APPLICATION OF HF COASTAL OCEAN RADAR TO TSUNAMI OBSERVATIONS

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ABSTRACT: When tsunami waves propagate across open ocean they are steered by Coriolis force and refraction due to gentle gradients in the bathymetry on scales longer than the wavelength. When the wave encounters steep gradients at the edges of continental shelves and at the coast, the wave becomes non-linear and conservation of momentum produces squirts of surface current at the head of submerged canyons and in coastal bays. HF coastal ocean radar is well-conditioned to observe the current bursts at the edge of the continental shelf and give a warning of 40 minutes to 2 hours when the shelf is 50-200km wide. The period of tsunami waves is invariant over changes in bathymetry and is in the range 2-30 minutes. Wavelengths for tsunamis (in 500-3000 m depth) are in the range 8.5 to over 200 km and on a shelf where the depth is about 50 m (as in the Great Barrier Reef) the wavelengths are in the range 2.5 - 30 km. It is shown that the phased array HF ocean surface radar being deployed in the Great Barrier Reef (GBR) and operating in a routine way for mapping surface currents, can resolve surface current squirts from tsunamis in the wave period range 20-30 minutes and in the wavelength range greater than about 6 km. There is a trade-off between resolution of surface current speed and time resolution. If the radar is actively managed with automatic intervention during a tsunami alert period (triggered from the global seismic network) then it is estimated that the time resolution of the GBR radar may be reduced to about 2 minutes, which corresponds to a capability to detect tsunamis at the shelf edge in the period range 5-30 minutes. It is estimated that the lower limit of squirt velocity detection at the shelf edge would correspond to a tsunami with water elevation of less than 5 cm in the open ocean. This means that the GBR HF radar is well-conditioned for use as a monitor of small and medium scale tsunamis, and has the potential to contribute to the understanding of tsunami genesis research.

KEY WORDS: HF Radar, Tsunamis, Surface Currents

1. INTRODUCTION

1.1 Tsunamis

Tsunamis are long wavelength, long period ocean gravity waves which are normally produced by earthquakes or underwater slumps. The wave period of tsunamis depends upon the scale size of the bathymetric event and water depth. Historical records of impacts on the shore indicate that the tsunami wave period (the so-called tsunami window) is in the range 2-30 minutes (Bryant, 2001). The connection between earthquakes and tsunamis is not well understood; some marine earthquakes do not produce tsunamis, and some tsunamis occur with quite small seismic activity. Once the pulse of elevated water is produced it can travel large distances with little attenuation and some dispersion. The elevation of the tsunami waves in the open ocean is generally less than about 0.5 m. This is amplified to tens of metres at the shore and can produce run-up of over 100 m (Bryant, 2001).

The Aceh tsunami on 26 December, 2004 produced wave height at the beach of the order of 10 m and the open ocean elevation was 0.5 metres as measured by the JASON 1 altimeter (Smith et al., 2005) in the Indian Ocean some 2 h after the earthquake in the Sumatra - Andaman Islands.

Barrick (1979) showed that HF radars have the potential to observe tsunamis up to 80 km offshore. In this paper we build on that concept to evaluate the feasibility of operational long-range HF radars to detect tsunami effects at the edge of continental shelves. Here we have the advantage of amplification of surface currents as the tsunami encounters the shelf edge, but the challenge is in the use long-range HF radar systems where time resolution is often sacrificed.

1.2. HF Radar Characteristics

HF ocean surface radar obtains high energy echoes from the Bragg waves on the ocean surface which have wavelengths equal to half the radar wavelength and are propagating radially towards and away from the radar site. The radial resolution scale for radars is set by the bandwidth of the transmitted signal. For HF radars this is typically 50 KHz and is controlled by international agreements on the use of the radio spectrum. Some systems are working with bandwidths up to 150 KHz. The radar being installed on the Great Barrier Reef in Australia has a bandwidth of 50 KHz, which corresponds to a radial range resolution of 3 km. The azimuthal resolution is controlled by the width of the lobe of a beam-forming radar, and by the accuracy of amplitude or phase measurements for direction-finding radars.

A typical radar spectrum is shown in Fig. 1. The offset of the Bragg peaks from the reference (which is the transmitter frequency) is due to Doppler shift, caused predominantly by the phase speed of the Bragg waves (which offset is indicated by the dashed lines in Fig.1) but

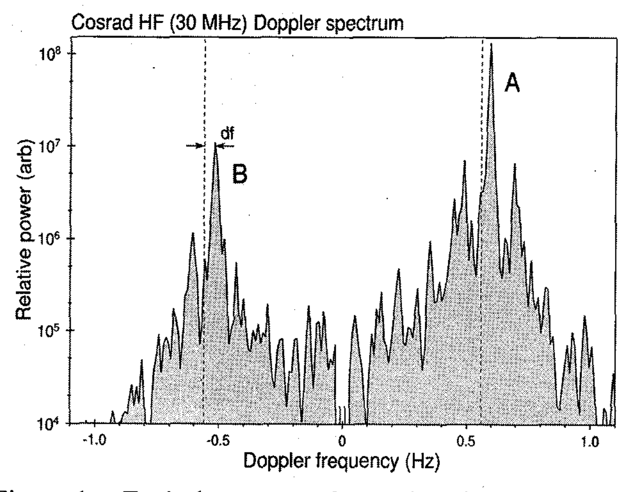


Figure 1. Typical spectrum for a phased array ocean surface radar. The peaks labelled A and B are the first-order Bragg scatter. The sideband peaks on each side of the Bragg lines are produced by wind waves. When swell is present it appears as sidebands on the Bragg lines and separated from them by a frequency equal to the frequency of the swell; in this record the swell lines are barely significant compared with fluctuations in the spectrum.

also caused by radial surface current which is indicated by df in Fig.1. The question arises here about the differentiation between swell (which shows up as sidebands on the spectrum) and tides which are effectively very long waves. We step around tsunamis here because they fall in between the two extremes and we will discuss them later. Tides have return periods much greater than the time it takes to record the time series (which is the basis of the spectrum shown in Fig. 1), and the pixel size. for the radar is much smaller than the length scale of the tidal cycles. Therefore the tidal currents are resolved in time and space by the HF radars. Swell (which has period of order 15 s) is not resolved in time or space and is manifest in the spectra as sidebands on the first-order Bragg lines. In this paper we investigate the time and space scales for tsunamis.

2. TSUNAMI CHARACTERISTICS

2.1 Tsunami Period and Wavelength

Tsunamis are shallow water gravity waves and we can use linear wave theory at least while the waves are in open water with gentle gradients in the bathymetry. The expression for phase speed, c, for gravity waves is

$$c^2 = \frac{g}{k}\tanh(kH) \qquad (1)$$

(Kinsman, 1965) where k is the wave number and H is the water depth. In shallow water a good approximation is

$$c = \sqrt{gH} \qquad (2)$$

which is illustrated in Fig. 2 by a family of curves with H varying from 500 to 3000 m. The flat sections for wave periods above about 300 seconds confirm the applicability

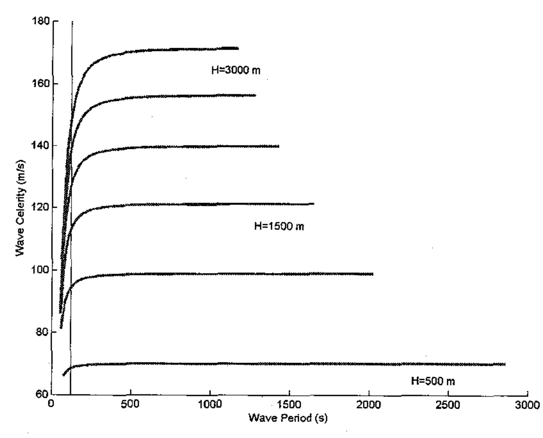


Figure 2. Properties of gravity waves derived from (1) for tsunami conditions. Celerity versus wave period shows the transition from dispersive conditions at short wave periods to no dispersion at long periods. The vertical marker at 120 seconds marks the shortest observed tsunami period (Bryant, 2001).

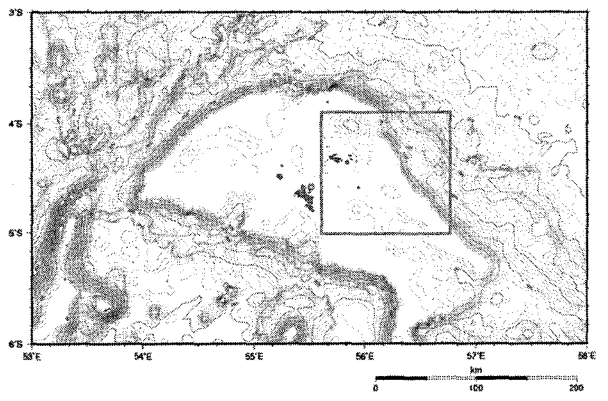
of (2), while for smaller wave periods (1) becomes more appropriate. One important indication from Fig. 2 is that for a pulse of water elevation at the source, the subsequent propagation will be subject to some dispersion for the higher frequency components (wave periods less than about 300 seconds) which means that for distant sources the first ocean waves will have periods longer than about 300 seconds.

2.3. Tsunami propagation

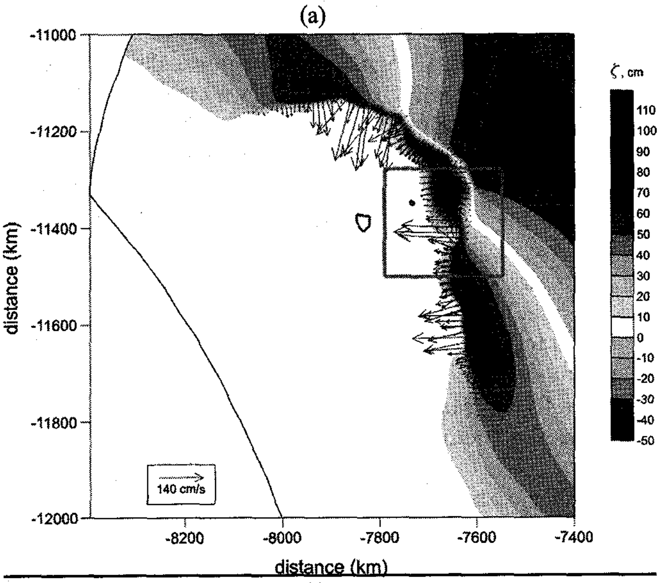
In open water, with gentle gradients in bathymetry, the tsunami direction is controlled by refraction and Coriolis rotation off great circle paths. When tsunamis approach steep gradients in the bathymetry (and at the coast) the scale of the physical features is usually much less than the wavelength of the tsunami and we get non-linear growth in water elevation and surface currents. This is different from swell waves which are still being refracted as they approach the coast. The main difference between swell and tsunamis is that swell focuses onto headlands away from coastal bays, whereas tsunamis have the biggest effect in the bays and at the head of submerged canyons.

Non-linear effects arise from the conservation of momentum in the water column which tends to drive water parallel to the depth contours to produce enhancement of surface elevation at the head of submerged canyons, which in turn produces the greatest effect at the coast in bays and estuaries rather than at headlands. If the bathymetry changes significantly within one wavelength then momentum effects dominate over refraction. Wavelengths for tsunamis in the open ocean (500-3000 m depth) are in the range 8.5 to over 200 km, and because bathymetric gradients are typically gentle, we may assume that bathymetric steering in the open ocean is controlled by refraction. As a tsunami propagates across a basin it will normally be refracted so that it approaches the coast orthogonally to the large-scale bathymetric contours. Near the coast however, where the scale of the spatial features is less than the wavelength, refraction does not work because the wave is changing (non-linearly) within the space of a wavelength. Here the water elevation (and run-up) is controlled by momentum transfer as the water becomes shallower.

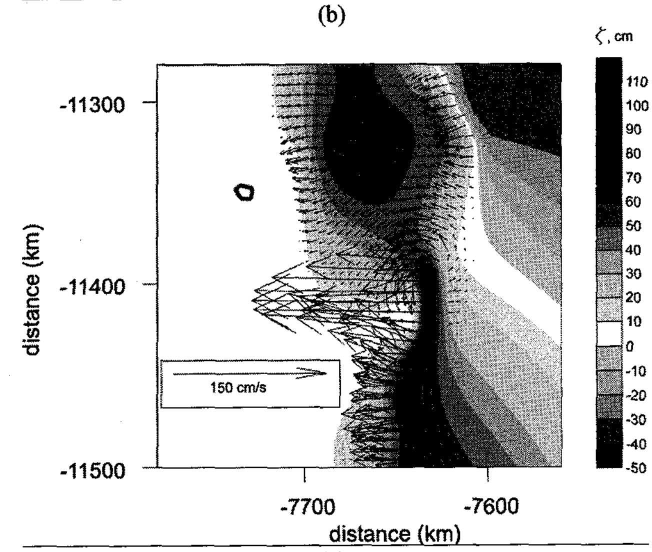
The momentum transfer is a result of non-linear processes and is better modelled by numerical calculations than by an analytic wave propagation approach. An example of such a model is shown in Fig 3b where the surface currents at the edge of the continental shelf near the Seychelles are shown for the first wave of the 26 Dec 2004 tsunami which originated near Aceh in Indonesia. The relevant bathymetry is shown in Fig 3a. It can be seen that the higher current surges are produced at the heads of submerged canyons while the submerged ridges experience smaller surges.



(a)



(b)



(c)

Fig. 3. Model calculations for the impact of the 26 Dec 2004 Tsunami on the Seychelles. (a) The bathymetry around the Seychelles shows a platform in otherwise deep water; (b) the large scale amplification of surface current when the first tsunami wave encounters the edge of the continental shelf; and (c) a section of the shelf-edge which could be monitored by an HF ocean radar.

3. TIME AND SPACE SCALES

3.1 Space scales for radar observation

The spatial resolution of HF radars in the radial dimension depends on the bandwidth of the radar in the radio spectrum. The range is determined by a time delay and the phase velocity of the electromagnetic wave as

$$r = c\tau/2 \qquad (3)$$

where c = 3x10^8 ms^{-1}, and τ is the out and back time delay between the transmitter and the receiver. For resolution we have

$$\Delta r = (c/2)\Delta\tau \qquad (4)$$

where $\Delta\tau$ is the resolution in time. This requires a bandwidth of Δf where

$$\Delta f = c/(2\Delta r). \qquad (5)$$

The GBR radar has a bandwidth of 50 KHz. This means that the shortest tsunami wavelength that can be observed is 6 km.

3.2 Time scales for radar observation

The time resolution for HF radars depends on the observation time required to make one surface current determination. For the GBR radar, the routine operation is to sample surface currents every 10 minutes, because each of the two stations operates for 5 minutes and then waits for 5 minutes. Some systems operate with surface currents produced routinely every 3 hours. To reduce the sampling time we have to trade off accuracy. For the GBR radar, which is a phased array system, the trade-off is between time resolution and accuracy of the surface current values. Fig. 5 shows the relationship where at a radar operating frequency of 8 MHz and a time-series length of 5 minutes the surface current resolution is about 7 cm/s.

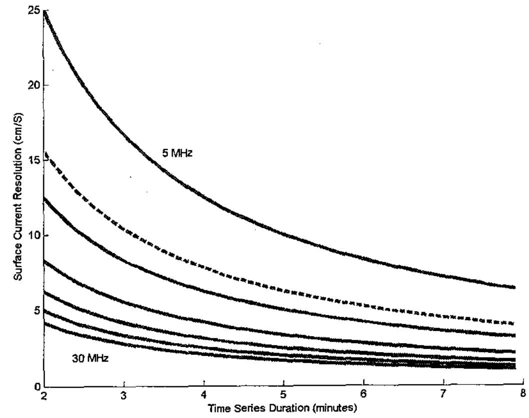


Fig. 5. The resolution for surface current is degraded as the radar transmitter frequency decreases. The parameter for the family of curves is for radar frequencies 5, 8 (dashed), 10, 15, 20, 25 and 30 MHz. To achieve resolution of 15 cm/s at an operating frequency of 5 MHz one would need to integrate for a period of not less than 200 seconds.

4. CONCLUSION: AN OPTIMISED TSUNAMI OBSERVING INSTRUMENT

The parameters discussed above can be re-arranged to form an optimised tsunami observing instrument. Firstly, one of the two radar stations will always be better positioned to observe surface current bursts orthogonal to the bathymetry contours along the edge of the shelf. During a tsunami alert (potentially triggered by a seismic alert) we would turn off the radar station that was less effective and run the other one repeatedly for time series lengths of, say, 2 minutes. According to Fig. 5 this would provide surface current resolution of about 12 cm/sec. This is a worst-case estimate because there are sophisticated analyses that can be used to combine spatial data to improve the detection resolution.

For the 26 Dec 2004 tsunami, the magnitudes of the surface currents at a shelf edge at the Seychelles were up to 150 cm/sec, and this corresponded to an open water elevation of about 50 cm. Assuming that we can scale linearly, the present indication is that the GBR phased array radar can detect tsunamis of less than one-tenth the magnitude of the 26 Dec 2004 tsunami with wave periods greater than about 4 minutes; under active management. Under routine operation (which means a huge archive of data) the GBR radar will detect tsunamis less than one-tenth the magnitude of the 26 Dec 2004 event but for wave periods longer than 20 minutes. This means that we would have access to many tsunamis which are not destructive at the coast, but which provides a useful research archive for studying tsunami occurrence and genesis. HF radar also promises to be a useful model verification tool.

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