

TRAVEL TIMES AND STATION CORRECTIONS FOR P WAVES
AT TELESEISMIC DISTANCES

Adam M. Dziewonski

Department of Geological Sciences, Harvard University, Cambridge, Massachusetts 02138

Don L. Anderson

Seismological Laboratory, California Institute of Technology
Pasadena, California 91125

Abstract. Approximately 3300 shallow focus earthquakes and 1000 seismic stations have been used in a study of P wave travel times and station residuals, including azimuthal effects. The events were selected from a catalog containing 160,000 earthquakes, and those having uniform distance and azimuthal coverage were systematically relocated and used to refine P wave travel times and station corrections. Station corrections are provided for 994 seismic stations. The station corrections involve three terms: the static effect and two cosine terms with appropriate phase shifts. They exhibit general consistency over broad geographic areas and, where coverage is dense, often show abrupt changes from one geological province to another. The $\cos 2\theta$ terms appear to be due to upper mantle anisotropy, and they correlate with the stress direction in the crust.

Introduction

This paper treats a rather traditional topic, one that attracted much attention during the late 1960's and early 1970's, when the high-quality data from the rapidly expanding global seismograph network and fast computers became readily available. Although several papers pointed out inadequacies in the Jeffreys and Bullen [1940] travel times, there were important differences among the travel time curves derived in these studies. Partly for this reason, the National Earthquake Information Service of the U. S. Geological Survey and the International Seismological Centre still use the J-B tables for estimation of hypocentral parameters.

In addition to the practical applications of travel time curves, such as in earthquake location, the inferences drawn from observations of travel times cover a broad range of basic problems in geophysics. Detection of discontinuities and resolution of lateral variations in earth structure are important in understanding dynamic processes within the earth. Station anomalies, including azimuthal effects, can be used to study heterogeneity and anisotropy.

Since arrival times of P waves are the most frequently reported functional of the earth's structure, our ability to process and interpret these data is, in a certain way, a measure of progress in seismology.

The travel times for P waves in the J-B

tables are an expansion of the results published by Jeffreys [1939]. In that paper he combined his results on travel times and velocity distribution in the earth which were published in a series of articles dating from 1936 to 1939 in the Geophysical Supplements to the Monthly Notices of the Royal Astronomical Society. The fact that these results are still used as the standard testifies to the excellence of this effort.

The existence of regional variations in travel times had been recognized by Gutenberg [1953], who attributed them primarily to regional changes in the depth of the Mohorovicic discontinuity. Herrin and Taggart [1962] studied regional variations in the P_n travel times in the United States. Carder [1964] used travel times from nuclear explosions and showed that there are significant deviations from the J-B travel times. His results indicated the need for a baseline correction of about -2 s and the existence of distance-dependent residuals. However, Carder used only data from explosions in the Marshall Islands, and the question of a regional bias remained open.

It was the study by Cleary and Hales [1966] that set the stage for most of the later investigations, including this one, that involved globally distributed sets of sources and receivers. The approach adopted by Cleary and Hales is an application of the 'time term' analysis previously used in interpretation of seismic refraction studies [cf. Willmore and Bancroft, 1960]. The objective of this analysis is to represent an observed set of travel times (or residuals) as a sum of source, propagation, and receiver terms. As a result, one obtains an improved set of epicentral coordinates, a new travel time curve, and a set of station corrections. Cleary and Hales were the first to publish an extensive set of station corrections, a parameter that they found to be closely related to the tectonic nature of the receiver region.

Other important papers that involved analysis similar to that of Cleary and Hales are the 1968 Tables [Herrin et al., 1968], the joint epicentre method of Lilwall and Douglas [1970], and the study of travel times for deep earthquakes by Sengupta and Julian [1976].

Dziewonski and Anderson [1981] used a very large set of data (1,500,000 P wave arrival times for 26,000 earthquakes) from the Bulletins of the International Seismological Centre (ISC) for the years 1964 to 1975 to obtain a travel time curve which, together with other subsets of data, was used to derive the Preliminary Reference Earth Model (PREM). However, the epicenters were not relocated in that study, and, as suggested by

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Paper number 2B1668.
0148-0227/83/002B-1668\$05.00

A. Douglas (personal communication, 1981), some residual effect of the J-B curve might have remained.

Other studies included $dT/d\Delta$ measurements using seismic arrays [cf. Niazi and Anderson, 1965; Johnson, 1967, 1969; Toksöz et al., 1967; Chinnery, 1969] and the search for the lateral heterogeneity in the mantle [cf. Julian and Sengupta, 1973; Dziewonski et al., 1977]. These papers have a bearing on the subject of this report, but our present purpose is to use the data for shallow earthquakes contained in the bulletins for the years 1964-1978 in order to clarify several issues that are still controversial.

The shape of the travel time curve. The principal difference between the results of Cleary and Hales [1966], Herrin et al. [1968], and Lilwall and Douglas [1970] was the slope of the travel time curve. The largest difference exists between the first two studies: between 30° and 90° the slowness of the 1968 tables is greater by 0.023 s/deg than that inferred by Cleary and Hales. There are also differences in the details of the travel time curves. Even though these differences do not exceed 0.3 s, they ought to be resolvable.

The question of the source bias. Most of the earthquakes ($\sim 80\%$) occur at or near subduction zones. Because of the velocity structure in the subducted slab, travel time residuals tend to have very particular patterns [cf. Davies and McKenzie, 1969]. A desire to avoid this effect motivated the study of Sengupta and Julian [1976]. However, if the effect of the slab extends below 700 km [Jordan, 1977], then the use of only deep-focus earthquakes not only does not free one from source bias but also severely restricts the geographical coverage.

Azimuthal dependence of station corrections. Herrin and Taggart [1968] and Lilwall and Douglas [1970] evaluated station correction terms that, in addition to the constant term, provided for a sinusoidal variation with azimuth (plus a constant phase term). Sengupta and Julian [1976] found that the correlation coefficient between these two sets of azimuth-dependent corrections was 0.41. This figure indicates contamination by noise or uneven coverage. Figure 3 of Herrin and Taggart [1968] indicates that while there is overall consistency among the slow direction vectors in the eastern United States, the magnitude and direction of these vectors are highly incoherent in the central and western United States.

Our objective is to demonstrate that by using a significantly larger set of data and an approach to estimation of the azimuth-dependent terms that accounts for the uneven distribution of observations as a function of azimuth, it is possible to demonstrate regional coherence of not only the first but also the second azimuthal term.

The Data Set

The ISC Bulletins for the years 1964-1978 contain entries for approximately 160,000 earthquakes. Most of these events are small with relatively few reports of arrival times by individual stations. For the purpose of our study

we require events for which stable epicenters can be determined using only the teleseismic travel times. Thus we have scanned the ISC tapes for earthquakes that had at least 30 arrival times between 30° and 90° of epicentral distance and at least five readings in each azimuthal quadrant. This search yielded 3270 events.

As discussed in the introduction, the effect of source bias in the vicinity of subducted slabs is a subject of concern. Subduction-related events dominate the catalog even though trenches cover a relatively small fraction of the earth's surface. In order to avoid as well as to study this bias, we divide the selected sources into two classes: class I, events away from subduction zones; and class II, events in the vicinity of subduction zones.

Figure 1 shows the geographical distribution of the two classes of events. The symbols indicate that at least one event of a particular class occurred within a given $5^\circ \times 5^\circ$ cell; circles designate class I events; stars, class II. Even though there are 3 times more class II events, class I earthquakes occupy roughly twice as many cells and provide more uniform coverage of the earth's surface. For this reason the class I events are considered the primary data; class II earthquakes are processed separately.

In the last stage of the entire process, when the station corrections are determined, the condition of the azimuthal coverage of an event is relaxed: three azimuthal quadrants must contain at least five observations. This triples the number of available events and improves the azimuthal coverage of the individual stations. A number of precautions are taken that this does not introduce a bias.

There are over 1000 stations that reported 30 or more arrival times within the established range of epicentral distance. At least one component of the station correction can be established for most of these stations.

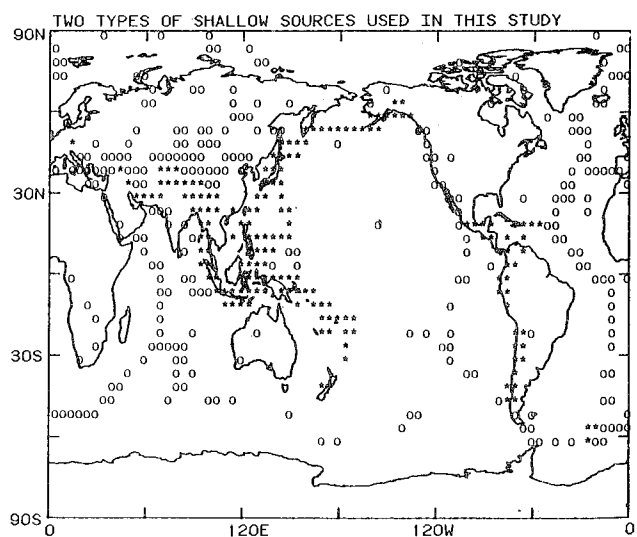


Fig. 1. Distribution of shallow sources used in this study. A symbol indicates that at least one event for a given $5^\circ \times 5^\circ$ cell was used in the analysis. Circles designate class I events; stars, class II events: earthquakes that occur near or at subduction zones.

The Iterative Time Term Analysis

The time term method, as described by Cleary and Hales [1966], is based on the following equation of condition:

$$\delta t_{ij} = a_{ij} + b_j + d_i \quad (1)$$

where for the *i*th station and *j*th earthquake, δt is the observed residual from the reference travel time curve; *a* is the perturbation to the travel time curve for the particular source-receiver pair; *b* is the source term and *d* is the station term.

Cleary and Hales assumed that a_{ij} is a function of distance only: $a_{ij} = a_k(\Delta_{ij})$, where *k* is an integer selected such that it covers a 2° cell. The same division has been assumed by Lilwall and Douglas [1970] and Sengupta and Julian [1976]; a 1° cell size was adopted by Herrin et al. [1968] and in this study. In general, the term *a* may depend on the geographical coordinates of the source and receiver; this formulation was used by Dziewonski et al. [1977] in a search for large scale velocity heterogeneities in the mantle. In this case a_{ij} is represented by a product of unknown structural parameters with the appropriate kernels.

Cleary and Hales represented the source term *b* by a constant, essentially a perturbation in the origin time. The events were relocated after each iteration. In general, the source term can be associated with a perturbation in the hypocentral coordinates of the source. In this case, *b* is equivalent to the product of a matrix of partial derivatives with the vector of perturbations in hypocentral parameters. Lilwall and Douglas [1970] adopted this approach with the source depth fixed.

Cleary and Hales represented the station term *d* by a constant. In general it may be azimuth dependent:

$$d_{ij} = \sum_{n=0}^N C_{in} \cdot \cos n\xi_{ij} + S_{in} \cdot \sin n\xi_{ij}$$

where ξ_{ij} is the azimuth from the *i*th station to the *j*th source. Herrin et al. [1968] and Lilwall and Douglas [1970] considered *N* = 1; in this study, *N* varies between 0 and 2, depending on the azimuthal coverage at the station.

There is a natural nonuniqueness in (1); clearly, the same amount of time may be added to all a_{ij} and subtracted from all d_i and the observed value δt_{ij} would remain unchanged. Therefore it is necessary to fix arbitrarily one value of the *a* parameter and one value of the station correction. Provided that this is done, application of the least squares condition to the set of equations of condition (1) leads to a system of normal equations:

$$\underline{A} \cdot \underline{x} = \underline{B} \quad (2)$$

The matrix \underline{A} has the structure

$$\underline{A} = \begin{bmatrix} C & H & G \\ H & D & F \\ C & F & E \end{bmatrix} \quad (3)$$

where *C*, *D*, and *E* are diagonal or block diagonal matrices. If *C* is associated with the travel time corrections that depend only on epicentral distance, then it is diagonal; if *D* is associated with the source term, then it is diagonal in the formulation of Cleary and Hales or (3 x 3) block diagonal in the case of Lilwall and Douglas. Matrix *E* is block diagonal with block dimensions of (2*N* + 1); *E* is diagonal for *N* = 0. Matrices *F*, *G*, and *H* are the cross-product matrices; in the case considered by Cleary and Hales they have the property that all their elements are positive integers and the sum of a given row or column is equal to the appropriate diagonal element.

If, as in this study, we consider 2000 events to be relocated, 1000 stations with an average of three unknowns per station to be determined and travel time corrections to be determined for 76 epicentral distance cells, the matrix *A* would have dimensions of the order of 9000 x 9000. Computation of the inverse of such a matrix is impractical even using the largest available computers.

However, the fact that matrix *A* is dominated by its diagonal submatrices *C*, *D*, and *E*, allows us to initiate an iterative procedure that does not require an explicit evaluation of the inverse of *A*.

If we approximate the inverse of *A* by

$$\tilde{\underline{A}}^{-1} = \begin{bmatrix} C^{-1} & 0 \\ & D^{-1} \\ 0 & & E^{-1} \end{bmatrix} \quad (4)$$

then the first approximation of the unknown vector \underline{x} is

$$\underline{x}_{(0)} = \tilde{\underline{A}}^{-1}_{(0)} \cdot \underline{B}_{(0)} \quad (5)$$

This is equivalent to (1) estimation of the correction to the travel times by averaging the residuals with respect to the starting travel time curve, (2) relocation of the individual events using the starting travel time curve, and (3) estimation of station corrections for the individual stations.

In the next iteration, we evaluate

$$\delta \underline{x}_{(1)} = \tilde{\underline{A}}^{-1}_{(1)} \cdot (\underline{B}_{(0)} - \underline{A}_{(1)} \cdot \underline{x}_{(0)}) = \tilde{\underline{A}}^{-1}_{(1)} \cdot \underline{B}_{(1)} \quad (6)$$

$$\underline{x}_{(1)} = \underline{x}_{(0)} + \delta \underline{x}_{(1)}$$

or, in general,

$$\delta \underline{x}_{(n)} = \tilde{\underline{A}}^{-1}_{(n)} \cdot \underline{B}_{(n)} \quad (7)$$

$$\underline{x}_{(n)} = \underline{x}_{(n-1)} + \delta \underline{x}_{(n)}$$

If the inverse of matrix *A* is stable, then the iterative process described above should converge. This means that, formally, the procedure adopted by Lilwall and Douglas [1970] on the one hand and Herrin et al. [1968], Sengupta and Julian [1976], and this study on the other are equivalent; the procedure of Cleary and Hales [1966] falls between the two approaches. Thus the differences in the final results must be explained by the selection of events and stations rather than by the specific algorithm adopted.

For practical reasons (the process of estimation of station residuals is very time consuming) we depart somewhat from the iterative procedure described above. Each cycle consists of (1) relocation of events, (2) determination of improved travel time curve, and (3) estimation of station corrections. We repeat steps 1 and 2 until convergence is achieved and only then evaluate station correction. The cycle is then repeated. To verify that this departure from the ideal procedure does not introduce a bias, we have carried out the entire process having in the first step calculated station corrections using the J-B travel times and the original ISC locations. The difference in the final results was negligible.

The Procedure and Results

Given a travel time curve and starting locations, relocation of a hypocenter is routine: an exhaustive description can be found, for example, by Tucker et al. [1968]. Ellipticity corrections were introduced using the formula and tabular values given by Dziewonski and Gilbert [1976].

Because the azimuthal distribution of stations with respect to the epicenter tends to be uneven, we have tested a weighting scheme in which, regardless of the number of observations, each azimuthal quadrant was given equal weight. In terms of the average travel time curve for the total set of either class I or class II events, that weighting approach did not change the results by more than 0.02-0.03 s. We have not investigated what effect this weighting scheme had on location of the individual events, but, in general, individual events were not of interest in this study. Another test was performed by relaxing the condition on the azimuthal coverage: only three azimuthal quadrants were required to have not less than five observations. This allowed us to triple the amount of available data. Again, the effect on the average travel time curve was negligible; other than that the increased amount of data resulted in a somewhat smoother residual curve.

In the relocation procedure it is necessary to have a differentiable travel time curve. The perturbations obtained in each iteration were fit with cubic splines with three knots. This representation was sufficient to avoid any significant long-term deviations between the observed and smoothed curves.

Figure 2 shows, in terms of deviations from the 1968 travel times, the starting curve, one obtained in the first iteration and the final one (after four iterations). It is clear that most of the change occurs in the first iteration. As stated earlier, the class I events are considered the principal data set. Therefore class II events are relocated also using the class I travel time curve. It is clear from Figure 2 that there are some differences between the travel times for class I and class II events. Both curves, however, are relatively close to the 1968 tables, with the curve for class II events being closer and showing only a slight offset in the baseline and slope. This is not surprising, since some 80% of the events used in the derivation of the 1968

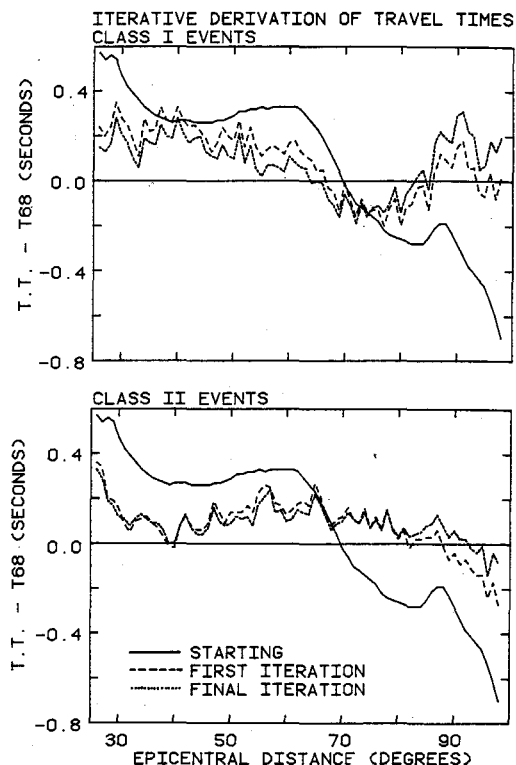


Fig. 2. Illustration of the convergence of iterative derivation of teleseismic travel times for P waves. The results are displayed as residuals with respect to 1968 tables. Station corrections are not yet included.

tables were class II events [see Herrin et al., 1968, Figure 3].

After the convergence of travel times is achieved, the residuals are sorted by station, and the process of determination of station corrections begins. The full azimuth range is divided into 18 windows, 20° wide. An average residual is calculated if there are at least five readings in a given window. From now on, these average residuals are treated with equal weight; this reduces the bias due to unequal distribution of events [see Herrin and Taggart, 1968, Figure 1]. In general, the least squares approach is used to determine the coefficients A and B in the equation

$$\delta t = A_0 + A_1 \cos \xi + B_1 \sin \xi + A_2 \cos 2\xi + B_2 \sin 2\xi \quad (8)$$

However, the number of terms to be determined depends on the azimuthal coverage at a given station. All terms are determined if the data exist for 15 or more windows. Below that number, decisions are made depending on the distribution of the missing windows and the pattern of deviations; admittedly, some decisions are subjective. Station corrections are not determined if data are available for fewer than two windows.

Figure 3 shows a plot of station residuals as a function of azimuth. It is one of the final plots, but it serves well to explain the interactive process of determination of station corrections. The squares are residuals for the class I events; stars are for class II. Averages

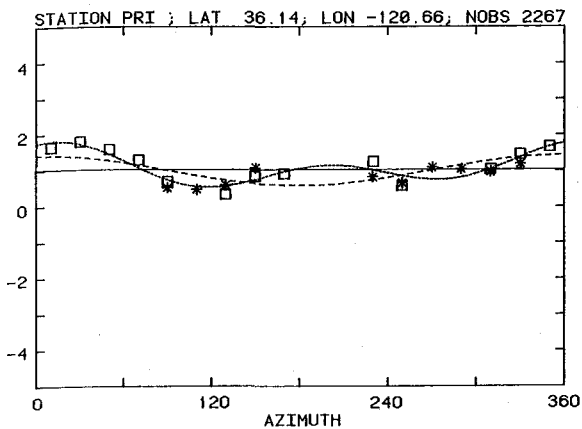


Fig. 3. Station residuals as a function of azimuth. Square represents an average of all residuals for class I events within a given 20° window; stars correspond to class II events. Solid line is the least squares fit of the constant term; long-dashed line, constant and first azimuthal term; short-dashed line, constant and two azimuthal terms.

for individual windows are given equal weight. The straight, solid line is the A_0 term: the average of all residuals. The long-dashed line corresponds to the least squares fit of the first three terms in equation (8). Finally, the short-dashed line represents all five terms. The coverage by class I events is more complete; it would have been possible to determine all five terms even if the data for class II events were not included, and the derived parameters would differ little from those shown. On the other hand, even though there are class II data for 10 out of 18 windows, fitting of terms other than A_0 would not be justified, as the data are very unevenly distributed. Overall, in windows for which both kinds of data exist, the average residuals for both classes of events are very close, indicating that the source term, at least for this station, is not very important or has been absorbed in the relocation of the epicenters. This seems to be the case for most stations.

The stations corrections were computed using three sets of relocated earthquakes: (1) class I data: four quadrants with at least five observations, (2) class I and II data: four quadrants with at least five observations, and (3) class I and II data: three quadrants with at least five observations.

Clearly, the number of data for each of the subsequent sets increases, and yet, whenever a comparison could be made, the results obtained using sets 1 and 3 were fully compatible. For this reason we use set (3) to evaluate station corrections, since it allows us to investigate more stations with a better azimuthal coverage.

In the next cycle, the events are relocated using station corrections, and after two more iterations, convergence of the travel time curves is achieved. The next set of station corrections differs insignificantly from the previous one, and we conclude that the process has converged.

Our main interest in this study was the derivation of station residuals and a new average

travel time curve for distances beyond 25° . However, with the relocated earthquakes and available readings at shorter distances we can also construct an average travel time curve from 1° to 25° . Although regional variations become important at these distances, the usefulness of the new travel times is enhanced by having available values for the full range of distances. We remind the reader that only the data between 30° and 90° were used in the process of relocation of events and estimation of station corrections.

As before, we construct average travel times for 1° cells and then smooth them with a set of polynomials with various conditions of continuity at the knots. The smoothed travel times are continuous throughout, but the first derivative (slowness) was allowed to be discontinuous at 18° and 24° . Except for the distance from 18° to 24° and from 90° to 100° , where a quadratic was fitted, cubic polynomials were used in the remaining segments (1° - 18° , 24° - 40° , 40° - 65° , 65° - 90°). The rms of the fit improves with increasing epicentral distance. For the class I travel time curve it is 0.44 s between 1° and 18° , 0.11 s between 18° and 24° , 0.06 s between 24° and 40° , and beyond this distance it ranges from 0.04 to 0.05 s.

Our final average travel times are given in Table 1. Tabulated are epicentral distance (DEL), number of observations (NOBS), average travel time (TOBS), standard deviation of single observation (S.D.), smoothed travel time (TCOM), difference between observed and smoothed values (DIFF), and slowness (DT/DD) in seconds/degree. Travel times for both class I and class II events are given; the maximum difference between smoothed values for the two sets is 0.12 s between 90° and 94° .

Table 2 contains stations corrections for 994 stations. Most of the entries are self-explanatory. NW is the number of 20° azimuth windows for which data were available; RMS0 is the rms error calculated after the A_0 term is removed from averages for individual windows; RMS1 is the error after the higher-order terms (if any) have been removed. The station correction term (to be added to the theoretical travel time) has the form

$$\delta t = A_0 + A_1 \cdot \cos(\xi - E_1) + A_2 \cdot \cos^2(\xi - E_2)$$

Thus the azimuths E_1 and E_2 represent the slow directions. For the second azimuthal term there is another slow direction: $E_2 + 180^\circ$.

Discussion

Figure 4 shows a comparison of the residual travel time curves for the class I and class II events before and after introduction of station corrections. There are several important conclusions that can be drawn from this figure.

The average number of observations for a 1° cell is about 1600 for the class I data and 4000 for the class II data. With an rms of single observation of 1.2 s, the standard error of the mean (sem) for the class I data should be 0.03 s and 0.02 for the class II data. The roughness in the uncorrected class I curve exceeds only slightly what could be considered random scatter.

TABLE 2. Station Corrections

Table with columns: CODE, LAT, LONG, ELEV, NOBS, NW, RMS0, RMS1, A0, A1, E1, A2, E2. Rows include stations AAA through BHC.

TABLE 2. (continued)

Table with columns: CODE, LAT, LONG, ELEV, NOBS, NW, RMS0, RMS1, A0, A1, E1, A2, E2. Rows include stations BHK through COB.

TABLE 2. (continued)

Table with columns: CODE, LAT, LONG, ELEV, P-WAVE STATION CORRECTION (NOBS, NW, RMS0, RMS1, A0, A1, E1, A2, E2). Rows include station identifiers like TAV, TBZ, TCF, etc.

TABLE 2. (continued)

Table with columns: CODE, LAT, LONG, ELEV, P-WAVE STATION CORRECTION (NOBS, NW, RMS0, RMS1, A0, A1, E1, A2, E2). Rows include station identifiers like VIC, VIE, VIS, etc.

See text for explanation of headings and contents of the table.

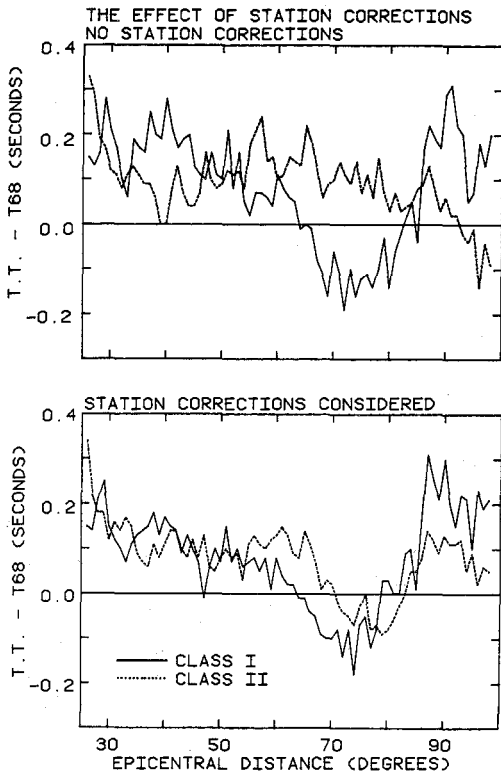


Fig. 4. The effect of introduction of station corrections on the travel time curves presented, as in Figure 2, in terms of deviations from 1968 tables. Station corrections not only reduce the roughness, but also are capable of changing the trend over a large range of distances, decreasing the difference between the travel times for class I and class II events.

Because of the better geographical coverage, and with the appropriate reservations, we recommend the travel times for the class I events, corrected for station effects, as the best estimate of teleseismic travel times for surface focus P waves.

Figures 5a-5h show examples of the azimuthal dependence of station residuals. The principal point is to demonstrate the variety and range of the station terms as well as the consistency of the residuals for the class I and class II events. Although these figures show stations reporting a relatively large number of observations with good azimuthal coverage, the selection in terms of similarity between the two types of residuals is typical. From this we conclude that the source effect is either not as important as might have been thought, or as already mentioned, it has been, in part, absorbed by the process of relocation of events. In either case, the station residuals listed in Table 2 would seem to be dominated by the upper mantle and, perhaps, crustal structure in the vicinity of the station.

Figure 5a represents a station with nearly no azimuthal dependence of residuals and also a very small constant term. Figure 5b, on the other hand, shows a peak-to-peak variation of nearly 4 s. The residuals for class I and II events differ in the window centered on an azimuth of 330° by more than 2 s, but in the other five windows for

which observations of both types exist, the data are very close to each other. Figure 5b represents a rare example of a station dominated by class II events.

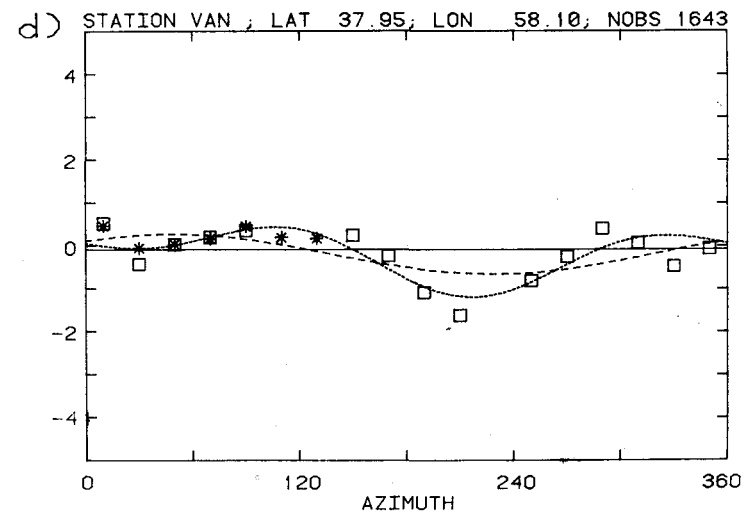
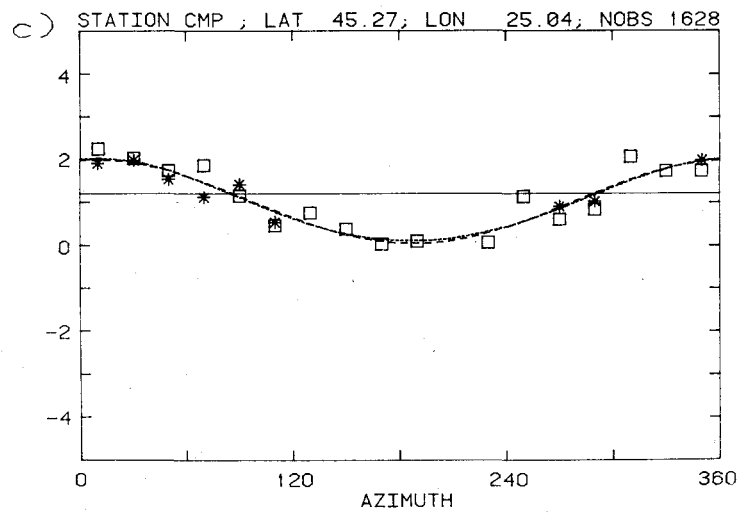
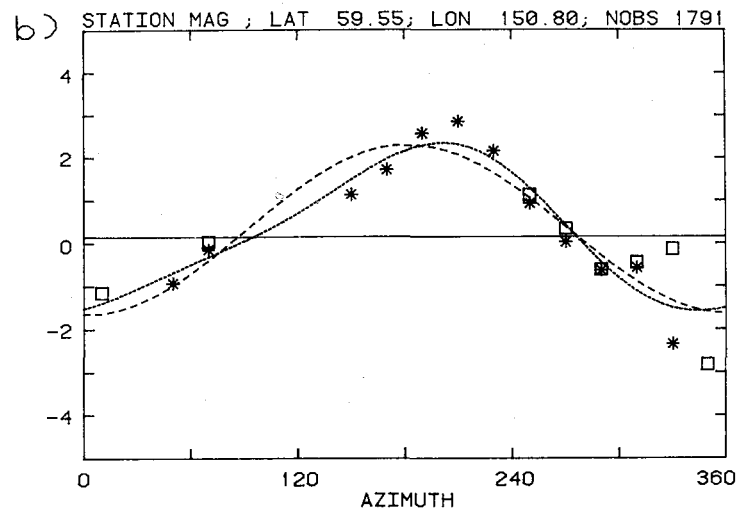
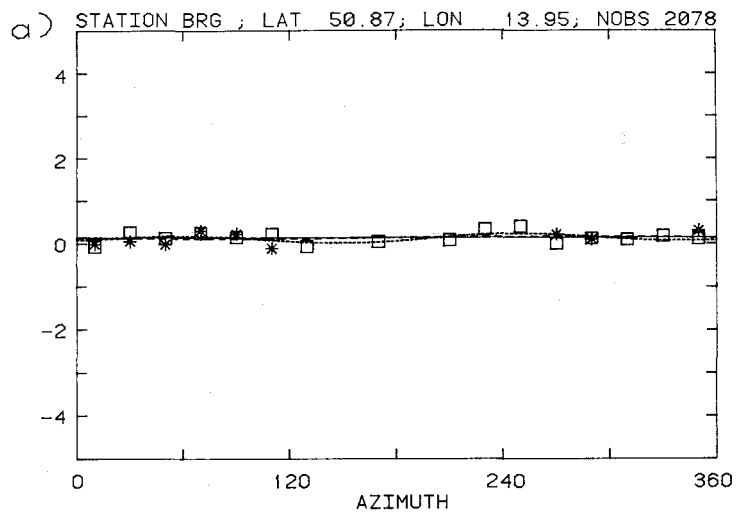
Figure 5c is an example of a station dominated by the A_0 and A_1 terms, with the coefficient A_2 being virtually zero, while in Figure 5d the latter term dominates. Figures 5e and 5f are examples of stations with relatively large second azimuthal terms. Figures 5g and 5h are shown to demonstrate the consistency of the residual pattern over fairly large distances: the two stations in the Canadian Arctic are nearly 900 km apart, and yet the magnitude and the phase of the anomalies are very similar.

The consistency of different terms in our table of station corrections can be more easily judged from a map display. Figure 6a shows the size and distribution of the azimuth-independent term over the large part of the North American continent. The stations displayed reported at least 700 observations distributed in at least 10 azimuth windows. Squares designate late, or slow, stations; triangles represent early, or fast, stations. The size of the symbol is proportional to that of the anomaly; the scale for +1 and -1 s is shown.

The arrivals are systematically late for stations in the Appalachians. Most of the stations in the Great Plains and in Canada east of the Rocky Mountains are fast. Nevada and California are very slow. This picture is in excellent agreement with the results of Herrin and Taggart [1968, Figure 2].

Figure 6b displays the variation in the first azimuthal term. The arrow points toward the slow direction, and its length is proportional to the A_1 coefficient (see inset for an arrow corresponding to 1 s anomaly). To be included in this figure a station had to report at least 1000 observations for at least 15 azimuth windows. While the stations in the east and north (including Alaska) show consistency over large distances, the situation in the west is more complex. If we discard the coastal stations, which show great variability in their pattern, we notice that there are two populations of arrows that represent a very consistent pattern. One group of relatively large anomalies (~ 0.5 s) points nearly due north, and the other, consisting of anomalies of about 0.25 s, points in the direction $N45^\circ E$. The two types of arrows are intermixed, but the populations are well defined. Clearly, an attempt should be made to correlate this observation with other kinds of geophysical and geological data.

Figure 6c shows the second azimuthal term. The criteria for displaying a station are the same as in the case of Figure 6b and remain so for all the figures to follow. The small square designates the position of the station and the line represents the slow azimuth. There is a great deal of regional consistency in the second azimuthal term. In the western United States and Canada the slow direction remains similar over an area about 2000 km long in the NS direction and 1000 km wide from east to west. There is some indication of a gradual rotation toward north-south slow directions as one progresses from Mexico to Canada. A group of stations near the coast of southern California represents an



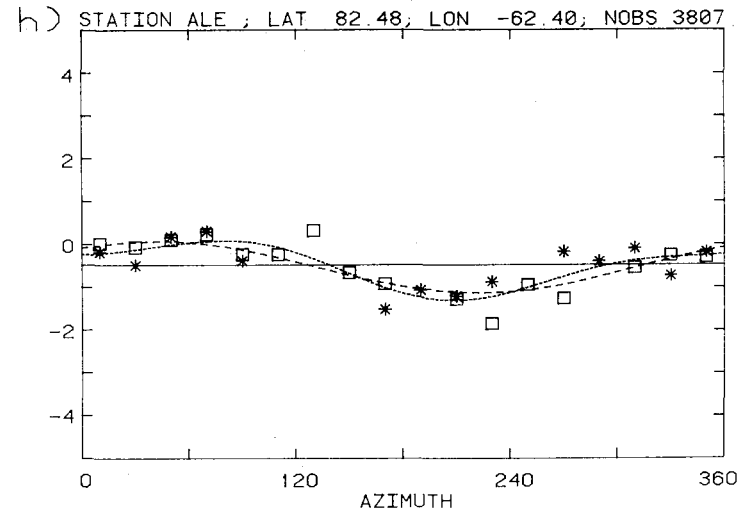
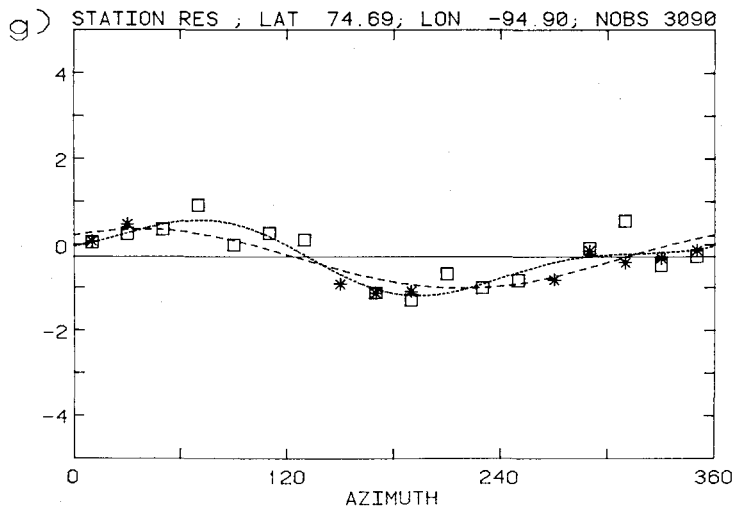
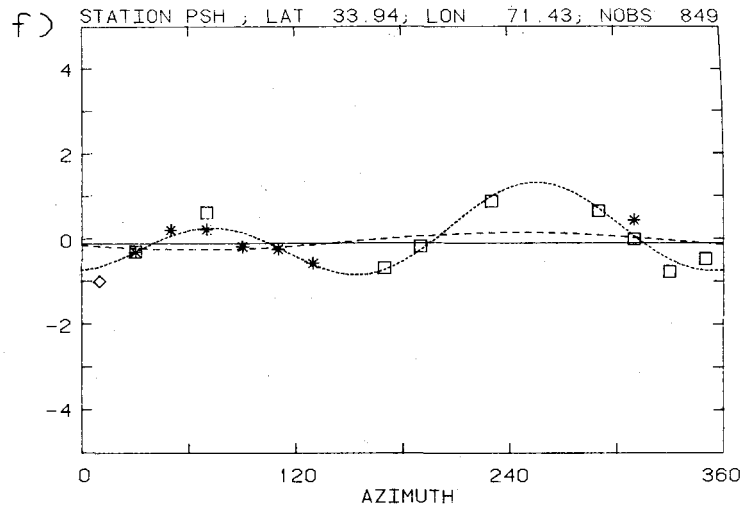
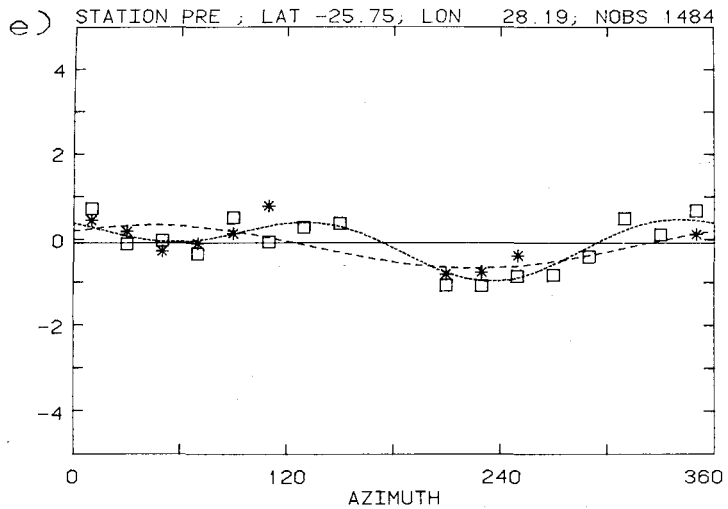


Fig. 5. Examples of station residuals as a function of azimuth illustrating variability in the relative importance of different station correction terms. For details see caption to Figure 3.

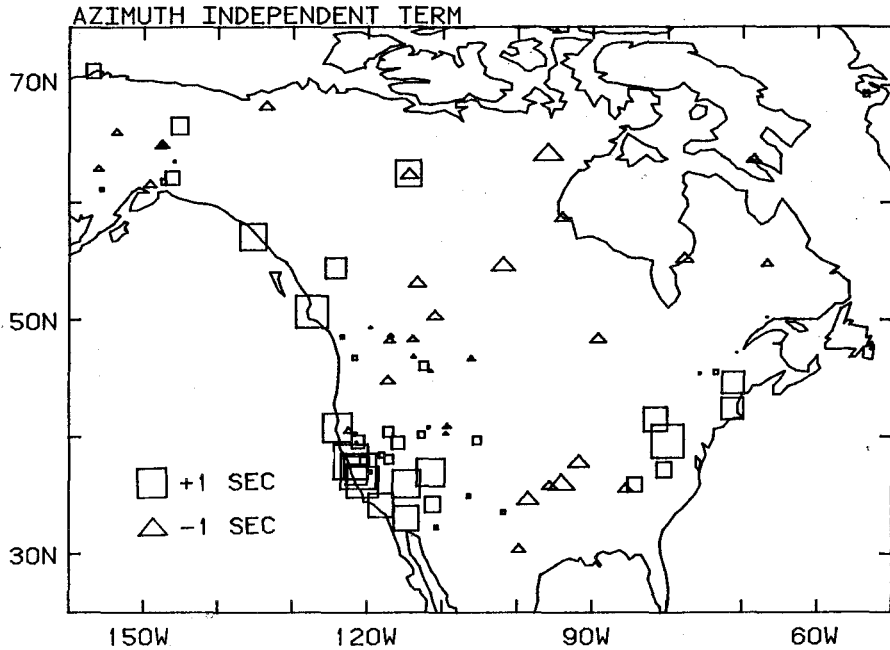


Fig. 6a. Distribution of the constant station correction term for stations in the North American continent. Only stations with at least 700 observations distributed among at least 14 azimuthal windows are displayed. The size of a symbol is proportional to the anomaly.

exception: the change in direction by about 60° is very abrupt. One of the sources of a perceptible second azimuthal term could be anisotropy. The fast direction in southern California is parallel to the direction of motion of the Pacific plate. This has also been found in

studies of P_n in southern California and has been attributed to upper mantle anisotropy [Vetter and Minster, 1981]. In studies of the anisotropy of olivine the fast axis is parallel to the flow direction [Christensen and Salisbury, 1980]. This raises the possibility that the second azimuthal

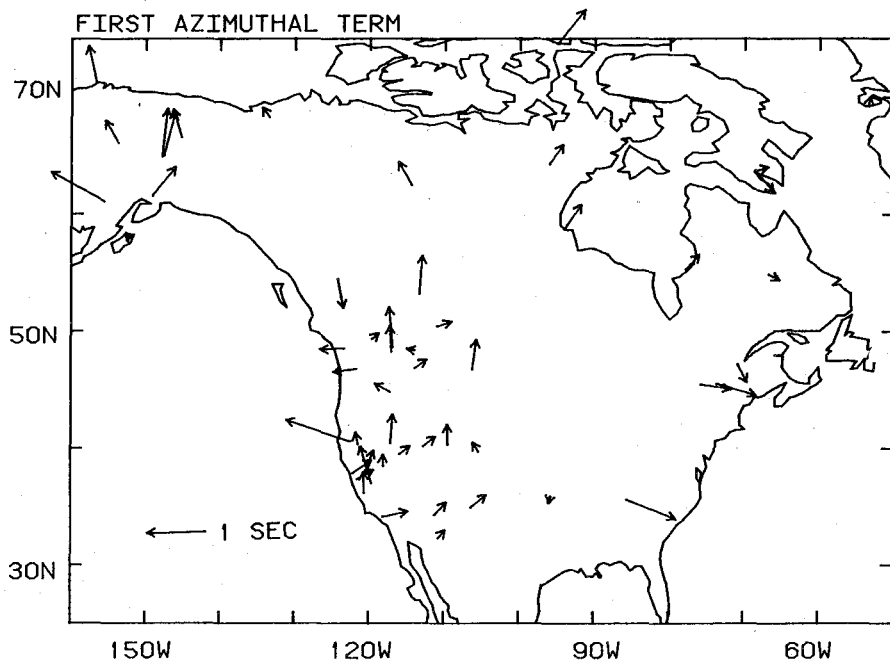


Fig. 6b. Distribution of the first azimuthal term in station anomalies for stations in the North American continent. Arrows point toward the slow direction, and their length is proportional to the anomaly. Only stations with at least 1000 observations distributed among at least 15 azimuth windows are displayed.

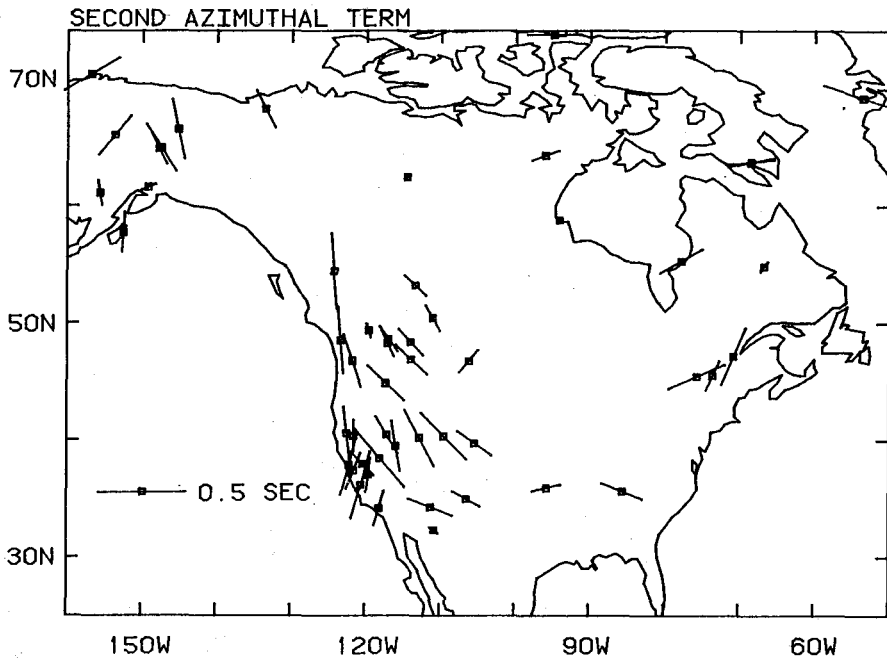


Fig. 6c. Distribution of the second azimuthal term in station anomalies for stations in the North American continent. The line centered over a station is aligned in the slow direction and its length is proportional to the anomaly. Observational requirements are the same as in Figure 6b.

term can be used to map the direction of flow in the mantle.

Figure 7a shows the azimuth-independent term for Europe. Stations distributed along the Alpidic belt are generally slow; station AKU on Iceland has a positive residual of over 2 s. Stations on the Baltic Shield are generally fast. Thus the

familiar pattern, first detected by Cleary and Hales [1966], is well reproduced. The most striking feature of Figure 7b is the fanlike pattern of the arrows pointing away from the Alpidic belt for stations between longitudes 0 and 30°E. For the Scandinavian stations the slow direction is toward the west, although the magnitude of the arrows there is rather small. Association of the fast-slow direction with the

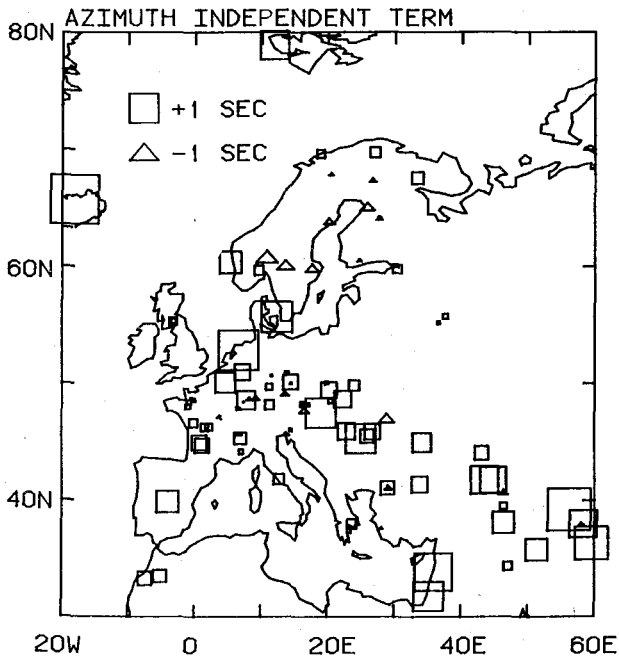


Fig. 7a. Same as Figure 6a but for Europe. Observational threshold: 1000 observations in 15 azimuthal windows.

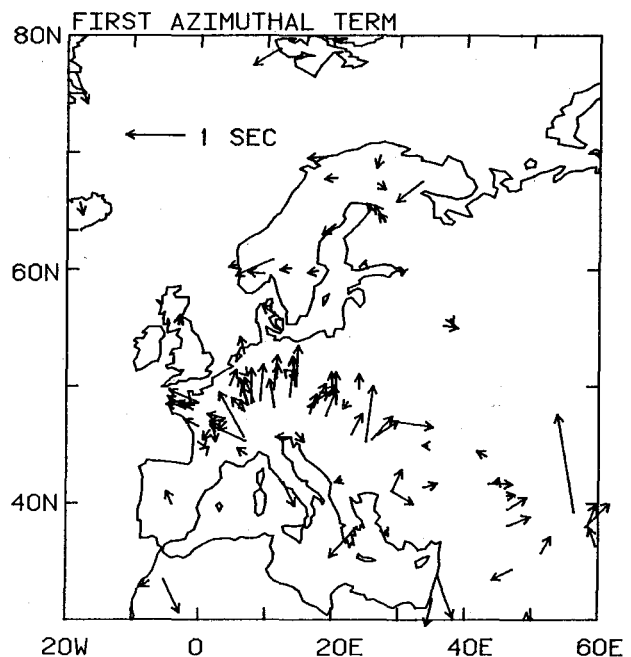


Fig. 7b. Same as Figure 6b but for Europe.

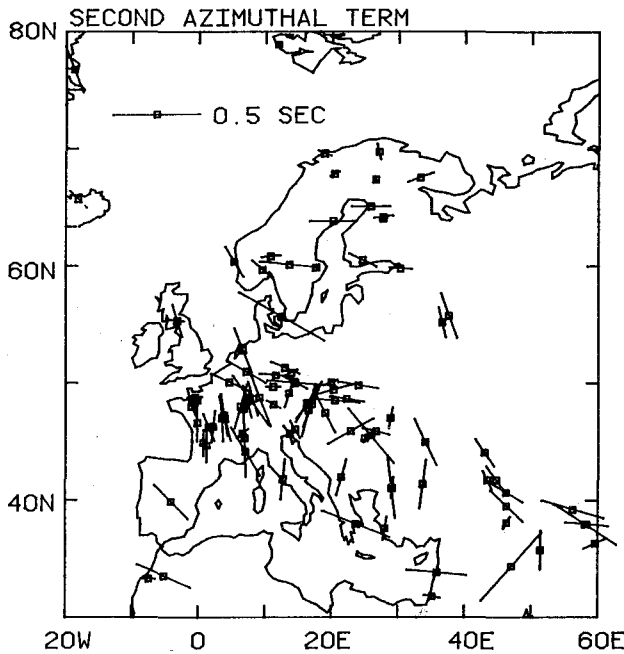


Fig. 7c. Same as Figure 6c but for Europe.

major tectonic feature of Europe can also be seen in Figure 7c, which displays the second azimuthal term. The slow direction here is mostly parallel to the Alpid belt, with the trend continuing to the east, although the picture in western Europe is less clear than in Figure 7b. Thus the fast direction is perpendicular to the Alpid arc, and this reinforces the negative anomaly produced by the first azimuthal term. If the proposed association of the azimuthal variations of the

travel time anomalies with the Alpid belt is correct, the structure associated with the Alps would have to extend well into the upper mantle.

On a global scale, the azimuth independent term (Figure 8a) correlates with the tectonic nature of a given region, as pointed out by Cleary and Hales [1966]: the shields are fast (western and central Australia, Siberia, South Africa, Canada), and tectonically active regions are slow (Tasmania, the west coast of North America, the Middle East, Iceland).

There seems to be a tendency among the coastal stations for the slow direction to point toward the sea: this might be associated with the relative slowness of the downdip propagation of the waves that encounter the thickening crust and possible depression of upper mantle iso-velocity surfaces. In any case, the first azimuthal term has to do with major structures in the crust and upper mantle. A simpler picture emerges from the distribution of the second azimuthal term. There are clusters of stations that show very similar behavior: Tasmania and southern and western Australia, southern Africa, a continuation of the Alpid belt into Asia, most of the Siberian stations and India, for example.

The density of station distribution is, on the global scale, much lower than that for Europe or the western United States, and therefore we do not have an opportunity to observe in comparable detail some of the gradual and sometimes abrupt but consistent changes in the pattern of different terms in station anomalies.

Establishment of the fact that under favorable circumstances, the azimuth-dependent terms show a high degree of consistency and relate to tectonic features brings the promise of an additional tool in studying the heterogeneity and anisotropy of the mantle.

The slow directions found from the second

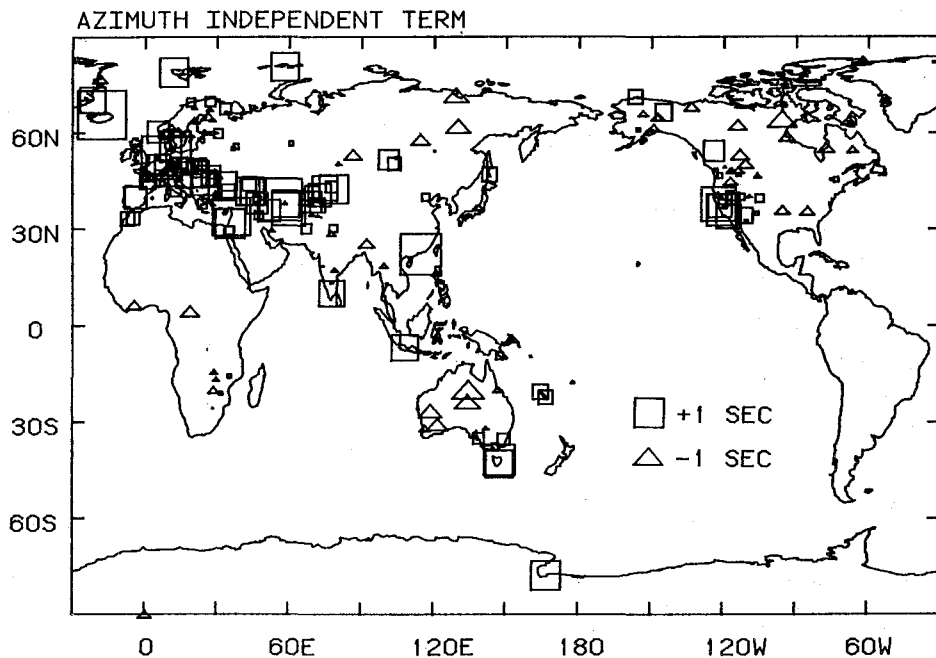


Fig. 8a. Global distribution of the constant correction term. For details see caption to Figure 6a. Observational threshold 1000 observations in 15 azimuthal windows.

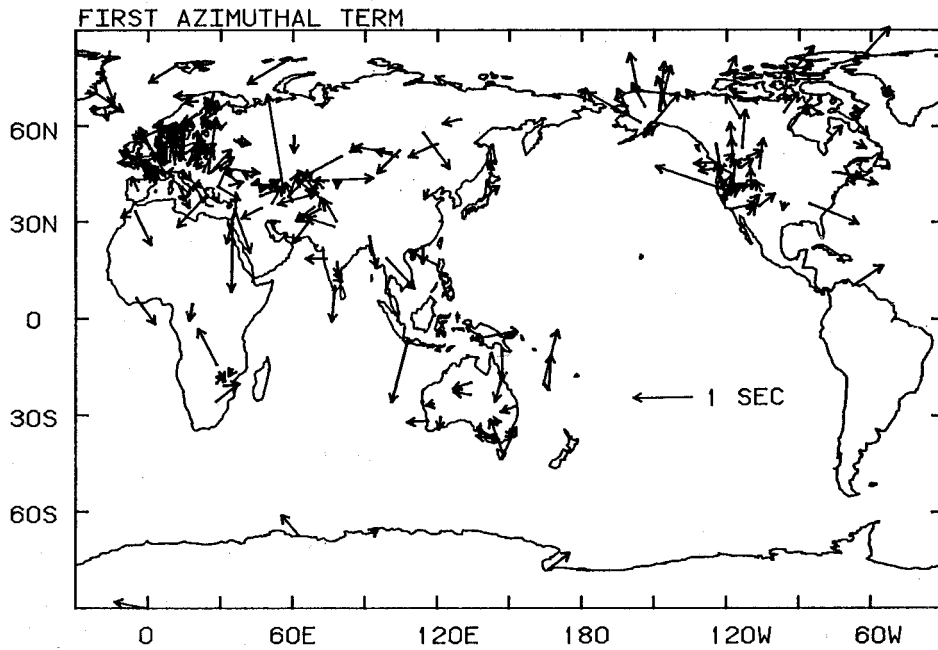


Fig. 8b. Global distribution of the first azimuthal term. For details see caption to Figure 6b.

azimuthal term bear an interesting and potentially important relationship to the direction of principal stresses in the crust. For example, the direction of greatest principal compression is NNE-SSW in southern California, EW in the southern midcontinent and NW-SE in the Colorado Plateau [Zoback and Zoback, 1980]. The axes of horizontal extension are NW-SE in the Basin and Range province and SW-NE in the Colorado Rocky Mountains. These directions are

roughly parallel with the slow directions of the second azimuthal station term. The principal stress axes are close to NS in the Pacific Northwest, swinging to NW-SE in the northern Rocky Mountains and to SW-NE in the northwestern midcontinent. The station terms show the same trend. The slow directions in South Africa and southern Australia are more-or-less normal to the African Rift and the Tertiary basalt trend in Australia, respectively. Crustal stress presum-

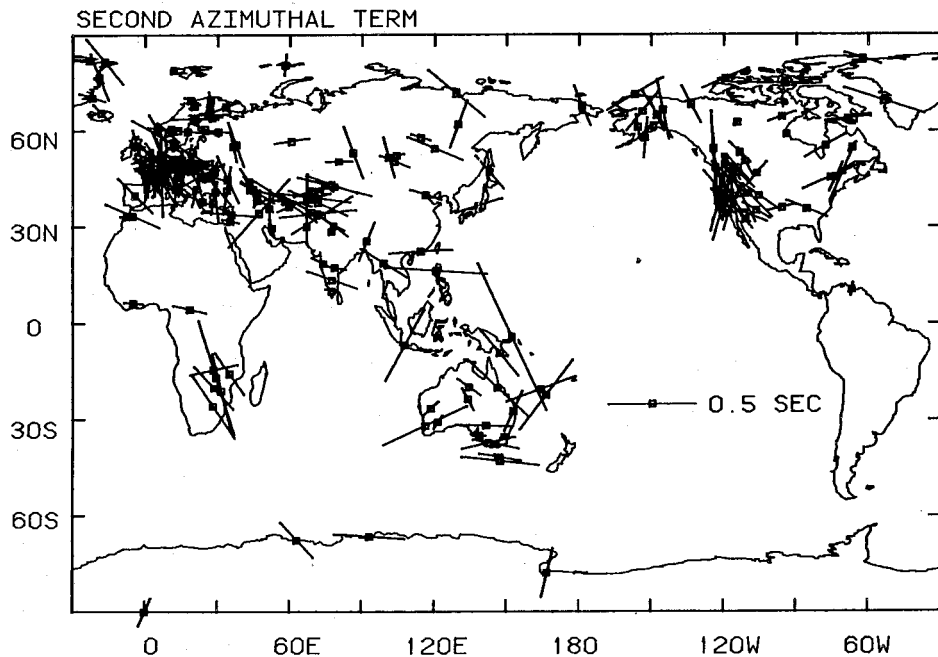


Fig. 8c. Global distribution of the second azimuthal term. For details see caption to Figure 6c.

ably is at least partially the result of mantle flow. The stress or flow-induced orientation of crystals in the upper mantle should therefore be related to stress directions in the overlying crust.

Acknowledgments. It is a great pleasure for us to dedicate this paper to Anton Hales who, in large measure, is responsible for its existence. Anton's dedication to the understanding of all of the uncertainties and subtleties in the travel time and location problems was a constant source of inspiration to us, and his personal advice and vast experience were tapped on many occasions. His studies with John Cleary provided many insights which we have profited from. Alan Douglas provided us with a detailed critique of our earlier travel time studies and convinced us that relocation of many events was a prerequisite for further progress in this field. This research was supported by National Science Foundation grants EAR81-20944 (Harvard) and EAR77-14675 (Caltech). Contribution 3801 of the Division of Geological and Planetary Sciences, California Institute of Technology, Pasadena, California 91125.

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(Received June 14, 1982;
accepted October 22, 1982.)