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Seismology: ancient and modern

Seismology is a stone age science
Sir (later Lord) William G. Penney,
c.1959 (attributed)

1.1 The long march begins

The AWE scientists who were assigned to the forensic seismology programme in the late 1950s found that although there was much theoretical work on seismic-wave propagation, experimental and observational seismology were poorly developed. The main practical interest in seismology was in the destructive effects of large earthquakes. The only useful information routinely obtained from seismograms was arrival times of the most significant seismic phases, from which epicentres and focal depths could be estimated and Earth's structure derived. Most if not all stations used drum recorders operating at low magnification with compressed time bases (maximum speed around 1 mm s^{-1}).

The continuous background of seismic noise has little effect on observations of the waves from large earthquakes as their amplitude is much greater than that of the noise. If seismology is to provide an effective way of detecting (and ideally identifying) explosions, the recording systems in use in the 1950s were inadequate for the detection of the weak signals from tests of a few kilotons (ground motions in the 30–90° range of around 10 nm at 1 Hz). Many of the seismometers did not have the required sensitivity, and responded to unwanted non-seismic disturbances such as changes of air temperature and atmospheric pressure; seismometers that had the required sensitivity were heavy and bulky and so were unsuitable for deployment as arrays. Further, if more information was to be obtained from seismograms than mere lists of arrival times of the various seismic waves, the seismograms had to be recorded in electronic form, for example, on magnetic tape. The recordings of the ground motion from earthquakes and explosions can then be seen for what they are, records of signals radiated as elastic waves from seismic sources. As with any signal, time-series analysis can be applied to those from such sources. Frequency filtering and the estimation of spectra through the Fourier transform can be used to search for and apply criteria to identify possible explosions, and array recordings processed to enhance signals and suppress noise. Automation of the detection and processing of seismic signals also becomes possible.

Having detected the seismic waves from earthquakes and explosions it was hoped criteria would be found (variously described as discriminants, discrimination criteria and identification criteria) that would allow the two types of source to be distinguished. Criteria were needed to supplement that of first motion, which is unreliable. The search for ways of using seismology to verify a test ban stimulated interest in the amplitudes of seismic waves. In addition to times read from the x axis, amplitudes (the y axis) became increasingly important. Before about 1960, only Gutenberg and Richter (Richter, 1935, Gutenberg and Richter, 1956) appear to have shown much interest in the amplitudes of seismic waves, their aim being to use the amplitudes to set up and apply a seismic-magnitude scale to measure earthquake size.

The relationship between explosion size (yield) and the amplitudes of the seismic waves and hence magnitudes had to be investigated. Simple theory of an explosion modelled as a pressure step on the wall of a hollow sphere around the explosion was known – but this turns out to be oversimplified. Despite all the research that has been carried out since the early 1960s how explosions generate seismic waves remains one of the great unknowns in forensic seismology.

The AWE scientists began to work on all the major areas of ignorance with programmes to: improve seismometer design; record ground motion on magnetic tape; evaluate the effectiveness of seismometer arrays; look at methods of distinguishing earthquakes from underground explosions; investigate the relation between magnitude and yield; and improve methods of epicentre estimation. It seemed self-evident as the programme got underway that detection and identification of seismic sources would be easiest at short range, distances say of less than a few hundred kilometres. At the Conference of Experts the observing range was divided into three zones: first zone, 0–700 km, where wave propagation is mainly controlled by crustal structure; second zone, 700–2000 km, where wave propagation is mainly in the upper mantle; and third zone, >2000 km, where much of the ray path for P and S waves is in the lower mantle. First motion it was thought would be most easily observed in the first zone.

Before the Conference of Experts the observing range was divided into two (Richter, 1958): local distances out to 1000 km – earthquakes in this range from a station were referred to as near or local earthquakes – and teleseismic distances, all distances beyond 1000 km. Earthquake signals observed at distances beyond 1000 km were referred to as teleseisms. In forensic seismology the observing range is divided into three with boundaries only slightly different from those specified by the Experts, thus: local distances, 0–1000 km; regional distances, 1000–2000 km; and teleseismic distances, >2000 km.

Initially most research by the USA and UK was on local and regional seismograms, but in 1962 the UK shifted much of its effort to teleseismic recordings and particularly recordings in the range 3000–10 000 km. In the USA, although recordings in all zones are used, most of the research continues to be on local and regional seismograms.

1.2 Seismic signals and noise

Implicit in any visual analysis of seismograms is that they are the sum of harmonic waves of the form $A(\omega)\cos\{\omega t + \phi(\omega)\}$, where $A(\omega)$ is the amplitude and $\phi(\omega)$ the phase at angular frequency ω (=$2\pi f$ or $2\pi/\mathcal{T}$, where f is frequency and $\mathcal{T}$ period). Signals are assumed to approximate to such waves and rough amplitudes and periods are measured. On the basis of these measurements, signals are described as being of high- or low-frequency compared to some average. Noise is often described as having a predominant frequency.

More detailed descriptions of signals are given using the Fourier transform to determine $A(\omega)$ and $\phi(\omega)$ against frequency – the amplitude and the phase spectrum respectively. Noise properties can also be expressed in the frequency domain.

The frequencies of ground motion of most value in forensic seismology range from around 0.01 Hz to about 20 Hz but a seismograph that records all frequencies of ground displacement in this range at constant magnification – that is, a broad-band seismograph – produces seismograms that are dominated by noise at 0.125–0.167 Hz (8–6 s period); the seismic-noise spectrum has a peak in this band. The noise peak is generated by water-wave action in the oceans and hence is referred to as the oceanic-microseism peak. During storms the amplitude of the microseisms can be 10 000 times the amplitude of the smallest detectable P wave. On broad-band recordings only signals from earthquakes of large magnitude (or explosions of high yield) are visible above noise. Figure 1.1(a) illustrates how the oceanic microseisms can swamp the P signal from an underground explosion. For, although the probable yield of the explosion is several tens of kilotons, the observed amplitude is less than a quarter of that of the noise. Yet signals from much smaller sources can be seen above noise at frequencies around 0.05 Hz (20 s period) and 1 Hz (1 s period). Consequently, until the 1990s most seismological stations recorded in two passbands: the long-period (LP) band with a peak magnification around 0.05 Hz and the short-period (SP) band where the magnification falls-off rapidly below 1 Hz.

The advantage of recording in the SP band is illustrated in Figure 1.1, for whereas P is small relative to the noise on the broad-band recording (Figure 1.1(a)), on the SP (Figure 1.1(b)) the signal is at least 4× the peak amplitude of the preceding noise – and this is a noisy station. At some stations the noise at 1 Hz can be as low as 1 nm, about 50 times smaller than that shown in Figure 1.1(b).

The passbands of LP and SP recording systems and the wave types of most interest in forensic seismology in each band, are shown diagrammatically in Figure 1.2. The most widely recorded waves from earthquakes and underground explosions are the SP P waves. From the onset times of these waves the best estimate of an epicentre can be obtained. P-wave motion is confined to the plane containing the epicentre, the observing point and the centre of Earth and, as motion is along the ray path, gives rise to vertical and radial components. Positive radial motion is away from the source. As the largest component of ground motion due to a P wave at long range is the vertical, most of the early SP arrays installed for forensic seismology recorded only the vertical component.

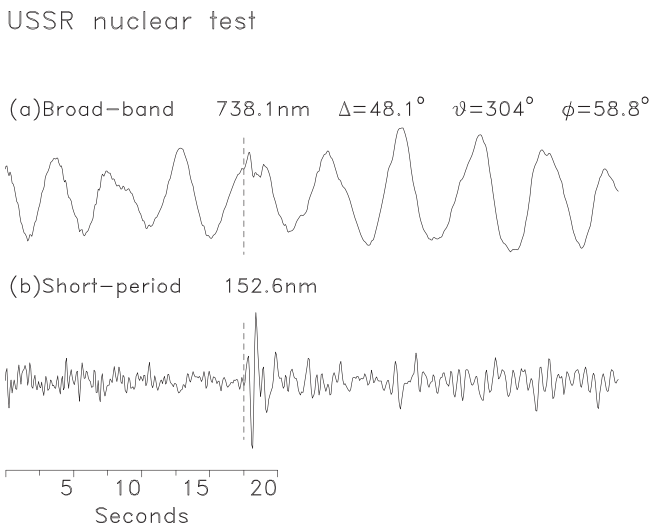

Figure 1.1 P from the explosion of the 7 August 1975 at 03:56:57.6 at the Degelen Mountain Test Site of the USSR in Kazakhstan, recorded in the south of England. Δ, distance, is the angle subtended at the centre of Earth by the arc of the great circle joining epicentre and station. ϑ, azimuth, is the direction from epicentre to station. ϕ, back azimuth, is the direction from station to epicentre. (a) Broad-band (0.1–5.0 Hz) seismogram. (b) SP seismogram ($\sim$1 Hz). The maximum peak-to-trough amplitude is shown in nanometres. The dashed line marks signal onset.

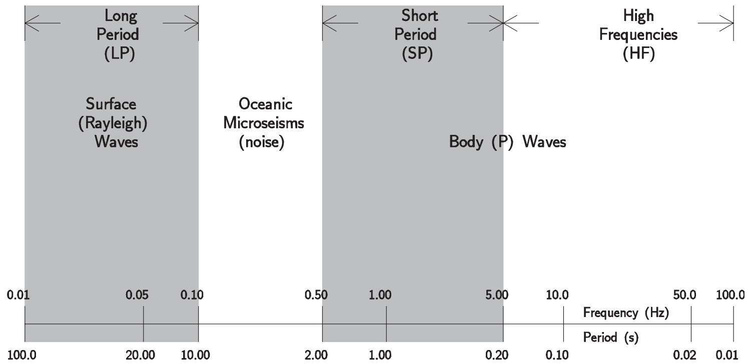

Figure 1.2 Recording bands and wave types used in forensic seismology. (From Douglas (2007).)

At local distances the predominant frequency of SP P waves can be 10 Hz or more. The high frequencies are preferentially attenuated with distance, so that at distances of more than a few hundred kilometres the predominant frequency is 1–2 Hz. S waves at local distances have frequencies similar to those of P. However, SP S waves attenuate more rapidly with distance than P and are usually below the noise level at teleseismic distances.

S waves are polarized: their motion is at right angles to the ray path and consequently in general S motion has vertical, radial and transverse components. SH, the transverse motion, is in the horizontal plane at the observing point, at right angles to the direction of propagation. The vertical and radial S motions are the components of SV. On reflection and refraction part of the SV energy converts to reflected and refracted P and conversely incident P gives rise to SV. When wave speeds are purely a function of depth, SH is uncoupled from P and SV.

As SV-wave motion is in the vertical plane at right angles to the direction of propagation, then at distances beyond about 200 km the largest amplitude is on the horizontal component. The use in the past of principally vertical-component instruments means that P was detected preferentially. As SP S waves are rarely seen at regional and teleseismic distances, it was argued that little is lost by not recording horizontal motion. Nevertheless, if there are paths through Earth of low S-wave attenuation, SP S waves, particularly the SH component, may have passed unnoticed because of the recording bias in favour of vertical-component systems.

When a plane body wave (P or S) strikes a plane free surface the boundary conditions are satisfied assuming all the energy is reflected as body waves. When body waves from a source at finite depth and thus with a curved wave front strike a plane free surface the boundary conditions cannot be satisfied without introducing surface waves. Conversely, if a plane body wave strikes an irregular free surface, the boundary conditions again cannot be satisfied without introducing surface waves. Surface waves generated in both these ways are observed: those generated at the free surface (assumed plane) above a source are the most widely observed and most valuable for the study of Earth's structure. Although a wide range of frequencies is generated, SP surface waves are only seen at local distances. At long range surface waves are most easily recorded in the frequency range 0.01–0.1 Hz. Surface waves generated by plane waves striking topographic features at a recording station are usually of high frequency ($\sim$1 Hz) and are sometimes seen on SP seismograms; they are treated as noise (Chapter 5).

As the name implies surface waves have their maximum displacement in the vicinity of the free surface. They are of two types: Rayleigh and Love waves. During the passage of a Love wave the particle motion is in the horizontal plane at right angles to the direction of propagation (Figure 1.3). Consequently, Love waves are recorded by horizontal-component seismometers only.

The only surface wave used extensively in forensic seismology is the LP Rayleigh wave. Ground motion due to Rayleigh waves has vertical and horizontal components of similar amplitude, but it is principally the vertical component that is used. During the passage of a Rayleigh wave a particle at the free surface describes a retrograde ellipse in the plane containing the direction of propagation and the vertical (Figure 1.4). The variation in amplitude with h/λ in a uniform medium (Poisson solid[1]) is shown in Figure 1.4; h is depth and λ is horizontal wavelength. The vertical component has a maximum at a depth of

[1] For a Poisson solid Lamé's constant and the modulus of rigidity are equal. Poisson's ratio is then 0.25, the value assumed for most material in Earth's continental crust.

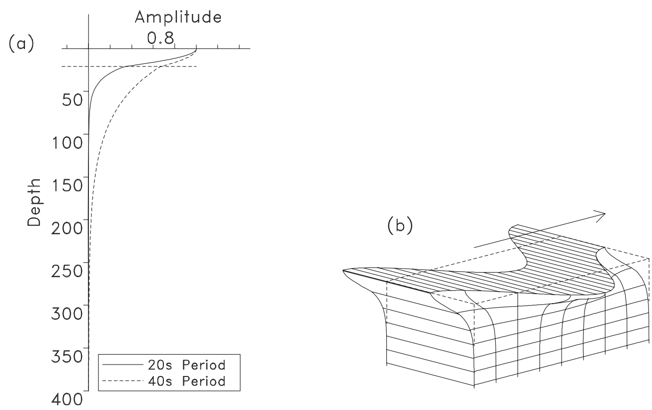

Figure 1.3 (a) Love-wave displacements with depth at frequencies of 0.05 Hz (20 s period) and 0.025 Hz (40 s period). The structure is a 20 km thick layer (S-wave speed 2.60 km s^{-1}, density 2.7 g cm^{-3}) over a half-space (S-wave speed 3.94 km s^{-1}, density 2.9 g cm^{-3}). As frequency decreases the Love-wave speed increases from around the S-wave speed in the layer to that of the half-space. For the two frequencies shown, as most of the motion at 0.05 Hz is in the layer, the Love-wave speed (3.0 km s^{-1}) depends principally on the S-wave speed (and density) in the layer. At 0.025 Hz most of the motion is in the half-space, so that the Love-wave speed (3.6 km s^{-1}) depends principally on the properties of the half-space. (b) Displacement in a parallelepiped during the passage of a Love wave.

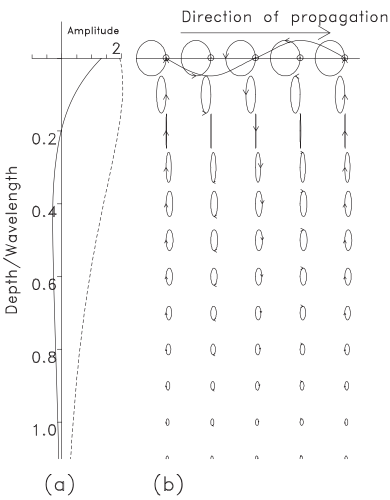

Figure 1.4 (a) Amplitude of a Rayleigh wave as a function of depth/wavelength in a uniform medium (fundamental mode): full line, horizontal component; dashed line, vertical component. (b) Particle motion caused by the passage of a Rayleigh wave.

0.076λ and then falls off so that at a depth of λ, the amplitude is only 0.19 of the amplitude at the surface. The amplitude of the horizontal component is a maximum at the free surface and has zero amplitude at a depth of 0.193λ; below this depth the displacements have the opposite sign to those at shallower depth, so that the motion is prograde.

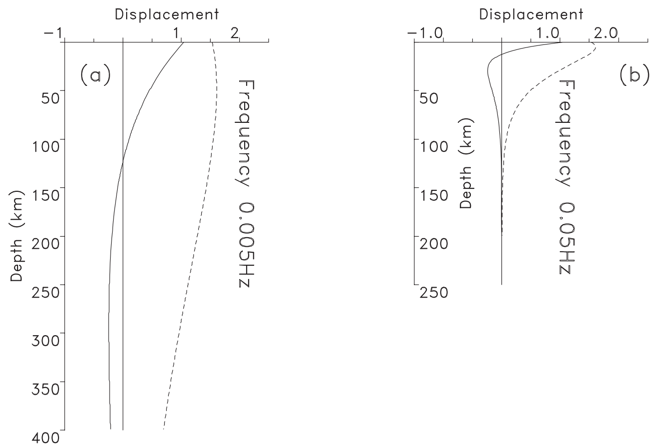

Figure 1.5 Variation in the amplitude of Rayleigh waves (fundamental mode) with depth in a uniform half-space ($\alpha = 8.1$ km s^{-1}; $\beta = \alpha/\sqrt{3}$) at two periods: (a) 200 s (0.005 Hz) and (b) 20 s (0.05 Hz). Full line, horizontal component; dashed line, vertical component.

The variation in Rayleigh wave amplitude with depth depends on wavelength and hence on frequency. At low frequency (long wavelength) the decay of amplitude with depth is slow; at high frequencies it is rapid. The variation with depth in a half-space for two frequencies is shown in Figure 1.5. For a layered half-space the variation in amplitude with depth has the same general form, with amplitude decaying more rapidly with depth at high frequencies than at low frequencies. A consequence of this is that at high frequencies and short wavelengths the bulk of the motion is in the topmost layers and it is the P- and S-wave speeds (α and β) and densities (ρ) in these layers that largely determine the Rayleigh wave speed. At long periods where a large proportion of the energy in the wave is travelling below the surface layers the speed is determined principally by the values of α, β and ρ at depth. The phase speed of Rayleigh waves is thus a function of frequency.

For Rayleigh waves in a uniform half-space, the phase speed is independent of frequency, and depends on the wave speeds (α and β) and density of the half-space (ρ); for a Poisson solid, where $\beta = \alpha/\sqrt{3}$, this speed is $\sim 0.92\beta$. For the simple model of a uniform crust over a uniform half-space the phase speed at low frequencies tends to $\sim 0.92\beta_1$ and at high frequencies to $\sim 0.92\beta_0$, where β_1 is the S-wave speed in the half-space and β_0 the S-wave speed in the crust.

For Love waves, for which the phase speed depends on the variation of β and ρ with depth, the phase speed is also a function of frequency: the lower the frequency the longer the wavelength, and the more the speed depends on β and ρ at depth (Figure 1.3); in a half-space if β does not somewhere increase with depth, Love waves cannot exist.

The expression that relates ω (angular frequency) and the variation of α, β, ρ with depth to wave number $\kappa(\omega)$ ($= \omega/c$) and hence phase speed c is the period equation and is rather complex. Note that at high frequencies (except in a uniform half-space) there are several modes of propagation; each mode is characterized by its own speed–frequency curve and the distribution of amplitudes with depth differs for each mode. For the lowest-speed Rayleigh mode, the fundamental mode, the particle motion at the surface is a retrograde ellipse. For the next (first) higher mode the particle motion is a prograde ellipse and for successive higher modes the particle motion is alternately retrograde and prograde. In practice, it is only the fundamental mode that is well observed. Love waves also propagate as a series of modes but again like Rayleigh waves it is only the fundamental mode that is well observed.

SP Rayleigh waves (called Rg) are observed from shallow sources (focal depths of 0–5 km) but these rapidly attenuate with distance and thus are usually seen only within a few hundred kilometres of the source. A second type of SP surface wave, Lg, propagates to greater distances than Rg, and has been used to estimate the yield of explosions (see Section 8.5).

The most important quantity derived routinely from observed amplitudes is the magnitude of a seismic source; this is intended to be a measure of the seismic energy released by an earthquake or explosion. The concept of magnitude was introduced in 1935 by Richter (1935) specifically to measure the relative sizes of Californian earthquakes from observations made within a few hundred kilometres of the epicentre. The magnitude scale is logarithmic (base 10) and magnitude zero is defined as a source that would give (if it could be detected) a maximum amplitude on the seismogram of 10^{-3} mm when recorded at a distance of 100 km on a Wood–Anderson seismograph (a type of seismograph for detecting the horizontal component of ground motion, formerly in common use in California) operating at a magnification of 2800. By observing the decay of amplitude with distance a table of corrections for distance was derived so that recordings made at any distance out to a few hundred kilometres could be used. The zero level of this magnitude scale is arbitrary but was chosen low enough for all felt earthquakes to have a positive magnitude.

Since its first definition by Richter several other magnitude scales have been introduced (see, for example, Båth (1966), Hanks and Kanamori (1979)). The expression for magnitude $\mathcal{M}$, for all the widely used scales has the general form:

$$\mathcal{M} = \log(\mathcal{A}/\mathcal{T}) + B(\Delta, h),$$

where $\mathcal{A}$ is the amplitude of the ground motion, $\mathcal{T}$ the period and $B(\Delta, h)$ a correction for the decay of amplitude with distance, Δ, and depth of focus h. As originally defined $\mathcal{A}$ is in micrometres, however, as forensic seismologists usually deal with weak signals with amplitudes less than a micrometre, AWE Blacknest, in common with others, works in nanometres which are more convenient to use than micrometres. Consequently, the baseline, of the Blacknest curves for correction for distance and depth of focus are the micrometre curves minus three.

The scales have grown haphazardly by empirical observation supplemented from time to time by theoretical considerations. Formulae for computing magnitudes are constantly being revised in an attempt to produce magnitudes that are independent not only of distance between source and receiver but also of, amongst other things, the type of seismograph, the structure under the recording stations and the form of the signal. The aim of these revisions is to produce formulae that are easy to apply so that a magnitude can be worked out from one or two simple measurements made by hand from a seismogram and using nothing more complex than a calculator. By keeping the computation simple, more stations are able to report a magnitude than would be possible if a detailed analysis of each seismogram was needed. The draw-back to using magnitudes is that any physical basis is obscure. The computation of magnitude becomes the routine application of a formula whose principal justification is that, as it has been widely used over several decades, it is useful for, amongst other things, statistical studies of the variation in numbers of earthquakes with size.

There are two widely used magnitude scales, one the body-wave magnitude, m_b, calculated from the SP P waves, and the other the surface-wave magnitude, M_s, calculated from the LP Rayleigh waves. Again these are arbitrary scales and most seismologists would like to have a measure of source size based on some physical property of the source, and some progress has been made towards this with the introduction of 'seismic moment' (see Section 6.2). Nevertheless, magnitude is still a useful guide to how widely a seismic source will be recorded. For example, an explosion of 1 kt has an m_b of 4–4½ and will be detected out to epicentral distances of about 90°; for such an explosion M_s may be as low as 2 and the Rayleigh waves are then detected only out to distances of a few degrees. For what is possibly the largest underground explosion ever, Cannikin (6 November 1971, Table L.1), yield ~4 megatons, m_b is ~7 and M_s about 5.6. Destructive earthquakes usually have magnitudes above M_s5.5 and are detected worldwide. There are estimated to be about 8000 earthquakes annually with $m_b > 4$ (Lilwall and Douglas, 1984).

One explosion that had a particularly important influence on the development of seismological methods of test ban verification is code-named Long Shot,[2] an 80 kt explosion fired underground by the USA in 1965 at Amchitka Island in the Aleutian Islands (Table L.1). The epicentre, time and yield of the explosion were announced in advance of the firing time. The amplitudes of the P signals at many stations were much larger than expected from NTS experience. (Station magnitudes, m_b, range from 4.75 at Prince George, British Columbia, PG-BC, to 6.64 at Fort Churchill, Manitoba, FCC.) When Long Shot was fired the largest station magnitude that had been recorded from an explosion at the NTS with a published yield was m_b6.3 recorded at Arequip, Peru (ARE) from Bilby, a 249 kt explosion (13 September 1963, Table L.1); this was an early indication that the magnitude for a given yield is test site dependent. The P signals from Long Shot were reported from over 300 stations making it at that time one of the most widely reported seismic disturbances.

The elastic waves radiated by seismic sources – earthquakes and explosions – are not band limited: the radiated frequencies range from zero upwards with no obvious upper limit. At teleseismic distances, however, amplitudes above a few hertz are low relative to the amplitude at 1 Hz and only rarely are frequencies above a few hertz of value in forensic seismology. The claim by Evernden *et al.* (1986) that conventional LP and SP recordings

[2] Most seismologists (including in the past the author) refer to the explosion as Longshot; the correct code name is Long Shot (Springer *et al.*, 2002).

are inadequate for test ban verification and that all the problems of verification could be solved with high-frequency (5–100 Hz) recording has failed to be substantiated.

If the only method of recording is on paper or film, then the ground motion must be filtered before the seismogram is written. With recordings in machine readable form, on the other hand, there is no reason in principle why ground displacement should not be recorded broad-band and filters applied to extract signals over as wide a band as possible. Despite this, forensic seismologists in the Western Bloc – including those at the AWE – initially followed standard practice and recorded only the LP and SP bands. In the Eastern Bloc broad-band systems were used but most of these seem to have operated at low magnification and recorded on paper or film, so the only useful seismograms were from large earthquakes.

Examination of many SP signals shows that signal-to-noise ratio (SNR) can often be improved by further filtering to pick out just the narrow band of frequencies where the ratio is a maximum. Such additional filtering lowers the detection threshold at many stations but has a disadvantage – it removes information. Any difference between the spectra of earthquakes and explosions, for example, is obscured, so increasing the difficulty of distinguishing between the two types of source. The conflict between the need for narrow-band recordings for detection and broad-band recording for identification has been a constant problem in forensic seismology.

In late 1969 AWE Blacknest began recording ground displacement in the broad-band, 0.01–10 Hz. Conventional SP and LP seismograms could then be derived from the broad-band recordings. The original reason for starting to record broad-band was to try to reconcile disagreements in m_b between seismologists in the Eastern Bloc who measure magnitudes on broad-band seismograms and those in the West who measure such magnitudes on SP seismograms. Once recording started, however, it became clear to the seismologists at Blacknest that they had discovered a new seismology – or more correctly had ‘reinvented the wheel’ – for the seismograms were often like those of classical textbooks, showing many of the standard phases. But, whereas early seismologists had to make do with drum recordings made on fixed (and compressed) time bases, magnifications and passbands, AWE Blacknest seismologists had all the flexibility provided by magnetic-tape recordings to get the most out of the broad-band seismograms. A striking feature of the broad-band seismograms for those used to the highly oscillatory SP seismograms is how clear the source functions of earthquakes often are when observed broad-band (Figure 1.6). Also pP and sP are often more clearly seen on broad-band than on SP seismograms; these arrivals result from upward travelling P and SV waves respectively, which are reflected as P from Earth’s solid free surface. From the time between P and the surface reflections source depth can be estimated.

AWE Blacknest was one of the first groups in the West to encourage the use of broad-band systems to extract as wide a band of signal frequencies from the recordings as possible (Marshall *et al.*, 1972). For various reasons the advantages of such recordings were only slowly (and sometimes grudgingly) acknowledged and instrumentation to make such recordings developed. Nevertheless, modern seismological stations now use broad-band systems from which LP and SP (displacement) seismograms can be derived as required.