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The level of deterrence provided by data from the SPITS seismometer array to possible violations of the Comprehensive Test Ban in the Novaya Zemlya region

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SUMMARY

The yield threshold at which a fully decoupled explosion can be identified has been a recurring issue in the debate on whether the Comprehensive Nuclear Test Ban (CTB) can be adequately verified. Here, we assess this yield threshold for the Novaya Zemlya (NZ) and Kola Peninsula regions by analysing seismograms from six small body wave magnitude ($m_b \leq 3.5$) seismic disturbances recorded at regional distances ($1050 < \Delta < 1300$ km) by the seismometer array at Spitsbergen (SPITS). Multiple filter analysis of the seismograms shows clear high-frequency P_n ($f \geq 14$ Hz), except from a calibration explosion on the Kola Peninsula. From four of the disturbances studied we observe clear high-frequency S_n; the explosion showed no clear high-frequency S_n and the data from the remaining disturbance was potentially contaminated by a data glitch. Frequency-domain analysis indicates that the P_n and S_n attenuation across the Barents Sea is similar to that observed across stable tectonic regions (shields). We define a spectral magnitude for the 2.5–3.5 Hz passband that is tied to teleseismic m_b from NZ explosions; the six disturbances considered have $2.3 \leq m_b \leq 3.5$. Three-component data are available from SPITS for four of the disturbances considered (including the explosion). From the explosion the S/P ratios on the vertical (Z), radial (R) and tangential (T) components (in the 3.0–6.0 Hz passband) are all less than unity. The S/P ratios for the same passband on the Z component from the remaining three disturbances are less than unity, but the ratios on the R and T components are significantly greater than unity. We argue that S/P ratios (3.0–6.0 Hz passband) of less than unity on all of the Z, R and T components at SPITS may indicate a potential treaty violation in the Kola Peninsula and NZ regions. The temporal variation of seismic noise, in the 3.0–6.0 Hz passband, at SPITS suggests that our three-component S/P criterion will be effective 95 per cent of the time for disturbances with $m_b \geq 2.8$. We suggest that $m_b = 4.25 + b \log_{10} W$, where W is the explosive yield in kilotons (kt), with $b = 0.75$ for $W \geq 1$, and $b = 1.0$ for $W < 1$, is suitable for conservatively estimating the yield threshold of a potential violation of the CTB in the NZ region. From this we infer that a 35 ton fully coupled explosion in the NZ region is likely to be identified as suspicious under the CTB using the three-component S/P criterion. Simulations show that the low-frequency decoupling factor (DF) for a fully decoupled nuclear explosion in hard rock is about 40, suggesting that such an explosion with a yield of 1.6 kt in the NZ region is likely to be identified using data from SPITS. The conservatism likely to be employed by a potential violator and uncertainties in the DFs for nuclear explosions in hard rock cavities, together with data from stations other than SPITS within 2000 km of the NZ region, suggest that the yield at which a potential violator of the CTB could confidently escape detection (using decoupling) in the NZ region is in reality probably less than 0.5 kt.

Key words: attenuation, Comprehensive Test Ban, decoupling, Novaya Zemlya region, seismology, yield estimation.

1 INTRODUCTION

It is generally accepted that modern global seismograph networks are capable of detecting seismic waves from a 1 kiloton (kt) fully coupled explosion (van der Vink & Park 1994). Recently, Douglas *et al.* (1999) showed that seismic data from the present International Monitoring System (IMS) can be used to detect and identify as a suspicious disturbance, and hence a possible violation of the Comprehensive Nuclear Test Ban Treaty (CTBT), a small fully coupled 0.1 kt chemical explosion (equivalent to 0.2 kt nuclear) in East Kazakhstan. Decoupling, by muffling the seismic waves generated by an explosion by detonation within a gas-filled cavity, will increase the yield at which a potential violator of the CTBT could escape detection. Here, we analyse data recorded by the IMS auxiliary seismic station at Spitsbergen (SPITS) from four seismic disturbances in the Novaya Zemlya (NZ) region and two disturbances on the Kola Peninsula (Fig. 1 and Table 1). Our aim is to assess

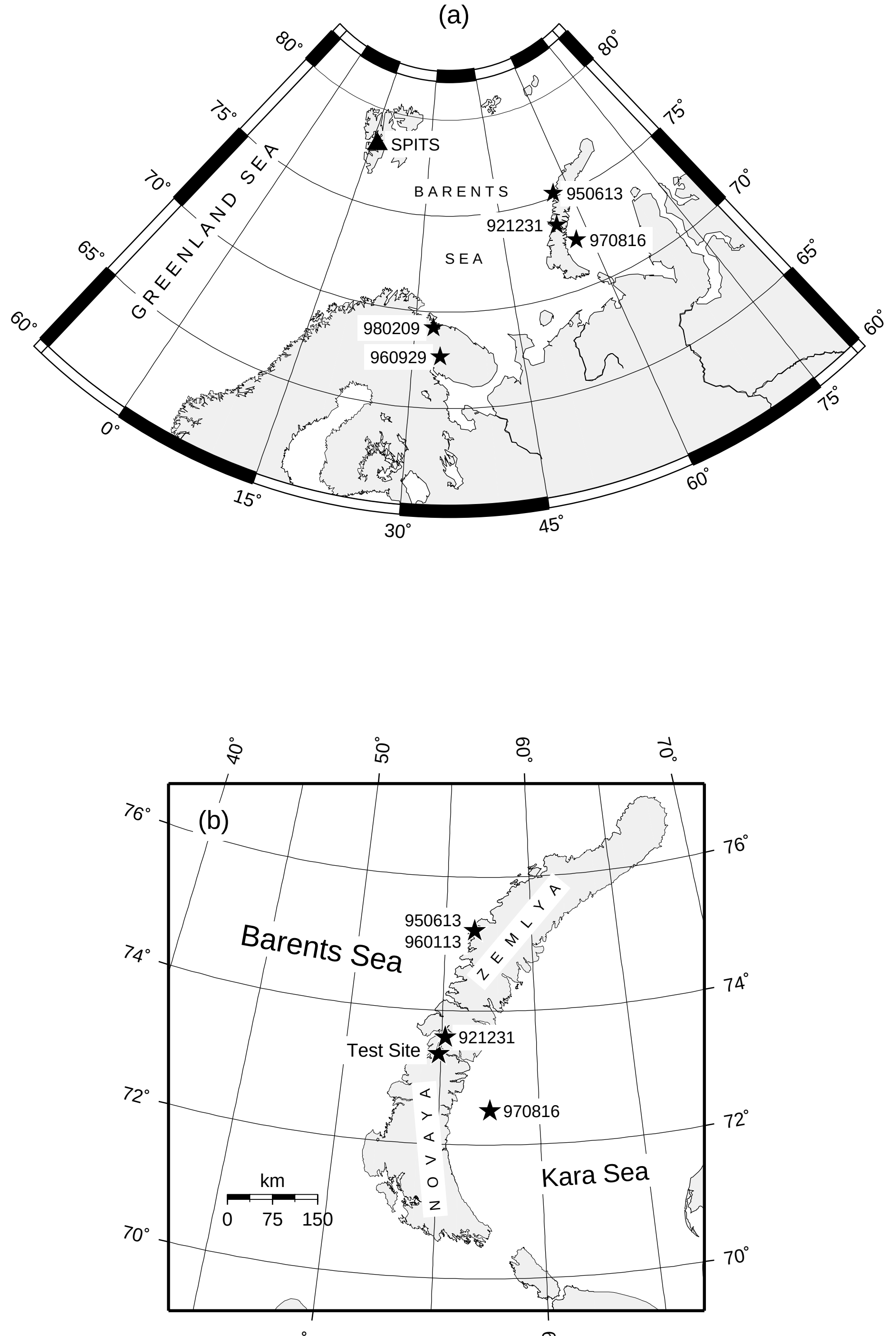

Figure 1. Location maps showing (a) the seismometer array SPITS and the epicentres of selected seismic disturbances considered in this paper and (b) the epicentres of seismic disturbances considered in this paper in the vicinity of the northern nuclear test site on Novaya Zemlya.

Table 1. Source parameters for the seismic disturbances considered in this study.

Event Code	Date	Origin Time	Latitude	Longitude	Magnitude	To SPITS		Reference
						Δ	Azimuth	
921231	31 December 1992	09:29:24.0	73.614° N	55.179° E	2.7 m_b	1151 km	314°	*a*
950613	13 June 1995	19:22:37.9	75.262° N	56.877° E	3.5 m_b	1070 km	307°	*a*
960113	13 January 1996	17:17:23.0	75.132° N	56.727° E	2.4 m_b	1076 km	307°	*a*
960929	29 September 1996	06:05:46.2	67.675° N	33.728° E	2.9 M_L	1298 km	342°	*b*
970816	16 August 1997	02:11:00.0	72.441° N	57.595° E	3.5 m_b	1300 km	318°	*a*
980209	9 February 1998	16:51:07.0	69.18° N	32.63° E	–	1124 km	341°	*c*

[a] NORSAR group "best" locations and magnitudes (Frode Ringdal 1998), [b] Ringdal *et al.* (1996), [c] Kremenetskaya & Asming (1998).

the level of deterrence provided by SPITS to possible CTBT violations (fully coupled and fully decoupled) in the NZ and Kola Peninsula regions.

The SPITS station has (1) an array of nine vertical-component seismometers, (2) a three-component sensor and (3) an acquisition system with a Nyquist frequency of 20 Hz. SPITS has not recorded signals from any known nuclear explosion at NZ as a Russian moratorium on nuclear testing has been in place since the station became operational in 1992. Hence, the level of deterrence provided by SPITS for the NZ region can only be assessed by analysing presumed non-nuclear seismic disturbances.

Of the seismic disturbances in Table 1, the 1996 September 29 disturbance in the Khibiny massif is a calibration explosion (total yield 350 t of chemical explosive) carried out by the Kola Regional Seismology Centre and the Ministry of Defence of the Russian Federation (Ringdal *et al.* 1996). The disturbance of 1997 August 16, in the vicinity of the northernmost nuclear test site on NZ in the Russian Arctic has been identified as an earthquake (Richards & Kim 1997; Hartse 1998; Baumgardt 1998; Bowers *et al.* 1998) using high-frequency *S*/*P* ratios recorded at regional distances (<2000 km). The other seismic disturbances in Table 1 have not been positively identified (Ringdal 1997; Kremenetskaya & Asming 1998), although Ryall *et al.* (1996) argued that the disturbance of 1992 December 31 is most probably an earthquake.

2 THE SPITS SEISMOGRAMS

2.1 Time-domain analysis

Fig. 2 shows the results of passing the seismograms, recorded by a single vertical-component channel of the array, from each of the six seismic disturbances, through a bank of two-pass filters. The filters are two-pole Butterworth filters. Where possible we have used the central element of the array SPA0, but where this was not working another channel has been selected. The multiple-filter analysis has been shown to be a useful tool for examining the frequency–time characteristics of regional phases (e.g. Bennett *et al.* 1994).

In general the filtered waveforms in Fig. 2 show a reasonably constant signal-to-noise ratio (SNR) with frequency, suggesting that the spectral amplitude of the seismic noise at high frequencies is falling off at a similar rate to that of the P_n signal. The exceptions are the filtered waveforms from the 1996 September 29 explosion (Fig. 2d) which show only a weak P_n signal in the 8.0–16.0 Hz passband, and no evidence of a P_n signal for frequencies greater than 14.0 Hz.

Fig. 2 also shows clear S_n above the noise (P_n coda) at least up to frequencies of 14 Hz for all but the 1996 September 29 explosion. However, the interpretation of S_n from the 1992 December 31 disturbance is complicated by a glitch (removed by linear interpolation prior to filtering) in the data at about the S_n arrival time. Also, the interpretation of S_n on the filtered traces from the 1996 January 13 disturbance (Fig. 2c) is complicated by an impulsive arrival at about 142 s visible in the very high-frequency passbands—Ringdal (1997) uses frequency-wavenumber analysis to demonstrate that this arrival is from another seismic disturbance. There is no evidence of L_g in any of the seismograms in Fig. 2, consistent with the hypothesis that L_g propagation is blocked by thick sediments under the Barents Sea (Baumgardt 1990).

2.2 Frequency-domain analysis

For each disturbance in Table 1 we calculate P_n and noise power spectra (e.g. Chael 1987) for the instrument-corrected data from each channel of the array where a P_n signal is clear (often the data from some channels are dominated by system noise and/or glitches and cannot be used). The power spectrum is the energy spectrum of a time window normalized by the window duration. We then calculate array-averaged spectra by taking the square root of the mean of the power spectra from each channel. Since the signal coherence generally decreases with increasing frequency, if conventional beamforming is applied the array effectively acts as a low-pass filter (e.g. Bache *et al.* 1985). Hence, we prefer stacking power spectra to the alternative of calculating the spectra from the beamformed seismogram. Fig. 3 shows the resulting array-averaged P_n and noise spectra.

We calculate array-averaged S_n and noise spectra for the 1995 June 13 and 1997 August 16 disturbances in a similar way to P_n, using a noise window containing P_n coda preceding the S_n onset. Fig. 2(e) shows the P_n, S_n and noise windows used to calculate the power spectrum for the SPA0 channel from the 1997 August 16 disturbance. The resulting array-averaged S_n and noise spectra are shown in Fig. 4. Fig. 5 shows that the *S*/*P* spectral ratios are reasonably stable across the array.

Excepting the 1996 September 29 explosion, the large SNR of P_n and S_n above 3 Hz in Figs 3 and 4 suggests first that SPITS is a low-noise site at high frequencies, and second that there appears to be little attenuation of P_n and S_n across the Barents Sea to SPITS. Fig. 6 shows that the noise spectra in Fig. 3 are typical for SPITS; while SPITS is a noisy station at low frequencies (above the high-noise model for $f<0.25$ Hz), at higher frequencies ($f>2.5$ Hz) the noise at SPITS is typically well below

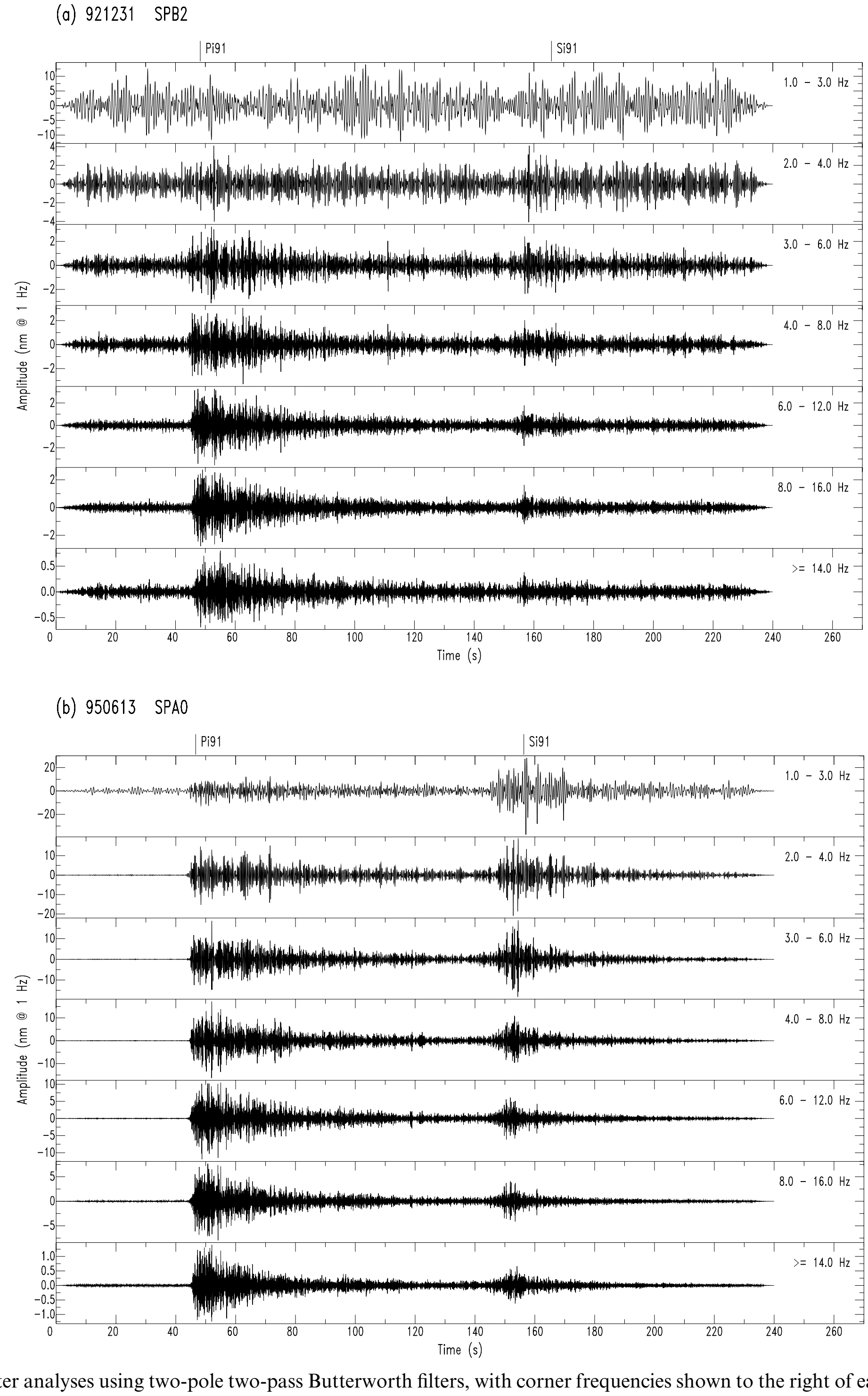

Figure 2. Multiple filter analyses using two-pole two-pass Butterworth filters, with corner frequencies shown to the right of each filtered seismogram. (a) The 1992 December 31 disturbance near the Novaya Zemlya test site, (b) the 1995 June 13, (c) 1996 January 13 disturbances about 250 km north of the Novaya Zemlya test site, (d) the 1996 September 29 Khibiny calibration explosion, (e) the 1997 August 16 Kara Sea disturbance and (f) the 1998 February 9 disturbance near Murmansk. Pi91 and Si91 mark the P_n and S_n arrival times respectively, predicted using the IASPEI 1991 model and the location parameters given in Table 1. (e) also shows representative P_n, S_n and noise time windows used to calculate the array-averaged normalized displacement spectra.

the low-noise model. We attempt to quantify the observed attenuation of P_n and S_n by comparing our calculated spectra with several published attenuation models derived assuming that the source spectrum can be adequately represented by the *w*-square source model (Chael 1987).

2.3 Regional attenuation models

Sereno *et al.* (1988) inverted P_n spectra recorded in Scandinavia assuming frequency-independent geometrical spreading and a frequency-dependent Q representing the 'apparent attenuation'

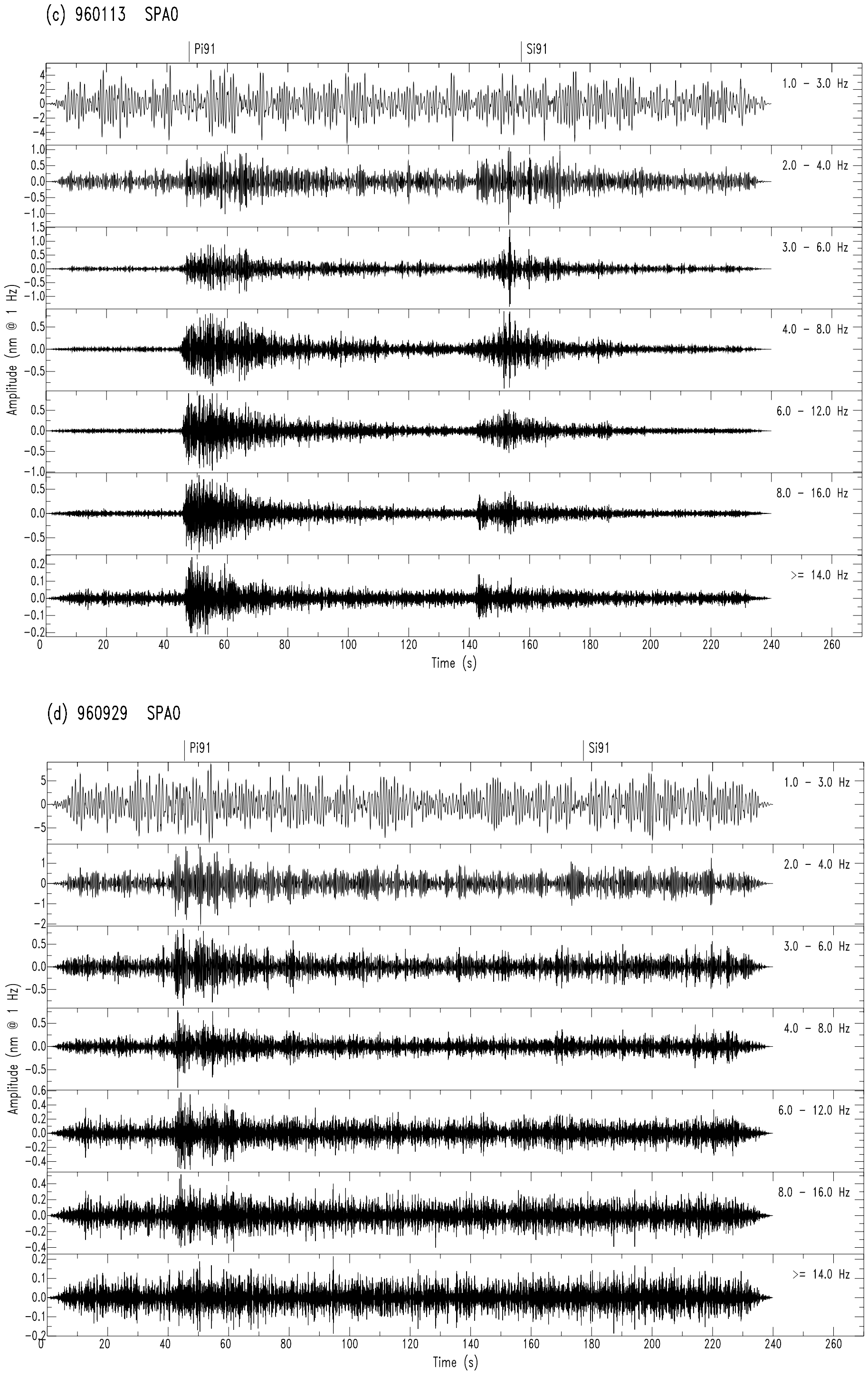

Figure 2. (*Continued.*)

(including the effects of anelasticity and scattering). The apparent attenuation operator determined by Sereno *et al.* (1988) is $Q_{P_n}(f) = 325f^{0.48}$.

Chun *et al.* (1989) noted that the 'total attenuation' experienced by P_n depends not only on the effects of anelasticity and scattering, but also on the effect of frequency-dependent geometrical spreading. Thus, the total attenuation should include a distance-dependent term (Δ). Chun *et al.* (1989) found that $\Delta^{-(2.17+0.022f)}$ provided a reasonable fit to the total attenuation of P_n observed across the Canadian shield.

Jenkins *et al.* (1998) determined an average model for the total attenuation of P_n for stable tectonic regions using 561 P_n waveforms recorded by stations reporting to the prototype International Data Centre (PIDC). Jenkins *et al.* (1998) described the frequency-dependent total attenuation of their observations by fitting a function of the form,

$$f^{-1}\exp(-\alpha f)\left(\frac{\Delta}{200}\right)^{-(af+b)}, \tag{1}$$

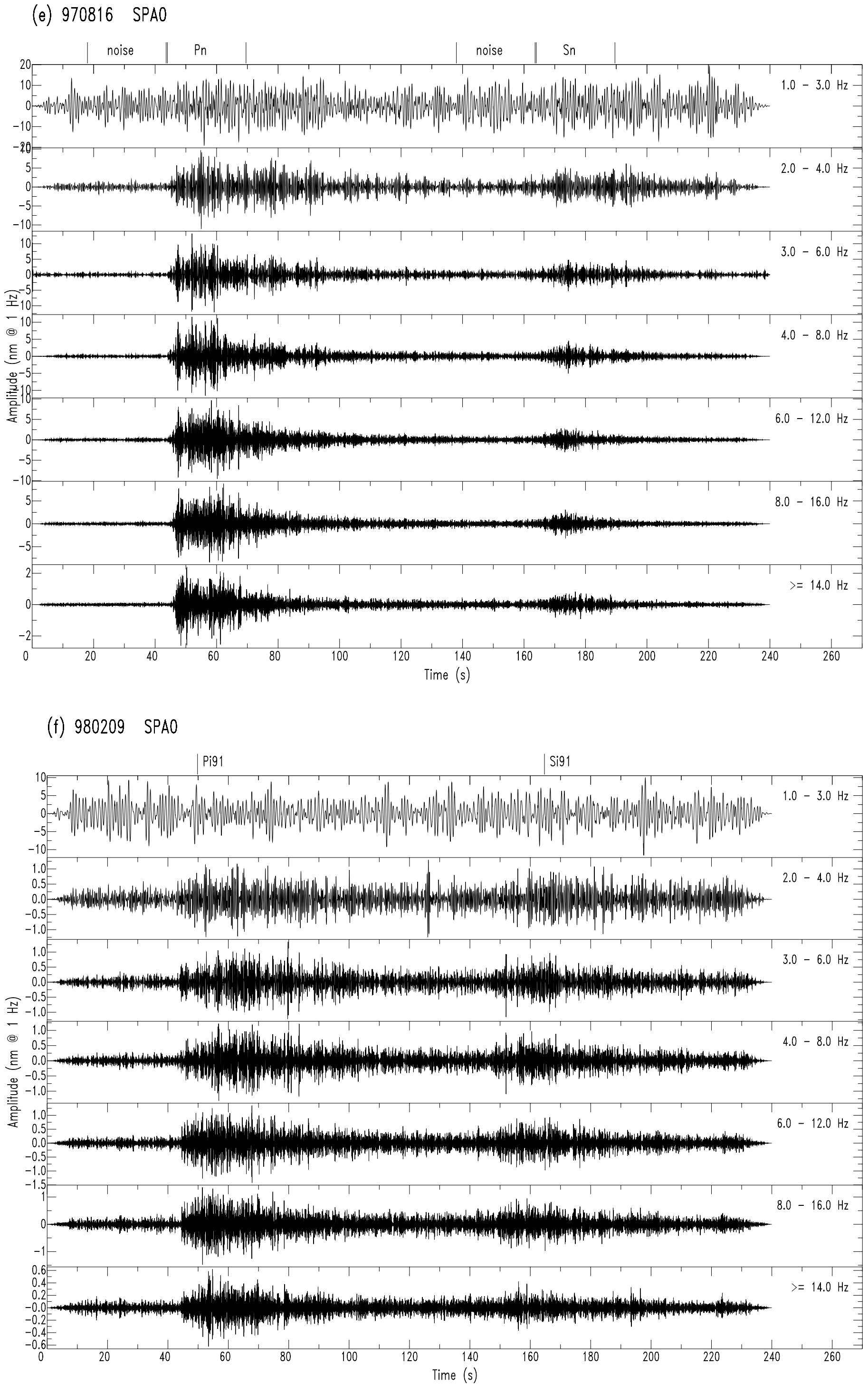

Figure 2. (*Continued.*)

where Δ is the distance in kilometres, and $\alpha = -0.25$, $a = 0.08$ and $b = 1.44$ for their stable-region P_n model ($1.0 \le f \le 15.0$ Hz). The term $f^{-1} \exp(-\alpha f)$ describes the frequency-dependent apparent attenuation to the reference distance of 200 km and the distance-dependent term is parametrized as a power law in a similar way to Chun *et al.* (1989).

α in eq. (1) is equivalent to the more familiar πt^*, where $t^* = T/Q$, with T representing the traveltime of P_n and Q the apparent attenuation. The f^{-1} term is a consequence of assuming that P_n propagates as a head wave to the reference distance. $\alpha = -0.25$ for the stable-region P_n model suggests that Q is negative (≈ -400). We suspect that in general P_n in stable tectonic regions does not propagate as a head wave to the reference distance, but as a turning wave. Indeed, at the mid-frequency of 8 Hz, $\exp(-\alpha f) = f^{0.96}$, very nearly cancelling the f^{-1} term. So, while interpretation of the model in eq. (1)

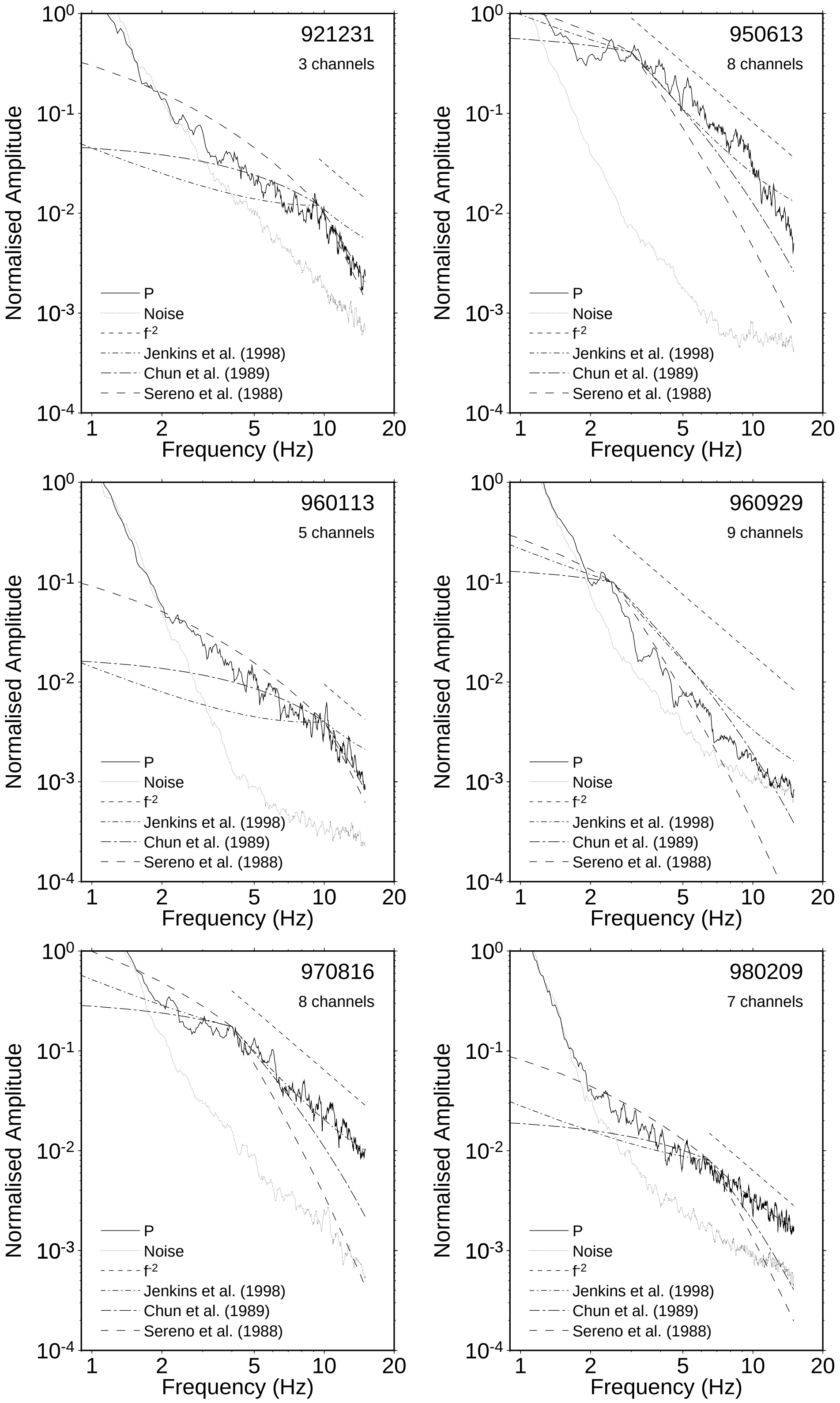

Figure 3. P_n array-averaged spectra from stacks of vertical-component channels at SPITS from the six seismic disturbances in Table 1.

may potentially be problematic, it does provide an empirical fit to the frequency-dependent variation of P_n amplitude with distance determined from a large number of observations in stable tectonic regions.

In order to compare the above attenuation models with our array-averaged P_n spectra in Fig. 3 we require a source model. We follow Chael (1987) and assume a *w*-square source model, and that source directivity effects are negligible. We determine

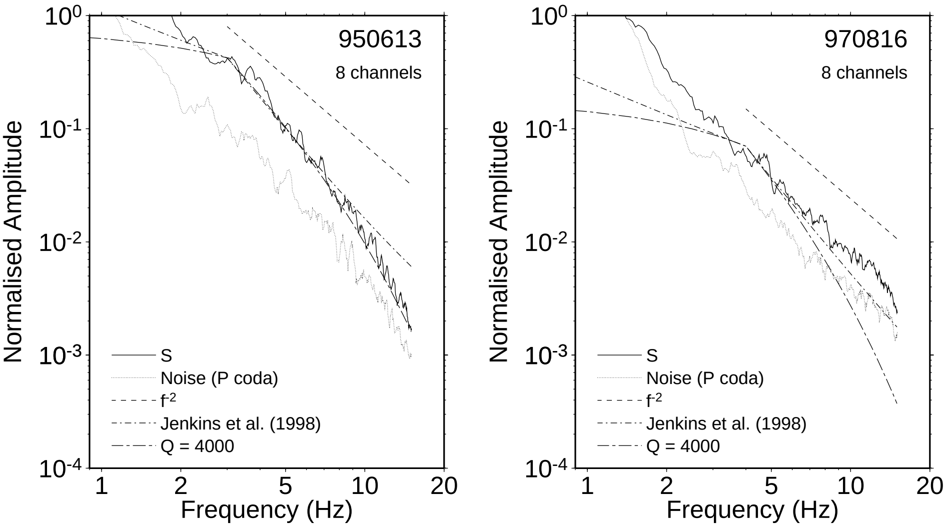

Figure 4. S_n array-averaged spectra from stacks of vertical-component channels at SPITS from the 1995 June 13 and 1997 August 16 seismic disturbances.

the corner frequency, f_c, by examining the P_n spectra in Fig. 3 for a rapid change in slope. Table 2 gives our estimated corner frequencies.

$f_c=2.5$ Hz for the 1996 September 29 Khibiny explosion is consistent with the reported total duration of the ripple-fired explosion of 0.4 s (Ringdal *et al.* 1996). However, since the 1996 September 29 explosion was ripple-fired the *w*-square source model may not be appropriate. The P_n spectrum from the 1996 September 29 explosion appears to fall-off at a higher rate than the spectra from the other disturbances, which is consistent with a source spectrum for the 1996 September 29 explosion with a fall off greater than the *w*-square model. However, the high fall-off of the 1996 September 29 P_n spectrum can also be explained by attenuation within the Kola Peninsula between the epicentre and the Barents Sea, since the P_n spectrum for the 1998 February 9 disturbance (Fig. 3) suggests that P_n propagates efficiently across the Barents Sea to SPITS.

$f_c=4.0$ Hz for the 1997 August 16 disturbance in the Kara Sea is consistent with the P_g spectrum at AMD (Amderma; $\Delta=360$ km, azimuth$=152°$) calculated by Ringdal *et al.* (1997a) (our interpretation of the AMD P_g spectrum assumes that the instrument response for the seismometer at AMD varies smoothly through our inferred corner frequency). $f_c=6.5$ Hz for the 1998 February 9 disturbance is supported by our analysis of the P_g spectrum calculated for the waveforms recorded by the ARCES array in northern Norway ($\Delta=285$ km, azimuth$=281°$).

Fig. 3 shows the combined effects of the *w*-square source model and the three P_n attenuation models considered above.

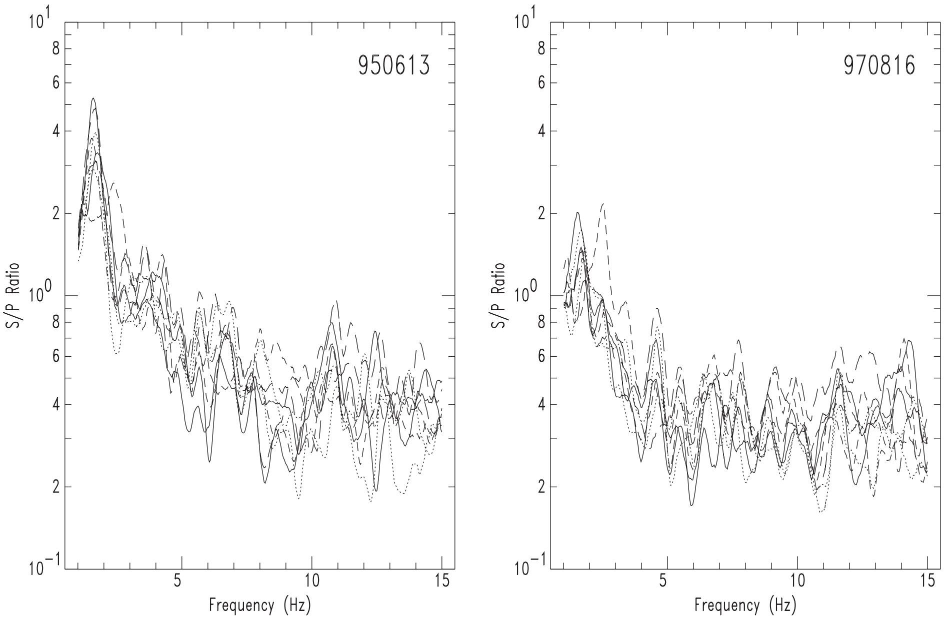

Figure 5. Variation of the vertical-component S/P ratio with frequency for each operational element of the SPITS array for the 1995 June 13 and 1997 August 16 disturbances.

Table 2. m_b estimates relative to the 1995 June 13 disturbance determined using array-averaged P_n spectra at SPITS in the 2.5–3.5 Hz passband.

Event Code	f_c (Hz)	Δ (km)	$B(\Delta)$	$\log \bar{A}$	m_b (this study)	m_b (previous studies)
921231	9.5	1151	0.053	−1.252	2.72	2.7 (Ringdal 1997)
950613	3.0	1070	0.000	−0.422	3.50*	3.5 (Ringdal 1997)
960113	10.0	1076	0.004	−1.574	2.35	2.4 (Ringdal 1997)
960929	2.5	1298	0.140	−1.461	2.60	–
970816	4.0	1300	0.142	−0.767	3.30	3.29 (Bowers *et al.* 1998)
980209	6.5	1124	0.035	−1.687	2.27	–

* fixed parameter, see text for definition of terms.

The amplitudes of the model P_n spectra are scaled to fit the observed spectra around the corner frequency. For the 1995 June 13, 1997 August 16 and 1998 February 9 disturbances the models of Sereno *et al.* (1988) and Chun *et al.* (1989) significantly overestimate the amount of attenuation. However, the Jenkins *et al.* (1998) model appears to fit the observed P_n spectra from these three disturbances remarkably well. For the 1992 December 31 and 1996 January 13 disturbances the Chun *et al.* (1989) model appears to be a reasonable fit.

We model the S_n spectra in Fig. 4 following a similar procedure to that adopted for the P_n spectra above. We assume that the corner frequencies for the S_n spectra are the same as for the P_n spectra (Table 2). Jenkins *et al.* (1998) derived an average total attenuation model for S_n propagating in stable tectonic regions using 241 S_n waveforms ($0.5 \leq f \leq 10.0$ Hz). The Jenkins *et al.* (1998) S_n model has $\alpha = -0.25$, $a = 0.12$ and $b = 1.85$ (see eq. 1). We also calculate the model spectra for a frequency-independent apparent $Q = 4000$ (Evernden *et al.* 1986).

Fig. 4 shows that the Jenkins *et al.* (1998) S_n model appears to fit the observed S_n spectra reasonably well up to 10 Hz (the highest frequency used to determine the model). Above 10 Hz the 1995 June 13 S_n spectrum seems to fall off at a rate consistent with an apparent $Q = 4000$, but the 1997 August 16 S_n spectrum falls off slower than even this high apparent Q value.

The similarity of vertical-component *S*/*P* ratios at SPITS from mid-Atlantic earthquakes and those from the 1997 August 16 disturbance led Baumgardt (1998) to argue that S_n at SPITS from earthquakes and explosions from the NZ region would be strongly attenuated. However, the spectra in Fig. 4 suggest weak attenuation of S_n for such paths, with apparent $Q \geq 4000$. Furthermore, the reasonable fit to the Jenkins *et al.* (1998) P_n and S_n models suggests that the average P_n and S_n attenuation in stable tectonic regions is remarkably low (at least up to 10 Hz), suggesting that observations such as those in Fig. 2 may not be confined to the paths across the Barents Sea to SPITS from the NZ and Kola Peninsula regions.

3 ESTIMATION OF m_b

In Section 5 we wish to infer the yield of an underground nuclear explosion at NZ using a magnitude yield relation, based on the teleseismic body wave magnitude m_b. It is well known that simply averaging station estimates of m_b results in a mean m_b that is biased upwards due to the effect of data censoring by noise thresholds at each station in the network. This bias increases with decreasing magnitude. Thus, m_b estimates routinely reported by agencies such as the International Seismological Centre for small disturbances are biased upwards. Maximum likelihood methods can be used to determine unbiased estimates of the true m_b (Ringdal 1976; Christoffersson 1980). However, since no signals from the seismic disturbances in Table 1 were recorded at teleseismic distances an alternative approach is required.

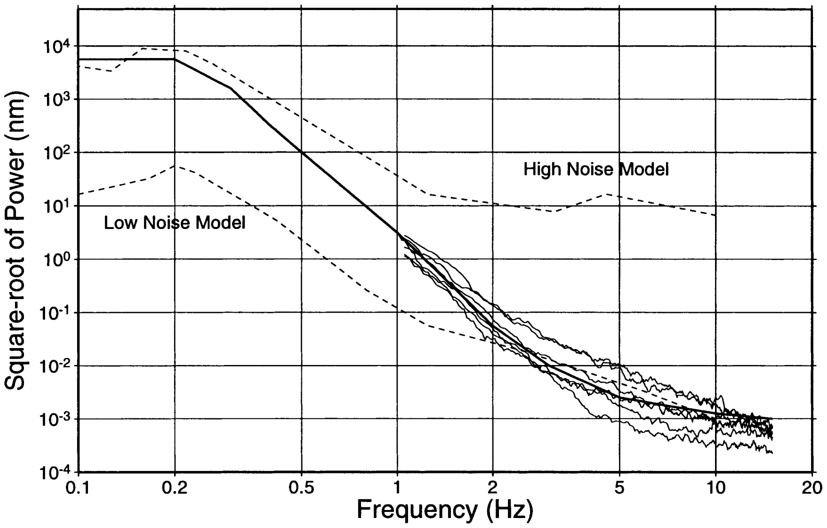

Figure 6. A comparison of the noise spectra shown in Fig. 3 (thin lines) with the median of 62 spectra recorded at SPA0 (thick line) reported by GSETT3 (1995), and the new high- and low-noise models of Peterson (1993).

Traditional teleseismic m_b estimates are from P signals with a dominant frequency of around 1 Hz. However, regional P_n signals generally have a higher dominant frequency than teleseismic P. Ringdal (1997) estimated the m_b for the 1992 December 31, 1995 June 13 and 1996 January 13 disturbances (Table 2) by comparing the amplitude of P_n recorded at ARCES (filtered with a passband of 2.0–4.0 Hz) with that of a 5.6 m_b NZ explosion on 24 October 1990 (which is of sufficient magnitude not to suffer from significant m_b bias). Bowers *et al.* (1998) followed a similar approach to Ringdal (1997) and estimate an m_b of 3.3 for the 1997 August 16 disturbance by comparing P_n amplitudes recorded at NORSAR (southwest Norway) with those from the teleseismically recorded Kara Sea earthquake of 1986 August 1 with a maximum likelihood estimate of 4.3 m_b (Marshall *et al.* 1989).

We require a measure of the relative size of the seismic signals recorded at SPITS from the seismic disturbances in Table 1 that is consistent with teleseismic m_b from NZ explosions. Since we have calculated the array-averaged P_n spectra (Fig. 3), we develop a magnitude based on the 2.5–3.5 Hz passband of these spectra, which utilizes the Jenkins *et al.* (1998) P_n attenuation model for stable tectonic regions to correct observations at different distances.

We take the 2.5–3.5 Hz passband of the array-averaged P_n spectra at SPITS and evenly resample the logarithm of the frequency; we then take the mean of the logarithm of the amplitude, $\log_{10} \bar{A}$ (Rodgers *et al.* 1997). We use the 1995 June 13 disturbance as a reference and calculate relative distance terms, $B(\Delta)$, using the stable-region P_n attenuation model of Jenkins *et al.* (1998) for a frequency of 2.96 Hz (central frequency of the 2.5–3.5 Hz passband transformed to $\log_{10} f$). [Our $B(\Delta)$ term is a regional equivalent to the distance term used in traditional teleseismic m_b calculations.] Table 2 shows our m_b estimates relative to the 3.5 m_b for the 1995 June 13 disturbance determined by Ringdal (1997). Encouragingly our estimates of m_b are consistent with those determined independently for the 1992 December 31, 1996 January 13 and 1997 August 16 disturbances (Table 2), suggesting that our m_b estimation procedure is reasonable.

4 THREE-COMPONENT SPITS SEISMOGRAMS

Fig. 7 shows three-component seismograms (filtered with a passband of 3.0–6.0 Hz) recorded at SPB4 (within the SPITS array), from the 1995 June 13, 1996 September 29, 1997 August 16 and 1998 February 9 disturbances. Unfortunately, the three-component sensor was not operational at the time of the 1992 December 31 and 1996 January 13 disturbances. The horizontal components have been rotated to the conventional radial (R) and transverse (T) directions assuming propagation along the great-circle path.

The three-component seismograms from the 1995 June 13 and 1997 August 16 disturbances (Figs 7a and c) show little P energy on the T component compared with that on the vertical (Z) and R components, suggesting that cross-component

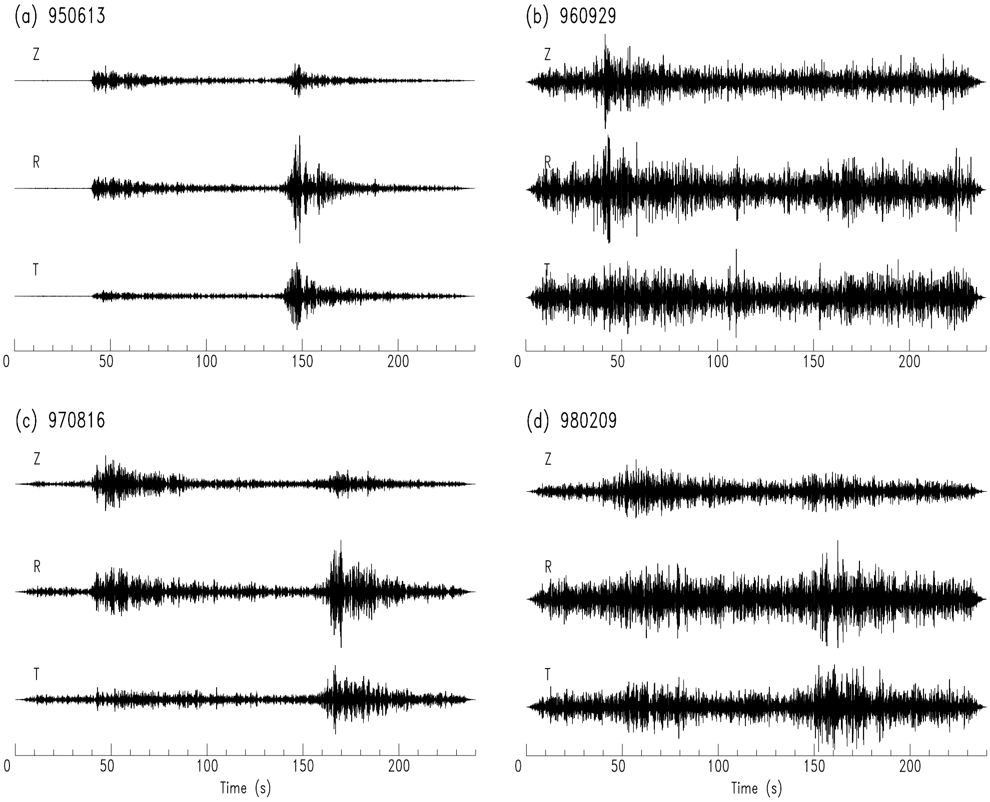

Figure 7. Three-component seismograms recorded at SPB4 of the SPITS array, filtered using a two-pass two-pole Butterworth filter with a passband of 3.0-6.0 Hz. (a) The 1995 June 13 NZ disturbance, (b) the 1996 September 29 Khibiny calibration explosion, (c) the 1997 August 16 Kara Sea disturbance, and (d) the 1998 February 9 disturbance near Murmansk.

scattering is weak. These same seismograms also show substantial S-energy on both the R and T components at SPB4 (Figs 7a and c).

Since no signals from known explosions in the NZ region have been recorded by SPITS, we cannot be certain whether the three-component S/P ratios from NZ explosions would resemble those from the 1995 June 13 and 1997 August 16 disturbances (Figs 7a and c). However, the observation that three-component S/P ratios differ between explosions and earthquakes was well known by Jeffreys (1959), who, commenting on a study by Willmore (1949) of seismic data recorded at distances of up to about 1000 km from explosions in north Germany, noted that the 'S waves were unsatisfactory, but this is a usual feature of explosions; it looks as if SV in near earthquakes is always less satisfactorily recorded than SH'.

Recent studies (e.g. Kim *et al.* 1997) show that using S/P measured from three components improves the separation of the explosion and earthquake populations relative to S/P measured solely from the Z-component. Such discrimination studies using S/P from SPITS would require at least one calibration explosion in the NZ region. However, an analysis of synthetic seismograms by Blandford (1993), generated for the Fennoscandian region, shows large S/P ratios on the R and T components from earthquake sources when S/P on the Z-component is small (explosion-like).

While the high-frequency S/P ratio for the 1995 June 13 and 1997 August 16 disturbances measured from the Z-component at SPITS (Fig. 5) appears explosion-like ($<$unity) (Baumgardt 1998), S/P on the R and T components is earthquake-like ($\gg$unity) (e.g. Kim *et al.* 1997). The observation of a small S/P ratio on the Z-component compared with large S/P on the R and T components is consistent with a small S_n emergent angle due to low-wave-speed sediments below the SPITS array (Kværna *et al.* 1999).

We argue because of (1) efficient propagation of P and S, (2) weak cross-component scattering and (3) observational and theoretical evidence that the high-frequency three-component S/P can potentially discriminate between explosions and earthquakes that observations of $S/P<$unity on the Z, R and T components at SPITS from seismic disturbances in the NZ region would be sufficient to indicate a possible CTBT violation and trigger an On-Site Inspection (OSI). Thus, high-frequency three-component S/P at SPITS is a deterrent to any potential evader of the treaty. Furthermore, since S from decoupled explosions is likely to be weaker than from fully coupled explosions (Blandford 1996), the deterrence of three-component S/P at SPITS is strengthened if the potential evader uses decoupling.

We can test our deterrent against possible CTBT violations by applying our high-frequency three-component S/P criterion for SPITS to the two seismic disturbances in the Kola Peninsula (Fig. 1). Figs 7(b) and (d) show the Z, R and T components for the 1996 September 29 Khibiny calibration explosion and those from the 1998 February 9 disturbance. There is no clear S energy from the 1996 September 29 explosion and $S/P<$unity on all components, indicating a possible explosion under our criterion, whereas $S/P>$unity on the R and T components from the 1998 February 9 disturbance, is consistent with an earthquake. Thus, our high-frequency three-component S/P criterion for SPITS appears to discriminate between one explosion and one presumed earthquake in the Kola Peninsula. However, since the 1996 September 29 explosion is about 175 km further away from SPITS than the 1998 February 9 disturbance the non-observation of S energy from the explosion may also be explained by S attenuation within the Kola Peninsula between the epicentre and the Barents Sea.

5 YIELD THRESHOLD

The observation of $S/P>$unity on the R and T components at SPITS from the 1998 February 9 disturbance suggests that our three-component S/P criterion is effective for disturbances as small as $2.3m_b$. However, the high-frequency noise (3.0–6.0 Hz passband) at SPITS preceding the P_n signal from the 1998 February 9 disturbance was below the median noise level (Fig. 6). 95 per cent of the time the noise at SPITS is below a level about 10 dB above that preceding the 1998 February 9 P_n signal (GSETT3 1995), suggesting that our three-component S/P criterion will be effective 95 per cent of the time for disturbances with $m_b \geq 2.8$.

5.1 Fully-coupled explosions

We now pose the hypothetical question what is the expected yield of a fully coupled explosion in the NZ region with $2.8m_b$? For this we need an appropriate m_b yield relation for the NZ region. Murphy (1996) suggested that the appropriate m_b yield relation for fully coupled nuclear explosions at the East Kazakhstan (EK) test site is

$$m_b = 4.45 + 0.75 \log_{10} W\,, \tag{2}$$

where W is the explosive yield in kilotons. The observations that support this relation are roughly over the range $4.5 \leq m_b \leq 6.5$. We make the common assumption that eq. (2) and subsequent modifications are valid down to $2.3m_b$ (e.g. Murphy 1996).

The P_n wave speed beneath EK is 8.3 km s^{-1} (Marshall *et al.* 1979), whereas for the NZ and Barents Sea regions the P_n wave speed is about 8.05 km s^{-1} (Ringdal *et al.* 1997b). Marshall *et al.* (1979) showed empirically that P_n wave speed is an indicator of attenuation of teleseismic P within the upper mantle; using the Marshall *et al.* (1979) relation we obtain a difference of about 0.20 units in m_b between EK and NZ. We adjust eq. (2) accordingly to obtain

$$m_b = 4.25 + 0.75 \log_{10} W\,. \tag{3}$$

Using eq. (2), a fully coupled nuclear explosion with a yield of about 6 t would cause a $2.8m_b$ seismic disturbance. If we use eq. (3) then a $2.8m_b$ is equivalent to an explosion with a yield of about 12 t. The observations of low-P_n attenuation in Fig. 3 may suggest that the attenuation in the upper mantle beneath the Barents Sea is more like that beneath EK than that inferred from the Marshall *et al.* (1979) correction. However, teleseismic m_b from explosions is typically measured at lower frequencies (1.0–2.0 Hz) than our P_n attenuation estimates, and the paths through the upper mantle of P_n and teleseismic P are different. Given these uncertainties in the appropriate m_b yield relation for NZ, we prefer eq. (3) as this is more conservative than eq. (2) in the context of estimating the yield at which a potential treaty violation in the NZ region could be identified.

The existing seismic station at SPITS can thus identify a suspected CTBT violation with exceptionally small yield providing the explosion is fully coupled. The detection threshold of the network of stations in the NZ region, required to locate any potential CTBT violation, is about $2.4m_b$ (e.g. Ringdal 1997); lower than our threshold of $2.8m_b$ for our three-component *S*/*P* criterion. Under the CTBT a detection could invoke an OSI.

5.2 Decoupled explosions

We now consider the effects of decoupling (e.g. Sykes 1996) on the ability of SPITS to identify a possible treaty violation. The effectiveness of decoupling is usually measured by the decoupling factor (DF), which is a frequency-dependent term that is the ratio of the displacement spectra from a fully coupled and a decoupled explosion of the same yield, in the same material, fired at the same depth. The highest DF is achieved at low frequencies when the cavity has sufficient volume for the walls to respond approximately linearly to the pressure step applied by the explosion (full decoupling). The minimum cavity volume required to achieve decoupling is controlled by the properties of the material surrounding the cavity. The low-frequency DF is not expected to increase if a larger cavity is used.

Early decoupling experiments, such as the 0.38 kt Sterling nuclear explosion in a salt cavity (Springer *et al.* 1968), suggest that a DF of about 70 can be achieved at frequencies around 1 Hz. Thus, modifying eq. (3) assuming a DF of 70, a 3.4 kt fully decoupled nuclear explosion at NZ would be expected to have an m_b of about 2.8.

The above argument uses DFs derived from the Sterling experiment, which was conducted in salt. However, there are no known salt deposits around NZ (Sykes 1996), so any attempt at decoupling is likely to be conducted using a cavity formed in hard rock. The solid geology of Novaya Zemlya reveals many limestone formations in the vicinity of the north test site (Marshall *et al.* 1994). Recently DFs as low as 10 (at 10 Hz) were reported from small (1 t) fully decoupled chemical explosions in limestone (Murphy *et al.* 1997). Murphy *et al.* (1997) also estimated the DF for a 1 t nuclear explosion in limestone, using simulation, to be about 40 at 10 Hz. Using a DF of 40 suggests that a 1.6 kt fully decoupled nuclear explosion in hard rock in the NZ region would produce a $2.8m_b$ seismic disturbance.

6 DISCUSSION

The slope of 0.75 in eq. (3) is a consequence of detonating underground nuclear explosions at a scaled depth of burial so that the radioactivity is contained (e.g. Sykes 1996); this depth (in metres) is usually assumed to be 122 $W^{-1/3}$, where *W* is in kilotons (e.g. Murphy 1996). Under the CTBT an evader would be likely to detonate the explosion deep enough to avoid any surface expression, this scaled depth is greater than that required for containment. In the context of a CTB it may be more conservative to assume a minimum depth for fully coupled explosions, for say $W<1$ kt. If we make this assumption then the slope of eq. (3) becomes unity for $W<1$, and our corresponding estimate of the yield from a $2.8m_b$ seismic disturbance is 35 t.

Since many geological materials are weak in tension, Latter *et al.* (1961) argued that to contain the radioactivity from a cavity-decoupled nuclear explosion the overburden pressure must exceed, by some factor of safety ($k=2$), the long-term pressure step at the cavity wall. The requirement for containment is (e.g. Sykes 1996)

$$k\rho gh \geq (\gamma-1)\,4.2\times10^{12}\;W/V\,, \tag{4}$$

where ρ is the average density (kg m^{-3}) of the material above the cavity, g is the acceleration due to gravity (9.81 m s^{-2}), h is the depth (m) to the top of the cavity, γ is the ratio of enthalpy to internal energy of the gas within the cavity (taken to be 1.2 for air at atmospheric pressure) and V is the cavity volume (m^3).

Murphy *et al.* (1997) determined the minimum scaled cavity radius (in metres) for full low-frequency decoupling (in air-filled cavities in limestone) to be 27 $W^{-1/3}$. Using this and assuming $k=1$ (Patterson 1966), $\rho=2400$ kg m^{-3} for limestone and a spherical cavity, then eq. (4) suggests that to ensure containment of a fully decoupled explosion one would require a cavity at least 425 m deep.

Since the minimum containment depth is inversely proportional to the cube of the scaled cavity radius, increasing the scaled cavity radius allows a shallower minimum containment depth. The main control on the minimum containment depth for a given yield is thus the volume of the cavity. Sykes (1996) argued that there are few free-standing hard rock cavities exceeding 280 000 m^3 in volume and that such cavities are difficult to construct. A volume of 280 000 m^3 corresponds to a minimum cavity depth of about 125 m for a 1 kt fully decoupled explosion, similar to that of a 1 kt contained fully coupled explosion (122 m). Thus, eq. (3), adjusted for the DF, appears to provide conservative estimates of the m_b from fully decoupled explosions in hard rock with $W\geq1$ kt. If the difficulty in the construction of underground hard-rock cavities results in cavities with a volume controlled by the minimum scaled cavity radius for full decoupling, then the minimum containment depth is yield-independent and eq. (3) adjusted for the DF with a slope of unity would be conservative for $W<1$ kt.

Above we have attempted to assess the maximum yield at which a potential violator of the treaty is likely to be caught at the approximate 95 per cent confidence level (assuming that a potential evader would not consider the noise conditions or operational status of SPITS when choosing to test). The deterrence of SPITS data may be better assessed by answering the question what is the yield at which a potential violator of the treaty would be confident of evading detection?. At the median noise level (equivalent to $2.55m_b$), assuming a DF of 40 and eq. (2) (with a slope of unity), our three-component *S*/*P* criterion could potentially identify a fully decoupled explosion with a yield of ≥0.5 kt.

Furthermore, since no country is known to have tested a decoupled nuclear explosion in hard rock (Sykes 1996), the DF for nuclear explosions in hard rock is necessarily determined by simulation (e.g. Murphy *et al.* 1997). The results of these simulations depend strongly on the equation of state of the material surrounding the cavity (Murphy *et al.* 1997). Since there is no known experimental data to support the results of the simulations of DFs for nuclear explosions in hard rock, there will always be some uncertainty in the DF (calculated by simulation) that a potential violator could rely on for effective decoupling in hard rock. Thus, the yield at which a potential violator of the CTBT could be confident of escaping detection (by decoupling) is probably less than 0.5 kt.

DFs are expected to decrease markedly at high frequencies (Murphy 1996). Blandford (1985) reports that the DF at 20 Hz for the Sterling experiment was about 7, compared with the DF of about 70 observed at 1 Hz. If the noise spectrum at SPITS (Fig. 6) continues to decrease as f^{-2}, then signals, with $f \geq 15$ Hz from $<2.8m_b$ disturbances around NZ, could potentially be recorded at SPITS if suitable instrumentation were deployed. Simulations suggest that the DF for nuclear explosions in hard rock decreases with increasing frequency, but the details depend strongly on the (generally poorly known) equation of state of the material surrounding the cavity. Thus if the DF for nuclear explosions in hard rock decreases with increasing frequency within the high-frequency limit we can potentially record at SPITS, then the yield at which a potential violator of the CTBT could be confident of escaping detection (by decoupling) is again probably less than 0.5 kt. Seismic data from other IMS stations will be required to locate the disturbance, but we expect that high-frequency data from these stations will further enhance the deterrence provided by SPITS.

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