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# Explosion Seismology

Capabilities and limitations of long-range methods for detecting and recognizing explosions are discussed.

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Take a good look at Fig. 1. The traces you see were recorded at distances of between 3000 and 10,000 kilometers from seismic events and represent the variation, with time, of vertical ground movement in a rather narrow frequency band between about 1 and 2 cycles per second. Three of the records are from underground nuclear explosions, the rest are from, I trust, earthquakes. The problem is to distinguish between the explosion and the earthquake records. Numbers 1, 7, and 9 are explosion records.

The problem you face is one which faces Western seismologists at the present day. They, of course, have access to many more records from each event, and the purpose of this article is to indicate ways in which present research is attempting to use such records to provide estimates of the capability of seismic detection systems.

In 1958 the Conference of Experts which met at Geneva proposed a seismic detection network of 180 stations distributed around the world. Since then, however, there has emerged a new approach based on national systems which rely on data obtained from outside the territory of the state in which the event occurs. The intercontinental distances between source and receiver, which such a system implies, provide a considerable number of geographical difficulties in the organization of controlled experiments. Consequently, the basic data from which system-capability estimates have to be made have been accumulated rather slowly, and considerable extrapolation is required.

In this article we examine some of the problems and techniques which arise with a national system. The seismological problems are those of detection, location, and identification, which, when taken together, form a mutually consistent scheme. Thus the seismologist can say when he has detected an event and where, within appropriate limits, it occurred. Although many earthquakes can be positively identified as such, it is not at present possible positively to identify explosions. However, given the proposition that an event could be an explosion, the seismologist can assess the probabilities. Note that he has to ignore anything he knows about relative numbers of earthquakes and explosions. He will also be asked: If it was an explosion, what was the yield? I now proceed to discuss these various aspects in more detail.

## Detection Problem and Definition of a Teleseismic System

A seismic event can be characterized by its magnitude—a position, on an arbitrarily defined scale, based on the size of the seismic signal it produces. If an event has been detected at enough stations to give even a crude estimate of its position, then each station can provide a magnitude according to the equation

$$m = \log_{10}(A/T) + B(\Delta)$$

where $A$, in microns, is half the peak-to-peak amplitude of the largest pulse within the first three or four cycles of the record; $T$, in seconds, is the apparent period of the pulse; and $B$ is a function of distance $\Delta$. The variation of $B$ with $\Delta$ and the corresponding (average) variation of amplitude at $T = 1$ sec for an event of $m = 4.5$ are shown in Fig. 2.

A magnitude may be ascribed to any event by taking the average of the individual magnitude determinations from stations at a variety of azimuths and distances. Few analyses of magnitude scatter have been made, although the consequences are of fundamental importance for the evaluation of the capabilities of a detection system. Part of the scatter results from the natural asymmetry of earthquakes. For explosions the asymmetry is minimized, yet Fig. 3 was prepared from explosion data. That there is scope for argument about magnitudes derived from different samples of records obtained at different azimuths and distances by different instruments needs no emphasis.

The character of a seismic record changes very considerably over the distance range shown in Fig. 2. When we concentrate on the first few seconds of the record we are considering only the compressional, or P, waves. Within an angular distance of, say, 10° of the source, in the so-called first zone, where the amplitude is decreasing rapidly with distance, there are many P-wave phases due to waves traveling complicated paths in the earth's crust. There are also many shear-wave, or S-wave, phases and surface- and

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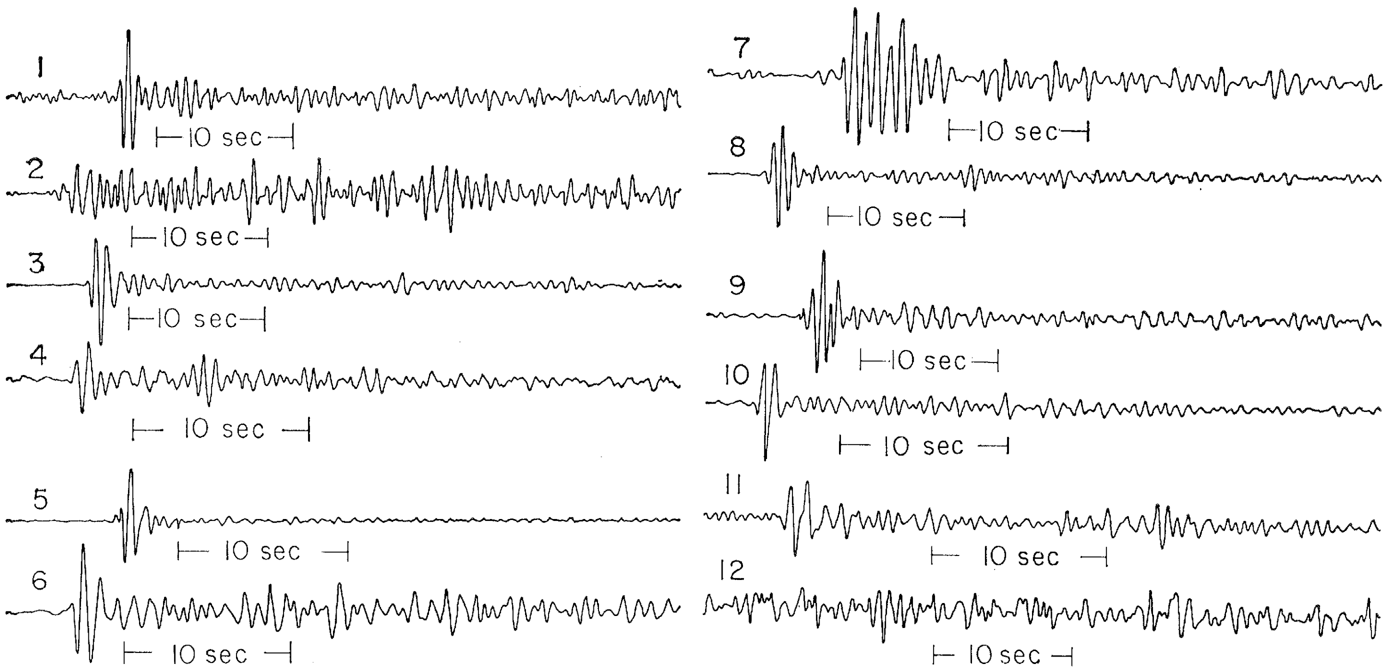



Fig. 1. A typical, but by no means fully representative, selection of teleseismic records.

guided-wave phases. Thus the record contains a lot of character, but most of it is governed by the propagation paths rather than by the source. Between 10° and about 25° of the source there is often a shadow zone, where the signals are small and diffuse. Not very much is known about this region, most of the available information coming from earthquakes in western North America and from explosions at the Nevada test site. Since contours of geophysical parameters seem to have a bad habit of circling Nevada, extrapolations to other regions are always a bit suspect. In the absence of relevant global data it seems safer to assume a shadow zone and accept any benefits that its absence in specific regions might imply. The most promising range is the so-called third zone, which begins at the end of the shadow zone and ends at a distance from the source of about 90°, where the characteristic upward refraction of energy due to the increasing velocity with depth within the earth is stopped by the earth's core.

In this third zone the seismic P-wave signals stand in their simplest relation to the waves emanating from the event. Thirlaway has coined the phrase "the seismic window" for the third zone, since it is through this window that we most clearly "see" the source. Both the shear and the surface waves are well separated from the P-waves in both frequency and time, although shear and surface waves from events in the magnitude range of interest—of say, 3.5 to 5—largely go undetected. Thus it is to the P-waves that we must turn for a teleseismic system based on stations remote from the seismic events, examining the P-wave character in the hope of finding parameters which distinguish earthquakes and explosions.

To complete the detection picture we have to consider the ambient noise which is always present. A very reasonable requirement for detection is that

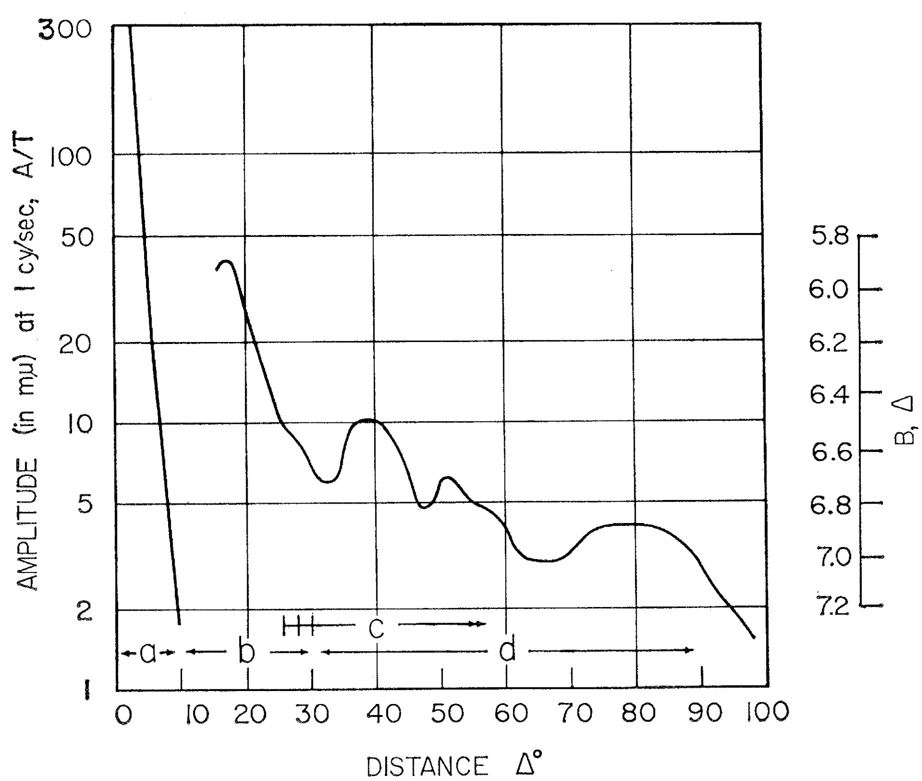



Fig. 2. Variation with distance of the average signal from a seismic event of $m = 4.5$. *a*, "First" zone; *b*, shadow zone for western United States and probably other areas; *c*, teleseismic zone; *d*, source window.

the signal should be twice the noise level, but to complicate matters the noise varies both from station to station and with time at each station. Figure 4 shows a typical method of presenting noise as the probability that, in a randomly chosen period of 2 minutes, the amplitude of the seismic noise will not exceed a specified level. Other noise presentations give the root-mean-square value, the peak value, or the average value, but it is often difficult to get strictly comparable data.

To be specific, suppose we want a system containing a relatively small number of stations, say about 20 to 30. Then we could probably find at least three or four sites where the peak noise amplitude at a frequency between 1 and 2 cycles per second is usually below 1 millimicron (there are seasonal variations) and several where it is below 2 millimicrons. Apart from island sites, introduced to cover areas like the Pacific, the remaining stations should have noise amplitudes below 5 millimicrons. At each station it is possible to employ noise-reducing techniques, such as putting seismometers down a borehole or employing a large number of seismometers suitably spaced on the surface as an array, and these techniques are in principle capable of providing improvement in the noise level by an order of magnitude without degrading the signal.

Once a system has been postulated it is a straightforward statistical problem to take the station distribution, the noise statistics for each station, an appropriate signal-to-noise requirement for detection, and a magnitude distribution curve such as that of Fig. 3 and to obtain estimates of the probability of detecting events of given magnitudes from a specific region at a specified number of stations. It would appear that with a global system of 20 to 30 stations, virtually all events of magnitude 4 and above should be detectable with a signal-to-noise ratio of 2 at a minimum of seven teleseismic stations. The important thing is the expectation of an *average* number of detections, ranging from, say, one at magnitude 2.5, through three at magnitude 3, to at least seven at magnitude 4 and above. For the larger events, extra information would accrue from the present worldwide system of seismological observatories, which, although individually autonomous, have a long-established custom of international cooperation.

## Source Location

When a seismic event has been detected at several stations the location of the source becomes a problem of prime importance. There are two reasons for this. First, the proof positive of a treaty violation would be the discovery of radioactive debris by an on-site inspection team, who obviously would require their search area to be closely defined. Secondly, accurate location is a valuable aid to earthquake identification.

Four parameters can be determined for each event—its latitude and longitude, which together define the "epicenter"; its depth; and the time at which it occurred. The method of determining these parameters is straightforward, and the use of modern computing machines has taken out the hard work. All that is required is knowledge of the arrival times of the P-waves at a number of accurately located stations and a set of tables giving travel times, as a function of distance. Such tables have been constructed from the earthquake data accumulated over the years. The method is as follows: assume a set of parameters, calculate the distance from the assumed epicenter to each station, obtain the travel times from the tables and add them to the assumed time of origin of the event. Then compare these computed arrival times with the actual arrival times. By a series of successive approximations the parameters are adjusted until the "best" solution is obtained—namely, that when the sum of the squares of the differences between observed and calculated arrival times is minimized.

In discussing the accuracy of a location it is convenient to treat location of the epicenter separately from the

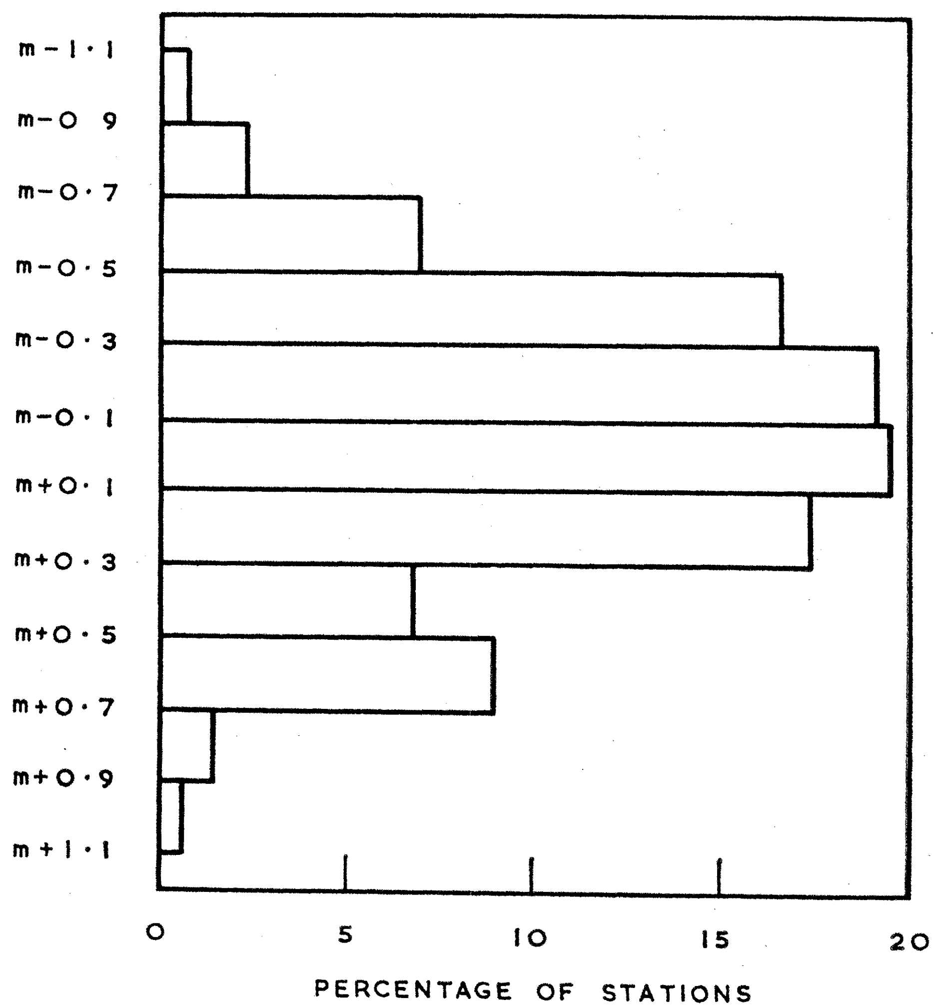



Fig. 3. Teleseismic magnitude scatter for typical explosions, from U.S. data. Earthquake magnitudes would probably be more scattered, as would explosion magnitudes recorded on the worldwide system. Some of the scatter probably reflects inherent variations in station sensitivity, and statistical analysis to determine correction factors for individual stations is required. Differences in magnitude scatter could prove to be of value as a diagnostic aid.

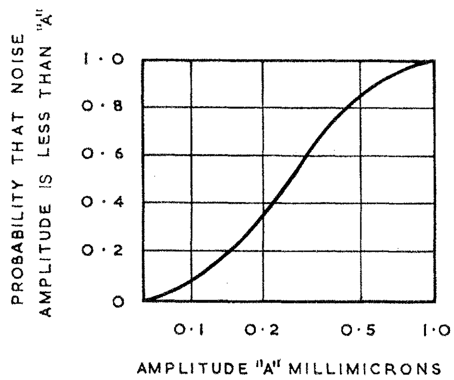


Fig. 4. A typical noise-probability curve.

determinations of depth and origin time. The foremost requirement for accurate location of the epicenter is a good (azimuthal) distribution of stations around the event. The distance of these stations from the event is not particularly important, although there is evidence, in fact, that epicenters are best determined from stations at angular distances of between 20° and 90°. The reason for this is that at shorter distances the first wave arrival has traveled entirely in the crust or upper mantle, where there are regional velocity variations. Unless regional correction factors are used to compensate for these variations, the consequent errors can put the true epicenter well outside the 500-square-kilometer area which has been suggested as the maximum area to be searched for radioactive debris following location of a suspicious event. The only method of checking location techniques is to apply them to known explosions and to those relatively rare earthquakes whose location is accurately known from very local data. Herrin (*1*) has made an exhaustive study of the problem and has concluded that with seven teleseismic stations well distributed around an event it should be possible to define areas of about 250 and 500 square kilometers for which the probabilities that one can find an epicenter located within these areas are 50 and 75 percent, respectively. As the number of stations is reduced below seven the corresponding areas increase rapidly until the process becomes impossible with less than three recordings, while with increases above seven in the number of stations the accuracy increases until a limit of something like 100 square kilometers for the size of the search area is reached.

The next step is an application of modern computing facilities to process the high-quality data from the worldwide system of standard stations set up under the Project Vela program. This should result in the production of an improved set of travel-time data and should make it possible to determine both local station corrections and small regional corrections of the same order of accuracy as the record readings. Jeffreys (*2*) has, characteristically, led the way with a refined statistical study of data from Pacific explosions. He found no evidence that the teleseismic travel times of P-waves were a function of anything other than distance from the epicenter, and he confirmed the findings of Gutenberg (*3*) and others that all the travel times should be reduced by about 2 seconds, a refinement which results directly from having independent data for the origin times of U.S. (and, so far, only U.S.) explosions.

Determination of the depth and the origin time of an event from the arrival times of P-waves is much more difficult. An essential requirement is that stations be well distributed in distance. It is easy to see why. Suppose we have a circle of stations with radius, say, 30°, and suppose that all the stations record an arrival at the same time. Then we could say that the source was at the center of the circle, but we would not know whether the event occurred at the surface at time $T$ or at depth $h$ an appropriate time later. For locating the epicenter this problem is trivial: just assume that the event took place essentially at the surface, where, by definition, it would

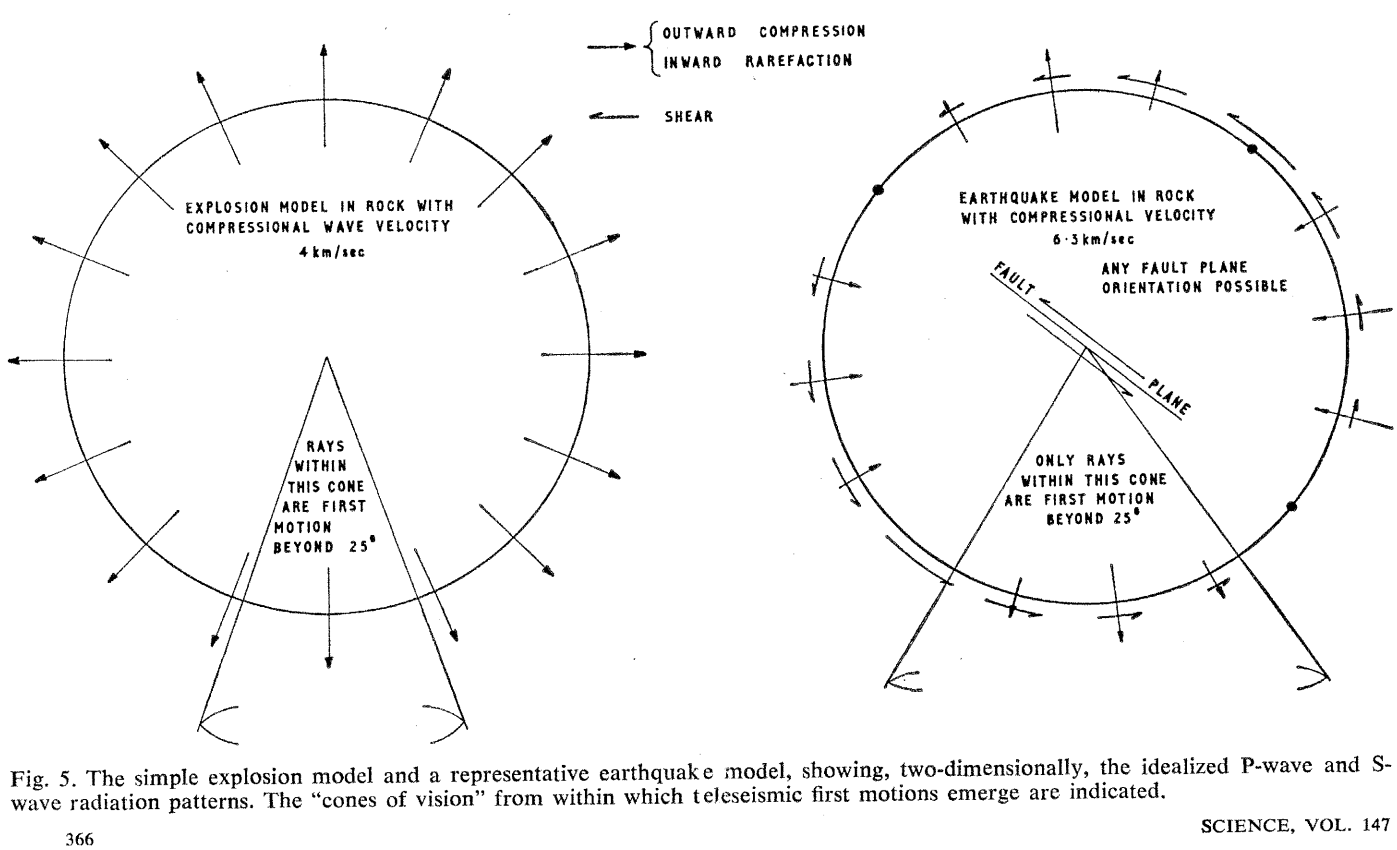


Fig. 5. The simple explosion model and a representative earthquake model, showing, two-dimensionally, the idealized P-wave and S-wave radiation patterns. The "cones of vision" from within which teleseismic first motions emerge are indicated.

occur if it were an explosion. However, the preceding sentence has a corollary—namely, that if an event takes place at considerable depth, say greater than 5 kilometers, it cannot be an explosion. Thus, determination of depth is important as an identification criterion.

Unfortunately, past experience with depth determinations from the arrival times of P-waves has not been very encouraging. The balance between depth and origin time is so fine that quite often earthquakes appear to originate above the surface! Clearly this is not satisfactory, and seismologists have tended to obscure the issue by labeling such events as shallow. Improvements are now being made in the computer programs, principally to work into the theory the premise that earthquakes must by definition occur within the earth and also to provide some estimate of the accuracy of the determinations. In particular the probability that a specific located event occurred at a depth greater than 5 kilometers can be explicitly provided.

Since determination of the depth of focus is such a potentially powerful discriminant, attempts have been made to utilize another technique, which in fact has historical precedence over the travel-time method. When an event occurs in the earth, the first P-wave arrival at teleseismic distances is the wave which leaves the event more or less vertically downward. In principle this is followed by the wave, designated pP, which goes up to the surface as a P-wave and is reflected down as a P-wave, and this in turn is followed by the wave, designated sP, which goes up to the surface as an S-wave and is reflected down as a P-wave. If these waves can be recognized, then the time separating them gives an accurate estimate of the depth; the problem is in recognizing them. Often the reflections are small and difficult to resolve above the noise, while in addition, and particularly for near-surface events, the separations are insufficient for the various phases to be clearly distinguished. No details of the effectiveness of different methods of determining depth of focus have been published, although it seems implicit from the congressional hearings (*1*) that events from depths greater than 60 kilometers are generally recognizable as such, either from analysis of arrival times or from surface reflections. Now we turn from the location of an event to a study of its character.

**Source Characteristics**

There has always been a possibility that the character of the source might impose itself on the seismic records sufficiently strongly for identification criteria to be evolved. There is in seismology a widely held concept that an earthquake consists of shearing motion along a fault plane (*4*), and this, at the risk of oversimplification, we will take as a convenient starting point. In Fig. 5 the P-wave starting out from an explosion and the P-wave starting out from an earthquake are represented. Whereas the initial motion from an explosion is outward, an earthquake exhibits four quadrants which alternately give outward (compressional) and inward (rarefactional) first motion. Now the first motion has the rather unique property of carrying its char-

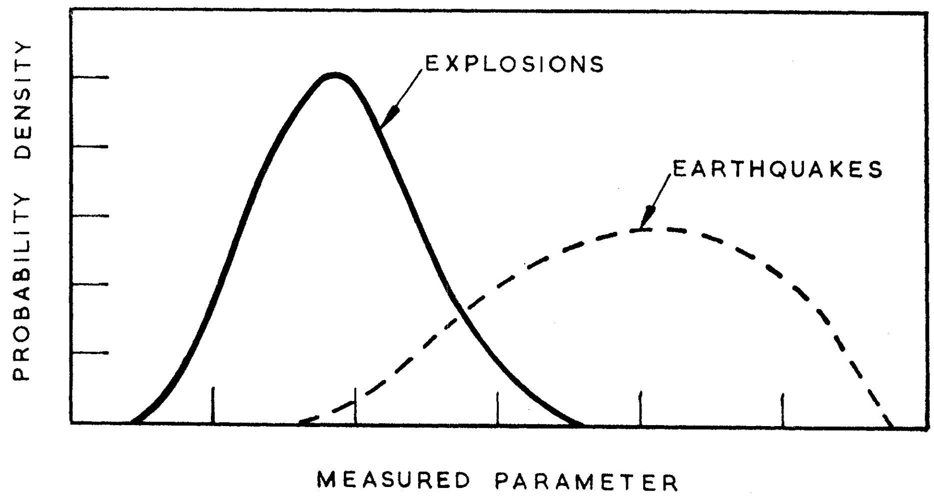



Fig. 6. Diagram showing the principle of overlapping distributions of shallow-earthquake and explosion properties (for example, surface-wave energy divided by body-wave amplitude).

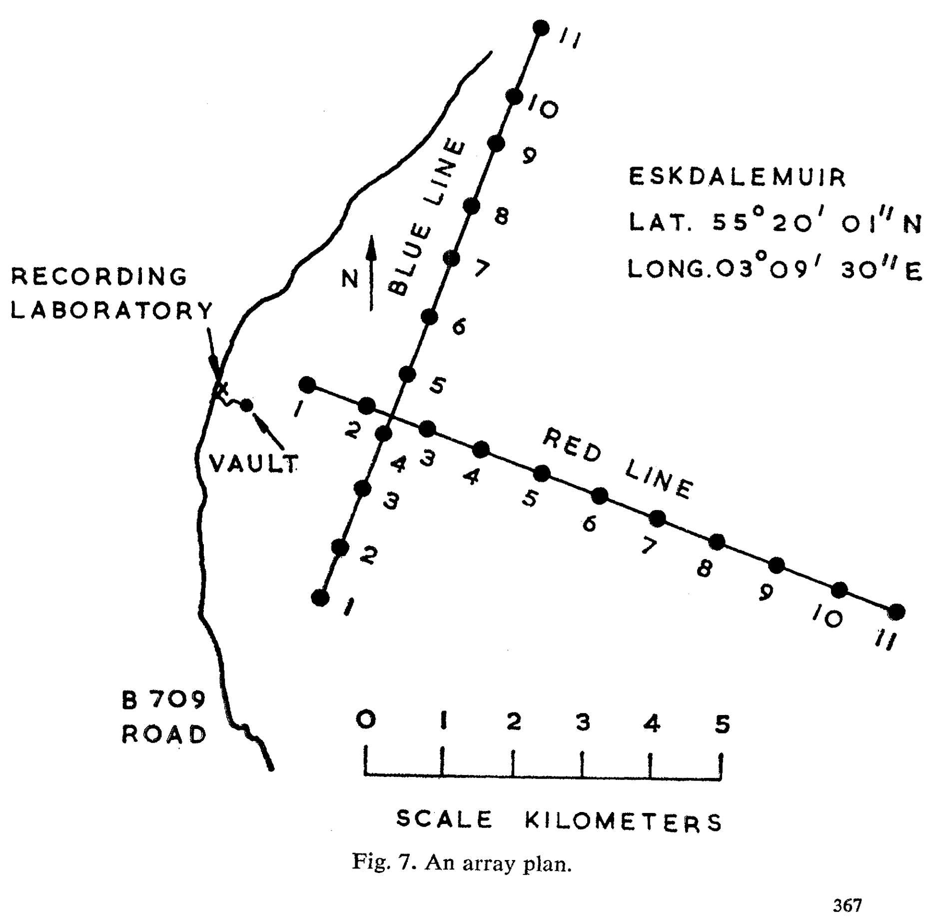



Fig. 7. An array plan.

acter all the way to the recording station, so that positive evidence of rarefactional first motions is conclusive evidence of an earthquake. Unfortunately this method of identifying earthquakes is not particularly helpful, for two reasons. First of all, the first motion tends to be very small relative to the maximum signal used for detection. Thus, whereas a (maximum) signal-to-noise ratio of 2 seemed a reasonable value for detection, a ratio of 10 may be necessary for confident recognition of the first motion as such.

The other difficulty concerns the "cone of vision." A system of stations only "sees" as first motion those rays which leave the source within a vertical cone (see Fig. 5). For stations at angular distances of up to 10°, the cone's semiangle, $\theta$, is greater than 45°, so that, if the number of stations is sufficient, rarefactional first motions are, in principle, always detectable. For stations at teleseismic distances, $\theta$ falls below 45° and there is consequently a finite probability that all first motions are compressive. A full evaluation of the probabilities of identifying earthquakes by first motion will require a considerable extrapolation of existing earthquake data, but it is clear that its applicability is limited.

As a positive method of earthquake identification the first-motion criterion occupies a unique position. We now consider less positive methods, which still have an important role to play. It is inconceivable that the mechanisms of earthquakes and explosions should not have more differences than a difference in first motion. What we have to ask is: What are these differences? Have they observable effects? How separable are they?

Several promising techniques immediately applicable at distances up to a thousand kilometers from the event have been investigated (5), with varying degrees of success. Probably the most promising has been the discovery that earthquakes and explosions which generate the same degree of compressional waves often generate quite different amounts of shear- and surface-wave energy. This difference is represented in Fig. 6, a figure which, to me, seems of quite outstanding significance. It shows an overlapping distribution of the properties of explosions and earthquakes. No longer are we faced with a clean-cut decision such as, for instance, that a rarefaction indicates an

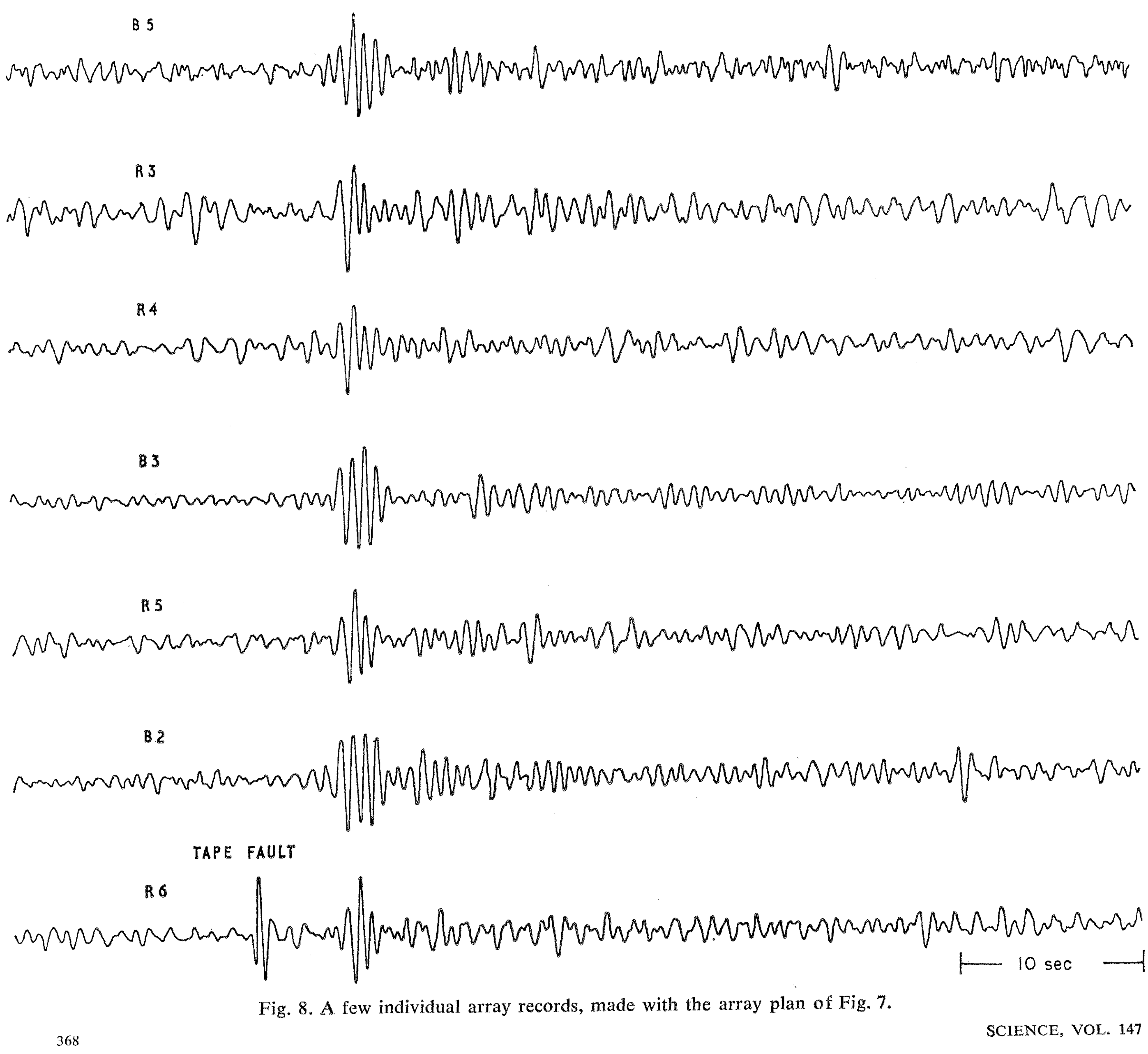



Fig. 8. A few individual array records, made with the array plan of Fig. 7.

earthquake; we are faced with a statement in terms of probability. By means of established statistical techniques, measurements of several different parameters, each having its own independent statistical distribution, can be combined to provide a probability that a specific event is an explosion. Attractive as this method appears, it must be remembered that the basic statistical data can only be acquired from actual events. The number of events at the Nevada test site has been ample for such studies, but how are the characteristics of the explosions and those of the Nevada site itself to be separated? One way of removing this limitation of the empirical approach is to provide a theoretical basis. For instance, the relation between compressional and surface waves, for explosions, seems in reasonable accord with simple theoretical considerations. If this is confirmed, our confidence in the technique is increased for all regions, irrespective of empirical evidence.

A number of teleseismic "diagnostic aids" and associated measurement techniques have been investigated at the United Kingdom Atomic Weapons Research Establishment, and I now devote a section to describing them.

## The Influence of Arrays

Consider again the simple picture of an explosion in a homogeneous rock. To a reasonable approximation, the explosion pushes the surrounding material outward and generates a very simple seismic signal. No matter from what direction we look at the source, the signal is the same, and we can deduce the motion at the source from it. In the actual earth the explosion takes place near a free surface in a more or less complicated crust. The teleseismic signals we record leave the crust within a small cone (generally smaller than in the case of earthquakes), are refracted upward by the velocity gradient in the mantle, and emerge from the mantle through another crustal section. Now the picture we have of the earth's mantle indicates that it would have little effect on the signal other than that of absorbing some of the high-frequency energy. The free surface above the explosion results in a reflected signal as large as the direct signal, but the time delay between the two signals is only of the order of a second, so they cannot readily be resolved. Thus we expect signals entering the mantle to be reasonably simple, with the am-

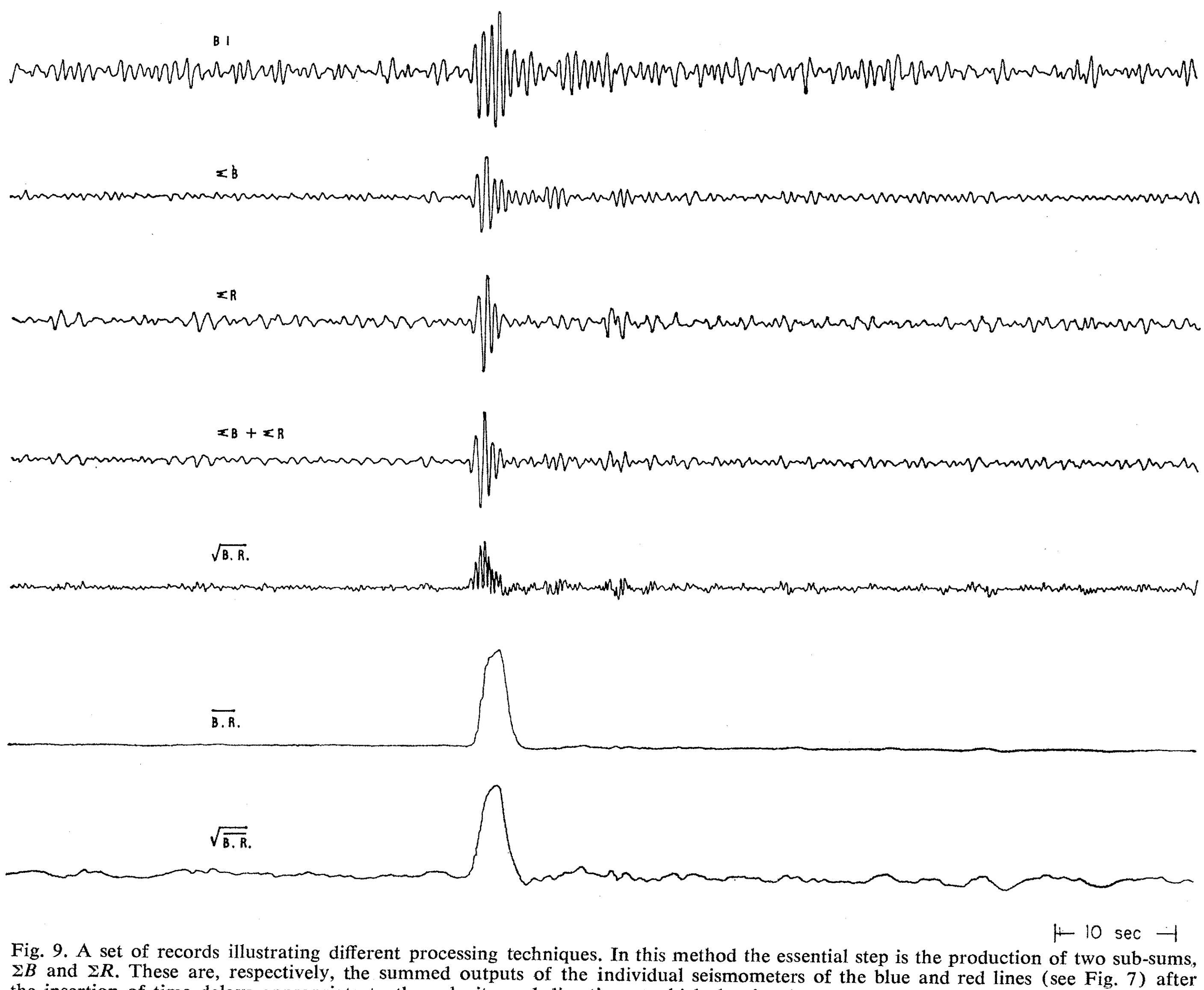



Fig. 9. A set of records illustrating different processing techniques. In this method the essential step is the production of two sub-sums, $\Sigma B$ and $\Sigma R$. These are, respectively, the summed outputs of the individual seismometers of the blue and red lines (see Fig. 7) after the insertion of time delays appropriate to the velocity and direction at which the signal crosses the array. Subsequent operations with these sub-sums include addition or multiplication. In the illustration, multiplication has been followed by smoothing and by taking the square root and smoothing.

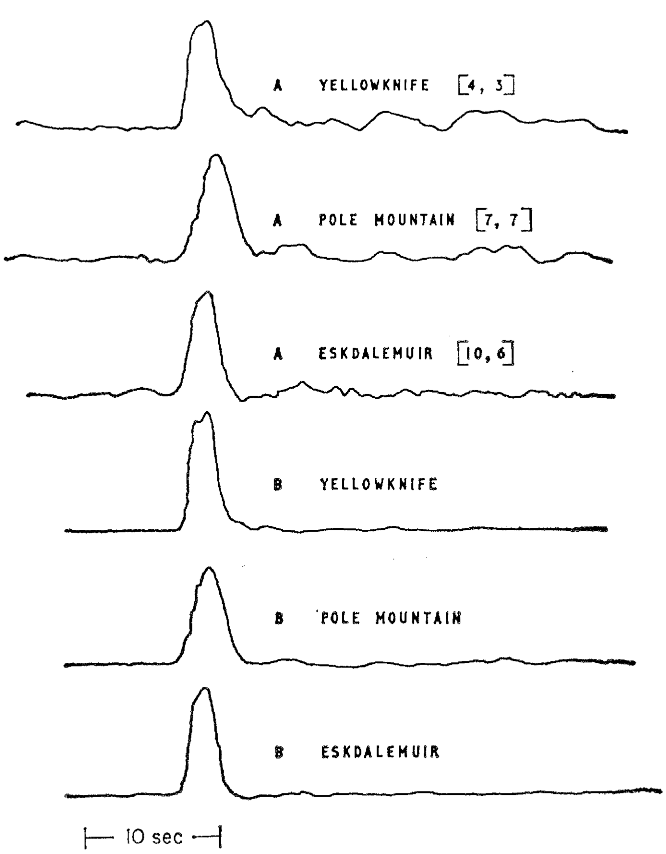
plitude dying rapidly away after a few seconds, either to zero or, if there are reflections within the crust, to levels of about one-tenth the peak amplitude.

However, when the simple P-wave signal arrives at the crust near the recorder it can create other signals whose amplitude and character depend upon the nature of the crustal inhomogeneities, such as faults or relief features. With a single vertical seismometer there is no way to differentiate the various parts of the signal, nor indeed can interfering signals from different events be distinguished. Three seismometers, two horizontal and one vertical, provide some improvement because they make it possible to distinguish particle orbits, but arrays of seismometers offer a very much greater capacity for resolving signals which have different velocities over the ground, which come from different directions, or which differ both in velocity and in direction.

There are many types of array, one of the simplest being the linear cross array, which consists of two lines of equally spaced seismometers, as shown in Fig. 7. The design is quite flexible, and the principle of operation is as follows. A signal from a distant event arrives at slightly different times at the various seismometers, and the time differences relative to the central point can be either measured or predicted. The original signals are all recorded on a multichannel magnetic tape, so that, with conventional analog or digital techniques, the appropriate time differences can be inserted to align the signals. The records are then added together so that the required signal is reinforced in proportion to the number, $N$, of the seismometers. If the noise is random, then on adding the records it increases on the average by $N^{1/2}$, with a net increase in the signal-to-noise ratio of $N^{1/2}$. Signals, including coherent noise, arriving from different directions or with different velocities will also tend to cancel in a way which can be predicted for each specific array.

The process is best illustrated by an example. Figure 8 shows the individual channels, and Fig. 9 shows their sum after the appropriate time delays have been inserted, together with a number of processed outputs. How you process the outputs is again a matter of choice; filtering is very important, and splitting the signals into two groups, multiplying, and smoothing the product has been found useful. The example in Figs. 8 and 9 is, in fact, the record from an underground nuclear explosion, and Fig. 10 shows the corresponding outputs from three arrays which employ identical instrumentations. These examples are typical of the (admittedly few) teleseismic underground explosions recorded at these sites and show a remarkable degree of similarity.

Explosion records from more limited arrays also tend to support this picture of the simplicity and similarity of explosion records. Nevada explosions have tended to produce somewhat more complicated records than have explosions in other areas, but not more complicated than would be expected from



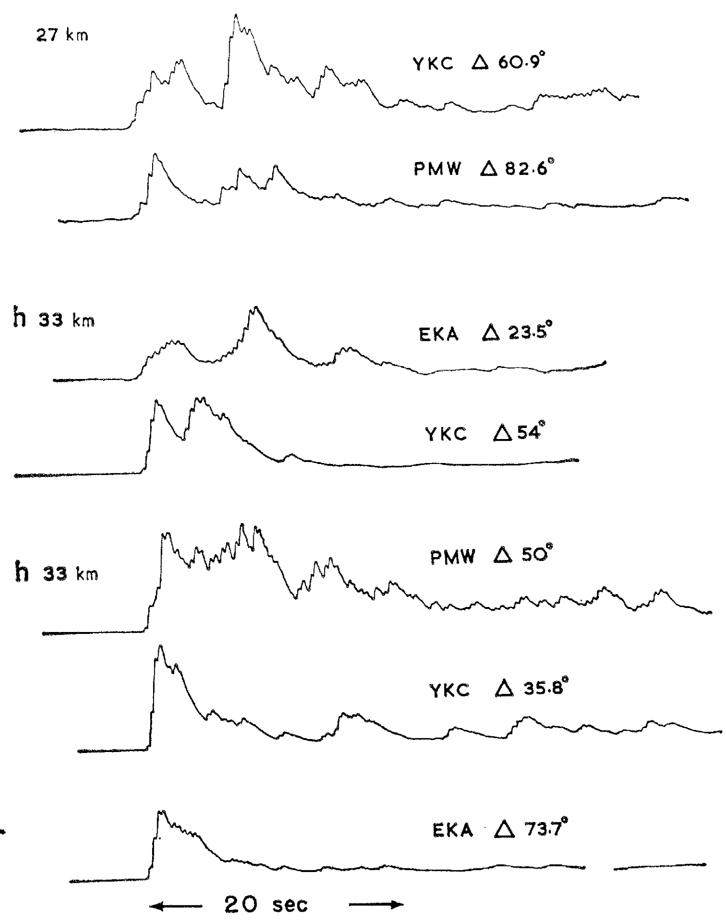




Fig. 10 (left). Three processed array records from an event in southern Algeria. *A*, The separate sums of the blue and red lines after multiplication, taking of the square root, and smoothing; *B*, the separate sums of the blue and red lines, multiplied and smoothed. Figures in brackets are the number of seismometers in each line. Fig. 11 (right). Typical records from three shallow earthquakes recorded at Yellowknife (YKC), Pole Mountain, Wyoming (PMW), and Eskdalemuir (EKA). The summed records of each of the two lines of an array have been summed after the insertion of appropriate time delays. The two summed records have then been multiplied and smoothed. The traces may be compared directly with the *B* traces of Fig. 10.

the geology, and the similarity between different records from the same event has remained high. Many records from single stations lend further support to the picture and, in the absence of positive evidence to the contrary, those records which do not can be assumed to be from "poor" sites, where there are complicating geological factors. The disturbing possibility is that the effects may be reciprocal in the sense that an explosion at a "poor" site could give complicated records at the distant "good" stations. In general, arrays of seismometers would be incapable of "cleaning" the records from explosions at such sites, but the complicated records would probably still be similar. This sort of effect would be expected to occur with an underwater shot in the deep ocean. The reflection coefficient is large at both the ocean floor and the free surface, so that a series of simple signals of slowly decreasing amplitude and constant time separation is transmitted into the mantle. From good sites the teleseismic signals would be expected to be virtually identical at all azimuths and ranges, and the available evidence certainly tends to confirm this view. The arguments apply equally to records from several shots fired at the same site.

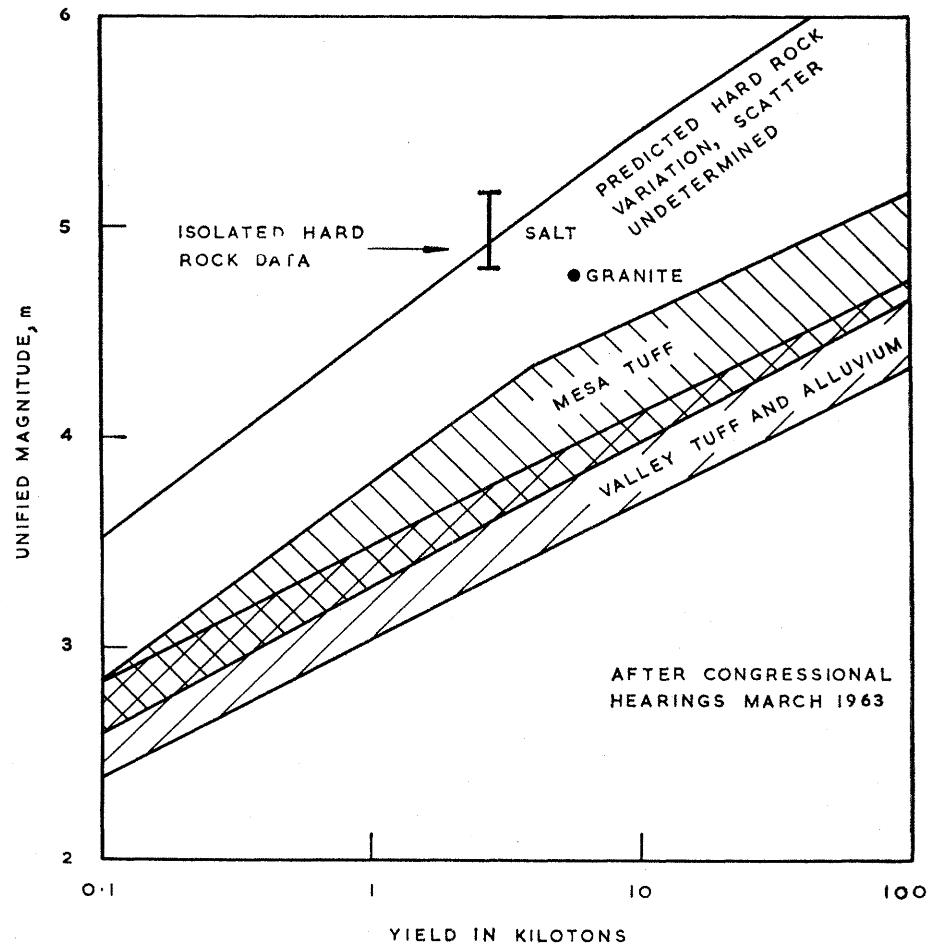



Fig. 12. Magnitude-versus-yield data derived from U.S. data presented before congressional hearings, 1963.

Consider now an earthquake within the crust. Most studies suggest that the majority of earthquakes occur at depths greater than 5 kilometers, but very little is known about the actual displacements they produce. What we would like to do is to draw a sphere around the event and define the motion on it. It seems clear that some earthquakes occur in a way that, pictorially at least, can be represented by prolonged movement along a fault plane (or along several faults in the same region), giving records which last for a long time. Reflection of the waves from the free surface also extends the records, by a time proportional to the depth. The large shear waves generated by earthquakes (see Fig. 5) also enter the picture, for they can create compressional waves by reflection in the crust. The conversion factors are small, because of the small angles of incidence, but the waves are often large enough to compensate for this, and since S-waves travel at a lower velocity than P-waves, the time scale is appropriately extended. In the most favorable situations, P, pP, and sP waves are distinguishable, and a depth of focus can be determined, but in general there results a complex extended P-wave train with the effects of complicated source motion and crustal reflection intermixed. However, one would expect not only complexity of individual records but also a lack of similarity between them because of the asymmetry of the source itself. To date, most of the shallow-earthquake data from linear arrays seem to fit the above picture of overall complexity. Figure 11 shows typical records for three shallow earthquakes recorded at array stations. In each case the records from each line of seismometers have been summed after the insertion of time delays. The summed records from each line have then been multiplied and smoothed so that the resulting traces can be compared directly with the *B* records of Fig. 10. The records illustrate both the complexity of earthquake signals and the differences between records of the same event recorded at different stations. Some earthquakes do give records which are simple in appearance, but most of the earthquakes for which the evidence is at all conclusive are deep-focus events. This also we would expect, for the deep earthquakes occur in the relatively homogeneous mantle, well away from the major contrasts of velocity and density which characterize the crust.

The general view that explosion records are characterized by either simplicity or similarity in all directions, whereas earthquake records are both complex in form and different in different directions, is of course very attractive as a diagnostic aid. It does not serve to identify explosions, but it provides a probability whose value depends upon statistics backed by theory. On the basis of present empirical data the discrimination appears to be good, but the data are limited, particularly the data from explosions in seismic areas whose very nature might imply sufficient complexity to complicate this simple picture.

A more objective method of looking for differences between earthquake records and explosion records would be examination of the Fourier spectra of the summed array output. Intuitively one would expect earthquakes to show relatively more low-frequency energy than explosions. One reason is that low-frequency energy from explosions is inhibited by the free-surface reflection. Another possible reason is that S-wave energy tends to have a lower frequency than P-wave energy, so that sP-waves have more low-frequency energy than P-waves—a point which could help in the identification of sP-waves.

Two further developments seem immediately possible. In the first place, array technique can be applied in the detection of surface waves. Data from Nevada explosions, explosion theory, and earthquake observation all indicate that shallow earthquakes generate significantly more surface waves than do explosions of the same magnitude. At present, detection of surface waves is the problem, but there does seem every indication that arrays of long-period seismometers are feasible. The amplitudes of surface waves decrease by only slightly more than the inverse of the distance traveled—an attenuation attributable to both geometric spreading and dispersion. If, as we think, correlation techniques improve detection by an amount proportional to the length of the record, then detection capability decreases only as the square root of the distance traveled, a slow rate of decay which is important if the recording stations are at great distances.

The desirability of using arrays for detecting surface waves from small earthquakes is emphasized by the possibility that, since the surface wave records last for a long time, much of the noise may be composed of the records from other events.

Another discriminant which requires only detection of energy is evidence of shear waves. The method has been applied with some success to records from events within 1000 kilometers, but the conversions of P-waves to S-waves within the crust tend to obscure the source-generated S-waves. At teleseismic distances these conversions are unimportant, and earthquakes and not explosions should produce S-waves. The practical problem is, again, one of signal-to-noise ratio. Shear waves in the period range 1 to 3 seconds probably behave much as P-waves in the range ⅓ to 1 second do, but the noise against which they have to be detected is likely to be higher by a factor of 20. Arrays of horizontal seismometers could well be the answer, and again we would know from the P-wave data just when to look for S-waves. It appears that the noise at periods above 1 second behaves more predictably than the noise at shorter periods, so that S-wave arrays could well provide signal-to-noise improvements very much better than the $N^{\frac{1}{2}}$ appropriate to random noise.

One final point about arrays which has always intrigued me is the possibility of using existing seismological observatories. The resolving power of an array increases with its size, provided the signal character is preserved across it. The separations between the P-wave and phases like pP and sP do vary with distance (deeper events showing more variation), and earthquake P-wave records must change rapidly between points on either side of a nodal line (see Fig. 5). Effects like these, which change the character of the records, impose upper limits on array dimensions, but there is certainly a case for using some of the many sensitive stations in the United States as a national array.

## System Statistics

So far we have considered the properties of earthquakes and explosions simply as an exercise in seismology. In our context, no one is very much interested in the magnitudes of explosions; a magnitude-yield relationship is required. The most comprehensive data available on this come from the congressional hearings and are summarized in Fig. 12. The easiest way of regarding this relationship is to pick a certain value of magnitude—say, $m = 4.0$. Then this can be equated with something between 0.3 and 1.5 kilotons in hard rock, 2 and 10 kilotons in tuff, or 6 and 30 kilotons in alluvium. On the average the amplitude of the seismic signal increases linearly with yield up to some limiting value beyond which it increases rather more slowly; the "harder" the rock, the higher the yield at which this change of dependence occurs.

Figure 12 highlights the very large variation in the yield-versus-magnitude relationship, which creates further uncertainties when we are trying to decide what identification and detection magnitude limits really mean. Also, a decoupling factor of the order of 100 may have to be introduced in connection with the hard-rock figures. Briefly this means that if a weapon were fired in a sufficiently large underground cavity in hard rock, a magnitude-4 event could correspond to something of the order of 100 kilotons.

Several points warrant mention. First, the magnitude-yield relationship is specifically derived from U.S. data, predominantly from shots in Nevada, a region whose unique character has already been alluded to. Some evidence does suggest that teleseismic magnitudes might be higher than first-zone magnitudes, and there is scope for a study of explosion-magnitude data specifically at teleseismic ranges. Unfortunately, most of the sensitive seismic stations are in North America and therefore do not contribute to teleseismic measurements from Nevada explosions, while neither Russia nor France have announced the yields of their explosions. Then, decoupling, although proved beyond doubt as a principle, involves formidable technical difficulties. The "medium decoupling" of alluvium relative to hard rock is well documented, but a possible violator of a test ban must beware the structural weakness of alluvium, which could result in the collapse of explosion cavities giving unmistakable surface evidence.

In the past the numbers of seismic events have been related to explosion yields. Although the step from magnitude to yield is an essential one, the two separate factors—that is, measured numbers of events versus magnitude for any geographical area, and magnitude versus yield for specific conditions—should be clearly recognized. Already the number of earthquakes equivalent to a given yield has been revised by a factor of 2½ without any alteration in the number-versus-magnitude relation. Unless the two factors are clearly separated, considerable confusion could result should either factor be subsequently revised.

As a pertinent example of magnitude and numbers of events, consider the data for Russia. On the average there are each year something like 170 shallow seismic events (defined as events which show no conclusive evidence of originating at depths greater than 60 kilometers) of magnitude 4 or greater (see *1*). For each 1.0 decrease in magnitude the number of events can be multiplied by 8—a seismic rule of thumb which always seems to work. Thus, there would be, on the average, 500 events of magnitude above 3.5 and 20 of magnitude above 5, the annual numbers varying about the mean by something like a factor of 2. Obviously, the higher you push the system sensitivity the larger become the problems of sheer numbers, particularly when each seismometer is detecting 10 times as many events from the rest of the world as from Russia.

The earthquakes from Russia are not

uniformly distributed geographically; they tend to occur along her borders. The majority, about 60 percent, come from the Kurile Islands and the Kamchatka Peninsula region, and that so many events are coastal is very important. First of all, by using records from ocean-bottom seismometers it is possible to apply identification techniques applicable to records obtained at angular distances of less than 10°. Secondly, many of the earthquakes probably occur sufficiently far out from land that, if there is no evidence of an underwater explosion, they can be eliminated as possible explosions. No evidence is available about location accuracies in this region, but there is nothing to suggest that teleseismic methods give less accurate results here than in other areas.

## Summary and Conclusions

I have tried to describe some current research trends in seismology which are specifically directed toward solving the problem of detecting, locating, and identifying underground nuclear explosions. Attention has been directed specifically toward problems which arise in efforts to obtain information at distances in excess of 2500 kilometers. The main scientific advantage which accrues from working at such distances is that the seismic signals suffer minimal distortion by the geological complexities of the earth. Extrapolation of the data to the question of an international test ban is not within the scope of this article. Suffice it to note that all of the parameters must, in the final resort, be resolved in terms of probabilities. In some cases the seismological probabilities can be estimated with reasonable degrees of accuracy, but the future of the test ban question depends not only on seismology but on such questions as inspection and what probabilities are acceptable.

The current research program has produced revolutionary advances in the science of seismology. By far the greater part of the work has been directed toward obtaining a fuller understanding of seismic propagation paths within the crust and upper mantle, and relatively little of it has been aimed at achieving a deeper understanding of earthquake mechanism. With a concentration of effort on interpreting the character of earthquake signals as seen through the "seismic window" it has become possible to think in terms of actually defining the motion at the source from the seismic records it creates. Despite the fact that only a small section of the source can be viewed through the "window," its definition would undoubtedly mark a big advance in our knowledge of how earthquakes occur—one which could lead, possibly, to a realization of the seismologist's dream of accurate prediction of earthquakes.

### References and Notes


1. A large amount of background material, including a relevant bibliography, is contained in the "Congressional Hearings," particularly in "Hearings before the Joint Committee on Atomic Energy, 88th Congress, First Session, on Developments in Technical Capabilities for Detecting and Identifying Nuclear Weapons Tests, March 1963."
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6. I am grateful for the help and guidance of my colleagues at Blacknest—in particular, Dr. H. I. S. Thirlaway. Nevertheless, the views expressed are my own and should not be taken as necessarily representing the views of the United Kingdom Atomic Energy Authority.


# Spandex Elastic Fibers

## Development of a new type of elastic fiber stimulates further work in the growing field of stretch fabrics


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Vulcanized rubber, from which the first elastic fibers were made, has found wide application in the manufacture of elastic fabrics. For many years fiber chemists have sought to develop materials superior to rubber in recovery force, resistance to abrasion, and chemical stability. Several such fibers have now been produced within the field of urethane chemistry and are being manufactured commercially, under various trade names. The generic name of these fibers is spandex, which is defined as a segmented polyurethane (*1*). In this article we discuss the synthesis and structure of segmented polyurethanes and the properties of spandex fibers made from these polymers.

The classical theory of elastic behavior is called the kinetic theory of elasticity (*2*). The theory requires kinetically active, long molecular chains with characteristics of liquids, joined by tie-points to provide for recovery from deformation (see cover).

In an ideal rubber there is no change in internal energy during stretching, and the retractive force, being solely dependent upon entropy change, is proportional to the absolute temperature. The retractive force of a deformed elastomer depends also on the molecular weight of the polymer and on the average distance between the tie-points. Statistical considerations (*3*) lead to the relationship

$$f = RTd\left(\frac{1}{M_c} - \frac{2}{M}\right)\left(\alpha - \frac{1}{\alpha^2}\right)$$

where $f$ is the force per unit initial area of cross section, $R$ is the gas constant, $T$ is the absolute temperature, $d$ is the density of the polymer, $M$ is the molecular weight of the polymer, $M_c$ is the average molecular weight between tie-points, and $\alpha$ is the


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