USING THE F-DETECTOR TO HELP INTERPRET P-SEISMOGRAMS RECORDED BY SEISMOMETER ARRAYS

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ABSTRACT

The F-detector was first suggested by Melton & Bailey (1957) and its effectiveness was demonstrated by Blandford (1974), using data from the short-period seismometer array at the Tonto Forest Observatory. For an array of N seismometers, the F-detector output is the time-averaged power on the beam divided by N-times the time-averaged variance of the individual channels about the beam. F is robust in that it is resistant to spikes and to signals arriving off the desired beam. However, while F is commonly used as a detector for infrasound signals, the short-term average divided by the long-term average (STA/LTA) has been preferred for the detection of seismic signals.

One reason why STA/LTA-type detectors are ubiquitous may be that when seismic-phase detection algorithms were originally developed the computing power was not available to calculate F. Another reason is that when the noise is correlated between array elements F is sub-optimal and the STA/LTA detector was found to be as effective as F.

Whether correlated noise is observed depends on the design of the seismometer array and on the properties of the noise. For small-aperture arrays ($\sim$ 2 km) 1 Hz noise is generally highly correlated between seismometers. However, for medium-aperture arrays ($\sim$ 10-20 km), such as EKA, YKA and WRA, the 1 Hz noise-correlation is generally low (average cross-correlation coefficient < 0.15). Under conditions of low noise-correlation the statistical distribution of F approximates to the non-central F-distribution. Thus, an approximate probability can be attributed to a postulated signal detection, under the hypothesis of a chosen signal-to-noise ratio (SNR). These appealing properties of F suggest that it may be more effective than STA/LTA-type detectors, at least for medium aperture arrays with low noise-correlation.

Here we consider an alternative application of F and demonstrate that this is a useful tool when interpreting array seismograms. We show examples where the F-trace and associated probability are helpful in confirming that a signal is real. We have found this to be particularly useful for 'special event' analysis, for example, detecting P at ASAR from the 100 tonne chemical explosion on August 22, 1998 in East Kazakhstan. We also show that F is an effective tool for helping to identify surface reflections (pP and sP) on teleseismic P-seismograms – phases that are important for determining the depth of a seismic source.

During the process of reviewing associated (and missed) detections from an automatic system, analysts are often faced with the difficult task of deciding whether a signal really exists, especially at low SNR. Clearly F and the associated probability have the potential to help analysts in the routine identification of seismic-phases during the review process.

Key Words: Seismometer Arrays, P-Seismograms, F-Detector, Special Event Analysis.

OBJECTIVE

To assess whether a trace based on the F-detector has the potential to help the analyst identify body wave phases, such as P, and surface reflections (pP and sP). Correct identification of P, pP and sP is of importance if the Comprehensive Nuclear-Test-Ban Treaty (CTBT) is to be verified using P seismograms recorded at long range.

RESEARCH ACCOMPLISHED

Introduction

Under the CTBT an analyst reviewed bulletin of seismic disturbances (REB) will be produced by the IDC (International Data Centre) using data from the seismic component of the International Monitoring System (IMS). The treaty also requires 'event screening' of the REB, that is producing a subset of seismic disturbances in the REB that can be identified as earthquakes with confidence.

One of the most effective methods of identifying an earthquake is to show that the disturbance is deeper than the maximum depth current emplacement technology will allow an explosive device to be buried (say > 10 km). This requires a reliable estimate of the depth, which is often not possible using travel times alone (especially for intra-crustal earthquakes). The time difference between teleseismic P and the associated surface reflections (pP and sP) is one of the most reliable methods for estimating the depth of a seismic source. However, the success of the method relies critically on identifying correctly P and associated surface reflections.

For deep sources (say ≥ 50 km) the moveout of $pP - P$ and/or $sP - P$ times with distance may be used to help confirm the identification of the surface reflections. For example, J.R. Murphy (*written comm.*) suggests at least 1.5 s moveout, between the $pP - P$ times at the nearest and farthest stations, to confirm that pP has been identified correctly, and that the source is really ≥ 50 km deep. However, this requires identification of candidate pP phases and accurate reading of the onset of P and pP – not an easy task as, (1) the onset of pP is often masked by the P coda, (2) pP can be weak, and (3) sP may be strong and confused with pP. Also, the pulse shape of pP and sP is often different from P for deep sources, making identification difficult on seismograms recorded by narrow band seismograph systems. Such problems may also affect the results of Woodgold (1999), who applied a wide-aperture beamforming technique to identify pP on data from the Canadian National Seismograph Network, using time scale contraction based on model $pP - P$ moveout.

One way to confirm that P, pP and sP have been identified correctly on a number of seismograms from the same seismic disturbance is to search for orientations of the double couple (earthquake) source that produce pP/P and sP/P ratios that are similar to those formed by the candidate phases (allowing for uncertainty in the observed amplitudes). Pearce (1977, 1980) formalised this procedure as the relative amplitude method. If after applying the method, orientations of a double couple are consistent with the observed amplitude ratios, then the hypotheses that the seismic disturbance is an earthquake, and that the candidate P and surface reflections have been identified correctly, are supported. The above relies on the fact that fortuitous compatibility of the amplitude ratios with a double-couple source, resulting from false identification of the candidate phases at even a small number of stations, is extremely unlikely (Pearce & Rogers 1995).

When a seismic disturbance is recorded at only a small number of stations (say less than four), then care must be taken to ensure that other seismic phases (such as S-to-P conversions at the Moho for intra-crustal sources) are not mistaken for surface reflections (Bowers *et al.* 2000). Maps showing the Moho depth and Poisson's ratio (which governs the $sP - pP$ time) for the structure in the epicentral region, can be used to increase confidence in the correct identification of the surface reflections (Bowers *et al.* 2000).

There will be many seismic disturbances that are too small to generate surface waves that can be detected by the IMS. For some of these disturbances regional discriminants may well be inconclusive. Then the only data available for 'special event' analysis will be teleseismic P, recorded by a few of the more sensitive seismometer arrays within the IMS. Douglas *et al.* (1999) demonstrated, for the 100 t chemical explosion in Kazakhstan on August 22, 1998, that the absence of surface reflections on a number of P seismograms, well distributed in distance and azimuth, can be used to infer that the seismic source is shallow. In the absence of any other information to indicate the non-nuclear origin of the disturbance, an on-site inspection is required to confirm that the source is not a violation of the CTBT.

The key factor in all of the above methods for identifying earthquakes and possible treaty violations is the ability to confirm the presence, or absence, of P and surface reflections. Here, we consider the F-detector output for an array of seismometers. Arrays are superior to single stations in that they can be 'steered' to enhance energy with a specific vector slowness. The most common array processing is 'delay-and-sum', or beamforming, which not only increases the SNR, but also suppresses signal-generated noise in the vicinity of the receiver (Key 1968). We argue that a trace showing some measure of the coherency of P energy across the array, such as F, displayed with the beamformed seismogram is of help in interpreting the seismogram in terms of candidate P, pP and sP.

Further, the F-trace may be of use to IDC analysts during the process of reviewing associated detections (and missed detections) from an automatic system (based on a tuned STA/LTA). Then analysts are often faced with the difficult task of deciding whether a signal really exists, especially at low SNR.

The F-Detector

The F-detector was first suggested by Melton & Bailey (1957). Representing the outputs, at time t, of an N-element seismometer array as $u_i(t)$, the beam can be written as,

$$\hat{u}(t) = (1/N)\sum_{i=1}^{N} u_i(t). \tag{1}$$

F is approximated by (Blandford 1974),

$$F \approx \frac{N-1}{N}\frac{\sum_{t=1}^{M}\hat{u}(t)^2}{\sum_{t=1}^{M}\sum_{i=1}^{N}(u_i(t)-\hat{u}(t))^2}, \tag{2}$$

where, the summation $t = 1, \ldots, M$ represents averaging over a 'box car' time window, T of M-samples. In words, F is approximately the time averaged power on the beam divided by N times the time averaged variance of the individual channels from the beam.

F can also be written as (e.g. Blandford 1974),

$$F \approx (N-1)\frac{\sum_{t=1}^{M}\hat{u}(t)^2}{\frac{1}{N}\sum_{t=1}^{M}\sum_{i=1}^{N}u_i(t)^2 - \sum_{t=1}^{M}\hat{u}(t)^2}, \tag{3}$$

which is simpler to compute. The expression is exact if the signal and noise are white. As Blandford (1974) points out, when a signal arrives with similar vector-slowness to the beam, F increases because, (1) the beam power in the numerator increases, and (2) the denominator in reduced to the residual noise (in this way F checks that the signal is recorded by several seismometers). If a signal arrives from a significantly different vector-slowness to the beam, F will be reduced as the denominator increases faster than the numerator. Also, F is robust in that it is resistant to data spikes, and so should be superior to a standard STA/LTA detector.

F Signal Hypothesis

The F signal hypothesis (Blandford 1974) assumes that on each channel of the array there is, (I) identical, deterministic signals, and (II) independent, Gaussian, and identically distributed noise with a stationary autocorrelation (the noise need only be stationary over the time window surrounding the signal). If (I) and (II) are satisfied then F has a non-central F-distribution $F(N_1, N_2, \lambda)$. With,

$$\begin{aligned} N_1 &= 2BT \\ N_2 &= N_1(N-1) \\ \lambda &= 2BTR^2, \end{aligned}$$

where, N_1 and N_2 are the degrees of freedom, λ is the non-centrality parameter, B is the bandwidth in Hz, T is the time window (s), and R^2 is the ratio of the signal and noise power on the beam.

Abramowitz & Stegun [p. 948, eqn 26.6.26] (1964) give an approximation for the probability of the non-central F-distribution,

$$P(F'|N_1, N_2, \lambda) \approx P(F|N_1^*, N_2), \tag{4}$$

where,

$$\begin{aligned} F &= \frac{N_1}{N_1 + \lambda} F' \\ N_1^* &= \frac{(N_1 + \lambda)^2}{N_1 + 2\lambda}. \end{aligned}$$

Thus, the probability of $F'(N_1, N_2, \lambda)$ can be easily computed using a numerical recipe for $F(N_1^*, N_2)$ and the approximation in Equation 4.

Relationship with Semblance

The semblance of an N-element array, over an M-sample window is defined as (e.g. Neidell & Taner 1971),

$$S = \frac{\sum_{t=1}^{M} \left[\sum_{i=1}^{N} u_i(t)\right]^2}{N \sum_{t=1}^{M} \sum_{i=1}^{N} u_i(t)^2}. \tag{5}$$

Thus, S is the power on the beam divided by the average power of the channels used to form the beam, each averaged over an M-sample time window.

Douze & Laster (1979) note that,

$$F = \frac{S}{1 - S}(N - 1), \tag{6}$$

thus F can be easily determined once S has been computed.

Practical Considerations

One of the most appealing properties of F is the ability to test a hypothesis that a signal exists with a given SNR. However, the interpretation of the probability associated with F from teleseismic P signals, recorded at seismometer arrays, depends on the validity of the F signal hypothesis assumptions (I) and (II) above.

For assumption (I) to be approximately satisfied the signal coherence across the array needs to be high. In general, the cross-correlation between signals from two seismometers in an array decreases with increasing distance, but this relationship also depends on the signal passband, the site conditions at each seismometer and the structure of the crust and upper mantle beneath the array. For assumption (II) to be approximately satisfied requires that the noise coherence across the array be low. In general, noise cross-correlation decreases with increasing distance between the seismometers in an array (e.g. Blandford & Clark 1975). Thus, there is a trade-off between the validity of assumptions (I) and (II) and the spacing between the seismometers in the array.

Early experiments using the F-detector, such as that at the 19-element, 2.9 km aperture array, at CPO (Cumberland Plateau Observatory) with a minimum seismometer spacing of 0.3 km, found that strongly correlated noise, at about 1 Hz, resulted in poor agreement between the false alarm rate predicted by theory and that observed (Edwards *et al.* 1967). Blandford (1974) reports on the implementation of the F-detector at the 37 element array at TFO (Tonto Forest Observatory) with a minimum seismometer spacing of 4 km. Blandford (1974) found low average noise cross-correlations between seismometers at TFO in the signal passband, and showed that the F-detector behaves as predicted by theory. Presumably the average signal cross-correlations at TFO are high to achieve agreement between observation and theory.

Based on the TFO experiment we might expect that large-aperture (> 30 km) seismometer arrays with a minimum seismometer spacing of 4 km to be near optimal to satisfy the F signal hypothesis assumptions.

	Example 1	Example 2
Source region	East of Honshu, Japan	East Kazakhstan
Date	1997 07 09	1998 08 22
Origin Time (UT)	09:36:08.7	05:00:18.7
Lat. (°N)	35.461	49.756
Lon. (°E)	140.174	77.827
Depth (km)	70.6±4.9	0
Az. Gap (°)	108	199
m_b	4.6	3.8

Table 1: REB source parameters for our two examples.

Unfortunately, the only large-aperture array in the IMS is at NORSAR (Norway), which while having low average noise cross-correlation, also has poor signal cross-correlation (Frazier 1972). Here, we show examples of signals recorded by medium-aperture (10–20 km) arrays, with a minimum seismometer spacing of 1–2 km, where the average signal cross-correlation is high. We show that while the average noise cross-correlation is not zero, it is low, suggesting that the probability of F may be interpreted semi-quantitatively, and used as guide to an analyst to help decide whether a signal exists.

We consider two seismic disturbances as examples (Table 1); the first to show how the F trace and the associated probability can help in the identification of surface reflections (pP and sP), and the second to show how these traces can help with 'special event' analysis. To control the bandwidth B we bandpass filter the time-shifted N-channel data (for the vector slowness under consideration) using a two-pass 4-pole Butterworth filter. The passband and window length T (also used for the STA window) seems intuitively to depend on the characteristics of the signal (Blandford 1974) – typically we use a 2 s window. The LTA noise window is set to 50 s. Since our algorithm for calculating the probability of F only allows integer values for the degrees of freedom, N_1 we follow Blandford (1974) and round down to the nearest integer. The values for the semblance, F, probability and STA/LTA traces refer to the mid-point of the time window T.

We report the maximum average cross-correlation for the STA window $\max(\hat{c})$, and the average noise cross-correlation for the LTA window, $\hat{c}$(LTA). Average cross-correlations are calculated using z-transformed cross-correlation coefficients for each independent pair of seismometers in the array (VanDecar & Crosson 1990, Lilwall 1991).

Example 1: July 9, 1997 East of Honshu, Japan

We process seismograms from three medium-aperture arrays (WRA, YKA and EKA) to show how the F-trace and probability can help in the identification of P and the surface reflections pP and sP. Table 2 shows the processing parameters used for the three array seismograms presented in Fig. 1. Each seismogram is formed by beamforming using the vector slowness calculated from the REB hypocentre and the IASPEI 1991 model (Kennett 1991). The average noise correlations, $\hat{c}$(LTA) are low (< 0.15), while for these seismograms because the SNR is high, the maximum of the average cross-correlation, $\max(\hat{c})$ is also high (> 0.90) indicating good signal coherence across the array in the 0.5–3.0 Hz passband. The probability trace in Fig. 1 indicates the probability of F under the hypothesis that R (the square root of the ratio of the signal and noise power on the beam) is that indicated in Table 2. Note that here we have normalised the WRA amplitudes to unity (and inverted one channel), as there are calibration problems for these data.

On the semblance, F, probability and STA/LTA traces there are precursors to the P onset. These are due to the time window T covering both noise and the first half-cycle, or so, of the signal. At YKA this first half cycle has the highest semblance, presumably because signal generated noise in the vicinity of the receiver degrades the semblance after the P onset. This effect becomes apparent under conditions of high SNR (when precursors due to the two-pass filtering also have to be considered).

At first sight, interpretation of these seismograms in terms of P, pP and sP is not straight forward if following the IDC analyst instructions (Anonymous 1998), as the highest amplitude arrivals after the P

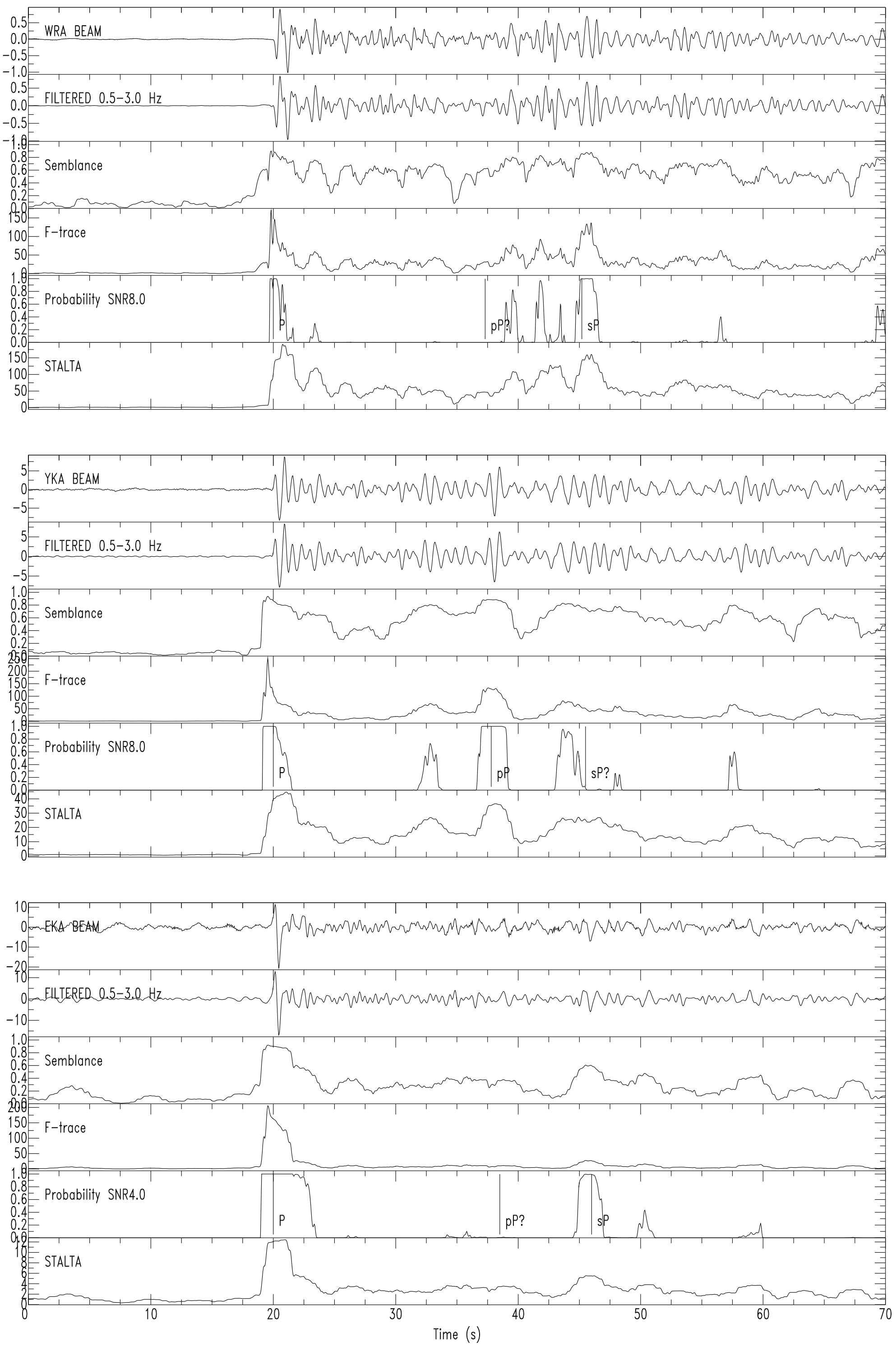

Figure 1: Seismograms and associated traces from the July 9, 1997 earthquake east of Honshu, Japan. The picks on the probability trace are the predicted times relative to the P onset for a depth of 67 km in the IASPEI 1991 model. The processing parameters are given in Table 2.

Station	Δ	Az	PV	BB	B	T	$\hat{c}$(LTA)	max($\hat{c}$)	R	$h = 67$ km	
	(°)	(°)	(km s^{-1})	(°)	(Hz)	(s)				$pP - P$	$sP - P$
WRA	55.4	186.7	15.5	5.8	$3.0 - 0.5$	2.0	0.047	0.933	8.0	17.3	25.2
YKA	65.7	29.5	17.3	300.2	$3.0 - 0.5$	2.0	0.082	0.951	8.0	17.8	25.5
EKA	84.3	339.9	22.0	29.3	$3.0 - 0.5$	2.0	0.128	0.929	4.0	18.5	26.0

Table 2: F processing parameters for Example 1. Δ: distance. Az: Azimuth. PV: phase velocity. BB: Backbearing. $\hat{c}$(LTA): average cross-correlation of the LTA time window. max($\hat{c}$): maximum of the average cross-correlation. h: depth used to predict the time difference between phases using the IASPEI 1991 model.

onset (at 20 s) show little similarity to the pulse shape of direct P. The exception is perhaps the arrival at about 38 s at YKA (which we interpret as pP with opposite polarity to P). The EKA seismogram appears simple, with small coda amplitude compared with direct P. WRA shows a series of pulses starting about 19 s after the P onset. The first of these could be pP, if so this suggests our interpretation of pP at YKA is incorrect as WRA is closer to the epicentre than YKA (Table 2), and 1-D earth models predict increasing $pP - P$ times with distance at long range.

Based on our initial interpretation that the pulse at about 38 s at YKA is pP, we calculate $pP - P$ and $sP - P$ times (using the IASPEI 1991 model) and find a good fit for a source depth of 67 km. The predicted relative $pP - P$ and $sP - P$ times are shown on the probability traces in Fig. 1. Where the probability is highest (approaching unity) corresponds closely to the predicted times for pP at YKA, and for sP at WRA and EKA. Close examination of the WRA and EKA seismograms around the predicted sP time does indeed show a weak, but distinct, arrival. Since the probability of F is high (say $> 90\%$) around the predicted sP time (and low surrounding this local maxima) suggests that this energy has the expected vector slowness for sP, and supports our initial identification of pP at YKA. We have indicated predicted arrivals that have a probability below 90% by appending a '?' to that pick. Of course, the apparent absence of pP at WRA and EKA may be due to either the orientation of the double couple (earthquake) source (pP could be near a minimum in the radiation pattern), or to defocusing of pP energy by above-source structure.

At WRA the $sP - P$ time may be about 1 s longer than predicted. Similarly, the probability trace for YKA suggests there may be a possible sP about 1.5 s before the predicted $sP - P$ time. Differences between observed and predicted $pP - P$ and $sP - P$ times of about 5% are perhaps not surprising given the heterogeneous structure expected above the slab in the subduction zone. However, the predicted moveout (Table 2) of $pP - P$ is 1.2 s between WRA and EKA, whereas the difference between the predicted $pP - P$ times and those inferred from the probability trace appears to be up to about 2 s. Thus, attempts to detect the diagnostic moveout of $pP - P$ (and $sP - P$) times may well fail as the moveout signal could be masked by heterogeneity in the above-source P and S wave speeds.

The $pP - P$ times in the REB are 24.5 s and 19.3 s, for YKA and WRA respectively. The $pP - P$ time for YKA in the REB is close to that inferred as possible sP from the probability trace, suggesting a misassociation. The REB $pP - P$ time at WRA corresponds to a local peak in the probability trace (of about 80%) that is about 2 s later than that predicted for a depth of 67 km. The probability trace for WRA in Fig. 1 shows a series of local maxima preceding the peak we have associated with sP. A similar unassociated probability peak is seen between P and pP on the YKA trace. Both WRA and YKA have azimuths that are sub-parallel to the subducting slab, so these unassociated arrivals may be due to effects such as S-to-P conversions, multi-pathed energy due to the slab, and focusing and scattering by heterogeneous structure in the source region. Analysis of F and the probability trace for signals recorded by medium-aperture arrays may help us to increase our understanding of these phases (and of scattered phases recorded by dense seismometer networks in general).

A comparison of the different traces suggests that the STA/LTA trace is qualitatively similar to F at WRA, but that F has better resolution than the STA/LTA for the YKA and EKA seismograms. However, it is clear that the probability trace is the most useful trace as it allows semi-quantitative statements to be made as to the statistical significance of each associated surface reflection. Our confidence that a probability peak from a given array is a real surface reflection is increased if peaks are found near predicted pP and sP times at other arrays.

Example 2: August 22, 1998 East Kazakhstan

Here we show how F and the associated probability can be used to test if a weak signal has been detected. P-waves from the 100 t chemical explosion in East Kazakhstan are reported in the REB from only six stations at long range ($> 30°$). The large azimuth gap (199°) is a result of no signals being reported from the south and east. For 'special event' analysis, seismic disturbances that may be considered possible treaty violations, good azimuthal coverage is desirable for, (1) reliable location, and (2) the application of the relative amplitude method to identify shallow (i.e. suspicious) disturbances (e.g. Douglas *et al.* 1999).

Fig. 2 shows the P seismograms and associated traces from two array stations, ILAR and ASAR. No detection is reported from ASAR in the REB. The ILAR filtered array beam shows a clear, simple P signal at the time reported in the REB, whereas the filtered beam at ASAR shows a possible signal at about 58 s (60 s is the time predicted using the REB location and origin time and the IASPEI 1991 model). However, there is a similar signal on the filtered beam at ASAR at about 34 s. The probability of the 34 s phase is about 40%, whereas that at 58 s is about 80%, suggesting that the 58 s signal is real (plus it is at about the expected time). Further, the probability traces in Fig. 2 confirm that the ILAR and ASAR seismograms from the East Kazakhstan explosion are simple (Douglas *et al.* 1999).

CONCLUSIONS AND RECOMMENDATIONS

Our main conclusions are listed below.

- For the medium-aperture seismometer arrays considered the average noise cross-correlations are low, and average signal cross-correlations are high, suggesting that the conditions for the F signal hypothesis are approximately met. Thus, the probability of F can be interpreted semi-quantitatively.
- The probability of F is shown to be useful for identifying weak P signals with a given vector slowness, recorded by medium-aperture arrays. This is found to be useful for identifying weak signals for 'special event' analysis, and for identifying surface reflections (pP and sP).
- Our example demonstrates that observed $pP-P$ and $sP-P$ times can differ from those predicted by the IASPEI 1991 model by up to about 2 s. Thus, the moveout of $pP-P$ and $sP-P$ times, expected to increase with increasing distance, may be masked by variations in the observed times from those predicted. The differences between observed and predicted times are consistent with variations in the above-source P and S wave speeds of up to about 10%.

We recommend that the F-trace and associated probability should be implemented as a tool to help analysts at the IDC identify P, pP and sP (and other body wave phases). Such a tool would also help analysts during the process of reviewing associated (and unassociated) detections from an automatic system, and in the process of identifying signals that have been missed by an automatic system.

Present computing power is such that there may be a case for replacing the current STA/LTA detector by F in the current IDC automatic detection processing for medium-aperture seismometer arrays in the IMS.

Further research is required to determine the P and S wave speed variations in the above-source region of deep earthquakes (the region above the subducting slab). If large variations are confirmed then the use of the moveout of $pP-P$ and $sP-P$ times as a method of confirming that surface reflections are real needs to be reconsidered.

Acknowledgments

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Station	Δ (°)	Az (°)	PV (km s^{-1})	BB (°)	B (Hz)	T (s)	$\hat{c}$(LTA)	max($\hat{c}$)	R
ILAR	60.7	20.2	16.3	328.4	1.0–3.0	2.0	0.073	0.690	3.0
ASAR	88.4	130.4	23.5	327.4	1.5–3.0	1.5	0.036	0.237	1.5

Table 3: F processing parameters for Example 2.

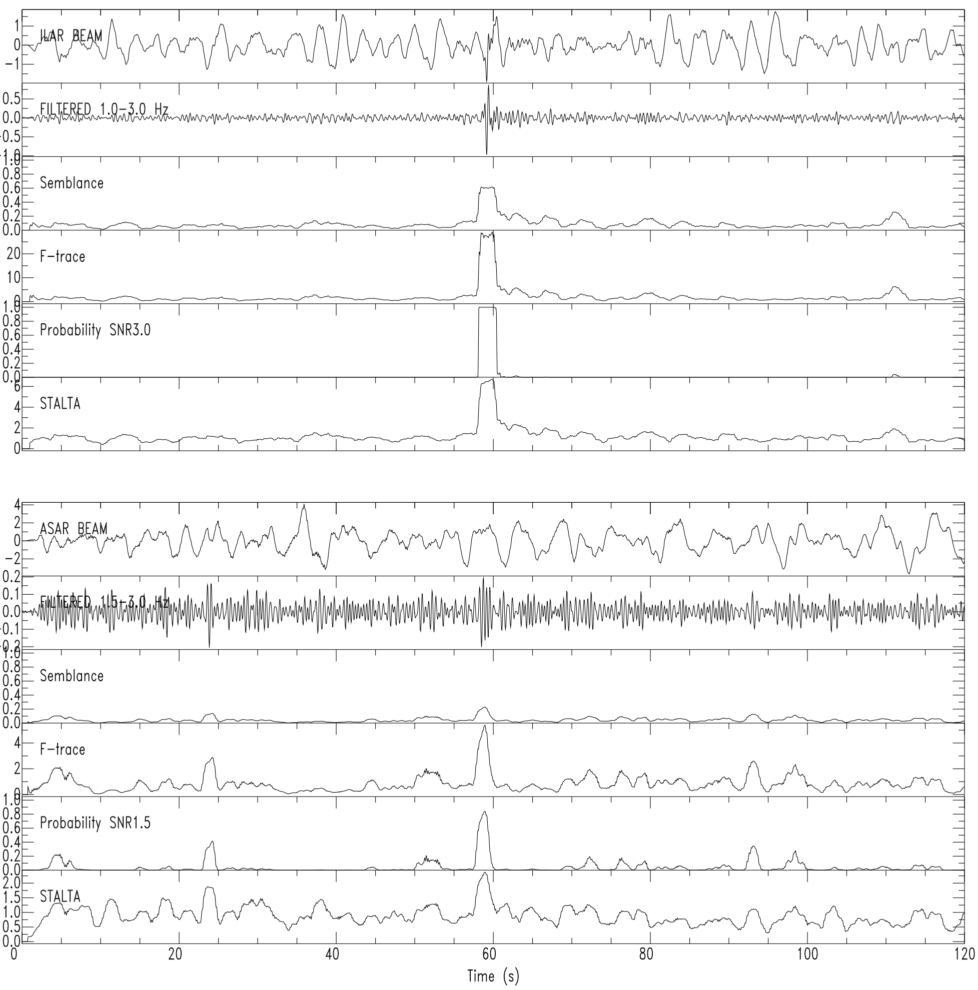

Figure 2: Seismograms and associated traces from the August 22, 1998 100 t chemical explosion in East Kazakhstan. The processing parameters used are given in Table 3.

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