Geophysica (1992), 28, 1-2, 1-14

Correcting of Location of Teleseismic Events Obtained From Recordings at FINET Seismic Stations in Central Finland

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(Received: November 1992; Accepted: January 1993)

Abstract

An algorithm is presented to improve the location accuracy of the three partite network in central Finland. It was found that systematic teleseismic slowness anomalies up to 0.01 s/km are observed in central Finland. These anomalies are probably caused by the inhomogenities in the crust and upper mantle near receiver. The resulting epicenter mislocations can exceed ten degrees at certain areas, the median value being 639 km.

When the observed slowness anomalies are taken into account it is possible to form a correction function to improve the location estimate.

The correction function algorithm as well as the results obtained from it are presented. The correction function is computed from 667 associated teleseismic epicenters provided by NOAA. After corrections the median error is 173 km.

1. Introduction

One of the most appealing topics of seismic arrays and networks is their suitability for the measuring the slowness and azimuth of the arriving signal. The location of a seismic event is usually based on the arrival of P-waves into a two dimensional network with *N* stations.

In the last three decades more effective locating and detection systems for arrays and networks have been published (*Vanderkulk* 1965, *Steward* 1977, *Mykkeltveit* and *Bungum* 1984, *Ingate et al.* 1985, *Bratt* and *Bache* 1988, *Husebye* and *Ruud* 1989, *Bratt et al.* 1990, *Stead et al.* 1990) and the deployment of 3-component stations has provided fast although less precise location methods (*Roberts et al.* 1989, *Ruud* and *Husebye* 1992, *Tarvainen* 1992a, 1992b). The 3-component stations can work adequately at local and regional distances and may in some cases match with network capability. In general, especially at teleseismic distances the accuracy of networks is always superior.

Published by the Finnish Geophysical Society, Helsinki

The location itself is based on the standard earth model with symmetric velocity distribution. From travel time tables like *Jeffreys* and *Bullen* (1967), *Herrin et al.* (1968) and *Kenneth et al.* (1991) the distance-slowness dependence can be found. The slowness as a function of distance according to *Jeffreys* and *Bullen* (1967) and *Kennett* (1991) are shown in Figure 1. The function works adequately in so-called teleseismic window at distances from 20° up to 90°. For distances less than 20° the function is unstable owing to various P-waves arriving first at the station. Beyond 90° the function $dT/d\Delta$ is almost constant and the determination of the distance is unreliable. Most of the location methods are based on the estimation of the slowness vector from the network stations (*Julian* 1973, *Buland* 1976).

The location accuracy depends strongly on the aperture of the array. If we are dealing with arrays having apertures less than 30 km, the teleseismic locations are not reliable.

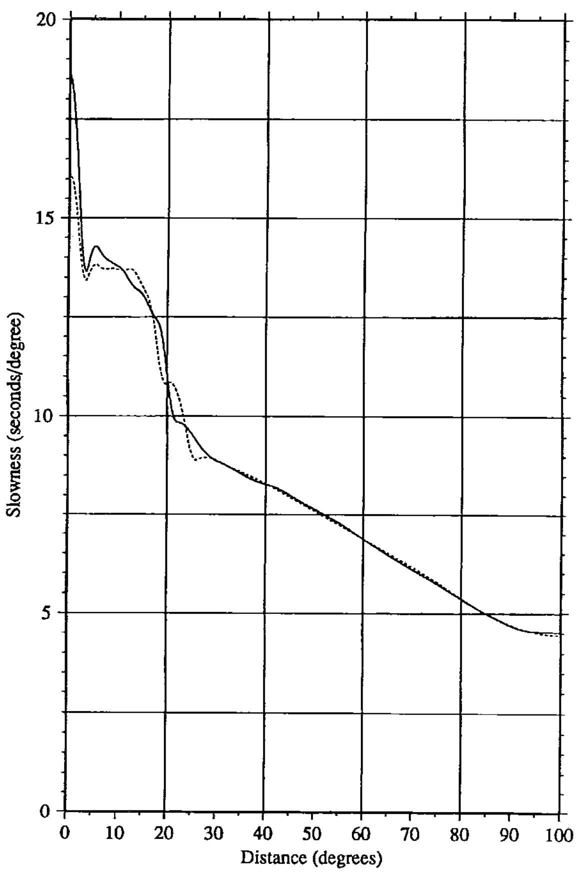

Fig. 1. The slowness of first arriving P-wave as a function of distance. The continuous line presents the Jeffreys-Bullen model (1967) for the source at the depth of 33 km and the dashed line is according to the IASPEI 1991 model (Kennett 1991) for depth 35 km.

There are numerous studies, which show, that teleseismic P-waves in general have slowness and azimuths which deviate from the values according to the spherical symmetric earth velocity models. Consequently these errors affect the location accuracy. The earthquake location errors can be divided into two distinct classes. There are systematic and random scattering (*Douglas* 1967, *Pavlis* and *Hokanson* 1985, *Pavlis* 1986, 1992). *Pavlis* and *Hokanson* (1985) showed different epicenter off-sets on opposite sides of faults. *Jordan* and *Sverdrup* (1981) studied locations of earthquake clusters at certain areas in the southern Pacific areas. *Engdahl* and *Lee* (1976) studied the location errors of local earthquakes by ray tracing. The structure effects on the location errors beneath large aperture arrays was also studied by *Aki et al.* (1976), which was one of the first tomographic studies in the whole world.

The topics of this paper is to introduce a procedure to improve the accuracy of the location of teleseismic events using the *FINET* seismic network in central Finland, consisting stations KEF (62.2°N, 24.9°E), SUF (62.1°N, 26.3°E) and KAF (62.7°N, 26.2°E). The precise station configuration and instrumentation are presented by *Teikari* and *Suvilinna* (1992).

The method is very similar to the one presented and used by *Noponen* (1971) for the seismic network in southern Finland around Helsinki. The slowness anomalies at the network in southern Finland differed remarkably from the values achieved in this study and the needs to improve the accuracy of daily locating analysis claimed to determine the correction function. The configuration of the network is shown in Figure 2. The aperture

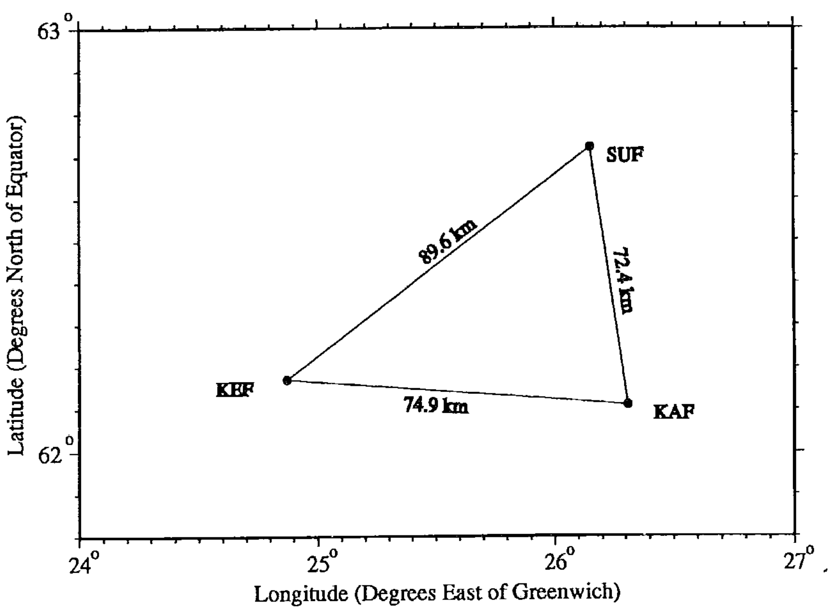

Fig. 2. The configuration of the *FINET* seismic network in central Finland.

of the network is roughly 70 km, for which a plane wave fitting can be applied for teleseismic events. The network was established in 1976 and upgraded in 1990. In the last few years as a reference station has been station KAF. Data from this network are transmitted via telephone links to Helsinki hub, where the data analysis is done. The correction study presented in this paper is performed using only short-period vertical recordings of the network. No 3-component analysis or long-period data of KAF is presented. The algorithm in detail is presented in Appendix.

2. *Determining the residual matrix*

The correction algorithm was computed using the preliminary epicenter bulletins (PDE) of 1987 distributed by NOAA. Because we studied teleseismic events, at distances more than 20°, we selected only those events which had bodywave magnitude equal to or greater than 4.5.

The event selection was done by computing the arrival time for every event by

$$ar_i = Ho_i + t(\Delta_i)$$

where ar_i is the computed arrival time Ho_i is the origin time according to NOAA epicenter bulletin and $t(\Delta_i)$ is the travel time depending on the distance and depth. For calculating the travel time values the bi-cubic spline fitting was used. The bi-cubic approach was chosen for more accurate arrival time estimations than using linear interpolation, especially for events at low or deeper focuses, when the depth allowances could cause several seconds bias. When this arrival time was calculated it was compared with P-phase readings at KAF station. If P-readings were found within a 12 seconds window the event was associated with that reading. If the event had also azimuth and velocity (slowness) data, computed via plane fitting, at the network, the corresponding residuals were computed. This was repeated for all the PDE data. The searching method is shown by the flow-chart in Figure 3. Finally a residual matrix which contained 667 events was formed.

According to this matrix the observed slowness values at the network in central Finland is dominantly smaller than the calculated values, so giving too long distances. The difference between observed and calculated slowness as a function of the azimuths is shown in Figure 4. The best fitting sinusoidal function is

$$\Delta\tau = 0.0475\, sin(\alpha + \varphi) - 0.015$$

where $\Delta\tau$ is the difference between observed and calculated slowness (in s/km), σ is the phase angle and φ is the station azimuth.

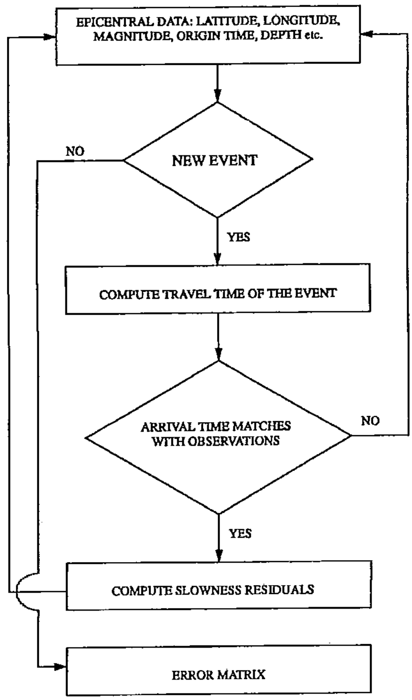

Fig. 3. The flow-chart expressing the formation of the residual matrix from the PDE information of NOAA.

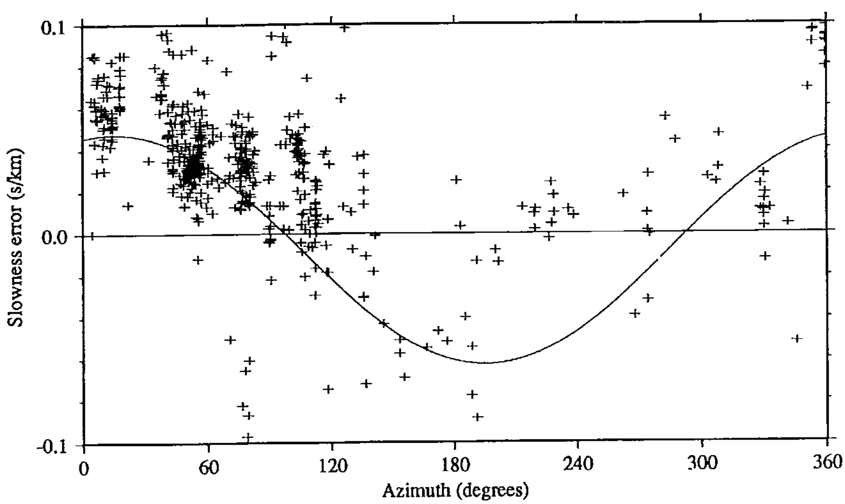

Fig. 4. The slowness errors (crosses) as a function of the true azimuth. The concentration at azimuths around 60° is caused by the island arc events in circum-Pacific areas. The solid line presents the best least square sinusoidal fit.

3. *The results of the correcting algorithm*

In Figure 5 the location errors obtained from non corrected slowness values are shown. The greatest location errors appear at island arcs in Pacific region. The median of the errors being 639 kilometers from the true epicenter.

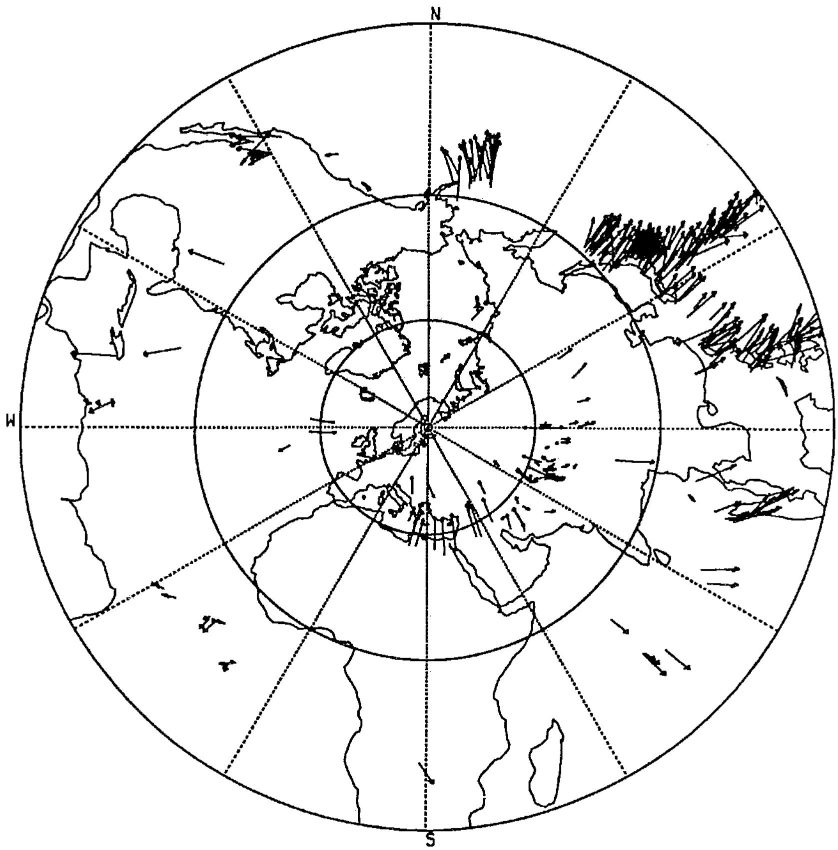

Fig. 5. Errors of locations without corrections. The base of the arrow locates at the PDE epicenter and the head points to the location obtained from non corrected readings. Events with deviations more than 10° are omitted for clarity. The cocentric circles present distance increments of 30° from the network in central Finland.

When the correction function was taken into account the error became remarkably smaller as shown in Figure 6. Because the main contribution to the correction matrix came from the areas of more frequent events, the correction function affected most efficiently at circum-Pacific areas. The median error of location was as small as 173 kilometers. In

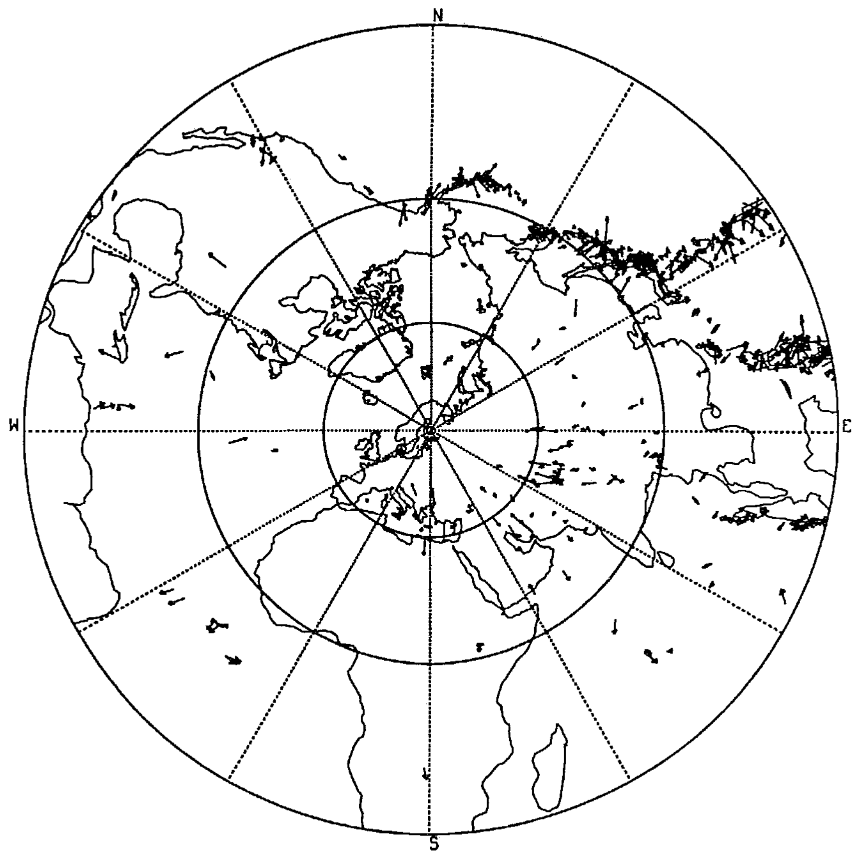

Fig. 6. Location errors after using the correction function. The explanation of the arrows as in Fig. 5. The errors have become remarkably smaller than without corrections. For areas with few events the correction function does not work adequately.

Figure 7 there are shown the cumulative error of location without correction and after using the correction algorithm. The location accuracy was improved strongly. After corrections only 67 events had a location error more than 439 kilometers. These events evidently are those having only little weight in forming the correction matrix.

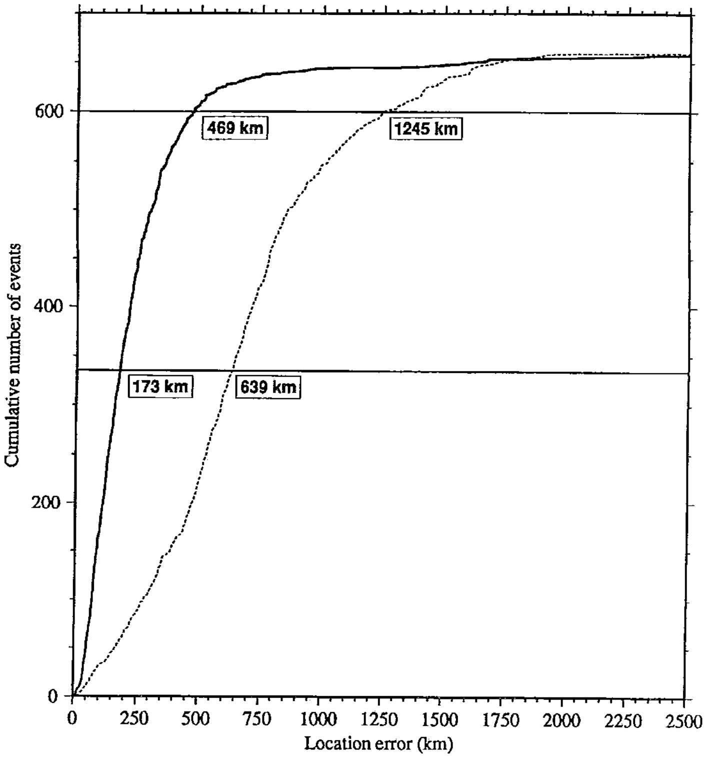

Fig. 7. Cumulative error functions prior to the corrections and after it. The dashed curve shows the distribution of the location errors without corrections. The horizontal lines present 50 % and 90 % percentiles and the kilometer values corresponding location errors respectively. The median error is 639 kilometers without corrections and 173 kilometers with corrections respectively.

4. *Discussion and conclusions*

We have presented an algorithm to improve the location accuracy of seismic events at teleseismic distances. The method is based on the residuals of observed and true velocities and azimuths. The structure of the crust and mantle beneath the source or receivers are not taken into account, even though some conclusions might be done. The total effect of some crustal velocity gradient on azimuthal misplacement in the direction perpendicular to the ray along the entire ray path can easily be shown to fulfill the following integral

$$X = \int_0^R r \frac{D_i v}{v} dr$$

where X is the azimuth misplacement, R is the length of the ray path and $D_i v$ is the partial derivative of the horizontal velocity perpendicular to the ray. Further dr is the distance element of the earthquake. Evidently any velocity gradient near the receiver, greater dr values, has larger effect on the mislocation than a corresponding gradient near the source. Possible near station scattering caused by the crustal and mantle inhomogeneties may affect strongly the slowness and azimuth estimations.

A complex Moho topography can cause the observed slowness anomalies. *Noponen* (1971) found the variation of Moho by 7 up to 10 kilometers beneath the network in southern Finland. We did not have enough station resolution for that kind of study, although some estimations of Moho depth variations which were done, match well with results of *Luosto* (1991).

It is possible that a region of lateral velocity anomaly layer locates north-east from the network. In practice the great majority of the events were observed at directions from 30° to 90°, which may cause bias to the deviations. In the other hand the distribution in any direction is systematic and such biases are hard to find.

In general the correction algorithm works well at any distances beyond 20°, at teleseismic distances. It was found to work in some cases also well beyond 85°, where the first arrival is the diffracted P-wave notations.

The method is based purely on statistical approach, we did not study the depths of events, which is neither possible to do by using array slowness estimations only.

Acknowledgments

The authors want to express their thanks to Professor Heikki Korhonen, the former director of the Institute of Seismology, for encouraging advice during the work. The comments of Dr. Eystein S. Husebye from NORSAR, Norway were well-come to improve the contents of the manuscript. The essential financial support was received from the Finnish Cultural Foundation (Maili Aution säätiö).

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Appendix

Determining the correcting algorithm

The arriving wavefield in an array far from its source can be assumed to be a plane wave. If the sensors are assumed to locate at the same altitude the arrival time at sensors N_i can be fitted by plane using the least square method. The equation of the best fitting plane is

$$Ax + By + Ct + D = 0$$

where x and y are the Cartesian coordinates of sensors and t is the time. The minimum number of stations is 3. The apparent velocity passing the sensor network can then be expressed by

$$V_{app} = \frac{\sqrt{A^2 + B^2}}{C}$$

where V_{app} is the apparent velocity. Corresponding azimuth φ referred to north can be evaluated by

$$\varphi = arc\ tan\ (A/B) \tag{3}$$

In the case of three stations network we can set the zero of coordinates in station with minimum time of the first arrival and the plane can be computed as the following set of linear equations

$$A = y_2t_3 - y_3t_2$$
$$B = t_2x_3 - t_3x_2$$
$$C = x_2y_3 - x_3y_2$$

where t_2 and t_3 are the time differences of first arrivals compared with station 1. The V_{app} is the function of distance and depth of the source. At distances less than 25° several different P-waves can be observed, which causes strong scattering of P-wave slowness at those distances. Because it is more convenient to use just the slowness components of observed velocity.

If a seismic signal passes through a network of 3 stations with an azimuth of φ as expressed in Figure (8), the observed slowness can be divided into x- and y-components by

$$S_x = -\frac{\cos\varphi}{\overrightarrow{V_{app}}}$$

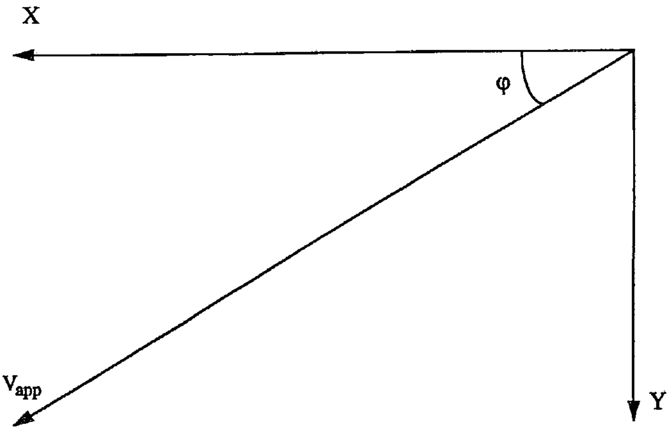

Fig. 8. The observed Cartesian components of slowness in a two-dimensional coordinate system.

$$S_y = -\frac{\sin\varphi}{\vec{V}_{app}}$$

The resultant vector of these components is the observed slowness:

$$\vec{S} = \sqrt{S_x^2 + s_y^2}$$

From the travel time tables for the referred events the components of the calculated slowness vector are:

$$\hat{S}_x = -\frac{\cos\varphi_{true}}{\hat{V}}$$

$$\hat{S}_y = -\frac{\sin\varphi_{true}}{\hat{V}}$$

where $\hat{V}$ is the apparent velocity corresponding a known distance. The components of the slowness vector of an indexed event are:

$$\Delta S_i^x = \hat{S}_i^x - \vec{S}_i^x$$

$$\Delta S_i^y = \hat{S}_i^y - \vec{S}_i^y$$

If we have a set of events for which we know the exact locations, we can estimate the systematic deviations as a function of slowness in the following way:

$$\Delta S_x(\vec{S}) = \frac{\sum_{i=1}^{N} w_i(\vec{S})\,\Delta S_i^x(\vec{S})}{\sum_{i=1}^{N} w_i(\vec{S})}$$

where

$$w_i = \frac{1}{|\vec{S_i} - \vec{S}|^2} = \frac{1}{\sqrt{(S_i^x - S_x)^2 + (S_i^y - S_y)^2}^{\,2}}$$

The weighing function wi will attenuate the effect of the events, for which the slowness vector $\vec{S_i} = (S_i^x, S_i^y)$ is strongly different, i.e. the events which are close to the studied value $\vec{S_i}$, give the most of the contribution. Further the corrected components for slowness vector can be calculated. This new slowness vector components are

$$S_x^{new} = S_x + \Delta S_x$$
$$S_y^{new} = S_y + \Delta S_y$$

The corrected slowness values can then be used in location procedure.