OCEAN WINDFIELD MAPPING AT LONG RANGES WITH AN HF RADAR

P.E. Dexter
Head Office, Bureau of Meteorology, Melbourne

and

R. Casey
Physics Department, James Cook University, Townsville

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ABSTRACT

Experiments have been conducted to test the potential value of the James Cook University HF ocean radar, operating in an ionospheric propagation mode, for deriving surface wind data over remote ocean areas as part of a meteorological synoptic data network. The paper discusses briefly radar instrumentation and background theory relevant to analysis of observational data. Radar derived surface wind directions agree fairly closely with surface winds deduced from coincident surface pressure analyses, despite uncertainties created by ionospheric motions and by the large spatial extent of the target scattering areas. In addition, good correlation has been obtained between a parameter of Doppler spectra of the backscattered radar signals and inferred surface wind speeds in the scattering areas, allowing an ultimate precision in estimating wind speeds of ±4.4 m/s for speeds of from 2 to 27 m/s.

INTRODUCTION

It has been known for some time now (Crombie 1955) that radio waves in the MF/HF band, when scattered from the sea surface, contain characteristics that may be interpreted uniquely in terms of features of the scattering surface. Various investigators (e.g. Tveten 1967, Ward 1969) have extended the observational work from backscatter echoes obtained at short range via surface wave propagation as in Crombie's (1955) original work to the use of an ionospheric propagation mode to examine signals backscattered from ranges out to nearly 4000 km.

In recent years, an increasing awareness of the potential of such a technique for obtaining meteorological and oceanographic data from remote ocean areas suitable for input to synoptic observational systems has led to the development of a variety of empirical methods for interpreting the backscattered signal simply and quickly. Such analysis was pioneered by Long and Trizna (1973), who produced a map of surface wind vectors over the North Atlantic, which coincided closely with the nearest (in time) available synoptic analysis. Similar methods for surface windfield mapping over the oceans have been developed by others for both bistatic (e.g. Tyler et al. 1974) and monostatic radars (e.g. Stewart and Barnum 1975, Bagwell et al. 1976). Very high resolution mapping on grids down to about 15 km by 10 km has been achieved in recent years, allowing for accurate location of fronts and other relatively fine-scale phenomena (Barnum et al. 1977). State of the art accuracies possible for surface wind measurements appear to be approximately ±16° for direction and ±2.4 m/s (up to 25 m/s) for speed (Stewart and Barnum 1975, Maresca and Barnum 1977).

HF radar observations of the sea surface at long ranges have been undertaken by the James Cook University Physics Department since 1966 with a radar facility located at Townsville, and some earlier results are described by Ward (1969) and
Ward and Dexter (1976). The facility has recently been upgraded to allow for windfield mapping similar to that achieved elsewhere, and this paper describes results from preliminary experiments with this new facility.

INSTRUMENTATION

The aerial array and transmitter have been described in detail by Ward and Dexter (1976). Briefly, the operating frequency of the radar is 21.840 MHz (wavelength $\lambda = 13.8$ m), the array being physically rotatable through a full 360° of azimuth, with a setting accuracy of ±0.25° and a final bearing accuracy of ±0.5°. The requirement for flexibility in selection of target areas necessitated limitations to the physical size of the array, resulting in a 3 dB beamwidth of ±12°. The present transmitter with the array yields an effective radiated power of up to 0.5 kW. Pulse lengths of 0.5 to 1.0 ms with repetition periods of 40 or 80 ms are available. The radar operates in a monostatic mode via a TR aerial switch.

The receiver and signal processing equipment in the facility have recently been extensively redesigned and reconstructed. A conventional Eddystone 840 receiver is used in the narrow bandwidth mode, with both its first and second local oscillators crystal locked. The receiver and signal processing equipment are located some 100 m from the aerial and transmitter, so the basic transmitter frequency has been unavailable as a reference for target data phase comparison, and it has been necessary to reconstitute the transmitted frequency at the receiver, with the ability to phase track the transmitted signal.

The TR switch theoretically shorts the receiver feeder during the transmitter pulse. However, the length of feeder line to the receiver enables sufficient transmitter pulse information to be present at the receiver IF output to generate a gate occurring at the same time as the transmitted pulse. A gate generator uses the first cycle in the radiated pulse to trigger a one-shot multivibrator, where a variable time constant can set an 'A gate', or reference gate, length from 0.1 to 1 ms. A 'B gate', or target gate, is generated from the positive edge of the A gate, and has the ability to be positioned at any range between successive A gate pulses. The B gate pulse width is also variable between 0.1 and 1 ms.

A and B gates are used to select required portions of the receiver IF for processing. The B gate selects any range window between transmitted pulses where target echoes occur, the selected information then being analysed in two channels. The first limits the data and phase compares it in a phase sensitive detector with the reference (A gate) data, thus giving a Doppler sense in the form of positive or negative ramps, each ramp corresponding to 360° of phase change. The second target channel is linear and maintains amplitude conscious at all times in order to generate an output suitable for power spectral analysis. Two-sided Doppler spectra may be obtained if the linear Doppler information in the target echo is synchronously detected with a reference carrier slightly offset from the CW carrier heterodyne. This phase locked offset is 2 Hz at the 100 kHz IF. A 4 Hz reference marker is also available from the 2 Hz oscillator. Ultimately, phase-amplitude information from the synchronous detector is recorded (via an audio frequency modulator) on magnetic tape with the lower level 4 Hz reference superimposed.

Following demodulation, the time series data are digitised at a rate of 10 samples per second on a PDP 8S minicomputer in blocks of 102.4 s length (1024 data points), and spectral analysis performed with a standard FFT package. A low pass filter (to 10 Hz) on the basic data eliminates aliasing problems. Spectral smoothing is effected through the averaging of non-overlapping sets of four raw spectral estimates, to give 90 per cent confidence limits of roughly +2.0 and -3.1 dB on individual spectral values. Spectral resolution is thus 0.04 Hz in the range 0 to 5 Hz, with the real zero Doppler occurring at 2.0 Hz because of the frequency offset. This spectral resolution is better than the 0.14 to 0.18
Hz resolution limit imposed by ionospheric motions, as originally applied by Ward (1972) and confirmed by Heron and Rose (1977). Finally, all spectra are normalised to the maximum spectral ordinate (usually that of the principal first order Bragg line). Examples of Doppler spectra are shown in Figs 1(a) to (c).

RADIO THEORY

In recent years the theory of radio wave scattering from the sea surface has become relatively well known, and full discussions may be found in Barrick (1972), Barrick et al. (1974), and Johnstone (1975). In particular, explicit formulations for the scattering cross-section, to both first and second order, have been derived. The application of such formulations, however, to the task of determining ocean wave spectra or simpler sea surface parameters from the backscattered signal has provided a significant problem. Generally, a theoretical wave spectrum has been assumed, the cross-section evaluated (to second order), and the result compared with observations. More recently, attempts at spectral inversion, to give empirical wave spectra direct from observed scattering cross-sections, have shown some limited success (Barrick 1977, Lipa 1977).

Nevertheless, such an approach is at best impracticable at present for the determination of simple sea wave and surface wind parameters on a synoptic scale. More direct, empirical techniques using first order spectral features enable reasonable estimates to be made of surface wind speeds. In simple terms, first order backscatter is interpreted as a resonant interaction between incident radio waves and ocean waves of one half the radio wavelength propagating along the radar radial, either towards or away from the radar. The backscattered radio wave has a Doppler shift in frequency imposed by the moving ocean waves, and its frequency spectrum contains discrete components at

\[ \omega = \omega_o \pm (2gk_o)^{\frac{1}{2}} = \omega_o \pm \omega_B \]

...1

where \( k_o \) and \( \omega_o \) are the wave number and frequency of the incident radiation (wavelength \( \lambda_o \)) and \( g \) is gravitational acceleration. These components, called Bragg lines, correspond to advancing and receding ocean waves. The ratio \( \tau \) of the amplitudes of the two Bragg lines is directly related to the relative proportions of advancing to receding waves and, through a directional model of the sea wave spectrum, may be interpreted in terms of surface wind direction. Examples of Doppler spectra are given in Fig 1, with positive and negative Bragg lines clearly visible (though with opposing sense) in both Figs 1(a) and 1(b). The value of \( \tau \) will vary from zero for wind blowing directly away from the radar, through 1 (cross winds at 90° to radar) to (theoretically) \( \infty \) for winds directed up the radar radial. A right/left ambiguity in wind direction about the radar radial is always present, but this is seldom likely to cause problems in meteorological interpretation, and could be removed by using two radars on a long baseline. For the James Cook radar, operating at 21.840 MHz, first order resonant (Bragg) scatter occurs from ocean waves of 6.9 m wavelength. In general, these waves will be in equilibrium with the local wind for all wind speeds above about 4 m/s, which is a necessary condition for the application of any technique using the Bragg lines.

Unfortunately, this equilibrium (or saturation) of the resonant ocean wave components means that the first order Bragg lines can supply no information on wave heights, i.e. the state of development of the total sea wave spectrum. For this information recourse must be made to second and higher order scattering terms. These produce lower-level sidebands in the frequency spectrum of the backscattered signal, which may sometimes be resolved as separate energy peaks but more generally act to broaden the Bragg lines. Among other empirical measures, the breadth (B) of the Bragg lines at a level 10 dB below the peak amplitude may be used to estimate wave heights. This will be discussed in more detail later.
Fig. 1. Doppler spectra of radio waves backscattered from the earth's surface. Data were recorded near 0000 GMT on 6 April 1977, on a bearing of 180° true from Townsville at ranges of 2100 km (a), 2500 km (b), and 2300 km (c). (a) and (b) both show sea Doppler with the peaks near ±0.5 Hz being the first-order Bragg lines, for a radar frequency of 1.84 MHz. (a) and (c) both exhibit significant multipath effects. (c) is purely a sea echo, with the peaks occurring at zero Doppler. Smaller peaks at zero Doppler are also present in (a) and (b).
OBSERVATIONS

In order to test both the performance of the newly constructed processing equipment and the potential of the radar for the collection of usable surface wind data in a quasi-operational situation, a set of experiments was conducted to observe seas surrounding Australia in April 1977. Radar observations were made on 6 and 13 April, primarily from scattering areas in the Tasman Sea region, from true bearing 120° (north of New Zealand) to true bearing 190° (west of Tasmania). Isolated observations were also attempted from the North Pacific Ocean and seas off the Australian northwest coast, with limited success. Ranges from 1500 km to 4000 km were sampled on most bearings. Observations on each day were made during the optimum time, roughly 1100 to 1500 local time, as determined by Ward (1972) to minimise mean ionospheric Doppler frequency shifts (not necessarily a problem, except when extreme: However, the ionosphere will consistently support radio transmission at 21.840 MHz for a period of only some 8 to 12 hours around local noon each day.) The radio propagation path was predominantly a single-hop F2 layer reflection, and ground ranges were calculated using an assumed reflection layer height of 300 km. These ranges could only be confirmed where the scattering cells included identifiable land features, e.g. in the vicinity of Tasmania.

Because of the large radar beam width (±12°), scattering cells can have considerable cross-beam dimensions - over 1000 km at 3000 km range. With a 1 ms pulse length, this gives average cell sizes around 150 km by 800 km. Since it is obviously unrealistic to expect sea state (and surface winds) to remain even approximately constant over such areas in most synoptic situations, this poor resolution of the radar-derived data limits its usefulness. Nevertheless, it will be noted that the results obtained from this set of experiments conform sufficiently closely to meteorological analyses determined from other sources as to point to some potential value for such large-scale averages, at least in defining synoptic scale systems over remote ocean areas.

A number of difficulties in data analysis and interpretation result from the use of an ionospheric propagation mode. Of most significance are:

(a) the frequent occurrence, for most bearings and ranges, of ionospheric multipath, i.e. more than one propagation path for a given radar range as determined by total-time-of-flight, implying scattering from different sea areas. Such a phenomenon is manifest as a splitting of the Bragg lines into two or more peaks, and examples may be seen in Figs 1(a) and 1(c). Figure 1(b) contains relatively clean Bragg lines;

(b) broadening of the Bragg lines due to small scale and random ionospheric movements, including O- and E-ray contributions. Such Doppler broadening imposes a resolution limit of the order of 0.14 Hz (for backscatter at 21.84 MHz) as found by Heron and Rose (1977), but may be considerably larger in occasional spectra. Neither of these phenomena affect the determination of wind directions, but are likely to cause substantial errors in wind speed computations.

Few direct observations of surface winds over the sea were available for comparison with the radar-derived data. The nearest available mean sea level (MSL) meteorological analyses were those for 0300 GMT (1300 local time) on each day. These are assumed to have been reasonably accurate, at least in terms of the location and strength of the major synoptic scale weather systems in the Tasman Sea region, since they also incorporate the daily visible and infrared satellite cloud pictures (NOAA-V, picture time near 2200 GMT daily). Thus the MSL analyses for 0300 GMT on 6 and 13 April afford the principal comparison for the radar data, and these are shown in Figs 2 and 3 respectively. Also included in the figures are such surface wind observations as were available and relevant to the experiments.
Fig 2 Surface meteorological analysis for 0300 GMT on 6 April 1977, with radar-derived surface wind directions superimposed. Solid arrows are radar winds obtained using the technique of Stewart and Barnum (1975) with a fixed mean wind speed; dotted arrows are radar winds from the same technique but with variable speeds inferred from the surface pressure contour spacings. Also shown are measured surface wind speeds, where available.
Fig 3 Surface meteorological analysis for 0300 GMT on 13 April 1977, with radar-derived surface wind directions superimposed. Solid arrows are radar winds obtained using the technique of Stewart and Barnum (1975) with a fixed mean wind speed; dotted arrows are radar winds from the same technique but with variable speeds inferred from the surface pressure contour spacings. Also shown are measured surface wind speeds, where available.
ANALYSIS AND DISCUSSION

In order to relate the two first order Bragg lines in the backscatter spectrum to a surface wind direction, it is necessary to assume some form for the ocean wave directional spectrum. Stewart and Barnum (1975) modified a technique due to Tyler et al. (1974) and assumed this spectral form to be

\[ g(\theta) = \cos^s(\theta/2) \]  ...2

where \( \theta \) is the angle between mean wind and radio propagation direction, and \( s \) is a parameter that Stewart and Barnum (1975) have related to wind speed through the parameter \( \mu \) of Tyler et al. (1974) (which characterises momentum transfer from wind to water) and a drag coefficient for surface wind over the sea:

\[ s = 0.4 \ (\mu - 0.1)^{-1} \quad \mu > 0.1 \]  ...3
\[ s = 4 \quad \mu < 0.1 \]

and

\[ \mu = 3.65 \times 10^{-3} \ U^{1/4} \ p^{1/2} \quad U < 15 \text{ m/s} \]  ...4
\[ \mu = 8.33 \times 10^{-3} \ U^{1/8} \ p^{1/8} \quad U > 15 \text{ m/s} \]

where \( U \) is surface wind speed (at 10 m) and \( F \) is radio frequency in MHz.

The ratio of advancing to receding waves is

\[ r = \frac{g(\theta)}{g(\theta + \pi)} = \tan^s(\theta/2) \]  ...5

However,

\[ r = \frac{A^+}{A^-} \]  ...6

where \( A^+ \) and \( A^- \) are the advance and recede Bragg line amplitudes respectively.

Thus \( \theta = 2 \arctan \left( \frac{A^+}{A^-} \right)^{1/s} \)  ...7

In practice, Stewart and Barnum (1975) claim that Eqn 5 will apply for all but very small \( \theta \).

Long and Trizna (1973), on the other hand, have employed the parameter

\[ \zeta = 10 \log \left( \frac{A^+}{A^-} \right) = 10 \log r \]  ...8

which varies (theoretically) from \( +\infty \) through 0 to \( -\infty \) for the full possible range of wind directions. By plotting normalised (to remove ionospheric loss effects) values of \( A^+ \) and \( A^- \) against \( \zeta \) they were able to determine an empirical ocean wave spreading factor, leading to the relationship

\[ \zeta = 20 \log \left( \frac{0.56 + 0.50 \cos 2\theta}{9} \right) + 34.02 \ (\text{dB}) \]  ...9

which may be solved for \( \theta \) in terms of \( \zeta \). This result is independent of surface wind speed.

In the present experiments we have used primarily the formulations of Stewart and Barnum (1975), Eqns 3, 4, and 5. However, in view of the fundamental uncertainty in \( \theta \) of up to \( \pm 12^\circ \) imposed by the wide aerial beamwidth, the large sea scattering areas involved and, in particular, the lack of direct observations of wind speed within the scattering areas, it was decided to use a single value of
U (and hence of \(u\) and \(s\)) throughout the radar computations. ...The value of 10 m/s was chosen as being a reasonable mean value in view of the surface pressure gradients existing in the Tasman Sea region at the time. This gave \(u = 0.305\), and \(s = 1.97\), very close to the \(s = 2\) that is still widely used, particularly in engineering type applications.

Based on these equations, and using spectra such as those in Fig 1, radar wind directions were computed, and full results are shown as the solid arrows in Figs 2 and 3 for the two days of the experiment. When allowance is made for cross-isobar flow (around 15° over the oceans), the agreement between radar derived winds and the surface pressure field is reasonable, especially on 6 April. (It should be noted that the left/right directional ambiguity in the radar winds has been resolved on the basis of other available meteorological information, and to maintain consistency, in all cases.)

Some features of the spectra in Fig 1 are relevant to the experiments in general, and should be considered. All these spectra were obtained from true bearing 180°, at ranges of 2700, 2500, and 2300 km, respectively. This placed the scattering area for Fig 1(a) across the southern half of Tasmania, that for Fig 1(b) through Bass Strait, while Fig 1(c) contains echoes from land areas only (see Fig 2 also). While the noise level is some 30 dB down from the strong zero Doppler peak in Fig 1(c), it is as high as -20 dB in Figs 1(a) and 1(b). This may be due to differing contributions to the Doppler spectrum from significantly varying sea states throughout the large scattering area.

The problem of multipath, evident in both Figs 1(a) and 1(c), has already been noted. While it does not seriously hinder wind direction computations, it is sufficiently bad in Fig 1(a) to preclude any assessment of peak width \(B\), and hence of windspeed.

Finally, the marked change in Doppler sense between Figs 1(a) and 1(b), i.e. over a radial distance of some 200 km, shows that a spatial resolution of this magnitude, which is near the minimum attainable with our system, is still adequate for the delineation at least of synoptic-scale meteorological systems.

An attempt has been made to improve the direction fit using Eqns 3, 4, and 5, and variable wind speeds estimated from mean sea level (MSL) pressure analyses, Figs 2 and 3. These estimates were simply geostrophic winds, corrected to sea surface (at 10 m) using the mean factors from Findlater et al. (1966), and as such are likely to have inaccuracies of ±50 per cent or greater, even in the Tasman Sea region. The radar wind directions so determined are shown as the dotted arrows in Figs 2 and 3. In general there is a slight improvement in the fit on 6 April, but the correlation is somewhat worse on 13 April, especially in the strong cross-wind situation in the southern Tasman. This is likely to be due as much to inaccuracies in analysed pressure gradients and derived geostrophic winds as to errors inherent in the radar analysis technique or resulting from the large radar beamwidth.

Radar wind directions were also determined using the Long and Trizna (1973) technique, Eqn 9, but in general the fit was significantly worse (of the order of 20°) than that for either of the result sets shown in Figs 2 and 3, except for the southern Tasman Sea on 13 April.

The analysed wind speeds were also used in an attempt to find some correlation with a radar-derived parameter. The technique employed was that of Stewart and Barnum (1975), as modified by Maresca and Barnum (1977), and used the principal Bragg line width \(B\). Maresca and Barnum adopted a simple normalisation of \(B\)

\[
B^1 = \left(\frac{F}{F_0}\right)^{\frac{1}{2}} B
\]  
...10
in order to allow for the varying transmitting frequencies (\$ F \$ MHz) of their experiment. Such a normalisation is unnecessary for our data, as we used only a single frequency (21.840 MHz), and we have plotted (in Fig 4) simply B against U, the surface wind speed (in m/s) in the centre of the scattering area. This was obtained from the relevant 0300 GMT MSL chart as detailed above. In all only ten values of B could be extracted because of multipath and other problems, but the results do show an obvious trend. A simple linear regression on the data yielded the relation

\[
U = (73.98 - 15.6) \quad \ldots 11
\]

with a standard deviation of \( \pm 3.6 \) m/s, and this is drawn as the solid line in Fig 4. Also included in the figure for comparison are the relationships

\[
U = 39.4 \ (B - 0.1) \quad \text{(Stewart and Barnum 1975)} \quad \ldots 12
\]

\[
U = 33 \ B^4 - 2 \quad \text{(Maresca and Barnum 1977)} \quad \ldots 13
\]

labelled SB and MB, respectively.

\[\begin{figure}
\centering
\includegraphics[width=\textwidth]{fig4.png}
\caption{Measured width (B) of the principal Bragg line in Doppler spectra such as Fig 1(b), at a level 10 dB below the peak, plotted against surface wind speed U deduced from the surface pressure gradient in the relevant scattering area. Solid line is a simple linear regression on the data with long dashes delineating \( \pm \) one standard deviation (\( \pm 3.6 \) m/s). Short dashed lines are the relations of Stewart and Barnum (1975) and Maresca and Barnum (1977), labelled SB and MB, respectively.}
\end{figure}\]

The correlation coefficient for the regression, Eqn 11, is 0.92, and the relationship has been derived for wind speeds of from 2 to 27 m/s. It represents a remarkably good fit, in view of the inaccuracies inherent in the values of both \( B \) and \( U \). It must be conceded, however, that these errors, particularly in \( U \), may be systematic. In addition, the fundamental uncertainty in \( B \) of around 0.14 Hz (Heron and Rose 1977), imposed by small-scale ionospheric motions, is seen from Eqns 11, 12, and 13 to represent a corresponding uncertainty in \( U \) of from 4.4 to 10.3 m/s (depending on the particular relationship adopted).
CONCLUSIONS

A set of experiments has been conducted to test the potential value of observations made by the James Cook University HF radar, operating in an ionospheric propagation mode, for deriving surface wind data over remote ocean areas of value to a meteorological synoptic data network. In view of the low spatial resolution imposed by the wide aerial beamwidth and problems associated with ionospheric variability and multipath, this potential is obviously limited. Nevertheless, the results achieved by the experiments discussed are at least encouraging, and point to some possible future role for the radar in large-scale meteorological ocean surveillance. Significant benefits could in turn accrue to both meteorological analyses over the ocean in the Australian region, and associated ocean activities such as ship-routing.

A generally close fit has been obtained between radar-derived surface-wind directions (averaged over the large radio scattering area) and those inferred from meteorological synoptic analyses for similar times. The majority of the radar wind observations were obtained from the Tasman Sea region, where such analyses are likely to be most accurate. Scattering ranges varied from 1500 to 3700 km from the radar site (Townsville). In addition, a parameter B was measured from the Bragg lines in the Doppler spectra of the backscattered radio signal, which could be used to infer wind speeds, from 2 to 27 m/s, with an ultimate precision down to 24.4 m/s.

Doppler spectra showing some coherence, indicating the possibility of meaningful data analyses, were also obtained for some two-hop modes to ocean areas of range 6000 km to 8000 km. Finally, further upgrading of the facility since these experiments were performed now enables the ocean radar to be used in a surface wave mode, for high-resolution observations at short ranges. Such observations may be used to estimate mean ocean surface currents (e.g. Stewart and Joy 1974), or to provide ground-truth data for ocean surveillance satellites such as SEASAT-A.

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REFERENCES


