Remote sensing of ocean wave spectra by interferometric synthetic aperture radar

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The two-dimensional power spectra of ocean waves are of great interest not only to oceanographers but also in practical applications such as wave forecasting, trans-oceanic ship routing, and design of coast and offshore installations. Remote sensing of ocean surface waves can be difficult using conventional synthetic aperture radar (SAR) techniques, but waves can be observed clearly by SAR in the interferometric configuration (INSAR)1,2. This improvement is due to the ability of INSAR to provide images of the local surface velocity field, in contrast to conventional SAR for which the imaging process is limited indirectly to the complex modulation of the surface reflectivity by longer waves and currents. Here we show that INSAR can be used to obtain wavenumber spectra that are in agreement with power spectra measured in situ. This new method thus has considerable potential to provide instantaneous spatial information about the structure of ocean wave fields.

The theory of microwave backscattering from a slightly rough surface was developed by Rice3. The dominant reflection for a side-looking radar is attributed to the resonance between the incident electromagnetic wave and the undulating surface. The resonant, or Bragg, condition is fulfilled when the crests of the surface waves are perpendicular to the radar viewing direction and the horizontal projection of the radar wavenumber is twice that of the resonant surface waves. Using a stationary Doppler radar, Crambolie observed this frequency shift from a moving ocean surface. For microwave radar, the resonant Bragg waves are relatively short (24.5 cm for L-band radar at 30° incidence angle). These waves are affected by the longer ocean surface waves and by currents, making the detection of the large-scale varying features possible.

SAR is a high-resolution coherent remote sensor carried either by an aircraft or by an orbiting platform. The along-track (or azimuth) resolution is achieved by using the phase history of the Doppler-shifted return signal over a finite integration time in which the moving radar forms a large aperture in the azimuth direction. This Doppler shift is due to the line-of-sight component of relative velocity between the target and the moving radar. The cross-track (or ground-range) resolution is obtained by modulating the transmitted signal5.

SAR imaging of a stationary area, as well as rigid moving targets, is well understood1. The imaging mechanism of a continuously varying ocean surface seems to be much more complicated and has been studied extensively for the past two decades4,6. Hasselmann et al.7 have clarified some of the fundamental principles governing the backscattering and SAR imaging of the ocean. The indirect nature of the SAR imaging mechanism of the ocean surface is, in our view, the main difficulty in correlating the radar return signal with the actual ocean reflectivity surface and is therefore an obstacle to obtaining quantitative information about the dominant ocean wave systems.

A modification of the conventional SAR sensor was suggested by Goldstein and Zebker1 and Goldstein et al.7. The technique consists of aligning two physically separated antennas located along the platform flight path. The backscattered echoes received by each antenna from the ocean surface are recorded and processed into two separate complex (magnitude and phase) maps. These maps are combined interferometrically into a single image. The SAR image of the ocean is a map of power reflectivity only, whereas each pixel in the INSAR image represents the line-of-sight velocity component of the surface. Such velocity is due to advancing currents, orbital velocity of long waves, and the phase velocity of the resonant waves. The INSAR image is thus directly related to the local scene by being proportional to the surface velocity. The ability of INSAR to sense ocean currents was demonstrated in refs 1 and 2. Here we show the potential of INSAR in imaging the complex ocean wave fields.

On 8 September 1989 remote imagery of the ocean surface in the nearshore region of Monterey Bay was carried out using an INSAR configuration installed on a DC-8 aircraft. The target area was ~12 × 6 km, centred on a wave pressure-resistant offshore at a depth of 16 m. The flight direction was northbound, roughly parallel to the coastline. The dominant wave propagation direction towards the shore was thus close to the radar line-of-sight. The environmental conditions were favourable, with light westerly winds of <4 knots and significant wave height <0.6 m.

SAR and INSAR images are shown in Fig. 1. The size of a pixel is 6.0 m in the range (x) direction and 12.1 m in the azimuth (y) direction in both images. The flight direction was right to left along the top of the image. The clearly visible coastline in the SAR image (Fig. 1a) separates the image into two different domains; the upper part clearly shows the countryside. The rest of the image represents moving ocean surface. In addition to the faint dominant waves, there are substantial local variations in the image intensity. These variations are probably

<table>
<thead>
<tr>
<th>Area</th>
<th>Mean depth (m)</th>
<th>L1 (m)</th>
<th>L2 (m)</th>
<th>a1</th>
<th>a2</th>
<th>Estimates from in situ frequency spectra</th>
</tr>
</thead>
<tbody>
<tr>
<td>a</td>
<td>40</td>
<td>283</td>
<td>128</td>
<td>255°</td>
<td>292°</td>
<td>l1 (m) (T = 15.9 s)</td>
</tr>
<tr>
<td>b</td>
<td>19</td>
<td>107</td>
<td>107</td>
<td>255°</td>
<td>285°</td>
<td>281</td>
</tr>
</tbody>
</table>

**TABLE 1** Characteristics of the dominant waves
this basin is in agreement with the phase velocity of the upwind resonant waves (~50 cm s⁻¹ for typical inclination angles), calculated from linear theory.

A bimodal wave system, which is rarely observed in the SAR image (Fig. 1a), is clearly seen in the INSAR image (Fig. 1b). The long-crested swell experiences refraction and shoaling as it approaches the shore. The shorter swell propagating to southeast is less sensitive to depth variations; some refraction is visible in regions of larger bathymetric gradients.

The well-identified wave patterns in the INSAR image allow us to obtain meaningful local two-dimensional wavenumber spectra of small subregions of the dominant wave propagating over a sloping bottom. Two domains along the coastline with nearly uniform depths were selected (Fig. 1b). The size of the deeper water domain (mean depth $H_d = 40$ m) is $3000 \times 1400$ m.

The shallow second domain, with $H_s = 19$ m, is $2200 \times 1200$ m. The two-dimensional wavenumber spectra for both domains are presented in Fig. 2. The mean line-of-sight velocity component associated with surface currents and Bragg-wave phase velocity has been removed. Two peaks can be identified in both spectra. The less energetic peak at lower wavenumber is well pronounced, whereas the energy density at higher wavenumber is more spread out. The 180° direction ambiguity, typical to all instantaneous spatial images, appears in the INSAR spectra as well. Here this ambiguity is resolved because waves naturally propagate towards the shore. The scanning distortion due to the relative velocity between the wave pattern and the radar platform is negligible for our experimental conditions.

Shoaling and refraction of waves can be characterized by the two-dimensional $k = (k_x, k_y)$ spectra. Wavelengths $L$ and propagation directions $\alpha$ corresponding to the peaks of Fig. 2 are summarized in Table 1. Both waves experience shoaling, becoming shorter with decreasing water depth. The variation of the wavelength with depth should satisfy the linear shoaling relation

$$L_0 \tanh (k_x H_s) = L_s \tanh (k_x H_s)$$

The results of Table 1 are in good agreement with equation (1). Refraction is more pronounced for the longer wave, where the relative water depth $kH$ is shallower. The effect of wavenumber alignment towards the direction of the bathymetry gradient $\alpha \approx 270^\circ$ is noticed for both wave systems.

The results obtained by the remote-sensing technique are verified by the in situ measurements. The surface-elevation power spectrum obtained by the wave sensor is shown in Fig. 3. The two peaks in the spectrum correspond to wave periods of $T_1 = 15.9$ s and $T_2 = 9.1$ s. The wavelengths corresponding to these periods (Table 1) are calculated at the appropriate depth using the linear dispersion relation

$$\omega = 2\pi / T = (k \tanh (kH))^{1/2}$$

where $\omega$ is the frequency and $g$ is the acceleration due to gravity.

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**Fig. 1.** Image of Marina Beach region in Monterey Bay. a. Conventional SAR image. b. Interferometric SAR image representing the velocity field.

**Fig. 2.** Two-dimensional power spectra of the velocity field calculated for subregions shown in Fig. 1b. a. Mean depth of 19 m. b. Mean depth of 40 m.
Evidence for thinning of the
Arctic ice cover north of Greenland

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In May 1987 a British submarine carried out an ice profiling
experiment in the Arctic Ocean in which the route closely approxi-
mated that of an earlier voyage in October 1976. Over a zone
extending more than 400 km to the north of Greenland there is
evidence of a significant decrease in mean ice thickness in 1987
relative to that found in 1976. This thinning amounts to a loss of
volume of at least 15% over an area of 300,000 km². The nature
of the ice thickness distributions suggests that the thinning is
primarily associated with the presence of a larger fraction of young
and first-year ice, a consequence of wind-driven divergence in the
ice cover during the months preceding the 1987 observations. Such
a large fluctuation in a region previously believed to possess
consistently high ice thicknesses illustrates the importance of
monitoring Arctic ice thickness more systematically to determine
whether such fluctuations fall within the limits of normal inter-
annual variability.

The 1987 expedition involved a submarine and two remote-
sensing aircraft. Results from various sensors are reported in
refs 1-3. The submarine carried a 48-kHz upward-looking echo
sounder of narrow beamwidth (<5°) and obtained a profile of
total length ~6,000 km from the Greenland Sea and the Arctic
Ocean. Data were recorded digitally and on a paper chart, which
was used for interpolation whenever the digitizer lost lock. Data
points were obtained at constant time intervals (~0.25 s) and so,
to obtain a uniformly spaced record, the profile had to be
corrected for varying submarine speed. Depth variations were
corrected by constructing a smoothed profile using data from
open water (open leads and cracks) encountered along the track
and subtracting it from the ice profile. The resulting draft values
were interpolated to a standard interval of 1 m and statistically
analysed in 50-km sections. Statistical results obtained in earlier
trips in both the 1976 and 1987 voyages and their voyages are
reported in the ref. 1. Figure 1a shows contours of mean ice
draft in 1976 constructed from values calculated over 100-km
track sections, corrected for beamwidth as described in ref. 5.
Figure 1b shows contours of mean ice draft from the 1987 data,
constructed from values calculated over 50-km track sections,
uncorrected for beamwidth. These results are conservative
because a beamwidth correction to the 1987 data would produce
a further reduction in apparent mean draft of some 1-2%. This
small effect should be borne in mind with respect to the figures
quoted below.

The significant differences between the two maps lie in the
region extending northwards from Greenland to the North Pole
(0-90°W at 82-90°N); in the eastern European Arctic (east of
0°) there were no significant differences between the mean drafts.
The 1976 map shows a steady build-up of mean ice thickness as
the coast of Greenland is approached. This effect was predic-
ted in a numerical model on the basis of the overall long-term
pattern of wind-driven surface circulation in the Arctic Ocean.
This consists of two main systems, the Beaufort Gyre in the
Canada Basin and the Trans Polar Drift Stream in the Eurasian
Basin. In both currents ice is driven from the general region of the
North Pole towards the coasts of Greenland and Ellesmere
Island. At a latitude of ~87°N the ice begins to feel the influence
of the coast, transmitted as internal stress through the ice from
the downstream land boundary. This causes pressure ridges to

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