Chapter 8

The Seismic Wavefield

So far in this book we have shown how we may calculate complete theoretical seismograms for a horizontally layered medium. Such calculations are most useful when the total time span of the seismic wavetrain is fairly short and there is no clear separation between different types of wave propagation processes.

As the distance between source and receiver increases, the wavetrain becomes longer and waves which have travelled mostly as $P$ waves arrive much earlier than those which propagate mostly as $S$. Also the surface waves which are principally sensitive to shallow $S$ wavespeed structure separate out from the $S$ body waves which are returned from the higher wavespeeds at depth. Once the seismic wavetrain begins to resemble a sequence of isolated phases it becomes worthwhile to develop approximate techniques designed to synthesise a particular phase. However, in order that such approximations can be made efficiently, with a due regard for the nature of the propagation process, we need to have a good idea of the character of the seismic wavefield.

In this chapter we will therefore survey the character of the seismograms which are recorded at different epicentral ranges.

8.1 Controlled source seismology

In the application of seismic techniques to the determination of geological structure, the source of seismic radiation is usually man-made, such as an explosive charge. In this case the origin time is known with precision and with high frequency recording the fine detail in the seismograms can be retained. Experiments of this type can be loosely divided into two classes characterised by the maximum range at which recordings are made and the density of recording points.

In reflection seismic studies attention is concentrated on $P$ waves reflected at depth and returned at small offsets from the source. The propagation paths are then close to the vertical, particularly for reflections from deep structure. The major use of the reflection method has been in prospecting for minerals and hydrocarbons. Here the features of interest usually lie shallower than 5 km and the array of geophones at the surface rarely extends to more than 5 km from the source. Many
receivers are used for each source (typically 48 or 96 in current practice), and the source point is then moved slightly and the recording repeated. The multiplicity of subsurface coverage can then be exploited to enhance the weak reflections from depth. Reflection methods are now also being used for investigations of the deep crust (Schilt et al., 1979) and here longer recording arrays are often used.

In *refraction* studies the receivers extend to horizontal ranges which are eight to ten times the depths of interest, and the density of observations is often fairly low. At the largest ranges the main features in the seismograms are refracted phases or wide-angle reflections from major structural boundaries. These can sometimes be traced back into reflections at steeper angles if adequate coverage is available at small ranges.

Early work in seismic refraction in Western countries used very limited numbers of receivers, but latterly the benefits of very much denser recording have been appreciated (see, e.g., Bamford et al., 1976). In Eastern Europe and the USSR, reflection and refraction techniques have been combined in ‘Deep Seismic Sounding’ (Kosminskaya, 1971), which has closely spaced receivers along profiles which can be hundreds of kilometres long. Such an arrangement gives a very detailed description of the seismic wavefield.

### 8.1.1 Reflection studies

For many years the major source of seismic radiation used in reflection work was small explosive charges, both at sea and on land. Now, however, a large proportion of the work on land uses an array of surface vibrators to generate the seismic waves. This avoids drilling shot holes and allows more control over the frequency content of the signal transmitted into the ground. For marine work the commonest source is now an array of airguns, which generate $P$ energy in water by the sudden release of high pressure air.

All these energy sources are at, or close to, the surface and so tend to excite significant amplitude arrivals travelling in the low wavespeed zone at the surface (water or weathered rock). In marine records these slowly travelling waves are mostly direct propagation in the water and multiple bottom reflections, but in very shallow water there may also be effects from the weak sediments at the bottom giving strong arrivals known as ‘mud-roll’. On land a surface vibrator is a very efficient generator of fundamental mode Rayleigh waves and such ‘ground-roll’ phases show up very strongly when single geophone recording is used (figure 1.3). When explosive charges are used they are usually fired beneath the weathered zone, this reduces the excitation of the ground-roll, but it can still have large amplitude.

The weak reflections from depth have very small apparent slownesses on a surface array and tend to be obscured in part by the shallow propagating phases. In order to remove the ground-roll and water phases, the seismic records are normally obtained not from a single sensor, but from an array of sensors. If such an array is chosen to span a wavelength of the ground-roll at the dominant frequency, the
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Figure 8.1. Seismic reflection gather showing prominent surface-reflected multiples indicated by markers. This set of traces has been selected to have a common midpoint between source and receiver and then has been corrected for the time shifts associated with source-receiver offset on reflection at depth, with the result that primary reflections appear nearly flat. (Courtesy of Western Geophysical Company).

resulting summed amplitude is substantially reduced. The residual ground-roll can be removed by exploiting the separation in slowness from the reflections and so designing a filter, e.g., in the frequency-slowness domain, to leave only the reflections with small slowness.

The use of arrays of sources and receivers makes it difficult to produce good theoretical models of the source radiation. For airgun arrays the guns are usually so close together that very complex interference effects occur. With surface vibrators the ground coupling can be highly variable and several vibrators in a similar location can have very different seismic efficiency.

The reflections from depth are weak compared with the early refracted arrivals from the shallow structure. These refractions are often forcibly removed from reflection records (‘muting’) and an attempt is then made to equalize the amplitude of the traces in time and distance to compensate for losses in propagation. As a result a set of reflection records will normally appear to become more ragged with increasing time, since noise is amplified along with the coherent signal.

In most reflection situations the reflection coefficient at the surface is larger than any of the reflection coefficients in the subsurface, particularly at near-normal incidence (cf. figure 5.4). Waves which have been reflected back from below the source can be reflected at the free surface, and then reflected again by the structure. A good example is shown in figure 8.1 where a prominent reflector (R) is mirrored at twice the time by its free-surface multiple (FR). This surface multiple obscures genuine reflections from greater depth. When there is a strong
contrast in properties at the base of the low wavespeed zone, multiple reverberations within this zone associated with each significant reflection lead to a very complex set of records. Often predictive deconvolution (Peacock & Treitel, 1969) will give help in ‘cleaning’ the records to leave only primary reflections. Sometimes internal multiples between major reflectors can also give significant interference with deeper reflections.

Most reflection recordings on land have been made with vertical component geophones and so $P$ waves are preferentially recorded. However, waves which have undergone conversion to $S$ at some stage of their path can sometimes be seen at the largest offsets on a single shot gather. Converted phases can occur at small offsets in the presence of strongly dipping reflectors. At sea, pressure sensors are used and the recording array is not usually long enough to pick up effects due to conversion. The standard data processing techniques, particularly stacking of traces with an estimated wavespeed distribution to attempt to simulate a normally incident wave, tend to suppress conversions.

Although the object of seismic reflection work is to delineate the lateral variations in subsurface structure, studies of wave propagation in stratified models can help in understanding the nature of the records, as for example in multiple and conversion problems.

### 8.1.2 Refraction studies

Whereas the object of a reflection experiment is to delineate the fine detail in geological structure, in refraction work resolution is sacrificed to penetration in depth. As a result the interpretation of a refraction profile, which does not cross any major vertical discontinuities such as deep faults, will give only the broad outline of the lateral variations in structure. The principal phases which can be correlated from record to record are refracted arrivals (quite often interference head waves, see Section 9.2.2) and wide angle reflections from major horizontal boundaries. As a result the portion of the wavefield which is studied is most sensitive to the wavespeed distribution near interfaces and strong gradient zones and reveals little information about the rest of the structure.

The source of seismic radiation for refraction work is normally an explosive charge recorded at an array of receivers. After a shot the receiver array is moved to a new location and a further shot fired. In this way a detailed profile can be built up to considerable range with only a limited number of recording stations. Since studies of deep structure require a detailed knowledge of the shallower regions, multiple shots at a variety of ranges are often fired into the same receiver array so that detailed results can be built up at both large and small ranges (Bamford et al., 1976). Such a procedure also allows a test of the degree of lateral homogeneity along a refraction profile. In work on land there is often considerable variability in amplitudes between nearby recorders, and in order to reduce local variations
The depth of the shot on land must be such as to contain the explosion. The resulting seismic wavefield is rich in $P$ waves, and some $S$ waves are generated by reflection at the free surface. These deep sources are not very efficient generators of Rayleigh waves, but a surface wavetrain is often seen late on refraction records. In figure 8.2a we show the close range seismograms from an experiment conducted by the United States Geological Survey in Saudi Arabia (Healy et al., 1981), which
8.2 Ranges less than 1500 km

show clearly the three major wave contributions. We display the beginning of the $P$ wavetrain at larger ranges in figure 8.2b, and we can see considerable detail in the phases with closely spaced receivers. At large ranges (300 km - 600 km) detailed refraction experiments have revealed character within the first arriving $P$ waves normally referred to as $Pn$ (Hirn et al., 1973) which could not have been observed with the coarse spacing available with permanent stations. These observations suggest the presence of fine structure in the uppermost part of the mantle which could not be resolved in previous studies.

Seismic refraction work at sea has significant differences from work on land, since it is difficult to use more than a few receivers and so a suite of observations is built up using multiple shots. At short ranges (< 20 km) airguns have been used to give high data density and provide continuous coverage from near-vertical incidence out to wide angle reflections (see, e.g., White, 1979). This data gives good control on the structure of the uppermost part of the oceanic crust and the longer range refractions fill in the picture at depth. For refraction experiments which have been shot along isochrons in the oceans, lateral variations in structure are not too severe, and stratified models give a good representation of the structure.

The interpretation of refraction records was initially based on the times of arrival of the main $P$ phases, but latterly this has been supplemented with amplitude information. Frequently the amplitude modelling has been done by computing theoretical seismograms for an assumed model, and then refining the model so that observations and theoretical predictions are brought into reasonable agreement. This approach has spurred on many of the developments in calculating theoretical seismograms both by generalized ray methods (Helmberger, 1968 - see Chapter 10) and reflectivity techniques (Fuchs & Müller, 1971 - see Section 9.3.1). The use of amplitude information has resulted in more detail in the postulated wavespeed distributions with depth. In particular this has led to a considerable change in our picture of the oceanic crust (cf. Kennett, 1977; Spudich & Orcutt, 1980). Braile & Smith (1975) have made a very useful compilation of theoretical seismograms for the continental crust, illustrating the effect of a variety of features in the wavespeed distribution.

In refraction work most attention is given to the $P$ arrivals, but on occasion very clear effects due to $S$ waves or conversion can be seen and these may be used to infer the $S$ wavespeed distribution. Even when such waves are not seen the $S$ wavespeed distribution can have significant influence on the character of the $P$ wavefield (White & Stephen, 1980).

8.2 Ranges less than 1500 km

For epicentral distances out to 1500 km the properties of the seismic wavefield are dominated by the wavespeed distribution in the crust and uppermost mantle.
8.2.1 **Strong ground-motion**

In the immediate neighbourhood of a large earthquake the Earth’s surface suffers very large displacements which are sufficient to overload many seismic instruments. However, specially emplaced accelerometer systems can record these very large motions. These systems are normally triggered by the *P* wavetrain and so the accelerograms consist almost entirely of *S* waves and surface waves. Normally the strong-motion instruments are somewhat haphazardly distributed relative to the fault trace since they are placed in major buildings. However, in the Imperial Valley in California, an array of accelerometers was installed across the trace of the Imperial fault. These stations recorded the major earthquake of 1979 October 15 with some accelerometers lying almost on top of the fault (Archuleta & Spudich, 1981).

In figure 8.3 we show the three components of velocity for a group of stations close to the fault obtained by numerical integration of the accelerograms. The earthquake rupture started at depth on the southern portion of the fault and propagated to the north-west. The major features of the velocity records are associated with the progression of the rupture, and show strong excitation of higher mode surface waves on the horizontal components. The early high frequency arrivals on the vertical component are probably multiple *P* phases. The oscillatory tails to the records arise from surface waves trapped in the sediments.

In order to understand such strong ground motion records we have to be able to calculate complete theoretical seismograms as in Section 7.3 and, in addition, need to simulate the effect of large-scale fault rupture. In a stratified medium this can be achieved by setting up a mesh of point sources on the fault plane with suitable weighting and time delays, and then summing the response at each receiver location. Such a representation will fail at the highest frequencies because the wavelengths will be smaller than the mesh spacing, but will describe the main character of the event. Point-source models are still useful since they allow the study of the effects of wave propagation in the crustal structure rather than source processes.

The aftershocks of major events are often quite small and these may be modelled quite well with equivalent point sources. In many areas the surface motion is strongly affected by the sedimentary cover, particularly where this is underlain by high wavespeed material. A detailed study of such amplification effects has been made by Johnson & Silva (1981) using an array of accelerometers at depth in a borehole.

8.2.2 **Local events**

Most seismic areas now have a fair density of short-period seismic stations which have been installed to allow detailed mapping of seismicity patterns and so have a high frequency response. A common features of seismograms at such sites are
8.2 Ranges less than 1500 km

Figure 8.3. Three-component velocity records for the 1979 earthquake in the Imperial Valley, California. The epicentre lay about 20 km to the south-east of the group of stations shown in map view. The surface fault break is also marked. (Courtesy of U.S. Geological Survey)

short bursts of energy associated with local earthquakes, less than 200 km or so from the station (see figures 8.4 and 8.5). These records are dominated by \( P \) and \( S \) body waves, although at larger ranges there are sometimes hints of surface waves which increase in importance on broad-band records.

In figure 8.4 we show seismograms recorded on the North Anatolian fault zone (Crampin et al., 1980), for an earthquake at 13 km depth at a hypocentral distance of 18 km. The \( P \) waveform is quite simple and is followed by very clear \( S \) onsets with a lower frequency content indicating significant attenuation in the fault zone. Close recordings of small aftershocks also show such a pattern of arrivals but the details of the waveform can be strongly influenced by near-surface structure, such as sediments.
At most permanent seismic stations with visual recording, it is difficult to separate the $P$ and $S$ wave arrivals from close events and so the records show strong excursion followed by swift decay. As the epicentral distance increases, the time separation between $P$ and $S$ waves is such that distinct phases are seen. In figure 8.5 we show vertical component records for two local events recorded at Jamestown in Northern California. The closer event (figure 8.5a) is about 100 km away from the station, and shows a clear crustal guided $P_g$ group which begins to die away before the $S_g$ waves which carry most of the energy. The $S$ wave coda has a generally
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The exponential envelope and the later arrivals may well be due to scattering in the neighbourhood of the recording station. The more distant event (figure 8.5b) is 200 km away from the station and now Pn waves propagating in the uppermost mantle have separated from the front of the Pg group. The corresponding Sn phase is often difficult to discern because of the amplitude of the P coda, but there is a hint of its presence before the large amplitude Sg group in figure 8.5b.

8.2.3 Regional events

In the distance range from 200-1500 km from the epicentre of an earthquake the pattern of behaviour varies noticeably from region to region. In the western United States there is a rapid drop in short period amplitude with distance with a strong minimum around 700 km (Helmberger, 1973) which seems to be associated with a significant wavespeed inversion in the upper mantle. Such a pattern is not seen as clearly in other regions, and for shield areas there is little evidence for an inversion. In general, however, the coverage of seismic stations in this distance interval is somewhat sparse and the structure of the top 200 km of the mantle is still imperfectly known.

At moderate ranges the character of the wavetrain is still similar to the behaviour we have seen for local events. In figure 8.6 we show a plot of the digital short-period channel of the SRO station at Mashad, Iran for a small earthquake at a range of 390 km. A very clear Pn phase is seen preceding the Pg phase, but once again it is difficult to pick the onset of the Sn phase, though there is the beginning of an apparent interference effect near the expected arrival time. The Sg waves grade at later times into a longer period disturbance composed of higher mode surface waves, the Lg phase. There is no clear distinction
between the $S_g$ and $L_g$ phases and so, to synthesise seismograms for this distance range, we again need to include as much of the response as possible. For the $S$ waves and their coda, this may be achieved by the modal summation techniques discussed in Chapter 11, in addition to the integration methods discussed in Chapter 7.

At longer ranges, the higher mode surface waves are only rarely seen on short-period records, but show up clearly with broad-band instruments. Figure 8.7 illustrates such a record at 980 km from a nuclear test in Nevada. The explosive source gives stronger $P$ waves than in the previous earthquake examples. The $P$ waves show high frequency effects modulating the long-period behaviour ($PL$) which has been studied by Helmberger & Engen (1980). Although $S_g$ is not very strong, a well developed $L_g$ wavetrain is seen. At the same range long-period earthquake records are dominated by fundamental mode Love and Rayleigh waves, which in areas with thick sedimentary sequences can have significant energy at very low group velocities.

### 8.3 Body waves and surface waves

Beyond about 1500 km from the source the $P$ and $S$ body waves are sufficiently well separated in time that we can study them individually. Out to 9000 km the earliest arriving waves are reflected back from the mantle, beyond this range the effect of the core is very significant. Reflected waves from the core $PcP$, $ScS$ arrive just behind $P$ and $S$ between 8000 and 9000 km. For $P$ waves the core generates a shadow zone and there is a delay before $PKP$ is returned. The $P$ wavespeed in the fluid core is higher than the $S$ wave speed in the mantle and so, beyond 9200 km, the $SKS$ phase penetrating into the core overtakes $S$. Simplified travel-time curves for the major phases seen on seismograms are illustrated in figure 8.8.

With increasing range the surface reflected phases such as $PP$, $PPP$ separate from the $P$ coda to become distinct arrivals. With each surface reflection the waves have passed through a caustic, and so the waveform is the Hilbert transform of the previous surface reflection. For $S$, such multiple reflections constitute a fair part
8.3 Body waves and surface waves

Figure 8.8. Simplified travel-time curves for the major seismic phases.

of what is commonly characterised as the surface wavetrain and so the $SS$ and $SSS$ phases appear to emerge from the travel-time curve for the surface waves.

8.3.1 Body waves

For epicentral ranges between 1500 and 3500 km the $P$ and $S$ waves are returned from the major transition zone in the upper mantle which occupies the depth interval from 300-800 km. In this region there are substantial wave speed gradients and near 400 km and 670 km very rapid changes in wavespeeds which act as discontinuities for large wavelengths. This complicated wavespeed distribution leads to travel-time curves for the $P$ and $S$ phases which consist of a number of overlapping branches, associated with variable amplitudes. As a result of interference phenomena the waveforms in this distance range are rather complex
and have mostly been studied in an attempt to elucidate the wavespeed distribution in the upper mantle (see, e.g., Burdick & Helmberger, 1978).

Beyond 3500 km the $P$ and $S$ waves pass steeply through the upper mantle transition zone and very little complication is introduced until the turning levels approach the core-mantle boundary. The resulting window from 3500-9000 km enables us to use $P$ and $S$ waveforms to study the characteristics of the source. For $P$ waves, the character of the beginning of the wavetrain is determined by the interference of the direct $P$ wave with the surface-reflected phases $pP$ and $sP$. The relative amplitude of these reflected phases varies with the take-off angle from the source and the nature of the source mechanism.

In figure 8.9 we show short-period records from a number of WWSSN stations for a shallow event in Iran. The stations lie in a narrow range of azimuths and allow us to see the stability of the direct $P$ wave shape over a considerable distance range. The ISC estimate of the focal depth of this event is 31 km, but the time interval between $P$ and $pP$ suggest a slightly smaller depth. The later parts of the seismograms are associated with crustal reverberations near source and receiver. The relative simplicity of mantle propagation illustrated by these records can be exploited to produce a scheme for calculating theoretical seismograms for teleseismic $P$ and $S$ phases discussed in Section 9.3.3. For teleseismic $S$ waves, a $P$ wave precursor can be generated by conversion at the base of the crust and on a vertical component record this can easily be misread as $S$. 

![Figure 8.9](image)

Figure 8.9. Short period WWSSN records at distant stations from an earthquake in Iran, focal depth 31 km.
8.3 Body waves and surface waves

On teleseismic long-period records the time resolution is normally insufficient to allow separation of the direct \( P \) from the surface reflections \( pP, sP \) for shallow events. However, the appearance of the onset of the \( P \) wavetrain provides a strong constraint on the depth of source particularly when many stations at different distances are available (Langston & Helmberger, 1975). This procedure relies on a very simple construction scheme for long-period records which is discussed in Section 9.3.3. To get depth estimates from these long-period records we need a model of the source time function. For small to moderate size events the far-field radiation can be modelled by a trapezoid in time. Figure 8.10 shows the long-period records from WWSSN stations for an Iranian event as a function of their position on the focal sphere. The simplicity of these long-period waveforms enables the sense of initial motion to be determined very reliably (Sykes, 1967) and so improves the estimate of the focal mechanism.

Once the earthquake focus lies well below the crust the surface reflections are seen as distinct phases, particularly for deep events. Surface reflections can also be returned as core reflections so that phases like \( pPcP, sPcP \) can often be found for intermediate or deep events. In figure 8.11 we show a vertical component broad band recording at Boulder, Colorado from an intermediate depth event (100 km) in
8.3.2 Surface waves

For all but deep earthquakes (focal depths > 300 km) the largest arrivals on long-period records occur after the $P$ and $S$ body waves. These surface waves have travelled with their energy confined to the crust and upper mantle and so have not suffered as much wavefront spreading as the body waves.

On the horizontal component oriented transverse to the path between the epicentre and the station, just behind the $S$ body phases a very long disturbance ($G$) appears which at later times is replaced by short-period oscillations often denoted $LQ$. These two features arise from the fundamental Love mode for which the group slowness normally increases with frequency. As a result the apparent frequency of the record increases with time (see figure 8.12). Superimposed on this wavetrain

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Figure 8.11. Broad-band record at Boulder, Colorado for an intermediate depth event, focal depth 100 km, epicentral distance 4200 km.
8.3 Body waves and surface waves

are smaller high frequency waves with small group slowness. These are higher mode Love waves whose excitation increases with increasing depth of focus.

On the vertical component and the horizontal component oriented along the path to the source, the principal disturbance $LR$ occurs some time after the commencement of the $LQ$ waves and arises from the fundamental Rayleigh mode. This is preceded by very long-period waves, arriving after the $G$ waves and with much lower amplitude. The wavetrain leads up to an abrupt diminuation of amplitude, followed by smaller late arrivals with high frequency. This Airy phase phenomena is associated with a maximum in the group slowness for the fundamental Rayleigh mode at 0.06 Hz (figure 11.10). Waves with frequencies both higher and lower then 0.06 Hz will arrive earlier than those for 0.06 Hz, but the frequency response of long-period instruments reduces the effect of the high frequency branch. The late arrivals following the Airy phase $Rg$ arise from scattering and sedimentary effects.

In figure 8.12 we illustrate long-period seismograms from the WWSSN station at Atlanta, Georgia for shallow events (focal depth 25 km) in the northern part of Baja California, with an epicentral distance of 3000 km. The path is such that the North-South component (figure 8.12a) is almost perfectly transverse and so displays only Love waves, whereas the East-West component is radial and shows only the Rayleigh wave contribution (figure 8.12b). The vertical component is illustrated in figure 11.9. There is a clear contrast between the Love and Rayleigh
waves; the Love waves show lower group slownesses but do not have the distinctive Airy phase. Higher frequency higher mode Rayleigh waves can often be seen in front of the main LR group and a good example can be seen in the broad-band Galitsin record shown in figure 1.1.

In oceanic regions, Rayleigh waves are affected by the presence of the low wavespeed material at the surface. The group slowness of the Rayleigh waves increases at higher frequencies when the wavelength is short enough to be influenced by the presence of the water layer. As a result the group slowness curve is very steep between frequencies of 0.05 and 0.1 Hz. This leads to a wavetrain with a clear long-period commencement followed by a long tail with nearly sinusoidal oscillations which is well displayed in figure 8.13.

In Chapter 11 we will discuss the dispersion of Love and Rayleigh waves and the way in which these influence the nature of the surface wave contribution to the seismograms. The dispersion of surface wave modes is controlled by the wavespeed structure along their path, whilst the excitation of the modes as a function of frequency and azimuth depends on the source mechanism. We are therefore able to achieve a partial separation of the problems of estimating the source properties and the structure of the Earth. The dispersion information can be inverted to give an estimate of the wavespeed distribution with depth and then with this information we have a linear inverse problem for the source moment tensor components.

8.4 Long range propagation

Beyond 10000 km P waves are diffracted along the core-mantle boundary and so their amplitude drops off with distance; this effect is particularly rapid at high frequencies. This leaves PP and core phases as the most prominent features on the early part of the seismogram.

The pattern of seismic phases is well illustrated by figure 8.14, a compilation of long-period WWSSN and CSN records made by Müller & Kind (1976). This event off the coast of Sumatra (1967 August 21) has a focal depth of 40 km. The focal
8.4 Long range propagation

Figure 8.14. Vertical component seismogram section of an earthquake near Sumatra, as recorded by long period WWSSN and CSN stations. The amplitude scale of all traces is the same (after Müller & Kind, 1976).
mechanism has one $P$ wave node nearly vertical and the other nearly horizontal; this leads to strong $S$ wave radiation horizontally and vertically.

The phases on the vertical component section in figure 8.14 mirror the travel-time curves in figure 8.8. The radiation pattern is not very favourable for the excitation of $PcP$ at short ranges, and near 9000 km there is insufficient time resolution to separate $P$ and $PcP$. The core reflection ScS is also obscured, but now because of the train of long-period waves following $S$. This shear-coupled PL phase arises when a $SV$ wave is incident at the base of the crust in the neighbourhood of the recording station at a slowness close to the $P$ wave speed in the uppermost mantle. Long-period $P$ disturbances excited by conversion at the crust-mantle interface then reverberate in the crust, losing energy only slowly by radiation loss into $S$ waves in the mantle. Such PL waves are associated with $S$ and its multiple reflections, and as in figure 8.14 can be the largest body wave phases on the record. No such effect occurs for $SH$ waves and so $S$ and ScS can be separated on the transverse component.

Around 9200 km SKS begins to arrive before $S$ but can only just be discerned on figure 8.14. However, the converted phase PKS is quite strong and appears just after PP. The shadow zone caused by the core gives a couple of minutes delay between diffracted $P$ and PKP. Near 15600 km there is a caustic for the PKP phase associated with very large amplitudes and this shows up very clearly on figure 8.14.