Wavenumber Spectra of Pacific Winds Measured by the Seasat Scatterometer

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ABSTRACT

Vector winds measured by the Seasat-A Satellite Scatterometer (SASS) are analyzed to determine the spatial structure of oceanic surface winds over wavelengths from 200 to 2200 km. The analysis is performed in four latitudinal bands in the Pacific Ocean. Sampling characteristics of SASS preclude the possibility of determining full two-dimensional spectra; the analysis is therefore limited to one-dimensional (along the satellite ground track) spectra of vector wind components and kinetic energy.

The salient features of the results are summarized as follows. (i) For each of the four geographic regions, the spectra of meridional and zonal wind components and of kinetic energy are consistent with a power-law dependence on wavenumber; for midlatitude regions in both the Northern and Southern hemispheres the wavenumber dependence of kinetic energy is $k^{-2.2}$, while for tropical regions in both hemispheres it is $k^{-1.5}$. (ii) For each individual region, the spectral dependence on wavenumber is nearly the same for both velocity components and for kinetic energy. (iii) Comparisons of zonal and meridional component spectra indicate that midlatitude winds may be isotropic, while tropical winds may be significantly anisotropic. (iv) The coherence between wind components is small everywhere.

1. Introduction

Characterization of the spatial variability of winds on scales from hundreds to thousands of kilometers is critical for many studies of oceanic and atmospheric processes. Limitations in conventional observation networks have restricted most field investigations to consideration of large-scale (greater than 2000 km) or very small-scale (less than 10 km) winds. However, it has long been recognized that atmospheric motions with intermediate spatial scales contain significant energy and can play an important role in the dynamics of the atmosphere and in atmospheric forcing of the ocean. These intermediate scales cannot be resolved by most measuring techniques. In the present study we use high resolution vector wind measurements from the Seasat-A Satellite Scatterometer (SASS) to examine, for the first time, the spatial structure of near-surface oceanic winds on scales from 200 to 2200 km.

Most studies of the temporal variability of the upper ocean require accurate, high resolution specification of surface winds. One application for which the present work is particularly relevant is the study of wind generation of mesoscale oceanic eddies. Frankignoul and Muller (1979), Willebrand et al. (1980), and Muller and Frankignoul (1981) have hypothesized that random (in time) atmospheric motions with scales of 50-4000 km can directly generate and maintain mesoscale oceanic eddies in midocean regions. These investigations have utilized the results of historical studies of surface winds. The extreme paucity of such data on the relatively small scales of interest has severely hampered modeling studies of the effects of winds on oceanic eddies (Schmitt et al., 1983). The works of Willebrand (1978) and Goldenberg and O'Brien (1981) are, to our knowledge, the only previous studies of the spatial variability of oceanic surface winds. However, as emphasized by Willebrand, his “winds” were calculated from synoptic maps of surface pressure using geostrophy. Due to smoothing in the production of the pressure maps, the resulting surface wind fields calculated by Willebrand were incapable of resolving wind variability with spatial scales less than approximately 900 km. In addition, as the pressure maps were initially based on ship and weather station reports which are sparse in the Southern Hemisphere, the analysis was restricted to the Northern Hemisphere (north of 20°). The work of Goldenberg and O'Brien (1981) used tropical Pacific wind fields constructed subjectively from in situ reports, and concentrated on latitudinal variations in the low zonal wavenumber portion of the spectrum.

Presently available global wind data are also inadequate for atmospheric applications. Studies of atmospheric predictability by Lorenz (1969), Leith (1971), and Charney (1971) have shown that small-
scale motions, unresolved by numerical models and conventional observation networks, can strongly influence the large-scale winds and weather systems in the real atmosphere. These studies have established the central role of turbulence models for describing the statistics of atmospheric variability. In addition, they have drawn attention to the importance of accurate knowledge of the kinetic energy spectrum of atmospheric motions on meso- and synoptic scales.

Recent measurements of upper tropospheric winds from commercial aircraft (Lilly and Peterson, 1983; Nastrom and Gage, 1983; Nastrom et al., 1984; Nastrom and Gage, 1985) have provided the first observations of winds over the oceans on scales from tens to thousands of kilometers. Spectra computed from these data appear to be consistent with several predictions of two-dimensional isotropic turbulence models. However, the limited geographical coverage (the measurements were restricted to selected Northern Hemisphere airline flight routes) and high altitudes (9–13 km) at which the measurements were acquired limit the usefulness of the data for studies requiring global knowledge of near-surface winds.

In this paper, we analyze near-surface oceanic wind measurements by SASS to examine the wavenumber spectra of component wind speeds and kinetic energy over wavelengths from 200 to 2200 km. Initially, our goal was to compute full two-dimensional spectra in order to provide a complete description of the horizontal variability of winds on these scales. However, the sampling characteristics of the SASS data precluded the possibility of such computations. Our analysis is thus restricted to one-dimensional (along the satellite ground track) spectra. As in nearly all past studies of atmospheric variability from conventional wind data, we use the kinematics of two-dimensional isotropic turbulence theory to infer properties of full two-dimensional spectra from one-dimensional spectra of zonal and meridional velocity components.

We begin with a brief description in section 2 of the SASS instrument and its sampling characteristics. Section 3 reviews models of atmospheric turbulence and summarizes the relationship between one-dimensional and two-dimensional spectra under the assumption of nondivergent and isotropic flow. A summary of past observational studies is given in section 4, and the results of our analysis of the SASS data are presented in section 5. A discussion of the results in light of the predictions of two-dimensional isotropic turbulence theory is given in section 6.

2. SASS instrument and data description

The Seasat-A Satellite Scatterometer measured near-surface vector winds over the oceans during the Seasat mission from 7 July–10 October 1978. Seasat was the first satellite dedicated to global observation of the oceans. It orbited at an altitude of about 800 km with an inclination angle of 108° and an orbital period of about 101 minutes. Several reviews have been written addressing aspects of the SASS flight hardware (Granham et al., 1977; Johnson et al., 1980), data reduction (Bracalante et al., 1980; Schroeder et al., 1982; Boggs, 1982), and calibration (Jones et al., 1982; Brown, 1983). This section briefly reviews the principles of satellite scatterometry and those aspects of the design of the SASS instrument relevant to the spatial and temporal sampling characteristics of the data used in the present study. We follow with a brief summary of the processing used to obtain unambiguous vector winds from the raw SASS backscatter data. The final subsection summarizes the spectral analysis used in this paper to extract information on the spatial variability of sea surface winds.

a. SASS instrument description

SASS was an active microwave radar instrument designed to make frequent global high-resolution measurements of near-surface vector winds over the oceans. Measurements of the power of transmitted and backscattered microwave radiation were used to calculate the normalized radar cross section ($\sigma_0$) of the ocean’s surface. Much empirical work since World War II (see Moore and Fung, 1979, for a review) establishes a strong correlation between $\sigma_0$ at microwave frequencies and local wind velocity. The empirically based relationship between $\sigma_0$ and local vector winds can be summarized in a “geophysical model function” of the form:

$$\sigma_0 = G(\theta, \chi, \epsilon) + H(\theta, \chi, \epsilon) \log_{10} V,$$

where $\sigma_0$ is expressed in dB, $V$ is the equivalent neutral stability wind speed at 19.5 m height, and the empirical coefficients $G$ and $H$ are functions only of the incidence angle $\theta$ (the angle, measured in the vertical plane, between the incident radiation and the normal to the mean sea surface), the azimuth angle $\chi$ (the angle, measured in the horizontal plane, between the incident radiation and the wind direction), and the polarization $\epsilon$ (horizontal or vertical) of the transmitted and received radiation.

Radar backscatter at small incidence angles ($|\theta| \leq 10^\circ$) is due to specular reflection from areas on the sea surface that have been tilted by ocean waves long compared with the wavelength of the incident radiation. As wind speed (and wave slope) increase, more of the incident radiation is reflected away from the satellite receiver so that the coefficient $H$ in the geophysical model function is negative. At these small incidence angles, both $G$ and $H$ are extremely weak functions of $\chi$. Thus, while measurements of $\sigma_0$ at small incidence angles can be used to measure wind speed, they cannot yield any information about wind direction.
For incidence angles in the range $10^\circ \leq \theta \leq 15^\circ$, $\sigma_0$ is relatively insensitive to both wind speed and wind direction, and thus measurements of backscattered power cannot be used to infer any information about the winds at all (cf. Fig. 1).

In the incidence angle range $20^\circ \leq \theta \leq 60^\circ$, backscattered radiation is due to resonant Bragg scattering from short ocean waves having wavelengths comparable to that of the incident electromagnetic radiation (approximately 2 cm for the SASS radar). The amplitudes of these short ocean waves increase with increasing wind speed so that the coefficient $H$ in the model function is positive. Figure 1a illustrates the variation of the coefficient $H$ with incidence angle at fixed azimuth angle $\chi$. Empirical studies have demonstrated that the power of the backscattered microwave radiation at these incidence angles is a sensitive function of both wind speed and wind direction (Jones et al., 1977). Figure 1b shows that the relationship between $\sigma_0$ and $\chi$ is nearly $\sigma_0 \sim \cos(2\chi)$. The cross-section $\sigma_0$ varies nearly harmonically with azimuth angle, having maxima at upwind ($\chi = 0$) and downwind ($\chi = \pi$), and minima at cross-wind angles ($\chi = \pm \pi/2$). The amplitudes of the modulations are functions of incidence angle, wind speed and polarization.

Given two or more independent measurements of $\sigma_0$ from the same ocean region, but with different azimuth angles, the geophysical model function can be inverted to determine both wind speed and direction (e.g., see Jones et al., 1982; Pierson, 1983). In practice, however, the inversion results in multiple solutions due to a combination of instrumental noise, geophysical noise, and the small upwind–downwind asymmetry in the model function. The multiple solutions all have nearly the same speed, but differ widely in direction.

SASS transmitted and received $K_u$-band radiation (14.6 GHz, 2.1 cm wavelength) using four fan-beam antennas arranged to provide ground illumination as shown in Fig. 2. As discussed above, vector winds can be determined only for incidence angles from $20^\circ$ to $60^\circ$. Consequently, vector winds could be obtained from SASS measurements over a pair of swaths, each approximately 500 km wide, separated by an approximately 450 km wide “nadir gap” centered on the subsatellite track. For each antenna beam, $\sigma_0$ was determined for 12 contiguous cells (corresponding to different incidence angles across the swath). Each $\sigma_0$ cell represented spatially averaged backscatter from a sea surface “footprint” with dimensions approximately $18 \times 60$ km, with the long axis aligned with the antenna beam. Approximately colocated measurements of $\sigma_0$ from different azimuth angles were obtained by combining measurements from the fore antenna beam (oriented $45^\circ$ relative to the satellite ground track) with corresponding measurements of the same ocean region using the aft antenna beam (oriented $135^\circ$ relative to the ground track). At the Seasat ground track speed of $6.6$ km s$^{-1}$ and altitude of $800$ km, the time separation between measurements from the fore and aft antennas ranged from about 1 minute at $20^\circ$ incidence angle to 4 minutes at $60^\circ$ incidence angle.

b. SASS data reduction

Processing of raw scatterometer $\sigma_0$ data to multiple solution wind vectors was conducted at the Atmospheric Environment Service (AES) of Canada by S. Petcherych and coworkers (see Baker et al., 1984) using $\sigma_0$ data provided by the Jet Propulsion Laboratory (JPL). Measurements of $\sigma_0$ were binned into $100$ km squares oriented in an earth-located cartesian grid aligned with the subsatellite track. All $\sigma_0$ measurements falling within a single square were treated as colocated for the purpose of vector wind retrieval. Multiple so-

![Fig. 1. Variation of normalized radar cross-section ($\sigma_0$) with wind speed and direction (from the SASS-I model function). (a) $\sigma_0$ (dB) vs wind speed (m s$^{-1}$) at several incidence angles for horizontal polarization and crosswind (90$^\circ$) azimuth angle. Vector winds for this study were calculated in the incidence angle range 20$^\circ$–55$^\circ$. (b) $\sigma_0$ (dB) vs azimuth angle at several wind speeds for horizontal polarization and 30$^\circ$ incidence angle. (Adapted from Bogg, 1982).](image-url)
Solution vector winds were obtained by inverting the so-called SASS-I geophysical model function using a least-squares retrieval algorithm (see Schroeder et al., 1982; Boggs, 1982). The coordinates of the wind measurement within each 100 km square were determined by averaging the latitudes and longitudes of all $\sigma_0$ measurements in the square. The multiple solution wind dataset was therefore irregularly spaced, with approximate 100 km resolution.

Ambiguity removal to obtain a single vector wind solution at each location from among the (up to four) multiple solutions was conducted by teams at JPL, at the University of California at Los Angeles, and at AES Canada. The basis of the technique is described by Wurtele et al. (1982). In order to draw attention to potential limitations inherent in the data analyzed in this study, we briefly summarize the technique here. The first step in ambiguity removal was the generation of plots of AES multiple solution wind vectors from an entire ocean basin. Such coverage required use of data from multiple, consecutive orbits of the satellite. In all cases, the data used in a single plot were acquired within a 12-hour period (P. Woiceshyn, personal communication, 1985). Two teams of analysts, each working independently, constructed streamline analyses of surface winds using subjective pattern-recognition techniques applied to the SASS plots. The subjective analyses were aided by the use of available conventional meteorological measurements, routine weather service forecasts and analyses, and additional data such as satellite cloud photos.

Based on the streamline analyses, regions were identified within the SASS swaths where the subjectively determined wind direction fell within a single quadrant (with axes defined by the instantaneous antenna illumination pattern). Polygons defining these regions were constructed and the vertices of the polygons were digitized for subsequent computer analysis. For each multiple wind solution inside the polygon, the vector with direction closest to the (nearly constant) direction indicated by the streamline analysis was objectively selected by computer. In this way, each analyst team produced a dataset containing unique vector winds. For much of the data thus processed, the results from the two independent teams were compared. In regions where the two analyses differed, the streamline analysis was repeated until a "consensus" dataset was produced in an iterative fashion (P. Woiceshyn, personal communication, 1985).

To date, 15 days of unique vector wind data (6–20 September 1978) have been produced using the technique as described. This dataset forms the basis for the present study. Although errors introduced by inaccuracies in the empirical geophysical model function, the model function inversion process, and the ambiguity removal technique have not been quantified, the 15-day vector wind dataset used in this study represents the best set of global, high-resolution, oceanic surface wind measurements presently in existence.

c. Spectral analysis

In this study, we examine spectra and cross-spectra of velocity components in four regions of the Pacific Ocean (see Fig. 3). Each region is 20° in latitudinal extent, and the boundaries of the regions were chosen to correspond approximately to boundaries between the climatic trade and westerly wind regimes. The choice of these regions allows latitudinal variations in the spectra to be examined. The 20° restriction on the latitudinal extent of each geographical area results in a wavenumber resolution in the latitudinal direction of approximately 0.00045 km$^{-1}$ (corresponding to a maximum resolvable wavelength of ~2200 km). Coupled with the ~100 km resolution of the scatterometer vector winds, the wavelengths that can be examined range from ~200 to 2200 km, corresponding to the "mesoscale-α" regime defined by Orlanski (1975). For motions on such scales, effects due to the spherical shape of the earth are small, and it is reasonable to analyze the data in a two-dimensional Cartesian coordinate system as in Leith (1971).

Figure 4 shows the locations of vector winds from a typical set of three consecutive orbits over the south tropical Pacific (region II). During a single orbit, the satellite traverses the 20° region in approximately 6 minutes. Over the spatial scales resolvable by SASS, variations in the wind field within such a short time period are negligible. Therefore, in the analysis that follows, all data from a single orbit in a given geographical region are treated as though they were acquired at the same instant in time.

Spatial and temporal gaps exist in the SASS data (e.g., see Fig. 4). In addition to the 450 km nadir gap between swaths (see section 2b), there are large gaps
between data obtained from consecutive orbits (which are separated in time by the orbital period of 101 minutes). The spatial extent of interorbit gaps is a function of latitude and varies from approximately 1300 km at the equator to zero at 51° latitude (Boggs, 1982). Poleward of 51° latitude, there is overlap in the measurements from successive orbits. Careful examination of Fig. 4 also shows that, within swaths, the data are irregularly sampled, as discussed earlier in section 2b.

Because of the wide nadir gap and the spatially irregular sampling within swaths, two-dimensional wavenumber spectra could not be computed directly using the usual fast-Fourier transform (FFT) techniques. The approach used here was first to interpolate objectively the irregularly spaced and gappy data to a uniform grid, and then to apply direct FFT techniques for the spectral calculations. The best (in a least-square-error sense) method of objective interpolation is optimal interpolation (Bretherton et al., 1976; Bretherton and McWilliams, 1980). The technique requires some a priori knowledge of the spatial covariance function of the wind field. Based on theoretical considerations (see section 3a), we anticipated that the spectra of surface vector wind components would be approximately isotropic in wavenumber space with a wavenumber dependence somewhere between $k^{-5/3}$ and $k^{-3}$ over the wavenumbers resolvable by the SASS data. The spatial covariance functions corresponding to these spectral shapes were obtained by inverse Fourier transforming the spectra, and we found that the resulting lag covariance functions could be very accurately approximated by exponentially damped cosine functions. These positive definite analytical approximations to the assumed covariance were then used initially to optimally interpolate the SASS vector wind components to a regularly spaced grid oriented parallel and perpendicular to the satellite subtrack. The interpolated area spanned the full cross-track distance from the outer edge of the port swath to the outer edge of the starboard swath for each orbital pass through the geographical region.

An advantage of optimal interpolation over other objective techniques is the ability to estimate the error of each interpolated value based on the statistics of the field. Within each swath where the vector wind field was densely sampled by SASS, the expected error of the optimally interpolated wind field was small. However, the expected errors of the interpolated winds in the nadir gap were large. We were therefore restricted to analysis of interpolated data within each sampled swath.

In practice, an objective Laplacian-spline interpolation scheme was used on each swath to calculate, for
each component separately, component speeds on a 6 \times 20 \text{ evenly spaced grid (111 km spacing)} aligned with the coordinate system. Test cases indicated that the values interpolated in this way did not differ significantly from interpolations obtained using the optimal interpolation technique summarized above. This favorable comparison reflects the fact that the interpolated value is relatively insensitive to the interpolation method in regions where the wind field is adequately sampled (i.e., within each swath). The computational efficiency and accuracy of the Laplacian-spline technique (within swaths) made it superior for the present study. The ability to interpolate accurately to a uniform grid made unnecessary the use of computationally complex techniques for the direct estimation of one-dimensional spectra from irregularly spaced data (Julian and Cline, 1974; Jones, 1975).

The limited spatial coverage afforded by a single swath did not allow calculation of full two-dimensional spectra with useful wavenumber resolution. Specifically, the narrow (~500 km) width of the swaths severely limited the cross-track resolution of the resulting spectra. The analysis presented here is therefore restricted to one-dimensional auto- and cross-spectra of component variances in the along-track direction. Spectra were computed separately for each individual swath containing at least 70 vector wind solutions, with at least one solution within 2° of each of the latitudinal extremes of the geographic region. Vector winds expressed as (speed, direction) were transformed to zonal (+ toward the east) and meridional (+ toward the north) component speeds. For each geographic region, a constant rotation was applied to the spatial coordinates (see Table 1) and the locations of the vector wind data were transformed from spherical latitude and longitude coordinates to cartesian coordinates in the new along-track, cross-track system.

A one-dimensional cosine-bell taper (Julian, 1971; Bendat and Piersol, 1971) was applied to the 10% of the data at each of the along-track extremes of the interpolated grid to reduce the effects of spectral leakage. One-dimensional FFTs of gridded component speeds with constant across-track coordinate were then computed. Autospectra of component speeds were calculated by squaring the Fourier coefficients and averaging in the across-track dimension. Cross-spectra between zonal and meridional components were constructed by multiplying Fourier coefficients and averaging in the cross-track direction. The auto- and cross-spectra were further smoothed by ensemble averaging the spectra from all individual swaths falling within the geographical region during the 15-day extent of the dataset.

### Table 1. Locations and quantities of SASS data used in the present study. Latitudes are in degrees North, longitudes are in degrees East. The fourth column shows the approximate angle between ascending orbits and true north–south.

<table>
<thead>
<tr>
<th>Geographic region</th>
<th>Latitude</th>
<th>Longitude</th>
<th>Approximate inclination</th>
<th>Number of realizations</th>
<th>Number passing ( \chi^2 ) test</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>-45°–-25°</td>
<td>160°–280°</td>
<td>25°</td>
<td>216</td>
<td>197</td>
</tr>
<tr>
<td>II</td>
<td>-25°–-5°</td>
<td>160°–280°</td>
<td>20°</td>
<td>251</td>
<td>239</td>
</tr>
<tr>
<td>III</td>
<td>5°–25°</td>
<td>140°–250°</td>
<td>20°</td>
<td>235</td>
<td>215</td>
</tr>
<tr>
<td>IV</td>
<td>25°–45°</td>
<td>150°–230°</td>
<td>25°</td>
<td>149</td>
<td>130</td>
</tr>
</tbody>
</table>

3. Turbulence models

The interrelationships between atmospheric motions on a wide range of temporal and spatial scales have been well established (e.g., see Lorenz, 1969). Errors in defining and predicting very small-scale motions propagate rapidly through deterministic atmospheric models, resulting in substantial errors in predicting large-scale quantities of interest. The time periods associated with this error propagation are similar, in fact, to the periods required for the evolution of the large-scale motions themselves. The development of accurate, physically based, deterministic models of large-scale atmospheric motions is thus overwhelmingly difficult.

Numerous observations (summarized and discussed in section 4 below) have demonstrated that the statistics (as functions of space and time) of atmospheric motions are similar from one observational study to another. A distinctive characteristic of “turbulent” flows is that although details of the motions are apparently random, statistics of various parameters are stable and can be used to characterize the flow (Hinze, 1959). Thus, considerable effort has been devoted to the construction of turbulence models to describe the space–time structure and evolution of the statistics of quantities used to characterize the atmosphere.

a. Overview of isotropic models

Three-dimensional isotropic turbulence models are among the oldest and simplest models used to describe turbulent atmospheric motions. Kolmogorov (1941) postulated the existence of an inertial subrange of wavenumbers in three-dimensional isotropic turbulence, within which no kinetic energy was dissipated. Dimensional arguments then led to an inertial subrange
spectrum with a \( k^{-5/3} \) dependence. In this model, kinetic energy supplied at low wavenumbers is cascaded by inertial forces through the subrange to higher wavenumbers, where it is eventually dissipated by viscosity (outside the inertial subrange). However, three-dimensional turbulence models are directly applicable to atmospheric motions only on scales less than about 1 km (Gage, 1979), due to stratification in the atmosphere that limits vertical velocities on larger scales.

The small ratio of vertical to horizontal velocities observed in large-scale winds (with horizontal scales greater than 10 km) has led to modeling of the atmosphere as a two-dimensional flow. Two-dimensional inviscid flow requires conservation of both energy and enstrophy (Ogura, 1952; Fjortoft, 1953; Leith, 1971), thus precluding the cascade of energy from large to small scales characteristic of three-dimensional turbulence. In fact, isotropic two-dimensional turbulence is characterized by a reverse cascade of kinetic energy from small scales to larger scales (Fjortoft, 1953; Leith, 1971; Lilly, 1972). Kraichnan (1967) postulated the existence of two inertial subranges in two-dimensional turbulence: one in which energy input at a given scale is transferred to larger scales, and a second in which enstrophy (input at the same scales as the energy) is cascaded to smaller scales. The reverse energy cascade leads, through dimensional arguments (Kraichnan, 1967), to a \( k^{-5/3} \) energy spectrum at scales larger than those at which energy is input, while the enstrophy cascade to smaller scales leads to a \( k^{-3} \) energy spectrum (Batchelor, 1969; Kraichnan, 1967; Lilly, 1972). By postulating a surface friction mechanism and random large-scale forcing, Lilly (1972) demonstrated numerically that a statistically steady-state condition with a \( k^{-3} \) spectrum was, in fact, possible.

Charney (1971) modified the two-dimensional turbulence theory to more realistically describe large-scale atmospheric motions. Charney's "geostrophic turbulence" model takes into account the physical reality that large-scale atmospheric motions are, to a certain extent, baroclinic, and thus a two-dimensional model is not strictly warranted. In addition, Charney recognized the geostrophic nature of large-scale atmospheric motions. By properly scaling the vertical coordinate, he postulated that motions (in the scaled coordinates) are three-dimensionally isotropic, conserving a single quantity identified as the "pseudo-potential vorticity." The geostrophic turbulence model predicts a \( k^{-3} \) kinetic energy spectrum at scales smaller than the forcing. Note that this wavenumber dependence is the same as that predicted for the enstrophy cascading inertial subrange of two-dimensional turbulence.

It is not our primary intent in this paper to use SASS data to differentiate between predictions of the two-dimensional and geostrophic turbulence models. Neither will we investigate the hypothesis that observed mesoscale wind spectra are due to a spectrum of atmospheric internal waves as proposed by Dewan (1979) and VanZandt (1982). The more complete two-dimensional turbulence theory will be used to establish a framework within which the SASS data can be analyzed and interpreted. Although the spectral analysis presented in sections 5 and 6 supports many of the predictions of two-dimensional turbulence theory, we recognize that the basic turbulence models must be significantly expanded in order to accurately describe near-surface winds over the ocean. Specifically, inclusion in the models of realistic terms describing the effects of the ocean surface are required.

b. Kinematics of two-dimensional turbulence

As shown in section 2 above, large gaps between swaths observed by SASS preclude the direct calculation of two-dimensional wavenumber spectra of near-surface winds, although one-dimensional spectra (in the along-track direction) can be easily calculated. However, if atmospheric motions are assumed to be two-dimensional (nondivergent) and isotropic, then one-dimensional spectra can be used to infer properties of the full two-dimensional flow field. Ogura (1952) and Leith (1971) have derived the kinematics of two-dimensional isotropic turbulence on a plane, and Fjortoft (1953) derives the results for motion on a sphere. The results are presented here in order to illustrate quantitatively the relationships between the one-dimensional spectra calculated in this paper (and presented historically in the literature), and the more complete two-dimensional descriptions of the flow that could be provided by an adequate observational system.

Let \( u_t(x, y, t) \) and \( u_x(x, y, t) \) represent the \( x \)-directed and \( y \)-directed components of velocity on a plane. (Numerical subscripts thus denote the components of the velocity field, while letter subscripts denote coordinate directions in physical or wavenumber space.) In the following, the subscript \( t \) representing time will be dropped; all derived quantities will be assumed to represent instantaneous statistics of the flow. The flow can be represented by its spatial Fourier coefficients

\[
\phi_t(k_x, k_y) = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} u_t(x, y)e^{-ik_xx+ik_yy}dx\,dy,
\]

(1)

where \( k_x \) and \( k_y \) are the wavenumbers in the \( x \) and \( y \) directions, respectively. The quantity

\[
|\phi_t|^2 = \phi_t(k_x, k_y)\phi_t^*(k_x, k_y)
\]

(2)

is the two-dimensional component spectrum. The two-dimensional kinetic energy spectrum \( E(k_x, k_y) \) is defined by

\[
E(k_x, k_y) = |\phi_t|^2 + |\phi_x|^2,
\]

(3)

so that

\[
\langle u_t^2 \rangle + \langle u_x^2 \rangle = \int_{-\infty}^{\infty} \int_{-\infty}^{\infty} E(k_x, k_y)dk_xdk_y.
\]

(4)

The angle brackets represent a spatial or ensemble average.
For isotropic turbulence, \( E(k_x, k_y) \) must be invariant under rotation of the coordinate system, and thus must be a function only of \( k = (k_x^2 + k_y^2)^{1/2} \). The two-dimensional isotropic energy spectrum, \( \tilde{E}(k) \), can be derived by expressing \( E(k_x, k_y) \) in terms of the polar coordinates
\[
k = (k_x^2 + k_y^2)^{1/2} \quad \theta = \tan^{-1}(k_y/k_x)
\]
and integrating over \( \theta \), yielding
\[
\tilde{E}(k) = \pi k E(k_x, k_y),
\]
(5)
where
\[
\frac{1}{2} \left( \langle u_1^2 \rangle + \langle u_2^2 \rangle \right) = \int_{0}^{\infty} \tilde{E}(k)dk.
\]
(6)

One-dimensional component spectra in the \( x \) or \( y \) directions can be obtained from two-dimensional spectra \( |\phi|^2 \) by integrating over the second dimension. For example, the one-dimensional component spectra in the \( x \)-direction are given by
\[
F_1(k_x) = \int_{-\infty}^{\infty} \phi_1^2 dk_y
\]
and the component variances can be expressed as
\[
\langle u_1^2 \rangle = 2 \int_{0}^{\infty} F_1(k_x)dk_x.
\]
The one-dimensional kinetic energy spectrum is determined from the one-dimensional component spectra by
\[
E(k_x) = F_1(k_x) + F_2(k_x),
\]
and the total kinetic energy is
\[
\frac{1}{2} \left( \langle u_1^2 \rangle + \langle u_2^2 \rangle \right) = \int_{0}^{\infty} E(k_x)dk_x.
\]
(9)

From (9) and (6) the one-dimensional energy spectrum, \( E(k_x) \), is related to the two-dimensional isotropic energy spectrum by
\[
E(k_x) = \frac{2}{\pi} \int_{k_x}^{\infty} \tilde{E}(k) \frac{1}{(k^2 - k_x^2)^{1/2}} dk.
\]
(10)

Equation (11) is satisfied identically by
\[
\phi_2(k_x, k_y) = \begin{cases} \frac{-k_x}{k_y} \phi_1(k_x, k_y), & k_y \neq 0 \\ 0, & k_y = 0 \end{cases}
\]
(13)

Equation (13) can be used to relate the one-dimensional component spectra \( F_i(k_x) \) to \( \tilde{E}(k) \), using the definitions (3), (7), and direct substitution:
\[
F_i(k_x) = \frac{2}{\pi} \int_{k_x}^{\infty} \tilde{E}(k) \frac{1}{(k^2 - k_x^2)^{1/2}} dk
\]
(14a)
\[
F_2(k_x) = \frac{2k_x^2}{\pi} \int_{k_x}^{\infty} \tilde{E}(k) \frac{1}{k^2(k^2 - k_x^2)^{1/2}} dk.
\]
(14b)

If the two-dimensional isotropic energy spectrum satisfies a power law \( \tilde{E}(k) = ak^{-b} \) as predicted by turbulence models (see section 3a), then the one-dimensional spectra \( F_1(k_x) \), \( F_2(k_x) \), and \( E(k_x) \) also obey a \( k^{-b} \) power law dependence (a result apparently first derived by Leith, 1971, for the special case \( b = 3 \)). Thus, if the atmospheric motions of interest are truly those of an incompressible two-dimensional fluid that is isotropically turbulent, then the calculation of one-dimensional spectra of components and/or kinetic energy can yield information on the shape of the two-dimensional component and energy spectra.

Substitution of the form \( \tilde{E}(k) = ak^{-b} \) into equations (14) and evaluation of the integrals also shows that
\[
F_2(k_x)/F_1(k_x) = b.
\]
(15)

[More generally, Ogura and Leith show that
\[
F_2(k_x) = -k_x dF_1(k_x)/dk_x.
\]
Physically, if the two-dimensional isotropic energy spectrum decreases monotonically with wavenumber, then the one-dimensional spectrum of the transverse component speed (relative to the direction along which the spectrum is calculated; in the present example, the \( y \)-directed speed) will be greater than the one-dimensional spectrum of longitudinal (\( x \)-directed) component speed.

The assumption of nondivergent flow can be tested by examining the cross-spectrum of zonal and meridional components. The two-dimensional cross-spectrum is defined by
\[
H(k_x, k_y) = \phi_1 \phi_2^*.
\]
(16)

Nondivergence (11) can be used to show that
\[
\text{Im}[H(k_x, k_y)] = 0,
\]
(17)
that is, the two-dimensional cross-spectrum must be real. The one-dimensional cross-spectrum can be obtained by integrating over \( k_y \) as in (7) to yield
\[
\tilde{H}(k_x) = \int_{-\infty}^{\infty} H(k_x, k_y)dk_y.
\]
(18)
Again using (11) and the fact that the field is real,
\[ \hat{H}(k_x) = 0. \]  
(19)

The assumption of nondivergence thus implies that the one-dimensional cross-spectrum is identically zero. Therefore, if the one-dimensional coherence
\[ \hat{H}(k_x)/(F_1(k_x)F_2(k_x)) \]
is significantly different from zero in any frequency range, then basic assumptions of two-dimensional turbulence theory are violated, and spectral slopes inferred from one-dimensional spectra may not correspond to the slopes that would be calculated from full two-dimensional spectra.

4. Review of past observations

a. Observations of large-scale atmospheric spectra

Wavenumber spectra of velocity components and kinetic energy in the troposphere and lower stratosphere have been reported many times in the past three decades (e.g., see the work and references in Ogura, 1958; Saltzman and Fleisher, 1962; Winn-Nielsen, 1967; Kao et al., 1970; Saltzman, 1970; Chen and Winn-Nielsen, 1978; Boer and Shepherd, 1983). While it is beyond the scope of the present work to review in detail and compare and contrast the results from each of the many studies, some general comments can summarize the methods and conclusions that have been reached by these analyses.

Most past investigations have been limited to analysis of one-dimensional wavenumber spectra along latitude circles, using data obtained from large-scale analyses produced by the National Meteorological Center. Daily or twice-daily streamfunction and/or isobaric height analyses on levels ranging from 1000 to 100 mb were converted to velocity components. These velocities were then interpolated and sampled along selected latitude circles (or portions of latitude circles) ranging from the tropics (see especially Julian, 1971) to high latitudes in the Northern Hemisphere (spectral analyses of Southern Hemisphere winds are scarce, although see Desbois, 1975). Few authors have attempted to utilize raw data directly in their calculations (although see Jones, 1975; Julian and Cline, 1974) due to the irregular spacing and generally sparse distribution of measurements. The duration of the datasets ranged from approximately 30 days to five years (Julian et al., 1970). Wavenumber spectra of both transient (after subtraction of temporal means) and stationary winds have been studied, and Kao and Wendell (1970) and Kao et al. (1970) have examined frequency–wavenumber spectra. Some authors (cf. Boer and Shepherd, 1983) have analyzed the spatial scales of variability of the wind field in terms of two-dimensional spherical harmonics.

The salient results of the many large-scale studies can be summarized as follows.

(i) Kinetic energy spectra have a maximum at planetary wavenumbers between 5 and 8, and fall off rapidly at higher wavenumbers. The location (in wavenumber space) of the peak is consistent with the scales associated with maximum baroclinic instability for large-scale atmospheric flows (Julian, 1971).

(ii) In the planetary wavenumber range 8–15, spectra of kinetic energy and components fall off with slopes between \( k^{-2} \) and \( k^{-3} \). These slopes are, at most, weak functions of latitude. There appears to be a slope dependence on altitude—as first noted by Horn and Bryson (1963), slopes increase in magnitude with increasing altitude. Thus, Horn and Bryson (1963) and later authors find that the relative variance in high wavenumbers is larger at low altitudes. Exceptions to the spectral flattening with decreasing altitude can be found in the works of Chen and Winn-Nielsen (1978) who show nearly constant slope in the range 700–200 mb, and Boer and Shepherd (1983) who show an opposite slope dependence on altitude. Note, however, that the total integrated variance over the full wavenumber range increases with increasing altitude.

(iii) Recent studies have investigated cross-spectra between zonal and meridional velocity components [see Eqs. (16–19)] to test the nondivergent nature of large-scale motions (Kao and Wendell, 1970; Boer and Shepherd, 1983). For wavenumbers greater than the kinetic energy maximum, cross-spectral densities have been found to be small, as predicted for two-dimensional (nondivergent) turbulent motions.

Evidence from past studies thus suggests the existence of an enstrophy-cascading inertial range for mid-level atmospheric motions having scales from a few thousand to about 10 000 km. Spectral wavenumber dependencies of about \( k^{-3} \) on these scales are consistent with predictions of anisotropic two-dimensional turbulence theory. Energy and enstrophy are thought to be input to the atmosphere at scales from about 5000–10 000 km by a baroclinic instability mechanism (Julian, 1971), and the observed spectral shapes are considered evidence of the resulting downscale enstrophy cascade.

Some care must be taken, however, when considering quantitatively the large-scale results in the literature. As discussed by Yang and Shapiro (1973), among others, the use of interpolated, analyzed data can substantially increase the apparent magnitude of spectral slopes. This effect is due both to the sparse and irregular nature of the observation system, and to the inherent (but nonquantifiable) smoothing imposed on the data in the production of analyzed maps.

b. Observations of smaller scale atmospheric spectra

Research aircraft and land-based instrument arrays have yielded data on atmospheric motions on very small scales from 100 m–100 km (see, for instance, Overland and Wilson, 1984). There have been few ob-
observations spanning the wavelength range from ~200 to 2200 km of interest in the present study. A remarkable dataset has recently been obtained as part of the Global Atmospheric Sampling Program (Lilly and Peterson, 1983; Nastrom and Gage, 1983; Nastrom et al., 1984; Nastrom and Gage, 1985). By recording flight-altitude (9–13 km) winds as determined by the inertial navigation systems of selected commercial aircraft, one-dimensional wavenumber spectra of components and kinetic energy spanning the wavelength range from 2.6–10 000 km have been obtained. "Eyeball" fits of the resulting spectra (Nastrom et al., 1984) show that in the range 1000–3000 km, slopes are approximately $k^{-3}$, and in the range 2.6–300 km, slopes are consistent with $k^{-5/3}$. The intervening wavelength range, from 300–1000 km, is a transition region, with spectra having wavenumber dependence between $k^{-3}$ and $k^{-5/3}$.

The aircraft small-scale spectra (wavelengths 2.6–300 km) are similar to other observed midlevel spectra on these scales (see Gage, 1979). Gage (1979) and Lilly (1983) attribute the spectral shape in this wavenumber band to the upscale energy cascade resulting from energy input by even smaller-scale convective motions. The data of Nastrom et al. (1984) thus are asymptotically consistent with the results of other studies at the high- and low-wavenumber extremes, and strongly support the hypothesis that at least two localized sources of energy and enstrophy are necessary in order to even schematically describe atmospheric motions on scales from 1 to $10^4$ km. These sources are a low-wavenumber source (discussed in section 4a above), leading to an enstrophy cascade to higher wavenumbers (and a $k^{-3}$ spectrum), and a high-wavenumber source, leading to a reverse energy cascade (and a $k^{-5/3}$ spectrum). These dual inertial ranges in two-dimensional turbulent fluids were first hypothesized by Kraichnan (1967) on dimensional grounds and their existence later was verified numerically by Lilly (1972).

A transition region at wavelengths from a few hundred to about a thousand kilometers as found by Nastrom et al. (1984) was also inferred by Brown and Robinson (1979) in their study of central European 500 mb winds. Although Brown and Robinson did not attempt to determine the actual wavenumber dependence for scales from 500–2000 km, they concluded that there was more observed variance in this band than could be accounted for by a $k^{-3}$ spectral shape, yet the evidence for a $k^{-5/3}$ spectrum on these scales was not convincing.

5. Spectral analysis of SASS data

a. Kinetic energy spectra

One-dimensional spectra (in the along-track direction) of zonal and meridional wind components were calculated for each SASS data swath falling within any of the four geographical regions shown in Fig. 3. One-dimensional kinetic energy spectra, $E(k_i)$, were constructed from the component spectra as in Eq. (8). Averaged kinetic energy spectra were then calculated separately for each geographical region by averaging raw [$\sim$ 2 degrees of freedom (dof)] spectral results from each individual data swath. Averaged spectra thus represent both spatial (longitudinal) and temporal (over the 15 day duration of the dataset) averages.

We note that the derivation leading to Eq. (8) assumed that the component speeds $u_1$ and $u_2$ were aligned with the axes of the coordinate system. The along-track direction for which one-dimensional spectra were calculated in the present cases is rotated (see Table 1) with respect to the coordinate system defined by the zonal and meridional directions. However, if (cross-track, along-track) component speeds $(u_1, u_2)$ are defined in terms of (zonal, meridional) component speeds $(u'_1, u'_2)$ by

$$u_1 = u'_1 \cos \theta - u'_2 \sin \theta$$
$$u_2 = u'_2 \cos \theta + u'_1 \sin \theta$$

where $\theta$ is the angle of rotation, then simple substitution yields

$$E(k_i) = F_1(k_i) + F_2(k_i) = F'_1(k_i) + F'_2(k_i)$$

[where primes indicate quantities calculated from Eqs. (1, 7) using unrotated component speeds in the rotated component system]. The kinetic energy spectra presented below are thus unaffected by the coordinate rotation.

Before examining the averaged one-dimensional kinetic energy spectra, we constructed normalized kinetic energy spectra for each swath in order to test the temporal and spatial homogeneity of the data, and thus the validity of the ensemble-averaging process itself. The use of normalized rather than raw spectra allows comparison of spectral shapes from individual realizations by eliminating variance as a factor in the overall spectral characteristics. Raw normalized spectra were constructed from raw spectra by dividing each spectral estimate by the total sample variance for the realization,

$$E_a(k_i) = E(k_i)/\sum_i E(k_i)$$

where $E_a(k_i)$ is the normalized one-dimensional kinetic energy (KE) spectral density in wavenumber band $i$, $E(k_i)$ is the unnormalized spectral density ($\sim$ 2 dof) in band $i$, and the sum in the denominator is over all wavenumbers of interest. For each geographical region, all raw normalized spectra $E_a(k)$ were ensemble-averaged to produce a single averaged normalized spectrum $\langle E_a(k) \rangle$ having many degrees of freedom.

As each ensemble-averaged spectral estimate $\langle E_a(k_i) \rangle$ is distributed as a $\chi^2_q$ random variable, where $q$ is the number of dof, there is little statistical variability in the averaged normalized spectral estimates since $q$ is large. Raw normalized spectral density estimates $E_a(k_i)$, on the other hand, are only $\chi^2_q$ distributed, and the
expected variability of each estimate is large. Following Haubrich (1965) and Bendat and Piersol (1971), 10% of the raw normalized spectral estimates $E_d(k)$ are expected to fall more than 10 dB below the ensemble average estimate $\langle E_d(k) \rangle$, and 10% are expected to fall more than 3.6 dB above the average. In the present study, each raw spectrum is composed of estimates at each of ten different wavenumbers. Thus, on the average, two spectral estimates from any raw normalized spectrum are expected to fall outside the 80% confidence limits about the high dof-averaged normalized spectrum. If more than four spectral estimates from a raw spectrum exceed the 80% confidence limits, it can be concluded with high probability that the raw spectrum is from a realization that is statistically different from the rest of the realizations used to construct the averaged normalized spectrum.

In our analysis of each geographical region, raw normalized spectra with four or more spectral estimates falling outside the 80% confidence limits about the averaged spectral estimate were considered to have been drawn from a different population and excluded from further processing. This screening criterion eliminated fewer than 9% of the raw normalized spectra in three of the four regions. In the fourth region, the north Pacific (region IV), 13% of the raw spectra were eliminated. The small number of realizations eliminated by this procedure indicates that one-dimensional component spectra within each individual geographical region have nearly uniform shapes, independent of total variance.

The averaged normalized kinetic energy spectra from the four geographical regions are shown in Fig. 5. In all cases, the normalized spectra are very nearly linear when plotted in log-log space. In other words, the spectra appear to obey a simple power law dependence on wavenumber which can be expressed as $k^{-b}$. Spectra from the midlatitude regions (regions I and IV of Fig. 3) have nearly identical power-law dependencies, with $b = 2.20 \pm 0.08$ for region I, and $b = 2.21 \pm 0.10$ for region IV (indicated exponent uncertainties are standard deviations derived from the regression of spectral density on frequency). Normalized spectra from the two tropical regions also have very similar power-law dependencies, with $b = 1.76 \pm 0.04$ for region II and $b = 1.85 \pm 0.06$ for region III. Although these normalized spectral slopes in the Pacific are independent of hemisphere, the differences between spectral slopes in tropical and midlatitude regions is statistically significant; the normalized kinetic energy spectra are flatter in tropical regions.

Averaged kinetic energy spectra (unnormalized) were also calculated for the four regions and are shown in Fig. 6. Spectral power-law dependencies are listed in Table 2. (Note that those wind fields that failed the chi-square test described above were excluded from these averages as well.) Both the shapes and the average total variances of the unnormalized average spectra from the midlatitude regions (I and IV) are extremely similar, and the wavenumber dependencies of these unnormalized spectra are not different from those calculated.
Table 2. Power law exponents and averaged variances for kinetic energy spectra from each geographic region. Uncertainties shown are standard deviations.

<table>
<thead>
<tr>
<th>Geographic region</th>
<th>Normalized power law</th>
<th>Unnormalized power law</th>
<th>Total variance (m² s⁻²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>-2.20 ± 0.08</td>
<td>-2.26 ± 0.09</td>
<td>17.7</td>
</tr>
<tr>
<td>II</td>
<td>-1.76 ± 0.04</td>
<td>-1.86 ± 0.05</td>
<td>5.6</td>
</tr>
<tr>
<td>III</td>
<td>-1.85 ± 0.06</td>
<td>-1.97 ± 0.07</td>
<td>8.8</td>
</tr>
<tr>
<td>IV</td>
<td>-2.21 ± 0.10</td>
<td>-2.21 ± 0.10</td>
<td>15.3</td>
</tr>
</tbody>
</table>

above using normalized spectra. Together with the results of the averaged normalized spectra, this implies a strong tendency for constant spectral shapes in both midlatitude regions.

The unnormalized spectra from the tropical regions II and III also exhibit similar wavenumber dependencies to those calculated using normalized spectra. The total kinetic energies in the tropical regions showed differences between the Northern and Southern hemispheres, with the total energy in the northern tropical Pacific (region III) being approximately 1.6 times larger than that in the southern tropical Pacific (region II). This disparity in total energy between similar latitudes in different hemispheres was not found in the midlatitude regions I and IV.

Comparisons between midlatitude and tropical unnormalized spectra show that the total energy in the midlatitude regions is higher than that in either of the tropical regions.

For comparison, the one-dimensional kinetic energy spectrum in this wavenumber band calculated by Nastrom et al. (1984) from aircraft data (see section 4b) is shown as the short dashed line in Fig. 6. The average latitude of these data was 50°N, corresponding roughly to region IV, and the flight altitudes for most of the measurements ranged from 9–13 km (approximately 200–300 mb). The power law exponent (b = -2.23 ± 0.03) calculated from the Nastrom et al. data agrees remarkably well with those calculated in the present study, notwithstanding the minor differences in latitude and the major difference due to the high altitude of the Nastrom et al. data as compared to the near-surface winds measured by SASS. As expected from the results of large-scale observations (see section 4a), the spectral densities of the Nastrom et al. high altitude measurements are larger (by about a factor of 4) than those from the near-surface SASS data.

b. Velocity component auto- and cross-spectra

Variance spectra of meridional and zonal component speeds and cross-spectra between zonal and meridional components can be used to test the assumptions of isotropy and nondivergence in models of atmospheric turbulence. As discussed in section 3, these assumptions are crucial in order to allow inferences regarding the full two-dimensional structure of horizontal motions to be drawn from one-dimensional data.

In section 3b it was shown [Eqs. (8, 14)] that two-dimensional turbulence models predict that one-dimensional kinetic energy and component variance spectra should have the same power law dependence on wavenumber as does the isotropic two-dimensional spectrum \( \tilde{E}(k) \). The averaged spectra of component speeds in each of the geographical regions are shown in Fig. 7, and the best-fit power law exponents are given in Table 3. In all cases, component variance spectra are nearly linear in log-log space. The slopes of the zonal component spectra are always slightly steeper than the corresponding slopes of the meridional spectra. The power law dependencies of the component spectra, however, are not statistically different from those of the kinetic energy spectra, notwithstanding the fact that the exponents vary as a function of geographical region.

The assumptions of isotropy and nondivergence in two-dimensional turbulence models require [Eq. (15)] that the one-dimensional spectral densities of zonal and meridional components not be equal. As the SASS swaths are not exactly aligned with either the zonal or meridional directions (due to the 108° inclination of the Seasat orbit), it is not possible to make a precise test of Eq. (15). However, the swaths are nearly aligned with the meridional direction in the four geographical regions examined here. Equation (15) thus predicts that one-dimensional spectra of zonal winds (approximately transverse to the along-track direction) should be larger than spectra of meridional winds (approximately aligned with the along-track direction).

The mean ratios of zonal to meridional spectral densities are given in Table 3. At midlatitudes, the ratio is greater than unity for both regions I and IV as predicted, although the observed ratio of 1.1 is considerably less than the predicted value of 2.2. This is to be expected since nonalignment of SASS swaths with the meridional direction would somewhat decrease the zonal/meridional variance ratio. In the tropical regions, however, the meridional spectra are uniformly larger than the zonal spectra. The assumptions of isotropy and nondivergence are thus suspect for near-surface winds on these scales in the tropics, based on this approximate test.

One-dimensional cross-spectra between zonal and meridional components can be examined to test the assumption of nondivergence alone. According to Eq. (19), the cross-spectrum is identically equal to zero in a nondivergent, two-dimensional turbulent fluid. Averaged co- and quadrature spectra from each ocean region are shown in Fig. (8). In nearly all cases, quadrature spectra are small compared with cospectral densities. Comparisons with Figs. 6 and 7 show that cross-spectral densities are small compared with autospectral densities of kinetic energy or component speeds.

Nondivergence in the lower atmosphere can be more
quantitatively tested through examination of coherence spectra. The one-dimensional coherence spectrum should be identically equal to zero at all wavenumbers.

**Table 3.** Power law exponents for averaged one-dimensional zonal and meridional component spectra. The fourth column shows the average (over wavenumber bands) ratio of zonal/meridional spectral densities.

<table>
<thead>
<tr>
<th>Geographic region</th>
<th>Zonal exponent</th>
<th>Meridional exponent</th>
<th>Mean ratio zonal/meridional</th>
</tr>
</thead>
<tbody>
<tr>
<td>I</td>
<td>-2.35 ± 0.08</td>
<td>-2.17 ± 0.09</td>
<td>1.11</td>
</tr>
<tr>
<td>II</td>
<td>-1.99 ± 0.04</td>
<td>-1.77 ± 0.06</td>
<td>0.71</td>
</tr>
<tr>
<td>III</td>
<td>-2.01 ± 0.05</td>
<td>-1.95 ± 0.08</td>
<td>0.69</td>
</tr>
<tr>
<td>IV</td>
<td>-2.31 ± 0.12</td>
<td>-2.11 ± 0.10</td>
<td>1.14</td>
</tr>
</tbody>
</table>

if the flow is truly nondivergent. Averaged coherence spectra were calculated using normalized Fourier coefficients, described in section 5a, in order to prevent high-variance realizations of the wind field from dominating the results. The coherence spectra and approximate 95% significance levels (Bendat and Piersol, 1971) are shown in Fig. 9 for each of the four regions in the Pacific. As expected from the relatively small co- and quadrature spectral values in Fig. 8, coherences are small at all wavenumbers in all geographical locations. In the Southern Hemisphere (regions I and II), coherences are essentially nonsignificant at all but the lowest wavenumber band. Coherences are small but significantly greater than zero everywhere in the trop-
A possible explanation for the differences between coherence spectra in the Northern and Southern hemispheres may lie in the observation that major large-scale divergence and convergence zones can be found in the Northern Hemisphere during September. Meteorological analyses of portions of the full 15-day SASS dataset by Peteherych et al. (1984) identify two major convergence features extending nearly latitudinally over major portions of the North Pacific. One of these zones has been identified as the Intertropical Convergence Zone, and is present in region III at approximately 10°N. The second major feature extends from 32°-42°N across region IV. Similar strong features do not appear to be present in the Southern Hemisphere. There is thus the possibility that, for the 15-day dataset investigated in the present study, the assumption of nondivergent flow may not be valid for large-scale (2200 to ~700 km) atmospheric motions in Northern Hemisphere regions. The assumption of nondivergence is not obviously violated for Southern Hemisphere flows and for smaller scale motions in the Northern Hemisphere.
6. Discussion and conclusions

Scatterometer measurements of horizontal wind velocity have been used to determine, for the first time, the spatial variability of near-surface oceanic winds on scales from 200 to 2200 km. Data used in this study were obtained from four latitude bands in the Pacific, two each in the Northern and Southern hemispheres. The spatial sampling characteristics of SASS precluded direct calculation of full two-dimensional wavenumber spectra. In this paper, we computed one-dimensional (along the satellite ground track) spectra of velocity components and kinetic energy to determine the spatial structure of near-surface winds and to test some predictions of two-dimensional turbulence models. Specifically, the models predict that 1) the one-dimensional spectra of each wind component and of kinetic energy all obey the same power law dependence, and this dependence is identical to that of the two-dimensional isotropic kinetic energy spectrum; 2) for one-dimensional spectra calculated in the meridional direction, zonal spectra should have greater variance than meridional spectra; and 3) the two vector wind component speeds should be spatially incoherent.

Our results show that within each geographic region, the one-dimensional wavenumber spectra of velocity components and kinetic energy have nearly identical shapes, as predicted by turbulence models. Although the spectral shapes of components and kinetic energy are similar within regions, they differ when spectra from different geographic regions are compared. In all regions, one-dimensional kinetic energy spectra from SASS winds were found to obey simple power law dependencies on wavenumber. In midlatitude regions of both hemispheres, spectra fell off approximately as $k^{-2.2}$, while in the tropical regions of both hemispheres, the falloff was a less steep $k^{-1.9}$.

In the midlatitudes, our analysis indicates that one-dimensional (in approximately the meridional direction) spectral densities of zonal component speeds were larger than the corresponding spectral densities of meridional component speeds, qualitatively consistent with expectations from two-dimensional turbulence models. In the tropics, however, meridional spectral densities were larger than zonal spectral densities—in apparent contradiction to the predictions of the models. As previously discussed, it appears that surface motions on these scales in the Southern Hemisphere tropics are nondivergent. We thus question the validity of the assumption of isotropy for Southern Hemisphere, tropical near-surface winds on these scales.

The possible anisotropic nature of tropical Southern Hemisphere winds is a surprising result. From observations of one-dimensional (in the zonal direction) spectra of 200 mb winds in the Southern Hemisphere, Desbois (1975) showed that the assumption of isotropy appeared valid on scales less than 3000 km. Because of the extreme paucity of surface wind data from the Southern Hemisphere for comparison with our results, we cannot dismiss the possibility that the subjective process by which unique vector winds were produced from multiple solution SASS data (see section 2b) may have introduced artifacts into the Southern Hemisphere data (few meteorological analysts possess the range of experience with Southern Hemisphere weather patterns that they have with Northern Hemisphere circulations).

Small values of the one-dimensional coherence between velocity components over all mesoscale wavenumbers in the Southern Hemisphere, and at high wavenumbers in the Northern Hemisphere support the assumption of a nondivergent atmosphere on these spatial scales. The relatively larger values of coherence (although always less than 0.25) observed at low wavenumbers in the northern hemisphere Pacific may be due to the existence of large regions of divergence and convergence such as the ITCZ.

The spectral results presented here represent the first direct evidence supporting the assumed spatial structure of the wind fields used to force the stochastic mesoscale eddy models of Frankignoul and Muller (1979) and Muller and Frankignoul (1981). By extrapolating published observations of large-scale atmospheric motions in the Northern Hemisphere, Frankignoul and Muller hypothesized that the surface wind field was isotropic and nondivergent, with a $k^{-2}$ wavenumber dependence on scales from 50 to 4000 km in midlatitudes. Although the SASS data do not uniformly support the assumptions of isotropy and nondivergence, one-dimensional spectra from SASS have nearly a $k^{-2}$ wavenumber dependence. The wavenumber dependencies of kinetic energy from SASS wind measurements are also consistent with previous observations of higher-level winds measured by aircraft.
The analysis presented here suggests that relatively simple models of atmospheric turbulence may be used to predict the general characteristics of one-dimensional autospectra of near-surface oceanic winds. At low wavenumbers, the models predict a $k^{-3}$ spectrum due to an enstrophy cascade from small to large wavenumbers. At high wavenumbers, the models predict a spectral region with a $k^{-5/3}$ dependence due to an upscale cascade of energy from large to small wavenumbers. Baroclinic instability at large scales (planetary wavenumbers 5–8) has been suggested as a possible source of energy and enstrophy at low wavenumbers (Julian, 1971; Gage, 1979). Gage (1979) and Lilly (1983) suggest that convective motions with length scales O(1 km) could be the source at high wavenumbers. The wavelength range 200–2200 km may be thought of as a transition zone between the two inertial regimes. In this transition region, one would expect a wavenumber dependence somewhere between $k^{-3}$ and $k^{-5/3}$.

The SASS data analyzed in this study have neither the spatial nor the wavenumber resolution necessary to identify and characterize either of the suggested source mechanisms. However, the SASS data do resolve the transition region from 200 to 2200 km. The wavenumber dependence of the spectra computed from the SASS data are approximately $k^{-2}$, which indeed lies between the $k^{-3}$ and $k^{-5/3}$ extremes. One can hypothesize that the slightly flatter spectra in tropical regions may be due to a stronger convective source (relative to the large-scale baroclinic source) than in midlatitude regions.

It should be emphasized that the approximate $k^{-2}$ wavenumber dependence of the one-dimensional spectra computed from SASS winds does not confirm the validity of the two-dimensional turbulence models. Charney (1971) has postulated that the ubiquitous presence of atmospheric fronts would result in a spectrum with $k^{-2}$ wavenumber dependence, approximately the same as the spectra presented here. Without detailed meteorological analysis of the SASS data, it is not possible to determine quantitatively the importance of fronts to our results. However, our qualitative impression from considering a number of examples is that, while fronts are present in the SASS data, their presence and intensity does not dominate the spatial structure of the wind field.

Care must also be taken when drawing conclusions about the full two-dimensional spatial structure of near-surface winds from the one-dimensional results presented here. Such extrapolations from one dimension to two dimensions are critically dependent on the validity of the turbulence models. In light of the apparent Northern Hemisphere divergence and tropical anisotropy identified above, some of the assumptions of the models may not be valid uniformly. Refined models of near-surface atmospheric turbulence may be necessary in order to accurately infer properties of the two-dimensional wind field from the basically one-dimensional data obtained by SASS.

The NASA Scatterometer (NSCAT) to be launched in 1989 as part of the Navy Remote Ocean Sensing System will represent a significant improvement over SASS. The NSCAT measurement swaths will be slightly wider than those of SASS, and backscatter measurements will be obtained with 25 km resolution (as opposed to 50 km resolution for SASS). Importantly, the addition of a third antenna on each side of NSCAT will greatly simplify the process of determining unique vector winds from multiple wind solution measurements. A ground data processing system, including an objective scheme for determining unique vector solutions, will be in place and operating at launch time. The higher resolution of NSCAT will allow studies of wind variability over a larger range of wavenumbers than was possible using SASS data, and the three-year mission of NSCAT will provide data on wind variability spanning all seasons.

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