W - D$ Sahelian composite (Fig. 4) is largest in the tropical Atlantic and Indian oceans.

Recent research has presented a strong case, using empirical evidence and GCM experiments, for SST patterns, including patterns outside the tropical Atlantic, forcing at least some and possibly most of the observed variation in seasonal Sahelian rainfall. This paper has described some of the near-surface worldwide atmospheric variations over the oceans that have accompanied the worldwide SST variations. For the remote SST to be important, as suggested by Folland et al. (1991), some of the remote worldwide atmospheric variations that have been identified must be related to the dynamics of the Sahelian atmosphere. A combination of GCM and empirical study is needed to identify the nature of the teleconnecting mechanisms. That different large-scale mechanisms are important in different years is supported by the fact that distinctive seasonal rainfall anomaly types have been identified over sub-Saharan Africa (Nicholson 1980; Janicot 1992a), and by the ability of all the major ocean basins to force Sahelian rainfall variations in GCMs (Palmer 1986; Folland et al. 1991). Furthermore, some regional-scale variations in rainfall anomalies within the Sahel itself appear related to ocean–atmosphere variability (Ward et al. 1990; and modeled by Druyan and Koster 1989). Further studies will need to consider which processes operate on the Sahel as a whole and which processes are regionally specific.

Acknowledgments. Jon Maybury helped devise and program the geostrophic wind analysis. Discussions with David Parker are always valued. This work is part of an ongoing Ph.D. study under the supervision of Brian Hoskins (Meteorological Department, Reading University) and Chris Folland (Hadley Centre for Climate Prediction and Research, Meteorological Office), Mike Hulme (Climatic Research Unit, University of East Anglia) kindly helped by producing the gridded rainfall dataset under UK DoE Contract PECD 7/12/78.

APPENDIX A

Calculating the Seasonal Mean Geostrophic Wind

To calculate geostrophic winds, estimates of pressure gradients are needed. These could be made by fitting a smooth surface to the SLP data, but the resulting gradients may be systematically influenced by data reliability, so that through time, as data become more reliable, a trend in the estimated gradients may be introduced. This needs to be avoided if the geostrophic wind trend is to be treated as “truth” in comparisons with observed wind trends. The basic strategy adopted involves applying the following process 120 times (once for every season 1949–88) to every $2 \times 2$ ocean box:

(i) Search for the pair of $2 \times 2$ seasonal SLP values that provide the most reliable estimate of seasonal mean SLP gradient across the $2 \times 2$ box.

(ii) Calculate the geostrophic wind that the SLP gradient implies.

(iii) Make approximate adjustments to the geostrophic wind for friction and other ageostrophic components.

In more detail, consider a pair of $2 \times 2$ boxes (see Fig. A1) yielding a seasonal pressure difference $\delta P$:

$$\delta P = P_a - P_b$$

(A1)

where $P_a$ is a seasonal mean SLP for a $2 \times 2$ box in COADS and is taken as an estimate of the seasonal mean SLP at the mean location $a$ of the observations through the season; $P_b$ is the mean SLP for the same season in a nearby $2 \times 2$ box. An estimate of the pressure gradient at the midpoint ($m$) of the line joining $a$ and $b$ and in the direction ($n$) defined by the direction of the line joining $a$ and $b$ is

$$\nabla_n P = \frac{\delta P}{D}$$

(A2)

where $D$ is the distance between points $a$ and $b$. For a particular $2 \times 2$ box $i$, surrounding pairs of boxes were analyzed in turn as candidates for providing a reliable estimate of the pressure gradient across box $i$. For a pair of boxes to qualify as candidates, the following criteria needed to be satisfied (see Fig. A1):

(i) $P_a$ and $P_b$ must both be based on at least five observations.

(ii) The angle $\theta$ between the line joining $a$ and $b$ and the “preferred axis” must be less than 20°. With

![Fig. A1. Identification of those pairs of boxes that qualify as candidates for providing an estimate of mean SLP gradient across box $i$ for a given season. Each pair of nearby $2^\circ \times 2^\circ$ boxes are considered. The point $c$ is the center of box $i$. $a$ and $b$ are the respective mean locations of the season’s SLP observations in the pair of boxes being considered, $m$ is the midpoint of the line joining $a$ and $b$, $\theta$ is the angle between the “preferred axis” and the line joining $a$ and $b.$](image)
the exception of the regions identified in Fig. A2, the “preferred axis” is always 15° clockwise from north–south in the Northern Hemisphere (as in Fig. A1), and 15° counterclockwise in the Southern Hemisphere, so the pressure gradient along the preferred axis yields a geostrophic wind, which can be compared with the observed zonal wind, assuming 15° backing of the geostrophic wind due to friction. The regions identified in Fig. A2 have data coverage mainly confined to narrow shipping lanes. To make best use of the SLP data in these regions, the preferred axis is orientated along the shipping lane.

(iii) The distance between $a$ and $b$ should be between 500–1000 km (effectively defining a length scale for the pressure gradient estimates).

(iv) The midpoint $m$ of the line joining $a$ and $b$ should be within 200 km of the center $c$ of box $i$.

(v) The mean date of the observations that yielded $P_a$ and $P_b$ must not differ by more than 15 days.

To identify which of the qualifying pairs would provide the most reliable estimate of $\nabla nP$, the standard error (SE) of $\nabla nP$ was estimated as

$$SE(\nabla nP) = \frac{\left[ SE(P_a)^2 + SE(P_b)^2 \right]^{1/2}}{D}.$$  \hspace{1cm} (A3)

The SE on the estimate of mean SLP for point $a$ is a function of the number and distribution of observations through the season, temporal SLP variability through the season, and spatial variability in the instantaneous SLP fields. The latter is the error introduced by assuming that all observations were taken at point $a$. Generally, the spatial variability over about 2° of latitude and longitude will be much less than temporal variability through the season, so spatial variability is ignored. In the absence of serial correlation and assuming $P_a$ is an estimate of the mean of an infinite population, $SE(P_a)$ is $\sigma(n)^{-0.5}$, where $\sigma$ is the standard deviation of the observations and $n$ is the number of observations. However, the SLP observations do contain serial correlation, and $P_a$ is an estimate of a mean over a restricted time period (the season). Assuming missing data are randomly distributed through the sample, following Parker (1984) and Kidson and Trenberth (1988):

$$SE = \left( \frac{\sigma}{\sqrt{n'}} \right) \left( 1 - \frac{n'}{N'} \right)^{1/2}$$ \hspace{1cm} (A4)

where $n'$ (the effective sample size) is a function of the number of SLP observations ($n$), the finite population size ($N$), and the serial correlation in the SLP observations ($r_i$, the serial correlation at lag $i$):

$$n' = \frac{n}{1 + \left( \frac{2n}{N^2} \right) ([N - 1] r_1 + [N - 2] r_2 + \cdots)}$$ \hspace{1cm} (A5)

$N'$ (the effective finite population size) is given by

$$N' = \frac{N}{1 + (2/N) ([N - 1] r_1 + [N - 2] r_2 + \cdots)}.$$ \hspace{1cm} (A6)

These equations cannot be applied immediately because they assume that the population is a time series sampled at regular time intervals. The ship observations are sampled at irregular time intervals, and they are continuously distributed in space. However, a number of assumptions can be made to allow an approximation.

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Fig. A2. Preferred axis along which a pressure gradient was sought. Arrow indicates the axis used for all $2^\circ \times 2^\circ$ boxes contained within the box marked by a solid line. Where a preferred axis is not specified, it was taken to be 15° clockwise from the north–south direction in the Northern Hemisphere, 15° counterclockwise in the Southern Hemisphere. In any given season, the two boxes used for the pressure gradient estimate were permitted to form an axis up to 20° from the “preferred axis.”
to the parameters needed for the above equations. First, though the serial correlation will be a function of sampling interval, a first-order autoregressive process with $r_1 = 0.8$ has been assumed throughout (consistent with Kidson and Trenberth 1988). This is a reasonable estimate for SLP in most ocean regions for a sampling interval of 12 hours. If the number of observations was less than 30, it was assumed that the finite population size was 60; $N$ was linearly increased to 150 as $n$ increased to 100. Any box with more than 100 observations was assumed to have had just 100. This practical choice was made to avoid the few boxes with many more than 100 observations dominating the results by always being selected to estimate the pressure gradients.

More sophisticated statistical treatment viewing the time series as a sample of a continuous process would be more accurate, but the aforementioned procedure was judged adequate for identifying the most reliable pressure gradient estimate. Errors associated with the individual observations themselves (e.g., instrumental, human) are ignored since we can only assume that these will on average have the same statistical properties (mean, variance) for all nearby $2 \times 2$ boxes.

The pair of boxes yielding the smallest value for $\text{SE}(\nabla_n P)$ was used to estimate the geostrophic wind:

$$G_{\alpha'} = \frac{-\nabla_n P}{f\rho},$$  \hspace{1cm} (A7)

where $f$ is the Coriolis force and $\rho$ is the density of air (standard sea level value of 1.225 kg m$^{-3}$ is assumed).

The true wind differs from the geostrophic wind due to various ageostrophic components. An adjustment to $G_{\alpha'}$ has been made to allow for these in two stages:

(i) Friction reduces the true near-surface wind speed below the geostrophic wind speed; the magnitude of the 10-m geostrophic wind including the effects of friction ($G_{\alpha'F}$) can be approximated as (Garratt 1977):

$$G_{\alpha'F} = r G_{\alpha'},$$

$$r = 1.91 f^{-0.074} w^{-0.139}$$  \hspace{1cm} (A8)

where $r$ is the ratio of the observed to geostrophic wind, and $w$ is taken here to be the mean reported wind speed for box $i$ in the season. The ratio of the geostrophic to observed wind is difficult to generalize for wind speeds less than about 4 m s$^{-1}$, when Eq. (A8) is inappropriate. For simplicity, when the mean wind speed is less than 4 m s$^{-1}$, the ratio $r$ is assumed to be that calculated for a wind speed of 4 m s$^{-1}$.

(ii) For other ageostrophic components, allowance was made for systematic departures of the $G_{\alpha'F}$ from the comparable observed vector wind by simply multiplying $G_{\alpha'F}$ by a constant to force the mean of the derived wind to be the same as the mean of the observed wind during the entire period 1949–88:

$$\mathbf{G}_{\alpha'} = G_{\alpha'F} \left[ \frac{\mathbf{V}_\alpha}{G_{\alpha'F}} \right]$$  \hspace{1cm} (A9)

where $G_{\alpha'F}$ is the mean of all the estimates of $G_{\alpha'F}$ that are made for box $i$ in the 40 seasons 1949–88, and $\mathbf{V}_\alpha$ is the mean of all the comparable observed winds in the 40 seasons, $\alpha$ differing from $\alpha'$ by 15° to make an approximate allowance for the impact of friction on the direction of the observed wind. Equation (A9) is obviously an approximation that is used to simplify the analysis. Future versions of the wind corrections may, in addition, partly force the mean of $V_\alpha$ and $G_{\alpha'F}$ to be the same by first adding a constant amount to each $V_\alpha$. Such a constant addition may be preferable in view of the nature of the bias in the conversion of Beaufort reports to wind speed (Cardone et al. 1990; Ismer and Hasse 1991). The use of (A9) may have slightly biased the estimates of $S$ in those regions where there is a large trend in the geostrophic wind.

**APPENDIX B**

**The Relationship between Trends in Seasonal Mean Vector Wind and Seasonal Mean Scalar Wind Speed**

Cardone et al. (1990) argue that the scalar wind speed (i.e., magnitude of the wind speed) reported by ships has increased due to changes in observational practice. It is important to identify if and how vector trends (e.g., trends in the seasonal mean zonal wind) can be used as a good approximation for trends in wind speeds, since it is vector wind trends that must be verified against trends in the derived vector wind $G_{\alpha'}$ in appendix A.

Consider sampling an identical season twice with an identically timed set of reports using 1) Beaufort estimates and 2) anemometer readings. Following Cardone et al. (1990), the difference between each pair of identically timed reports will be a nonlinear function of stability and wind speed. However, for simplicity, assume that each anemometer wind speed report $w_a$ is always larger than the Beaufort wind speed report $w_b$ by the same percentage amount $S$:

$$w_a = w_b + \frac{S}{100} w_b.$$  \hspace{1cm} (B1)

The corresponding zonal wind reports will be

$$u_a = w_a \sin \theta$$

$$u_b = w_b \sin \theta$$  \hspace{1cm} (B2)

where $\theta$ is the angle of the wind measured clockwise from north. Substituting between Eqs. (B1) and (B2), it follows that

$$u_a = u_b + \frac{S}{100} u_b.$$  \hspace{1cm} (B3)

Equation (B3) shows that the modulus of the anemometer vector wind has the same percentage increment as the anemometer wind speed. So over a period of time, an increase in the ratio of anemometer to Beaufort reports will yield the same percentage increase.
The relationship between the trend in a vector quantity and the trend in its modulus. The graph shows the trend in the modulus of $V$ as a function of the mean vector wind ($\bar{V}$).

In principle, trends in $|V|$ can therefore be used to yield $S$. However, trends in $|V|$ are strongly influenced by trends in data reliability whenever the mean seasonal vector wind ($\bar{V}$) is close to zero. In the early years of the historical record, less reliable data introduce erroneous outliers that raise the mean value of $|V|$. So reliable unbiased estimates of the trend in $|V|$ cannot be made. However, it is possible to make an unbiased estimate of the trend in the seasonal vector wind ($\bar{V}$) and use this trend as an approximation for the trend in $|V|$. In order to assess the accuracy of this approximation, consider a vector wind time series to be made up of trends in $|V|$, $\bar{V}$, and white noise. Figure B1 assumes the white noise to have a standard deviation of 0.75 m s$^{-1}$ (observed order of magnitude for seasonal zonal wind) and the trend of $|V|$ to be 0.005 m s$^{-1}$ yr$^{-1}$ [consistent with that proposed by Cardone et al. (1990)]. The trend in $\bar{V}$ tends to 0.005 m s$^{-1}$ yr$^{-1}$ as $\bar{V}$ becomes different from zero. For $\bar{V} = 1$ m s$^{-1}$, the trend in $\bar{V}$ is a good approximation of the trend in $|V|$, biased low by only about 15%. Series with $\bar{V}$ in the range $-1$ m s$^{-1}$ to $+1$ m s$^{-1}$ were judged unsuitable for estimating the trend in $|V|$. Series with $\bar{V} < -1$ m s$^{-1}$ were used after being multiplied by $-1$, so that trends in $\bar{V}$ and $|V|$ were of the same sign.

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