

# Measured Ionospheric Distortion of HF Ground-Backscatter Spectra

R. MICHAEL JONES, T. M. GEORGES, AND J. P. RILEY

**Abstract**—Skywave-radar measurements of backscatter from land are used to estimate some statistical properties of the ionospheric distortion encountered by sea-state radars. We confirm that samples of the path spectrum  $P(\omega)$  are normally distributed and uncorrelated in frequency, and we estimate the first two moments of  $P(\omega)$  and their spatial and temporal correlation, for quiet  $F$ -layer daytime propagation. An empirical relation connects the equivalent width of  $P(\omega)$  and expected errors in estimating ocean wave height.

## I. THE PROBLEM—IONOSPHERIC DISTORTION OF SEA-ECHO SPECTRA

TEMPORAL FLUCTUATIONS of ionization density broaden and split the frequency spectra of ionospheric radio transmissions, an effect long known to communicators [1]–[4]. Such distortions are particularly troublesome to remote sensors that use ionospheric paths, such as HF skywave sea-state radars [5], [6]. Extracting the desired information about the sea (wave height, for example) depends on resolving backscatter spectral features separated by less than 0.1 Hz in frequency and more than 20 dB in intensity [7], [8]. Such sensors can be successful only if we find ways to model and deal with ionospheric distortion [9], which in turn requires representative measurements of its spectrum and variability. This paper summarizes the results of our attempt to isolate and measure the spectrum and some statistical properties of ionospheric distortion over two-way skywave-radar paths. More detailed results can be found in six National Oceanic and Atmospheric Administration (NOAA) Technical Memoranda [10]–[15].

## II. A MODEL OF THE DISTORTION

The spectrum of skywave echoes backscattered from the land or sea depends not only on the characteristics of the scattering surface and the ionospheric path, but also on such radar characteristics as antenna beamwidth, range gate, and coherent integration time [16]. In general, the complex amplitude spectrum of skywave-radar echoes can be represented mathematically by a convolution in the frequency domain [17]:

$$R(\omega) = S(\omega) * P(\omega) * W(\omega) \quad (1)$$

where  $S(\omega)$  is the temporal Fourier spectrum of echoes from the scattering surface,  $P(\omega)$  is the Fourier spectrum of the effective ionospheric reflection coefficient,  $W(\omega)$  is the Fourier transform of a time window used to sample the echo,

$\omega$  is the frequency offset from the radar's carrier frequency, and the explicit dependence on other radar parameters has been suppressed in (1). If the scattering surface is land,  $S(\omega) = \delta(\omega)$ ; if it is sea,  $S(\omega)$  is the well-known double-peaked spectrum of Bragg scattering from the sea [18].  $P(\omega)$  is called an effective ionospheric reflection coefficient because it represents the combined effects of all the outgoing and returning ionospheric paths within the radar beam and range gate. Equation (1) is rigorously correct because scattering, reflection, and weighting can all be represented as products in the time domain [17]. (Notice, however, that the simple relation (1) does not apply to power spectra, except in the case of an infinite ensemble average.)

In remote sensing by sea-state radar, one measures  $R(\omega)$  and attempts to extract the spectrum of the scattering surface  $S(\omega)$  in the presence of the unknown path distortion  $P(\omega)$  and the known window spectrum  $W(\omega)$ . If  $P(\omega)$  were known or could be measured, the extraction would be a straightforward deconvolution. In practice, however,  $P(\omega)$  varies unpredictably in space and time and cannot be measured for the exact radar path and interrogation time. Furthermore, measurements differing from the radar path by a few minutes in time or a few kilometers in space give quite different estimates of  $P(\omega)$  [9]. The best we can usually do is estimate some statistical properties of  $P(\omega)$  with representative measurements and base processing strategies on those estimates.

## III. WHAT DO WE EXPECT $P(\omega)$ TO LOOK LIKE?

From our knowledge of the spectra of point-to-point ionospheric transmissions [1]–[4], [26], we expect radar returns to exhibit multipath caused by multiple ionospheric layers, traveling ionospheric disturbances (TID's) and other spatial inhomogeneities, as well as multipath caused by magnetoionic splitting. In general, each such path can impose a different Doppler shift, typically a few tenths of a hertz. It is not uncommon, for example, to observe the spectrum of a point-to-point signal split into six discrete lines by the combination of TID and magneto-ionic multipath [4, fig. 6–13], [3].

Within a skywave-radar beam that includes ground or sea backscatter, the number of possible paths is multiplied by the number of scatterers within the radar footprint and range gate, and by the number of combinations of outgoing and returning paths, so that the corresponding path spectrum is essentially a continuum. Ionospheric nonstationarity contributes still more spectral structure when echoes are integrated over a radar dwell time.

Assuming that space and time averages of  $P(\omega)$  have qualitatively the same effect, we can get an idea of what  $P(\omega)$

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The authors are with the Wave Propagation Laboratory, National Oceanic and Atmospheric Administration, Boulder, CO 80303.  
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for a skywave radar might look like by averaging point-to-point spectra over the time it takes typical ionospheric motions to pass through the radar beam. If our radar beam is 5 km wide where it passes through the ionosphere, and typical ionospheric disturbances travel at 150 m/s, then a representative averaging time would be about 300 s. One can obtain a visual impression of the effect of such temporal smearing for different ionospheric conditions by imagining it applied to the frequency-versus-time displays obtained with HF Doppler sounders over point-to-point paths [1]-[4].

#### IV. HOW CAN WE ISOLATE $P(\omega)$ ?

Because we cannot measure  $P(\omega)$  directly, we have to resort to indirect measurements from which  $P(\omega)$  can be estimated. We considered two possible ways to isolate measurements of  $P(\omega)$ : one using skywave-propagated backscatter echoes from the land (since  $S(\omega) = \delta(\omega)$  for ground scatter), and the other using transponder echoes over skywave paths. Ground backscatter would give an estimate of  $P(\omega)*W(\omega)$  that would be more relevant to sea-state radar measurements, because it would average  $P(\omega)$  over the radar beam and range windows in the same way. A disadvantage is that ground-scatter coefficients usually vary in space, biasing the measurements with an unknown spatial weighting function. Transponder measurements, on the other hand, would give a  $P(\omega)$  corresponding to all the paths connecting the radar and a point target, which would depend less on radar parameters but would be more difficult to interpret in terms of areal scattering. In our experiment, we attempted both kinds of measurement, but failed to obtain usable transponder echoes. Therefore, our interpretations are based on ground-backscatter measurements alone.

#### V. WHAT PROPERTIES OF $P(\omega)$ ARE USEFUL TO MEASURE?

First, it would be useful to know how  $P(\omega)$  is distributed statistically from sample to sample. There is strong evidence that  $S(\omega)$  is a Gaussian random variable for sea echoes [22], and some of our models of the statistics of ensembles of  $R(\omega)$  depend on the distribution of  $P(\omega)$  being Gaussian, too. Our measurements substantiate that assumption, and they also show that  $P(\omega)$  has zero mean and a normalized variance of one.

Because the smearing of sea-echo spectra depends mostly on the width of  $P(\omega)$ , we would like to know typical values of width and how it varies in space and time and with ionospheric conditions. During our measurements, equivalent width for  $F$ -layer reflections varied between 0.03 and 0.1 Hz, for our antenna beamwidth [16]. By direct measurement, a typical correlation time for changes in width is about 3 min, and a typical correlation distance is about 5 km.

We would also like to know something about the shape of  $P(\omega)$ , both individually and on the average. Is it smooth and single-peaked, or is it spiky like a discrete spectrum? And how does its shape change spatially and temporally? Our measurements, which were made using a 102 s integration time and a Blackman-Harris weighting [19], suggest something in between a smooth and discrete spectrum.

The mean frequency shift of  $P(\omega)$  and its spatial and

temporal variability are of interest, for example, in skywave measurements of ocean currents, where ionosphericly induced shifts are indistinguishable from those caused by ocean currents. Our measurements suggest that mean shifts of about 0.3 Hz are common for  $F$ -layer propagation, and that the shift decorrelates in a few kilometers spatially and a few minutes temporally.

Finally, is  $P(\omega)$  correlated in Doppler frequency? Some properties of our contamination model depend on the assumption that adjacent frequency values of  $P(\omega)$  are uncorrelated. Our measurements, which are really of  $P(\omega)*W(\omega)$ , suggest that there is no more correlation between adjacent frequencies than that introduced by  $W(\omega)$ .

The dependence of  $P(\omega)$  on radar parameters such as antenna beamwidth, range gate and integration time are not considered here, but were treated in another paper [16].

#### VI. DETAILS OF THE MEASUREMENTS

We used the FM-CW skywave research radar known as WARF (Wide Aperture Research Facility), operated by SRI International in central California and described in detail by Washburn *et al.* [20]. Most of the measurements discussed here were made on October 17, 1980, with the  $0.5^\circ$  radar beam pointed eastward and using a range gate at 9.25 ms, corresponding to a transponder location near Albuquerque, NM. The radar frequency was 18.3 MHz, and the ionospheric diagnostics [21] revealed  $F_2$ -layer propagation through a quiet ionosphere free of unusual disturbances and sporadic- $E$  layers. Although our results do not cover a wide variety of ionospheric conditions, we believe they are typical of quiet daytime  $F_2$ -layer propagation.

Data were collected for a 25.6 min period beginning at 1743 UT (midday over the path) and processed as we process sea echoes in an array of four azimuth cells by 21 range cells [9, fig. 5]. The received time series was processed every 25.6 s in 102.4 s "dwell-time" segments that overlap by 75 percent and are weighted by a minimum four-term Blackmann-Harris window [19]. An on-line array processor performed the required range and time FFT's (fast Fourier transforms), the result of which is 84 spectra every 25.6 s. Each spectrum represents the echo from a ground cell measuring approximately 3 km in range and 13 km in azimuth, with some cell overlap because of antenna beam shape and windowing of the range FFT.

The data presented here are from 15 nonoverlapping time cells (25.6 min). For some calculations, we average the 84 individual spectra, shifting in frequency and normalizing in power level, obtaining a single spectrum representing the  $63 \times 32$  km radar footprint. In another, we show how the 84 individual spectra vary over the footprint. Time histories of equivalent spectral width are calculated from both overlapping and nonoverlapping time series.

#### VII. REPRESENTATIVE AVERAGE GROUND-BACKSCATTER SPECTRA

Fig. 1 shows three average land-echo skywave spectra, each of which is derived from the 84 individual spectra obtained during one radar dwell time. The three spectra shown represent low, medium, and high amounts of ionospheric

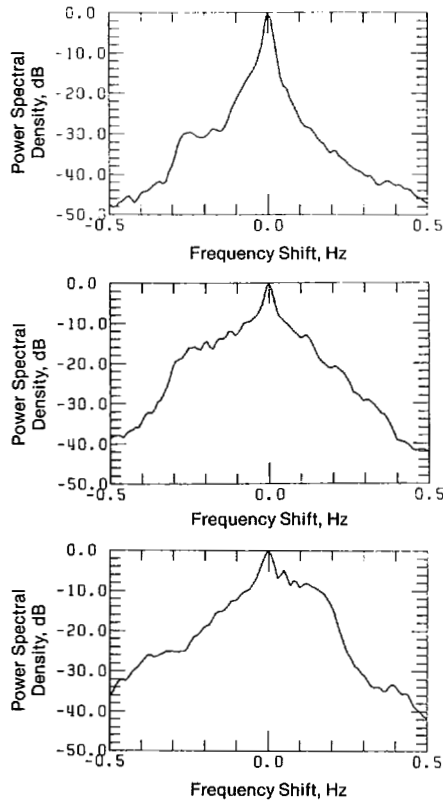


Fig. 1. Backscatter power spectra from three 102 s time windows showing the range of ionospheric distortion of land-echo spectra, from low (top, equivalent width,  $W = 0.037$  Hz) to medium (middle,  $W = 0.064$  Hz) to high (bottom,  $W = 0.083$  Hz). Each spectrum is the average of 84 FFT's from 21 range cells by four azimuth cells, in which individual spectra are frequency shifted to the same peak value and logarithmically normalized to equalize their contributions to the mean. In the absence of ionospheric distortion,  $W$  would be about 0.02 Hz with the time window we used.

distortion. Before averaging, however, the individual spectra are shifted in frequency to align their spectral peaks, and they are logarithmically normalized to equalize their contributions to the mean spectrum [9]. This is the same process we use to obtain sea-echo spectra for wave-height processing, so that land echoes thus processed represent the effective  $P(\omega) * W(\omega)$  that would have distorted sea echoes, had they been obtained over the same ionospheric paths. The corresponding errors in wave-height estimates are discussed in Section XIII.

### VIII. THE STATISTICAL DISTRIBUTION OF $P(\omega)$

Samples of sea-echo amplitude spectra have a Gaussian distribution and a normalized variance (the variance divided by the square of the mean) of unity [22]. Assuming that the ground consists of many small scatterers adding randomly, we guessed that the amplitude of scatter from the ground would be similarly distributed. For similar reasons, we guessed that the ionospheric path distortion  $P(\omega)$  might have the same statistics but a different spectrum. Theoretically [23], combining surface scatter having Gaussian statistics with a path distortion also having Gaussian statistics should yield a received signal  $R(\omega)$  whose power spectrum has a normalized variance of 3.0 and exhibits a Hankel distribution.

When we looked at the variance and distribution of skywave echoes from the land, we found variances significantly larger

than we expected. Fig. 2 shows some cumulative distributions of 20 different spectral lines, each distribution consisting of the set of the 84 spatial samples from time period 7. The normalized variance ranged between 2.0 and 17.0, with a mean of 6.2. Skywave echoes from the sea, however, showed much closer agreement with the Hankel distribution, having a normalized variance between 1.6 and 3.7 and a mean variance of 2.5.

These results suggest that our assumption about the statistics of  $P(\omega)$  are essentially correct, but that our assumptions about the statistics of the ground-backscatter echoes are incorrect. In other words, the large variances of the ground-backscatter echoes are caused by the statistics of the ground echoes, not the ionosphere. When we examined the spatial variation of the ground-backscatter intensity, we found consistent trends in range and azimuth, supporting the conjecture that the large variance of ground-backscatter amplitude comes from the ground itself [13, fig. 4]. We suspect that removing those trends would bring the distributions for ground scatter close to the Hankel distribution, but we did not check this before the project ended.

### IX. THE EQUIVALENT WIDTH OF $P(\omega)$ AND ITS VARIABILITY

We have determined that equivalent spectral width, defined in [16], is a useful index of ionospheric distortion, particularly its influence on wave-height estimates (Section XIII). (Briefly, equivalent width is computed by dividing the power in some spectral interval by the peak power spectral density.) Earl and Bourne [6], using a  $10^\circ$  radar beam, found that spectral widths of ground-backscatter signals were typically in the range 0.2 to 0.3 Hz for both  $E_s$ - and  $F$ -propagated echoes. Using a  $0.5^\circ$  beamwidth radar, however, Georges and Maresca [16] found average widths of about 0.1 Hz for  $F$ -layer-propagated echoes, and about 0.06 Hz for  $E_s$ -propagated echoes. Thus, one observed difference between backscatter spectral distortion and that measured on point-to-point paths is a dependence on the radar's beamwidth.

Fig. 3(b) shows how equivalent width varies with time over the 25.6 min time period of our measurements. This particular time history averages 84 width computations that cover the radar footprint in each overlapping 102 s dwell time. Two other such measurements on other days are shown in Figs. 3(a) (sea echoes) and 3(c) (land echoes). They tell us that such a time history is typical of quiet  $F$ -layer propagation, with mean width increasing by 0.02 Hz or more under disturbed conditions. Computations of the autocorrelation of the three histories [21] show that equivalent width decorrelates in about 3 min.

### X. THE SHAPE OF $P(\omega)$ AND ITS VARIABILITY

The spatial variability of  $P(\omega)$  is graphically shown in Fig. 4, which plots the individual received echo spectra for all 84 spatial cells that make up the  $63 \times 32$ -km radar footprint at about 1803 UT. This time period produced the sharpest spectra (smallest width) obtained during the test and was selected because this interval would normally be most suitable for wave-height computation. Spectra obtained at all other times were broader.

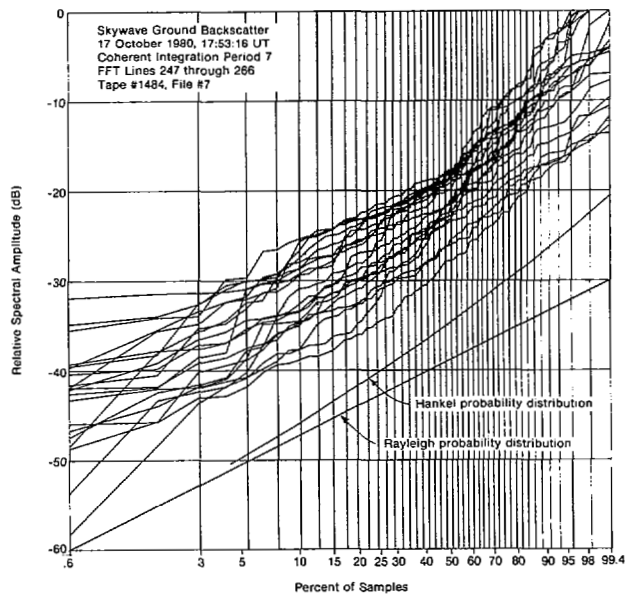


Fig. 2. Cumulative distribution functions (CDF's) for 84 skywave ground backscatter spectra for October 17, 1980, 1753:16 UT (dwell time 7). Cumulative distribution functions are also shown for a Rayleigh and a Hankel distribution. For the most part, the CDF curves are steeper than those for the Hankel or Rayleigh distribution, implying that the CDF's have a larger variance than that of the Hankel or Rayleigh distribution. That large a variance is not consistent with a normalized variance of one for both the ionospheric reflection coefficient and the ground backscatter coefficient.

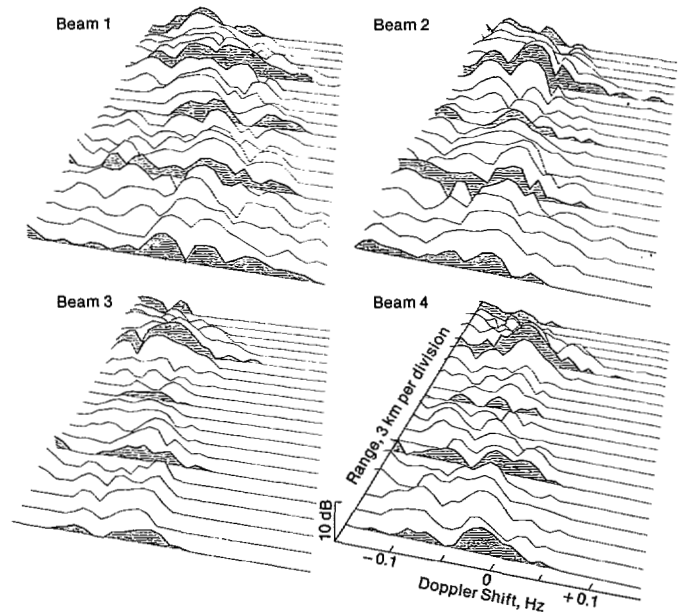


Fig. 4. Land echoes processed by WARF in the same way sea echoes are processed display the spatial variability of the ionospheric distortion. These four perspective plots show how land-echo spectra in a 0.3 Hz spectral interval about the peak vary in range and azimuth. The spatial cell size covered is roughly 63 km in range and 32 km in azimuth. Only the 10 dB range below the largest peak is shown for pictorial clarity. Land echoes exhibit spatial variations in strength, but produce no Doppler shifts, so the spectral shapes and frequency shifts can be attributed entirely to the ionosphere; however, variations in overall spectral level probably have a land-induced component. This figure gives an idea of the spatial variability of the ionospheric transmission function.

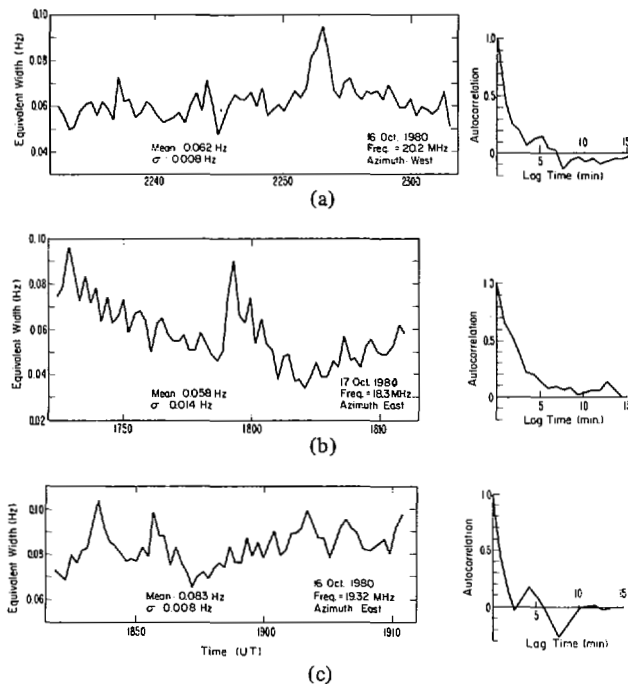


Fig. 3. Three examples of computations of equivalent spectral width over about 30 min time intervals and their autocorrelation coefficients, showing the character of temporal fluctuations in the spectral quality caused by ionospheric spreading. In (a) equivalent width for sea-echo spectra, whereas (b), (c) are for ground backscatter. Autocorrelation coefficients were calculated by the FFT method with a Hanning spectral window.

This display makes it obvious that  $P(\omega)$  changes drastically in two or three cells in azimuth or range. (The azimuth beams are approximately  $0.5^\circ$  wide and separated by about  $0.25^\circ$ , so that each cell covers a ground area of about 3 km in range, 13 km in azimuth, and 102 s in time.) Spectral changes must be caused by the ionosphere, because changes in the ground-reflection coefficient can cause only variations in echo power. Some of the apparent correlation between adjacent cells is caused by overlap in the radar's delay-time and azimuth windows, so the actual spatial correlation of  $P(\omega)$  is even less than this figure shows.

#### XI. THE MEAN FREQUENCY SHIFT AND ITS VARIABILITY

Fig. 4 also shows how the mean frequency of the echo spectrum drifts in range and azimuth. This is of concern in skywave measurements of ocean currents, which shift the sea-echo spectrum in the same way. Some claims have been made about ocean currents measured with skywave radar, in which the ionospheric shifts have been estimated and removed by observing the shift of nearby land echoes. Such procedures must be questioned, since a radial ocean current of 1 m/s (a large current) would produce a spectral shift of only 0.12 Hz (assuming a radar frequency of 18 MHz), smaller than a typical  $F$ -layer-induced variation of Doppler shift in a few tens of kilometers. Current measurements using more stable (lower) ionospheric layers and nearby land references may be practical, however.

Temporal variations of mean Doppler shift are of similar magnitude, as Fig. 5 shows. This 12 min drift of mean

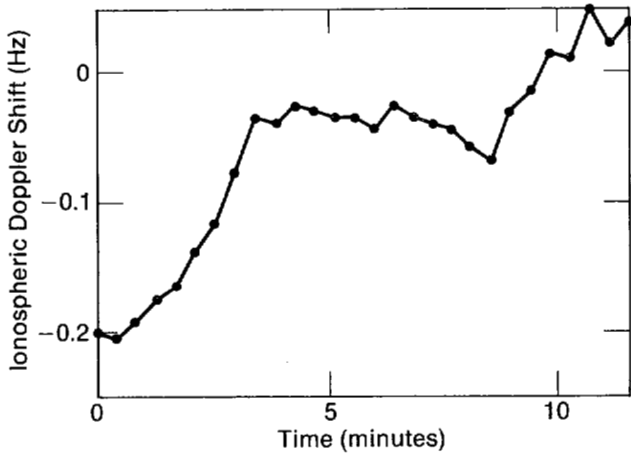


Fig. 5. A 12 min history of average frequency shift of 16.25 MHz sea echoes, obtained at WARF on April 5, 1978. Such variability is typical of quiet F2 layer propagation.

frequency was recorded on another measurement day (April 5, 1978), but the ionospheric conditions were similar and the echoes were sea echoes, which should not influence the short-term drift of the mean frequency.

XII. THE CORRELATION OF  $P(\omega)$  IN FREQUENCY

Barrick and Weber [24] showed that the spectra of direct sea echoes are theoretically uncorrelated in frequency. We verified this prediction by computing the correlation of sea-echo spectra obtained with the NOAA Coastal Ocean Dynamics Applications Radar (CODAR) system [25]. This justifies our assumptions in [10]–[15] and [17] that  $S(\omega)$  is uncorrelated in frequency and simplifies the mathematics of analyzing the properties of ensembles of such spectra.

To benefit from applying the same simplifying assumptions to skywave spectra, we would like to assume that  $P(\omega)$  is also uncorrelated in frequency, that is, that the ionosphere introduces no correlation between adjacent frequency components of skywave-propagated echo spectra. To verify this assumption, we measured the frequency correlation of the ground-backscattered skywave spectra obtained at WARF on October 17, 1980.

Because the received echo spectrum  $R(\omega)$  is complex, we have to look at the real and imaginary parts of the ensemble averages  $\langle R(\omega_1)R^*(\omega_2) \rangle$  and  $\langle R(\omega_1)R(\omega_2) \rangle$ , as functions of  $\omega_1 - \omega_2$ , where  $\omega_1$  and  $\omega_2$  are different values of spectral frequency. For brevity, we show in Fig. 6 only  $\text{Re} \{ \langle R(\omega_1)R^*(\omega_2) \rangle \}$  for one dwell interval; plots of the remaining functions for other intervals are shown in [15]. The ensembles used are, of course, not infinite, but consist of 16 frequency pairs and the 84 individual spectra obtained for each time period of interest. Because the ensemble is finite, we show also the one standard deviation limits.

The correlation exhibited by Fig. 6 is caused entirely by the time window used before the FFT is applied to the received time series, as shown by the heavy line in the figure, which was computed for a Gaussian window. A Blackmann–Harris window was actually used, but its correlation properties are close enough to those of a Gaussian window that we can be

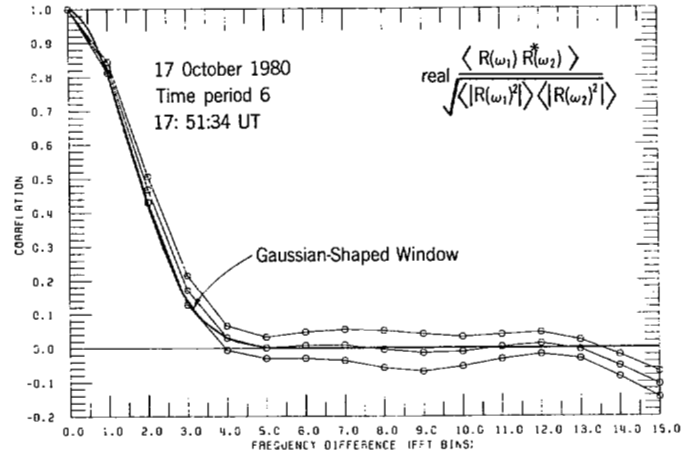


Fig. 6. Frequency correlation (real part of  $\langle R(\omega_1)R^*(\omega_2) \rangle$ ) for ionospherically reflected ground backscatter for October 17, 1980, coherent integration period 6. For each value of  $\omega_2 - \omega_1$ , we calculated 16 samples of this ensemble average for different values of  $\omega_1$ . The center curve in the figure shows the mean of the 16 samples. The upper and lower curves show the mean plus and minus one standard deviation for a single sample; they also show the mean plus and minus 0.25 standard deviation for the mean. The theoretical correlation introduced by the window is also shown. The mean frequency correlation is not significant because it lies within one standard deviation of the correlation introduced by the window.

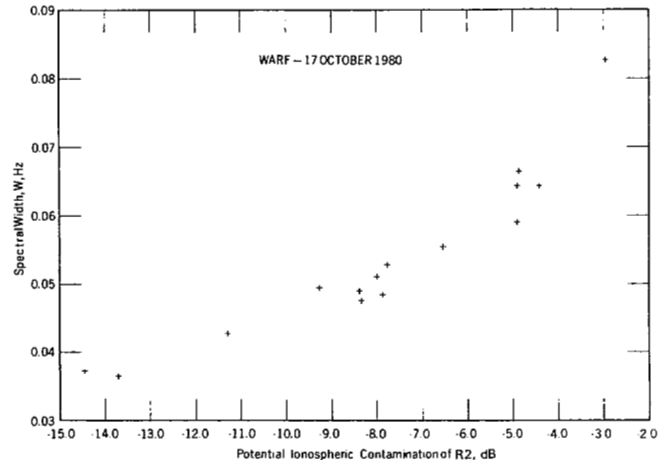


Fig. 7. A scatter plot showing that equivalent spectral width,  $W$ , is highly correlated with the error in estimating  $R_2$ , which is related to ocean wave height by a power law. The ionospheric contamination is expressed as the increase in  $R_2$  (in decibels) caused by ionospheric spreading when first-order sea echoes spill over into the spectral region normally interpreted as second-order scatter. The error in estimating  $R_2$  is computed from space-averaged land-echo spectra by taking the ratio of the power outside a band  $\pm 0.07$  Hz about zero to the power inside that band.

confident that no additional frequency correlation of  $P(\omega)$  exists, within the uncertainty of the correlation estimates.

XIII. IMPLICATIONS FOR OCEAN WAVE-HEIGHT MEASUREMENTS

One relatively simple way to estimate the height of short-period wind-driven waves on the ocean from sea-echo spectra is to measure the ratio of total power contained in the first- and second-order parts of the dominant half of the echo spectrum [8]. Simulations using theoretically computed spectra show that this ratio, which we call  $R_2$ , can be related to ocean wave height by a power law. Therefore, errors in wave-height estimates caused by the ionosphere can be traced to distortions

of  $R2$ . Furthermore, distortions of  $R2$  can be directly related to the equivalent width of  $P(\omega)$ . In other words, the wider  $P(\omega)$ , the larger the error in the wave-height estimate derived from  $R2$ .

Jones *et al.* [13, eq. D-4] have shown that the error in  $R2$  caused by a  $P(\omega)$  of any shape can be approximated by the ratio of the portion of  $W(\omega)*P(\omega)$  outside some frequency band  $B$ , to the portion inside that band. For practical use,  $B = \pm 0.07$  Hz. If we apply this formula to the 15 (nonoverlapping) measured average spectra (like the three shown in Fig. 1) obtained during the 17 October experiment, we get the expected errors in  $R2$  under those ionospheric conditions. The scatter plot of Fig. 7 shows that  $W$  is highly correlated with  $R2$  error, at least in this example. If, for example, the correct  $R2$  were  $-10$  dB, the  $R2$  errors would range between 35 and 51 percent. From [8, formula 7] the corresponding errors in wave height would be 20 and 193 percent.

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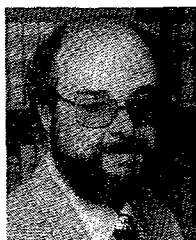
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**R. Michael Jones** was born in Los Angeles, CA, in 1937. He received the B.S. degree from the California Institute of Technology, Pasadena, in 1959 and the M.S. and Ph.D. degrees from the University of Colorado, Boulder, in 1963 and 1968, all in physics.

He has been employed by the U.S. Department of Commerce in Boulder, Colorado, since 1961, first at the National Bureau of Standards, later at the Institute for Telecommunication Sciences, and currently for the NOAA Wave Propagation Laboratory. He has worked on radio wave propagation in the ionosphere, radio groundwave propagation, and acoustic-gravity-wave propagation in the atmosphere and ocean.

Dr. Jones is a member of the Acoustic Society of America, the International Association of Mathematical Physics, the International Society on General Relativity and Gravitation, and the Rocky Mountain Peace Center.



**T. M. Georges** was born in Los Angeles, CA, on August 8, 1938. He received the B.S. degree from Loyola University, Los Angeles, CA, in 1961, the M.S. degree from the University of California, Los Angeles, in 1963, and the Ph.D. degree from the University of Colorado, Boulder, in 1967, all in engineering.

Since 1963, he has worked at the U.S. Department of Commerce Boulder Laboratories, first with the National Bureau of Standards, and now with the NOAA's Wave Propagation Laboratory. He is the author or coauthor of more than 60 papers on radio and acoustic-gravity-wave propagation, ionospheric dynamics, meteorology, atmospheric and ocean acoustics, and radio oceanography. He has also published a book on business and technical writing.

**J. P. Riley**, photograph and biography not available at time of publication.