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AN EVALUATION OF HYPOCENTER LOCATION TECHNIQUES WITH APPLICATIONS TO SOUTHERN UTAH: REGIONAL EARTHQUAKE DISTRIBUTIONS AND SEISMICITY OF GEOTHERMAL AREAS

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An Evaluation of Hypocenter Location Techniques With Applications to Southern Utah: Regional Earthquake Distributions and Seismicity of Geothermal Areas

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ABSTRACT

Three techniques for the computation of earthquake hypocenter locations were compared through empirical results of synthetic test cases utilizing four different computer programs. The three approaches: (1) the single event method, (2) the master event method, and (3) the joint hypocenter determination method, were analyzed with respect to their application to regional (epicenter-to-station distances from 10 to 500 km) and local (epicentral distances from 0 to 70 km) seismic network recording situations. It was found that the joint hypocenter technique can significantly correct for inadequacies in assumed velocity models by simultaneous computation of station adjustments. Earthquakes located using the joint hypocenter technique are located more precisely with respect to each other. The joint epicenter location technique was then applied to relocations of epicenters in southern Utah below 40°N latitude with the intent of resolving some spatial relationships of epicenter occurrence on this regional scale. Earthquakes in Utah occur in a diffuse zone 150 km to 200 km wide that coincides with the physiographic boundary of the Colorado Plateau. Swarm activity is prevalent where the zone changes from a north-south to a southwest orientation, but except for a cluster northwest of Cedar City, the only major alignments of epicenters are north-south.

On a local scale an objective was in determining some

relationships of epicenters to two known geothermal resource areas: Cove Fort and Roosevelt Hot Springs, Utah. The differences in the character of these two areas as expressed by the earthquake occurrence was emphasized by the jointly determined relocations. Areas of possible active east-west and northeast faulting near Cove Fort were delineated, but it was verified that seismic activity in the Milford area is sparse. Attention was given to possible effects of lateral inhomogeneities in the velocity model on earthquake locations. Evidence for substantial lateral variations in the velocity models used in the locations was contributed from an analysis of P-wave residuals from broadside refraction data. Arrival times of P-waves were found to be early near the Opal Mound fault, possibly due to local siliceous cementation of alluvium. Contacts between the gneissic basement and the granitic pluton could account for some of the observed delays in P-wave arrival times. Effects of lateral velocity variations on earthquake locations were minimized by precise relocation using joint hypocenter determination, thus aiding is defining seismically active areas. Results of earthquake relocation for the Cove Fort-Roosevelt Hot Springs KGRA's and adjacent areas indicate that methods of systematically and precisely locating earthquakes can better contribute to understanding the seismicity and velocity structure and the relationships of geothermal areas to regional and local stress fields and geologic characteristics.

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CHAPTER 1

INTRODUCTION

The accurate location of earthquake hypocenters has been the concern of many seismologists since Geiger first introduced the least squares approach to the problem in 1912 (Lee and Lahr, 1975). Most computer algorithms presently available are based upon similar assumptions concerning the formulation of the forward problem and the statistics involved, and work quite well for the ideal recording situation with simple flat-layered velocity structure. Problems, however, exist in the common situations of lateral inhomogeneities in the velocity structure, unknown velocity structure, or variable instrumental coverage ranging from good to poor over short distances. These problems are frequently ignored.

Considering the growing number of methods available to treat the hypocenter location problem, are some more suitable to specific regions than others? A primary objective of this thesis is an evaluation of three approaches to the computation of epicenters: (1) the single-event method, (2) the master-event method, and (3) joint epicenter determination. The fact that a numerical solution is obtained is not necessarily sufficient justification for using a particular technique. The real question is, do the results add significantly to the understanding of the problem? Application to earthquakes in southern Utah is of chief interest here, particularly the resolution of epicenters and the delineation of epicentral patterns to examine spatial associations and relations to geothermal areas.

Compared to other regions covered by the University of Utah seismic network, relatively little work has been done in southern Utah. Because of the sparse station distribution in this area, with an average station-spacing of greater than 100 km,the problems in earthquake location are compounded. This emphasizes the need for optimum techniques for relocating epicenters.

It has been noted (e.g., Ward, 1972) that there may be a correlation between seismicity and area of geothermal potential. Some workers (e.g. Smith, 1978; Anderson, 1978) have considered the possibility that the trends of epicenters and Quaternary deformation in southern Utah, which seem to show a change in orientation at about 38° N latitude, have a tectonic significance that may be related to the occurrence of several known geothermal resource areas. But because the quality of the solutions was relatively unknown it was difficult to make any definitive statements about possible trends observed. Local microearthquake surveys (e.g., Olson 1976), because of their limited areal coverage, have not clarified regional relationships.

Of supplementary consideration in accurately determining the seismicity of southern Utah is the recent interest in that area for the siting of U. S. strategic missile systems and the development of large-scale commercial power plants.

Further incentive for attempting this project was generated

during the course of revising earthquake locations in Utah for a 1962-1978 catalog (Arabasz, and McKee, 1979).
Relocations were initially computed with HYPO71, a program written by Lee and Lahr (1975). Subsequently, HYPOELLIPSE (Lahr, 1979) was adopted for the systematic relocations. The relative merits of using HYPOELLIPSE and discussions regarding the best procedure for locating regional seismic events suggested that a more careful analysis of the methods and problems specific to earthquake location in Utah was in order.

The analysis required extensive testing of several computer programs in order to quantitatively determine their respective applicability to the particular characteristics of earthquake recording in Utah. This involved the generation of synthetic data to simulate epicenters recorded by seismic arrays in southern Utah, both on a regional and a localized scale. The testing indicated the limitations in the accuracy that can be anticipated for locations in various areas of the state relative to the instrumental coverage. It also enabled an estimation of the sensitivities of particular location techniques to instabilities in solutions introduced by variable station coverage. The usefulness of the techniques can now be shown specifically for southern Utah, but the study incorporated enough generality to enable extrapolation of major conclusions to any regional or local seismograph stations array.

The basic theory and methods of earthquake location techniques are outlined in Chapter 2. More detailed and comprehensive explanations of the methods and programs are presented in Chapter 3, the body of which concerns the results of testing and comparison analysis. In Chapters 4 and 5, the methods and conclusions discussed in Chapter 3 are applied to specific regions in Utah, using relocation of epicenters to define more precisely any linear or clustering trends in the data. Characteristics of earthquake occurrence are then discussed in light of their structural and tectonic significance. The focus is first on a regional scale (Chapter 4), subsequently narrowing to specific localities around the Cove Fort-Roosevelt Hot Springs localities (Chapter 5). The issue is assessment of the value of regional and local seismicity studies in locating or studying potential geothermal areas.

CHAPTER 2

THEORETICAL CONSIDERATIONS AND NUMERICAL APPROACHES TO THE HYPOCENTER LOCATION PROBLEM

Presented here is a summary of the theoretical basis that all of the techniques to be considered share in common. A brief review of the fundamental aspects is followed by a short section on the commonly employed numerical techniques used in solving the problem. Implications of statistical results obtained by the least squares approach are discussed, as well as frequent misconceptions and possible problems related to some basic assumptions.

Theory

In the earthquake location problem we are interested in four parameters: the three spatial coordinates and the origin time. These are nonlinear functions of arrival times of various phases (usually P and/or S) observed at stations recording the event. The arrival times are dependent on the positions of the stations relative to the event as well as the velocity-depth relationships along the ray path. Essentially, then, this is an inverse problem based upon several simplifying assumptions about source mechanisms governing earthquakes. A point source of energy generates seismic waves that arrive at the recording stations; these rays must somehow be traced back to their source using information derived from their observed arrival times.

The major elements of the following discussion can be found in basic references on statistics or solution of parametric equations (e.g., Scheffé, 1959; Draper and Smith, 1966; Beck and Arnold, 1977). My purpose is to clarify certain assumptions and methods that have been implicitly or explicitly applied in earthquake location programs, often without proper explanation if not incorrectly. Also, comparison of the programs and methods requires at least a cursory understanding of the theoretical and statistical background used to develop the criteria by which the accuracy of the resulting solutions obtained is judged.

In the following discussion, all variables in capitals refer to matrices, small letters are vectors, and subscripted variables designate scalar quantities. Estimators, whether vector or scalar, are designed by a hat over the variable.

The problem as it is presently approached by seismologists is essentially an analysis of variance or a linear regression on four variables (Bolt and Freedman, 1968). Let x_j, y_j, z_j, and t°_j be the spatial coordinates and origin time, respectively, of the j^{th} earthquake. If n is the total number of arrival times observed for m earthquakes, then t is the (n x 1) vector of observations, y is the (4m x 1) vector of parameter changes (dx, dy, dz, dt°), and A is the (n x 4m) matrix referred to as the system, condition, or coefficient matrix. The vector t is expressed as a linear function of the parameters; then the hypocenter is adjusted until the difference between the observed and calculated arrival times, the residual r_{ij} for the i^{th} station and j^{th} earthquake, is minimized. The linearized function can be expressed as a Taylor series expansion of the travel time about some initial guess, and the estimate of the residual becomes

$$\hat{r}_{ij} = r_{ij} + \frac{\partial t_{ij}}{\partial x_j}dx_j + \frac{\partial t_{ij}}{\partial y_j}dy_j + \frac{\partial t_{ij}}{\partial z_j}dz_j + \frac{\partial t_{ij}}{\partial t^o_j}dt^o_j + e_{ij} \qquad (2.1)$$

neglecting higher order terms. In matrix notation,

$$\hat{r} = Ay + e. \qquad (2.2)$$

The vector of residuals, $\hat{r}$, is only an estimate because we can never know the true error in the solution. The true error would be $r = y - \hat{y}$, where $\hat{y}$ is the vector of estimated parameters. Instead we are actually solving the system $\hat{r} = Ay - A\hat{y}$, or $\hat{r} = Ar$, since $Ay = t$ (Conte and de Boor, 1972). To simplify notation, however, in subsequent discussion the hat over the r is dropped.

We seek to minimize the quantity

$$Q_j = \sum_{i=1}^{l} w_{ij}r^2_{ij},$$

or $$Q = \| t - Ay \|^2 \qquad (2.3)$$

where l is the number of observations for a particular earthquake j ($n = l \times m$) and w_{ij} are linear or non-linear weights applied to each station residual. Various weighting procedures will be discussed in sections treating specific programs. The matrix A in equation 2.3 has been redefined to include the weights w_{ij}. To retrieve the least squares unbiased estimates of y that will minimize 2.3 the normal equations are applied (Scheffé, 1959):

$$\frac{\partial Q_j}{\partial y_j} = 0 \ ,$$

or

$$A^T A y = A^T r \ ,$$

then,

$$y = (A^T A)^{-1} A^T r \quad . \qquad (2.4)$$

The problem is solved iteratively, with the new estimates of y added to the previous hypocenter parameters until Q_j is reduced to some specified minimum value. Alternative methods for terminating the iteration procedure are discussed in Chapter 3.

There are several assumptions which are implicit in the above development: i) the errors are additive, ii) E (e) = 0, or the expectation of the error vector e is zero, and iii) E $(e^T e)$ = V (e) = $\sigma^2 I$, or the variance of the errors is constant and uncorrelated (Beck and Arnold, 1977). In other words, the errors are assumed to be independent and normally distributed as N $(0, \sigma^2)$ (Bolt and Freedman, 1968). In general, these assumptions can probably be considered valid for random errors due to mistakes and inconsistency in the timing and picking of earthquake seismograms or to the misidentification of phases. However, for the residuals, which are estimates of the total error, the assumptions may not be justified. It has been pointed out that the total residual is composed of random errors plus non-random station and source terms, and that unless these are estimated separately the errors are not normally distributed (Bolt and Freedman, 1968). Buland (1976) has reported that certain characteristics of the residuals for a localized array are known from experience. Under these circumstances the ordinary least squares procedure is not recommended (Beck and Arnold, 1977; Draper and Smith, 1966), and application of the maximum likelihood estimation involving minimization of the sum of absolute errors rather than squares of errors may be more appropriate (Buland, 1976).

As previously mentioned, the residual gives an estimate of the error in the solution; therefore, careful analysis of the residuals may help to confirm or invalidate the assumptions made (e.g., tests for normal distribution). However, the magnitude of the residual does not always give an indication of the magnitude of the true error in the solution, as is often thought to be the case. (For a simple example see Conte and de Boor, 1972, p. 143.) How well the residual approximates the true error is dependent upon the "size" of the system matrix and its inverse. The "size" is measured in terms of some matrix norm, which is always greater than or equal to 1, and is denoted as $\|A\|$. A measure of the condition of A is the condition number,

$$C(A) = \|A\| \cdot \|A^{-1}\| \ . \tag{2.5}$$

We already have that

$At = y,$ and similarly,

$Ae = r.$

Then $e = \frac{r}{A}$,

(2.6)

and taking norms (Ralston and Rabinowitz, 1979),

$$\|e\| = \frac{\|r\|}{\|A\|} .$$

Also, $$\|t\| = \|A^{-1}\| \cdot \|y\| . \qquad (2.7)$$

Dividing (2.6) by (2.7), a lower bound for the relative error is

$$\frac{\|e\|}{\|t\|} \geq \frac{1}{\|A\| \cdot \|A^{-1}\|} \cdot \frac{\| \quad \|}{\| \quad \|} \qquad (2.8)$$

A similar derivation (Ralston and Rabinowitz, 1978) results in an upper bound, so that the relative error of the solution can be expressed in terms of the relative residual as

$$\frac{1}{C(A)} \cdot \frac{\|r\|}{\|y\|} \leq \frac{\|e\|}{\|t\|} \leq C(A) \frac{\|r\|}{\|y\|} . \qquad (2.9)$$

So if C(A) is close to unity, the relative error will be essentially equivalent to the relative residual, and the system is described as well-conditioned. But if the condition number is large, the relative residual places very little constraint on the possible values of the relative error (Conte and de Boor, 1972). Furthermore, it has been shown (Buland, 1976) that C(A) for the earthquake location problem can be very large quite near the array stations, and if the station distribution is poor, or if only P arrivals are used, the condition number can be large everywhere. That is, the system matrix A may always be ill-conditioned.

The size of the condition number can also indicate how much of the error in the solution vector (y) is contributed by error in the data vector (t) (Ralston and Rabinowitz, 1978). However, in most practical cases, it is very difficult to compute A^{-1} exactly, so C(A) may be quite inaccurate; in fact, it is seldom computed. The discussion here was to emphasize the inadequacy of the use of the residual as a direct estimate of the error in the solution.

Correlation of Residuals and Parameters

In general there are 4m parameters (recalling that m is the number of earthquakes) estimated by n observations. The residuals have only n - 4m degrees of freedom; thus they are not independent. It can be shown that the correlations in the residuals depend entirely on the elements of the A matrix (Draper and Smith, 1966, p. 93-94). Correlations of coefficients of A are of course correlations of parameters. It is known from empirical results that origin time and depth generally exhibit a high correlation coefficient. This may be seen by examining the elements of the system matrix in more detail. The α_{ij} are the coefficients of the unknown parameters, or the partial derivatives of the travel times with respect to each parameter, and can be written as:

$$\begin{aligned}
\alpha^1_{ij} &= \frac{\partial t_{ij}}{\partial \Delta_{ij}} \cdot \frac{\partial \Delta_{ij}}{\partial x_j} \approx \frac{1}{V_{app}} \cdot \frac{\partial \Delta_{ij}}{\partial x_j} \\
\alpha^2_{ij} &= \frac{\partial t_{ij}}{\partial \Delta_{ij}} \cdot \frac{\partial \Delta_{ij}}{\partial y_j} \approx \frac{1}{V_{app}} \cdot \frac{\partial \Delta_{ij}}{\partial y_j} \\
\alpha^3_{ij} &= \frac{\partial t_{ij}}{\partial \Delta_{ij}} \cdot \frac{\partial \Delta_{ij}}{\partial z_j} \approx \frac{1}{V_{app}} \cdot \frac{\partial \Delta_{ij}}{\partial z_j} \\
\alpha^4_{ij} &= \frac{\partial t_{ij}}{\partial t^\circ_j} = 1 \quad .
\end{aligned} \tag{2.10}$$

Since the least squares procedure for earthquake location as described above is originally attributed to Geiger, these are commonly referred to as Geiger's coefficients. Writing them in this fashion gives some insight as to the relative significance of the earthquake parameters in the stability of the problem and their differing degrees of resolution in the ultimate solution. All of the coefficients include the reciprocal of the slope of the travel-time curve ($1/V_{app}$). V_{app} is the apparent velocity at the particular distance Δ_{ij}, (i^{th} recording station for j^{th} hypocenter) and can either be read directly from travel time tables or calculated from the velocity model. It is intuitively reasonable to suppose, and further verified from geometrical considerations, that the change in Δ_{ij} with respect to a change in depth is small when compared to changes with respect to x and y (longitude and latitude, usually). This becomes more pronounced as Δ_{ij} for the nearest station increases. Figure 2.1 illustrates this effect, and also shows the beneficial effect of even one distant station to enhance the azimuthal control. For depth, the difference in magnitude of the partial derivatives for each station becomes dependent only on the change itself. The derivatives for depth do not have the extra azimuthal effect at separate stations, which helps to assure differing parameter changes in x and y. At a particular distance from the array, which is a function of the array geometry with respect to the event, the station residuals are no longer affected independently by a change in focal depth. (This can also occur if the array has a poor distribution of stations with respect to distance, even if the event is within the array.) For an event at or outside this distance, any adjustment in depth can be directly compensated for at all stations by an adjustment in the origin time,

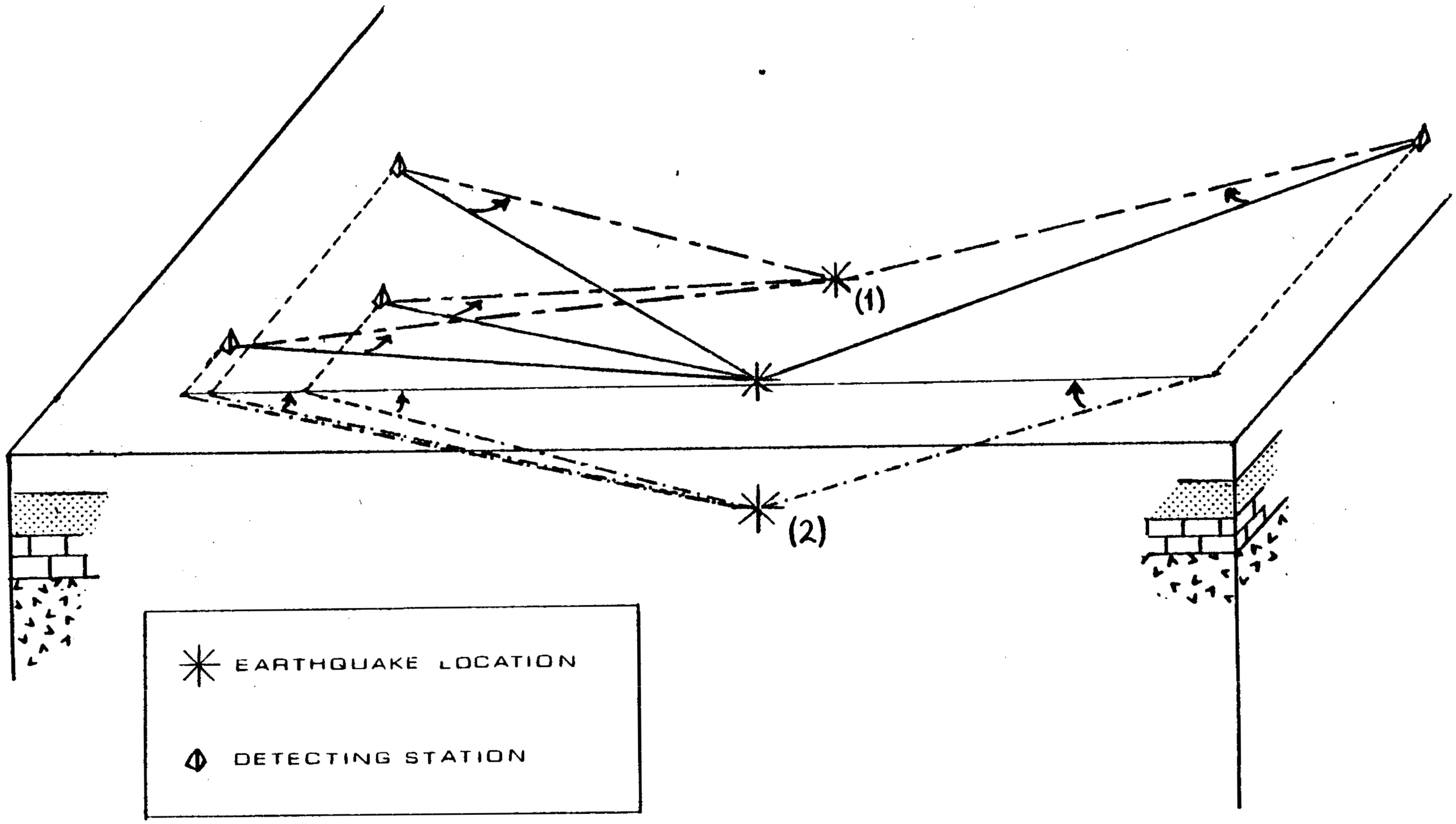


Figure 2.1. Diagram illustrating difference in magnitude of changes in hypocentral parameters with station distance. (1) change of epicenter. (2) change of focal depth (stations projected to line containing hypocenter). Changes in focal depth parameter will not be as well resolved.

since the corresponding change in distance is essentially equal for all stations and is therefore no longer a factor.

If there are no stations within a distance of 1 to 2 times the focal depth, the entire column of the system matrix associated with the depth parameter may approach zero, which will make the solution unstable if not unsolvable (reducing the rank of A by one). In addition, since the change in the residuals is relatively insensitive to depth (∂z_j), the control of the step-size for this parameter is not well-regulated and the depth may fluctuate wildly, seeming to converge to some meaningless and obviously erroneous values. The wrong value for depth then has an adverse affect on the epicenter location. For these reasons, some programs include a convention to fix the focal depths at some predetermined value and reduce the dimension of the condition matrix A. Focal depths are commonly fixed at some reasonable average value for a particular locality if the distance from the event to the nearest station is greater than 1 to 2 times the assumed focal depth or if the epicentral distance is greater than the array diameter.

Confidence Regions

The confidence ellipsoid commonly computed for earthquake location solutions is another statistical result that is often incorrectly interpreted. Flinn (1965) is usually credited with being the first to introduce the concept to seismologists concerned with the location of earthquakes. Flinn, and presumably most of those who have subsequently incorporated the techniques into their programs, did not misinterpret the meaning of the confidence regions, but the lack of adequate explanations left much room for later misconception.

The development of the equation for the joint confidence region as used by Flinn and other workers can be found in references on probability and statistics in problem solving (e.g. Draper and Smith, 1966; Beck and Arnold, 1977; Scheffé, 1959). The joint 100 $(1 - \alpha)\%$ confidence region for all parameters is expressed as

$$(y-\hat{y})^T A^T A(y-\hat{y}) = ps^2 F(p,n-p,1-\alpha) \; , \tag{2.11}$$

where p is the number of parameters (p = 4m) or elements of y, n is the number of observations, and s^2 is the estimate of error variance. This estimate is obtained from equation 2.3 as

$$s^2 = \frac{Q_j}{(n-p)} \; , \tag{2.12}$$

and is also referred to as the mean squared error or standard error (Draper and Smith, 1966). $F(p, n-p, 1-\alpha)$ is the $(1-\alpha)$ percentile point of the F distribution found from tables of the F or χ^2 distributions. Note that $(A^TA)/s^2$ is the inverse of the parameter covariance matrix, and from this the standard error for each parameter is $P_{ii}^{1/2}$, where $P = s^2 (A^TA)^{-1}$ (Flinn, 1965; Draper and Smith, 1966).

Confidence ellipsoids can be computed for the hypocenter (three spatial parameters), but are often computed just for the epicenter (two location parameters). Usually they are calculated during the final iteration. Sometimes an extra iteration is performed to obtain the statistics, since the elements of the computation are dependent on the magnitude of the last movement in parameter space.

The problem with presentation of the confidence ellipsoids has been pointed out by Evernden (1969a),and is summarized here. If the probability that the true hypocenter will be within the confidence region is X, then the correct interpretation of the 100% confidence region around the estimated hypocenter is as follows. In the frequency interpretation, in 100 estimates of hypocenters, 100(X) of the computed confidence ellipsoids will cover the corresponding true hypocenter. This does not say anything about the probable location error of the true hypocenter. In other words, what the confidence ellipsoid does <u>not</u> indicate is how far off the estimated solution is likely to be for a particular event. It does not place error bounds on the true hypocenter, which theoretically could lie anywhere within the confidence region or anywhere outside of it. The procedures for obtaining a probable error estimate are different, and involve computing solutions to a number of test cases and subjecting the results to methods of normal bivariate statistical analysis (Evernden, 1969a).

The value of the confidence ellipsoids lies in the fact that their orientation and size are dependent on their position with respect to the stations used to record the events (Flinn, 1965). Therefore they can give a qualitative indication of the capability of the seismic array in resolving hypocenters in a particular region. The orientation and elongation of ellipses can also show, for a specific area, whether trends seen in the epicenters could be due to a preferential dispersion of the solutions which is solely the result of the array's inability to locate them precisely.

Determinants

Another result that is obtained during the solution of the earthquake locations, and one that can give an estimate of the relative capability of the seismic array with respect to a given region, is the determinant of the matrix of normal equations, A^TA. This matrix is essentially the inverse of the covariance matrix for the parameters and hence the magnitude of its determinant can give an indication of how well the variance of the parameters is being restrained by the station distribution. In general, when the determinant is small, the variance and covariance terms of A^TA will be large (Flinn, 1965). The user could then control the amount of variance allowable in the parameters of the solution by documentation of an acceptable range of values for the determinant.

Numerical Methods

The purpose of this section is to point out some of the differences in the numerical techniques employed in each program to solve the normal equations and to assess these differences in terms of possible effects on the solutions. It is beyond the scope of this thesis to present a comprehensive analysis of the subtle numerical aspects; therefore, the following is a brief discussion of some characteristics that may be important considerations for users of the programs.

Each of the programs studied employs some variation of Gaussian elimination with back substitution to obtain the solution. Dewey's (1978) programs (JHD77 and SE77) use the classical form of Gaussian reduction for all but the last iteration. In GHYP1, Urhammer (Bolt, Okubo, and Urhammer, 1978) uses the Crout factorization; in HYPOELLIPSE (Lahr, 1979) and HYPO71 (Lee and Lahr, 1975) the authors use the Doolittle factorization. The latter two methods are variations of the compact Gauss elimination technique, and are special cases of the decomposition of the A matrix into unit lower (L) and upper (U) triangular matrices and a diagonal (D) matrix (Ralston and Rabinowitz, 1979):

$$A = LDU. \tag{2.13}$$

The Doolittle and Crout methods (like the abbreviated form of Gauss elimination) reduce roundoff errors by eliminating the need for storing intermediate results through the accumulation of inner products and the use of double precision arithmetic. The Crout algorithm has additional storage advantage over the Doolittle algorithm, and they are essentially equivalent in stability characteristics, so that for many cases the Crout method is selected (Ralston and Rabinowitz, 1978).

The above numerical techniques are regarded as direct methods (as opposed to iterative methods such as Jacobi and Gauss-Seidel), but they are often used iteratively, as is the case here. Use of direct methods is generally preferred for small (order < 100) or intermediate systems. For large (order > 1000), sparse systems of equations iterative techniques are more commonly employed (Ralston and Rabinowitz, 1978).

In arriving at the solutions through back substitution, one can obtain bounds on the errors. It should be noted that these error bounds are independent of the data vector t. (For derivations of error bounds for the Crout and Doolittle methods, see Ralston and Rabinowitz, 1978, p. 436-437.) Instability in direct methods can arise due to the roundoff errors, but if the A matrix can be manipulated into a form that is diagonally dominant and tridiagonal, these methods are stable with respect to the growth of roundoff errors (Varga, 1963).

For techniques such as these that rely on minimization of the residual r, Ralston and Rabinowitz state the requirement that computations involving Ay (e.g., equations 2.2 and 2.4) be carried out in double precision arithmetic. They indicate that this is because r will generally be of the same order of magnitude as the roundoff error. In the test runs of these computer programs (to be discussed in Chapter 3) perfect data was used; i.e., programs were required to generate their own data sets and then essentially work the same problem backwards to obtain the solution. Results computed by the Dewey programs, which did not use double precision, often exhibited station residuals on the order of 0.01 seconds, while HYPOELLIPSE sollutions generally showed station residuals of less than 0.01 seconds. I attributed this to development of roundoff errors in JHD77 and SE77 due to single precision arithmetic. Although this may have some significance for the test case results, for real data the difference becomes academic because the contribution of error due to record quality done is usually no smaller than 0.05 seconds.

Jordan's method (Dewey, 1970) is used for the final iteration for the computation of $(A^TA)^{-1}$ in JHD77 and SE77. This is necessary to obtain the necessary information for the computation of confidence ellipsoids.

CHAPTER 3

COMPARISON OF HYPOCENTER LOCATION TECHNIQUES

This chapter presents a discussion of three hypocenter location methods: (1) the single event method, (2) the master event method, and (3) the joint hypocenter determination method. The first has been most widely and routinely utilized. The relative merits of the last two have not been fully investigated to my knowledge, on the regional and local scales as defined in the first chapter. The first sections of the chapter involve generalized comparison and considerations of the theoretical basis for using different techniques. Brief histories of previous utilization of the methods are also presented. Later sections examine the results of synthetic test cases applied to the specific programs studied. These have explicit implications for the regional seismic array maintained by the University of Utah Seismograph Stations, from about 1962-1978, but conclusions are also presented that have a bearing on the relative applicability of the techniques for various situations and purposes of seismic event detection in general.

The Single Event Method

Most routine location of earthquake epicenters (or hypocenters) is presently accomplished using one of several computer programs available which obtain a solution for a single event. In this case the equations presented in the second chapter are modified accordingly, i.e., the number of events, m, becomes one and the subscript j is dropped. Thus the number of parameters (4m) equals four and the dimensions of the parameter vector y and the system matrix A are reduced. Because of its widespread use and familiarity it will not be discussed further here, except to state that the major advantage of the single event method is its comparative lack of complexity when used for initial studies and simple cataloguing of event occurrence. It will be seen that for more detailed and comprehensive work the other two methods give statistically better results.

The Master Event Technique

This method involves application of station adjustments computed from the location and travel-time residuals of some standard event (the master event or calibration event) to location of all other events within a specified region. Station adjustments are equivalent to the residuals computed for the stations recording the master event. The success of the method is dependent upon appropriate choice of the master event. It must be representative of all other events within the area; this implies that the area must be delineated so that all events are assumed to be subject to similar conditions of subsurface structure and geology. The extent to which a master event must represent the other events depends upon the scale of the study, and could vary from occurring in the same goelogical region to occurring on the same fault plane. In addition, the calibration event is usually of large enough magnitude that it was recorded by a large number of stations, preferably all of the stations that will be used to locate the other events. The reason for this is that all events located using a master event will have epicenters biased relative to that event. If stations are used that did not record the master event, the bias is no longer consistent between events, and the advantage of the technique is lessened considerably.

The bias introduced into the relative locations will be at a minimum with the proper choice of the master event, the location of which is thought to be more accurate either because it is very well recorded or perhaps has been verified by other means. For example, a large master event may be associated with an earthquake that produced ground breakage or it might be a mine blast. Use of the latter, however, presents problems since one must then assume similar source mechanisms for explosions and earthquakes. Problems may also arise if one considers the source mechanisms of large events, which may be extended in space and time, versus smaller events, for which the assumption of a point source may be more justifiable. Furthermore, because the quality of first arrivals is often dependent on the size of the event, a later phase may be picked for the smaller events as the first arrival. Possible effects of these discrepancies must be considered minimal because they are unknown, and so will not be considered further in this paper.

Another consideration in choosing the master event, one which will become more obvious in the discussion of the test cases, is the position of the earthquake relative to the detecting array. For a group of events in a poor location, recording stations will be subject to biased travel-times. Since this bias cannot be completely eliminated from the location of the master event, earthquakes located using this event as a reference should have similarly biased travel-times.

The master event method for locating earthquakes has been used with measurable success by a number of investigators. Evernden (1969b) discussed its applicability in the resolution of depths of teleseisms by improvement of estimations of P-wave travel times with use of station adjustments. In that study he noted a substantial decrease in the variance of average station residuals for master event controlled data. In his Ph.D. thesis J. Dewey (1971a) compared the master event technique to the joint hypocenter technique for teleseismic locations and nuclear blasts. But, as was previously mentioned, the usefulness of its application to regional or local recording situations is relatively unknown.

Joint Hypocenter Determination

The joint hypocenter determination (JHD) or joint hypocenter inversion technique has several major advantages over the single and master event methods. However, it is not as simple to use, requiring somewhat more consideration of possible inter-relationships between events before an attempt is made to locate them together.

The theory is as presented previously in Chapter 2. A major advantage the joint hypocenter location procedure has is that, by addition of appropriate terms to the equations of condition (equation

2.2), one can obtain station adjustments simultaneously with the hypocenter parameters. As first considered by Freedman (1967) and Douglas (1967), the method was developed to help minimize the effect of stations which are known to have biased travel time residuals, either due to local structure or some other reason. Freedman also made some conclusions about the improved precision of the hypocenter parameter estimates based upon an investigation of the variances of estimates and residuals. In her algorithm, Freedman alternately treated hypocenter and station adjustment parameters, holding one fixed while solving for the other and computing estimates for the other in the next iteration. Douglas solved for both simultaneously but found that, for events which were close together (on the scale of his investigation events spaced over a few degrees), the method often failed to converge because the station adjustments and travel times became essentially linearly dependent. He suggested that use of a fixed reference event might help to eliminate this problem. Dewey (1971a) wrote and tested a joint hypocenter location program and a master event program. His method follows that of Douglas in essential aspects, but he included the provision of fixing the location of a calibration event. The procedure incorporates into the equations of condition the additional side equations of the form

$$w_{i1}dg_i + w_{i1}e_{i1} = w_{i1}r_{i1}, \tag{3.1}$$

assuming the first event is the calibration (reference) event and letting dg_i be the station adjustment for station i, e_{i1} the error term due to timing and picking event one at station i, and r_{i1} the residual term. The w_{i1} term is, as before, the weight applied to the observation of event one at station i. There are as many of these equations as there are stations recording event one, and they add the necessary stability to the condition matrix (Dewey, 1971a).

Offsetting to some extent the advantages of this improvement of the system matrix is the fact that now the solutions are dependent on a well-located reference event, and will suffer a bias with respect to that event similar to the one introduced with use of the master event technique. The influence of the reference event in joint hypocenter computation is not as profound. Furthermore, there are not as many restrictions on the calibration event used with the joint hypocenter method. Dewey (1971a) found that there was not a significant difference in the location of known events with use of calibration events recorded by large versus small numbers of stations. This conclusion was supported by my own testing. Calculated station adjustments are similar whether or not the station recorded the calibration event. So the size of the event is no longer a main consideration; in fact, it may be preferable to choose a reference event of similar magnitude to ensure first arrival phase consistency. However, a very poorly recorded event will not restrain the equations of condition properly. The minimum number of observations of the reference event required to properly constrain the equations for a given situation can usually only be determined from testing with synthetic (known solution) data. The position of the calibration event relative to the other events and to the stations remains an important aspect to take into account.

Since the station adjustments are computed from the travel-time residuals, a more thorough examination of this subject than was previously presented in the second chapter is appropriate. Travel-time residuals have been assumed by most to be normally distributed with a low amplitude background which may be slowly varying but can be thought of as a superimposed uniform distribution (Buland, 1976). However, studies by Freedman (1967), Bolt and Freedman (1969) and Dewey (1971a) have shown that this is not the case rather that non-zero mean and skewness of the distribution characterize the residuals. These effects might be combinations of terms due to deviations from the crustal model below the receivers, at the source, and along the ray path, the position of receivers relative to the source, and even consistently poor readings at a particular station or group of stations. The distribution is thus biased, and the normal single event method works, through minimizing the sum of the squared uncorrected residuals, to incorporate any bias into the solution. By simultaneously estimating individual station corrections from the residuals some of the bias may be removed.

The first estimate of the station adjustments is of the form

$$dg_i = \sum_{j=1}^{m_i} \frac{w_{ij}^2 r_{ij}}{w_{ij}^2 m_i} \qquad (3.2)$$

(Dewey, 1971a; Freedman, 1967). Here m_i is the number of events recorded by station i. The concept of estimating station adjustment simultaneously is an important one. A single event program can be adapted to obtain station adjustments, but they will differ from those computed by the JHD technique in that they cannot be estimated simultaneously with the earthquake locations. This type of modification was applied to the single event program HYPOELLIPSE (Lahr, 1979). A representative group of events was chosen and locations for these were computed. Travel-time residuals for each station from each event were summed and the means were taken as station adjustments. In HYPOELLIPSE these adjustments could be used in the subsequent locating procedure by entering them as delays for each station, essentially modifying the velocity model. There is a basic error associated with the application of this method. Because single event location uses least squares to determine a solution, it necessarily incorporates extreme residuals into the minimization process. As previously noted, this can lead to a biased result, and the station adjustments computed from the averaged residuals include this source of error. This is in contrast to the true JHD technique, which is not a simple least squares procedure. Instead it does not require the assumption that the residuals are completely composed of error to be minimized, and modifying the station adjustments at each iteration can restrain the solution from moving successively toward an erroneous location. In particular, the present form of Dewey's (1978) program enables the user to have more control through previous experience with the data and application of initial weights for station phases.

Two other improvements that JHD has over both the single and master event techniques are connected to this idea of inclusion of the biased station residuals in the normal least squares location procedure. Because estimates of variance include the effects of the biased residuals if they are not removed, the weighting schemes based on the variance will be incorrectly computed. Since the weighting schemes are used to truncate or eliminate residuals, this may adversely affect the convergence rate as well as the final solution. The second improvement concerns the ultimate size and eccentricity of the confidence ellipses, which may be in error because their computation also includes the variance terms.

Comparison of Specific Programs

In this section the results of test runs on synthetic data sets are examined with the intent of discovering under what particular circumstances each method can best be applied. The test sets were generated to represent specific situations in southern Utah. For the regional study, the network maintained by the University of Utah is used; for the local study, stations used are part of a microearthquake array set up in 1974 and 1975 in the Cove Fort-Roosevelt Hot Springs area (Olson, 1976). Thus the results can aid in future earthquake studies in Utah because much supplementary information about the recording set-ups has been gathered. But the strengths and weaknesses of a particular array geometry, and conclusions regarding certain location methods, are applicable to many generalized situations and should not be considered as being solely relevant to the described localities in Utah.

Points of Comparison

There are several aspects of earthquake location that may be particularly amenable to treatment using special methods or programs. Some aspects that were considered in evaluation of the techniques and programs were: the resolution of hypocenters versus epicenters and the effects of fixing the depth parameter; responses to simplification of the velocity model; sensitivity of the solution to initial starting points; and the distribution of recording stations with respect to the events. Regarding this last aspect, it has been known for some time that a wide distribution of recording stations in distance and azimuth results in a better (more accurate) solution. But the sensitivities of various location methods to common inadequacies of station coverge are not often quantitatively understood, and although problems are acknowledged to exist, usually nothing is done other than perhaps assignment of an often poorly resolved or little understood quality factor to indicate the general quality of solutions in a particular region.

The capabilities of the programs with respect to these aspects of earthquake location can be assessed in several ways. One is comparison of the relative accuracy of the computed estimate in relation to the true position of known (synthetically generated) hypocenters. The precision of the estimate computed by various techniques can also be compared. Perhaps most importantly, how well the results of a specific program can reflect both precision and accuracy of the solution is a valuable point to consider.

For this study a total of four different programs were run with test cases. The single event technique is represented by the programs HYPO71 (Lee and Lahr, 1975) and its successor HYPOELLIPSE (Lahr, 1979), and SE77 (Dewey, 1978). The latter is a single event program which can be used with the master event technique also. A procedure was developed to use HYPOELLIPSE as a type of master event program as well, as was previously discussed. The joint hypocenter location techniques is represented by two programs, JHD77 (Dewey, 1978) and GHYP1 (Bolt et al., 1978).

Several modifications were necessary to get the programs GHYP1, JHD77, and SE77 operating on the University of Utah UNIVAC 1108 and to make them compatible with the input and output of programs already operating on the system. These adaptations, for the most part, do not have any direct relevance to the problem and results presented in the main body of this thesis; therefore, details are summarized in Appendix A. User's guides for these programs and for several supplementary programs are also found in Appendix A.

Synthetic Data Generation

A program (CREATED) was written to create an arrival time deck for any number of events with up to thirty stations for each event. Arrival times for P and S phases are read from travel-time tables which can be generated using a modified version of the program GM. This program was written to compute and plot travel-time curves for spherical or flat-earth models (Kaminski and Muller, 1978). The modified version of GM searches through calculated direct and refracted branches of the travel-time curves to obtain first arrivals at user-specified distance increments. Arrivals for the S phase are simply the P arrivals multiplied by an externally specified factor, the V_p/V_s ratio. These travel-time tables can then be easily manipulated into the same format utilized by the program CREATED as well as JHD77 and SE77. CREATED can generate a perfect data set or one contaminated with random noise of any specified amplitude.

The following account will proceed first with an examination of the regional network in southern Utah and its locating capabilities with respect to the three methods. Then the scale of interest will be reduced to the more localized situation of a microearthquake survey in the Roosevelt Hot Springs and Cove Fort areas.

Regional Recording Situation

To begin with, it should be noted that all of the depths were fixed in the solutions for the regional events. For most of southern Utah this is necessary to obtain a good solution for the epicentral coordinates, for reasons which were discussed in Chapter 2. For those localities with possibly sufficient station coverage to resolve the depth parameter, the depths were also fixed for test runs to maintain an internal consistency in the comparison with events in the whole region.

Figures 3.1 a, b, c, d, and e show five contour plots of a grid of twelve synthetic events. The five plots represent location error for five different programs. Stations used for the location of the events are stations from the University of Utah seismic network before October 1974. Figure 3.1 f shows the location of the grid relative to the stations used in the synthetic data set. This region in eastern Utah was chosen as representative of what might be a rather typical

Figure 3.1. Contour plots of location error for synthetic events; random noise with standard deviation of 0.2 sec added to arrival times. Value assigned to each grid point (true location) is magnitude (km) of mislocation by particular program. a) JHD77, b) SE77 used with a master event, c) SE77 used with station adjustments from JHD77, d) SE77 used as a single-event program, e) HYPOELLIPSE, f) location map of the grid relative to the stations used.

EXPLANATION

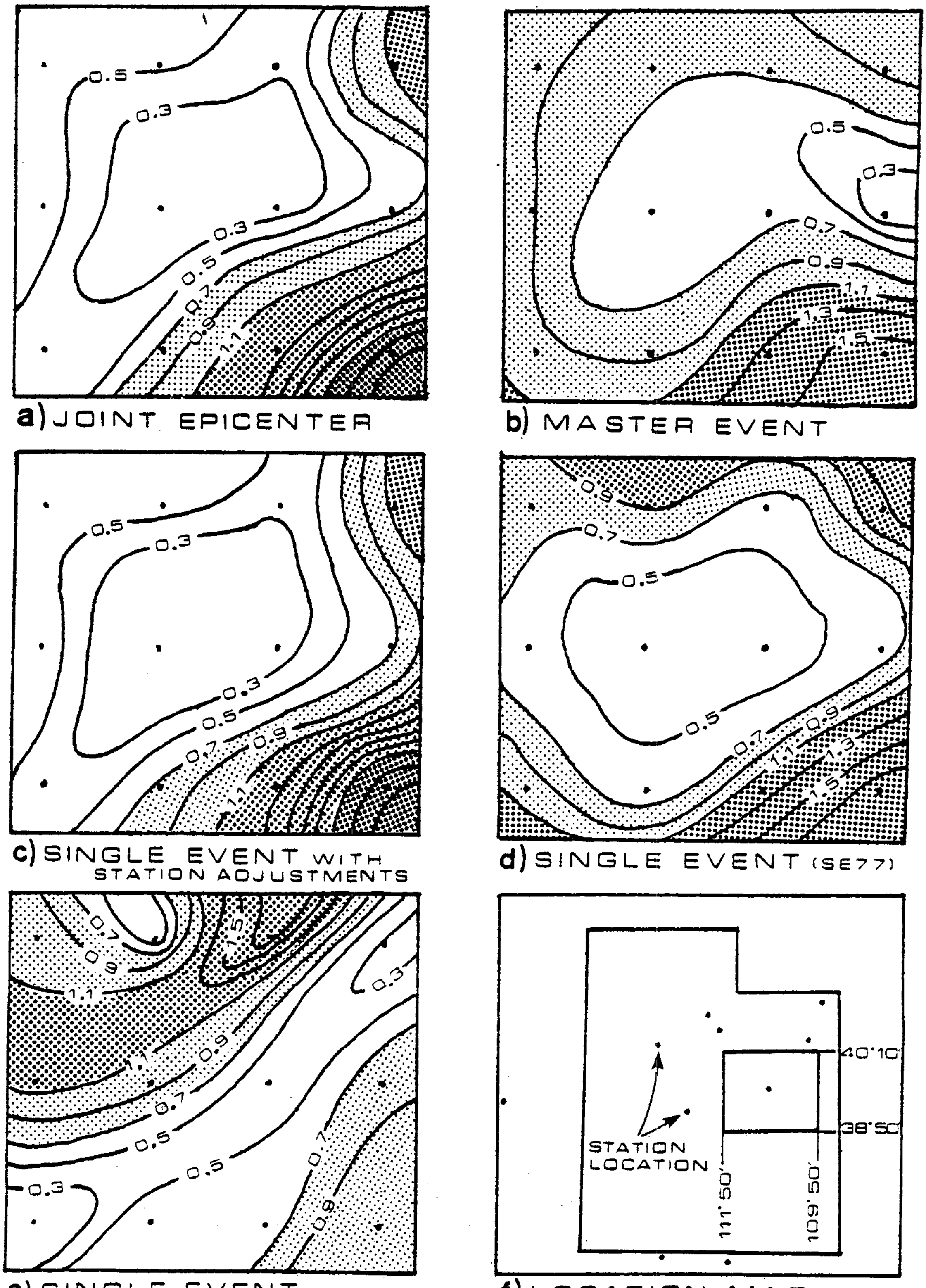
Contour interval 0.2 km

Mislocation less than 0.7 km

Mislocation 0.7 to 1.1 km

Mislocation 1.1 to 2.1 km

Mislocation greater than 2.1 km



a) JOINT EPICENTER

b) MASTER EVENT

c) SINGLE EVENT WITH STATION ADJUSTMENTS

d) SINGLE EVENT (SE77)

e) SINGLE EVENT (HYPOELLIPSE)

f) LOCATION MAP

regional data set. Distances to stations range from 20 to over 400 km, and the azimuthal distribution ranges from good (largest gap between event-to-station azimuths is approximately 80°) to poor (gap of approximately 200° or greater).

The data set for the tests with Dewey's (1978) program (3.1 a, b, c, and d) was generated using the previously described program CREATED. Arrival times included random noise with a standard deviation of 0.20 seconds. The data set for the HYPOELLIPSE single event test (3.1 e) was self-generated. That is, the option in HYPOELLIPSE to generate a corrected set of travel-times was used. Then the same values of the random noise were added to those travel times. Arrival times for S phases were generated for the four stations that often exhibit discernible S-wave arrivals: DUG, PCU, SLC, and GMU. All events for all test runs started at an initial location corresponding to the position of the calibration event.

The value assigned to each point in the figures is the amount of mislocation, in kilometers, of the estimated epicenter. Note that plots 3.1a and 3.1c are very similar; 3.1c is the plot of locations computed by the single event method (as in 3.1d) but utilizing station adjustments computed in the previous run of the joint epicenter determination (JED) program (3.1a). Also note that results of the two single event programs represented in plots 3.1d and e are substantially different, which I believe can be attributed in part to the different weighting schemes and damping of adjustments used in the two programs.

The patterns in the figures indicate that the area covered by the larger magnitude mislocations is greater for the single event programs than for the master event and JED techniques. This is partially due to the exact knowledge of the reference event location. But another important consideration is that the master event and JED methods seem to smooth out variations within the region, i.e., the contours are less contorted. This is especially apparent for the master event method (3.1b), but the magnitudes of the mislocations are in general greater than for JED. Another point to observe is that the joint location method appears to be especially sensitive to instabilities near the edge of the array. Although all of the techniques show increasing displacement toward the southeast corner of the test region, which corresponds to moving out of the recording station array, JED and the similarly adjusted single event program show the largest deviations for the event farthest from the array. The large relative magnitude of this location error can be attributed in part to the fact that station adjustments computed for the other events do not adequately represent that particular event. Because of its position, it is governed by different (and larger) adjustments then the others of the group, and the addition of an erroneous adjustment to the travel times for this event probably magnified the error. Another source of mislocation that may be a consideration, in all of Dewey's programs, is the lack of any distance weighting.

The major conclusion that can be drawn from Figure 3.1 is that the JED method has an advantage over other techniques in its more uniform location of events within a particular region. In other words, within the delineated region of events, all should be located to a similar degree of accuracy. Also, the magnitudes of the displacements of the estimated epicenters from the true locations are lower than for other techniques, taking into account the fact that the accuracy of the location of the reference event places a lower limit on the absolute accuracy of the others. With respect to the reference event and thus to each other, however, the events will have better solutions.

Applications to Southern Utah

The regional seismic network in southern Utah underwent major modifications throughout a period from approximately 1972 through 1974. Therefore, anyone interested in studying the historical seismicity must contend with essentially two different sets of seismograph stations with somewhat different coverage capabilities. For this reason all analysis was done for two networks in order to effect a comparison of relative location capabilities during different time periods as well as to test the methods under a variety of recording situations. One set of stations represents the period from July 1962 through September 1974; the other represents the major recording network for southern Utah from October 1974 to December 1978. The stations utilized for the test cases do not include all possible stations that could have recorded an event during either time block, but for the most part they represent those stations most often recording. No station that was critical to the distribution, either with respect to distance or azimuth, was left out. S-phase arrivals were also generated for those stations which often recorded S-arrivals in the real data set for each time block in a given region. The stations recording S arrivals in the synthetic data vary depending on the location of the event.

To obtain more information on the relative capabilities of the joint location technique represented by JHD77 and single event locations represented by HYPOELLIPSE, synthetic data were generated for a grid covering southern Utah below 40° N latitude. Results of the test runs are plotted on contour maps in Figures 3.2 through 3.6. These are similar to the contour diagrams of Figure 3.1 discussed previously but differ in that no noise was added to the arrival times, and the station sets are slightly different. These synthetic events, because of the number of stations recording and the lack of noise, represent optimum cases, and care must be taken in extrapolating the test results to actual Utah earthquakes. As a very qualitative estimate, an earthquake would have to be at least magnitude (M_L) 3.0 to be as widely recorded as these test cases. Figures 3.2 and 3.3 are results for the time block prior to October 1974, and 3.4 and 3.5 are the same events for the network after September 1974. Again, values assigned to the grid points are deviations, in kilometers, of the estimated solutions from the known epicenter.

To utilize the joint epicenter determination (JED) program, which can solve for only fifteen events simultaneously, Utah below 40°N latitude was divided into the regions delineated by the dashed lines. Regions were defined by considering previous test results from the single event method. Two essentially opposing considerations were in effect. One was to define an area relative to the network that would

Figures 3.2, 3.3, 3.4, 3.5, 3.6. Contour plots of location error for synthetic events in southern Utah recorded by representative regional stations; no noise added to arrival times. Value assigned to each grid point (true location) is magnitude (km) of mislocation by particular program.

EXPLANATION

Contour interval 0.10 km

Stations used for test events

Mislocation less than 0.3 km

Mislocation 0.3 to 0.6 km

Mislocation 0.6 to 0.9 km

Mislocation greater than 0.9 km

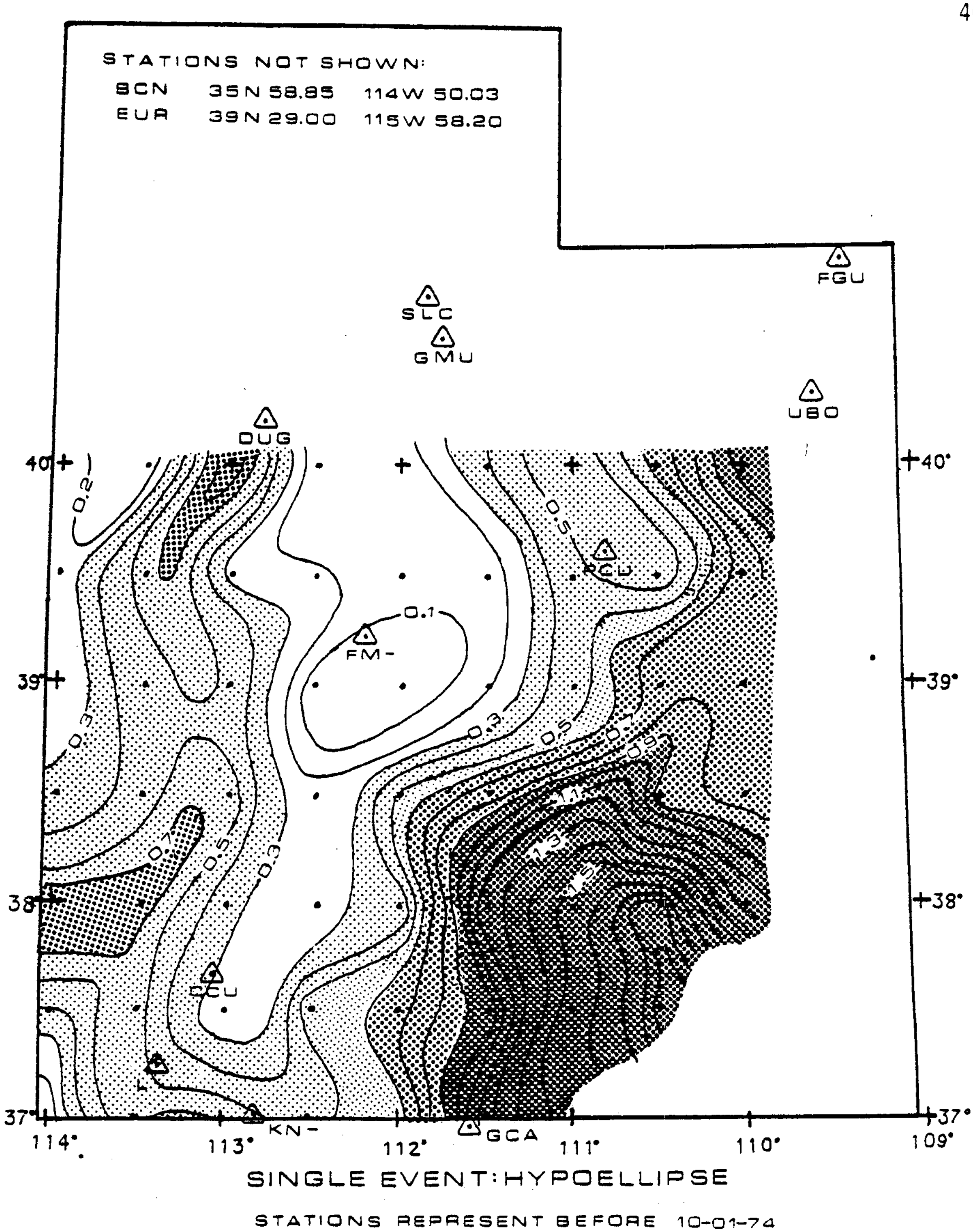

STATIONS NOT SHOWN:
BCN 35N 58.85 114W 50.03
EUR 39N 29.00 115W 58.20
FGU
SLC
GMU
UBO
DUG
PCU
FM-
CCU
KN-
GCA
0.1
0.2
0.3
0.5
0.7
0.9
40°
39°
38°
37°
114°
113°
112°
111°
110°
109°
SINGLE EVENT: HYPOELLIPSE
STATIONS REPRESENT BEFORE 10-01-74

Figure 3.2

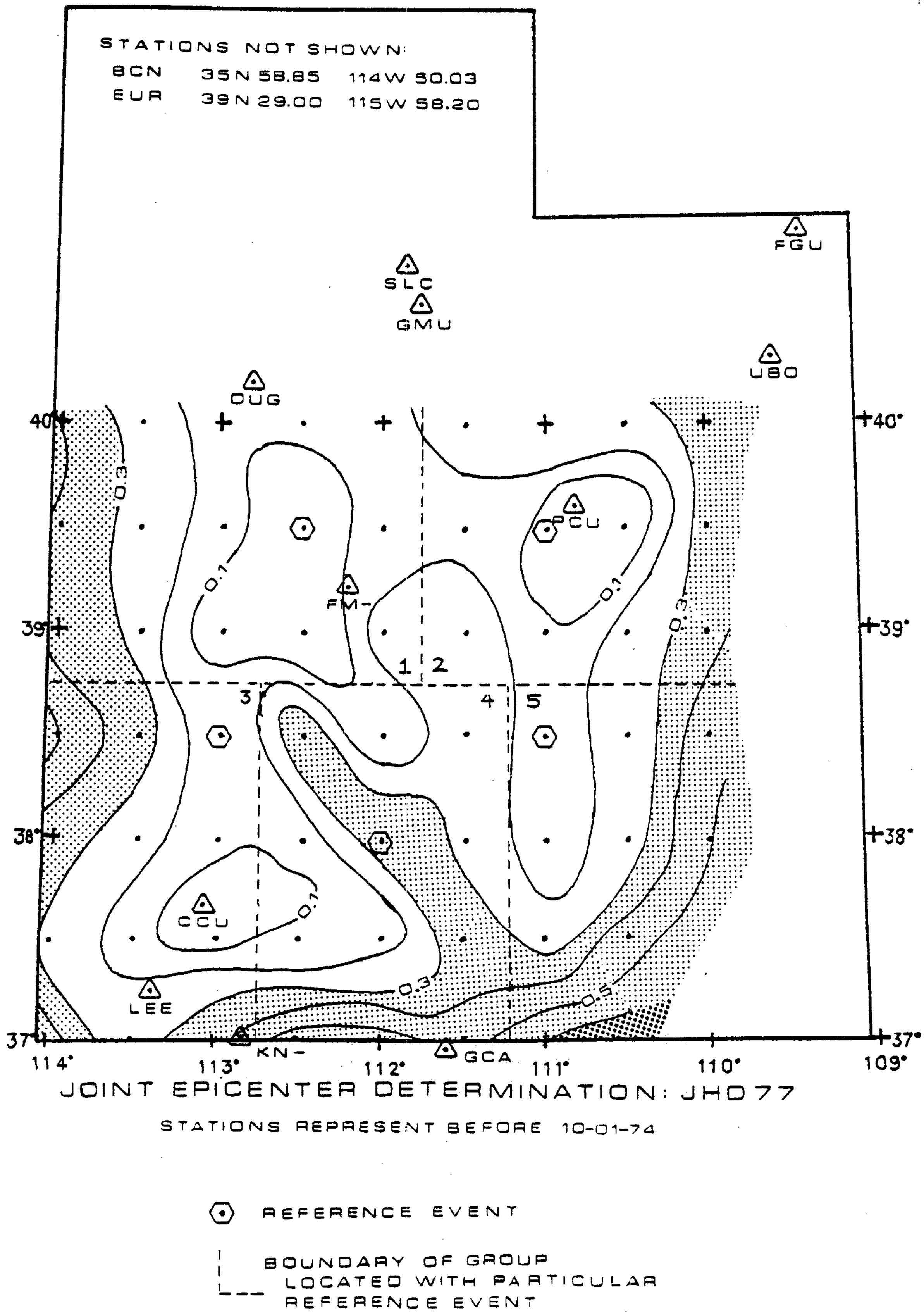
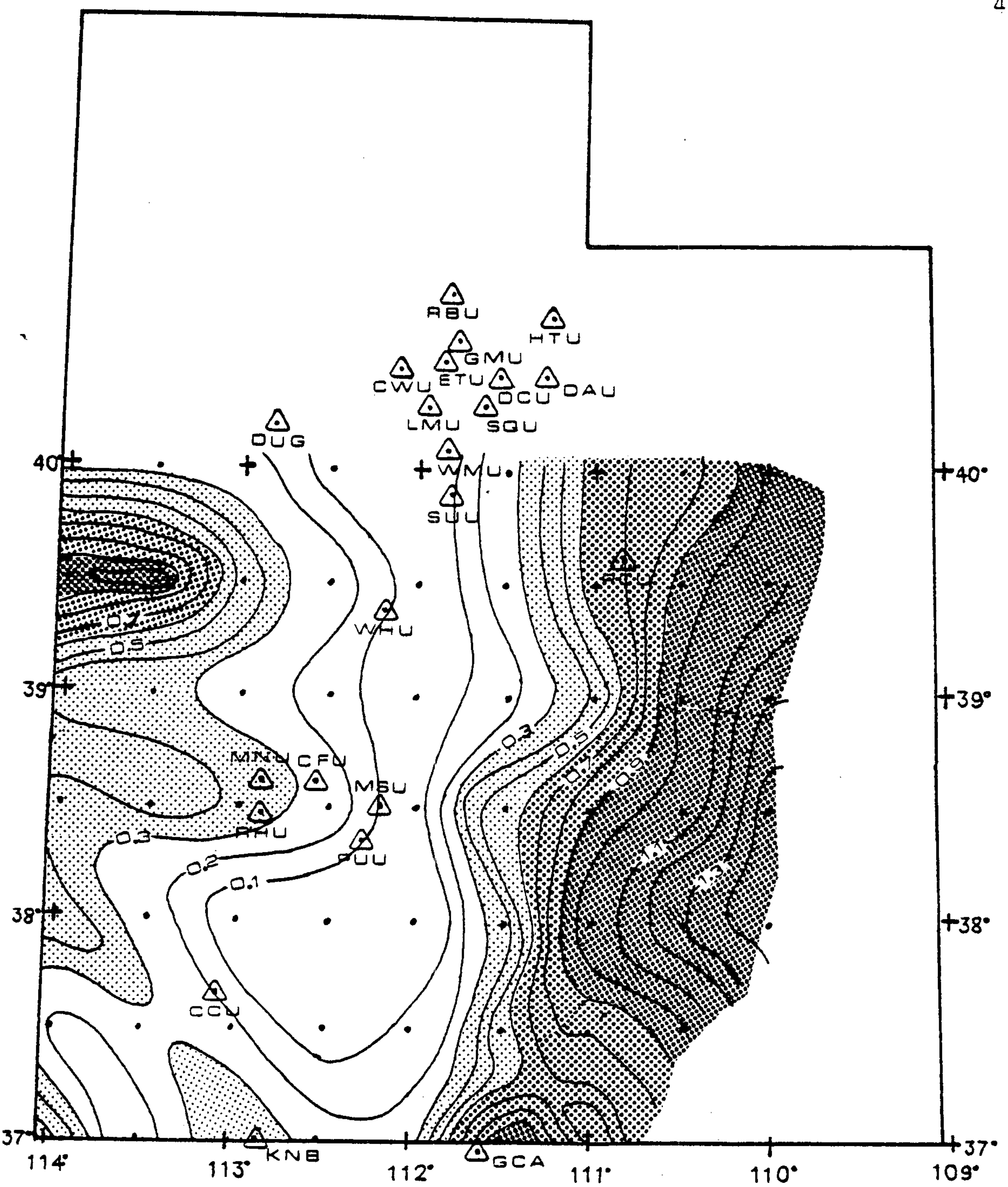

STATIONS NOT SHOWN:
BCN 35N 58.85 114W 50.03
EUR 39N 29.00 115W 58.20
FGU
SLC
GMU
UBO
DUG
PCU
FM-
CCU
LEE
KN-
GCA
0.1
0.3
0.5
1
2
3
4
5
40°
39°
38°
37°
114°
113°
112°
111°
110°
109°
JOINT EPICENTER DETERMINATION: JHD77
STATIONS REPRESENT BEFORE 10-01-74
REFERENCE EVENT
BOUNDARY OF GROUP
LOCATED WITH PARTICULAR
REFERENCE EVENT

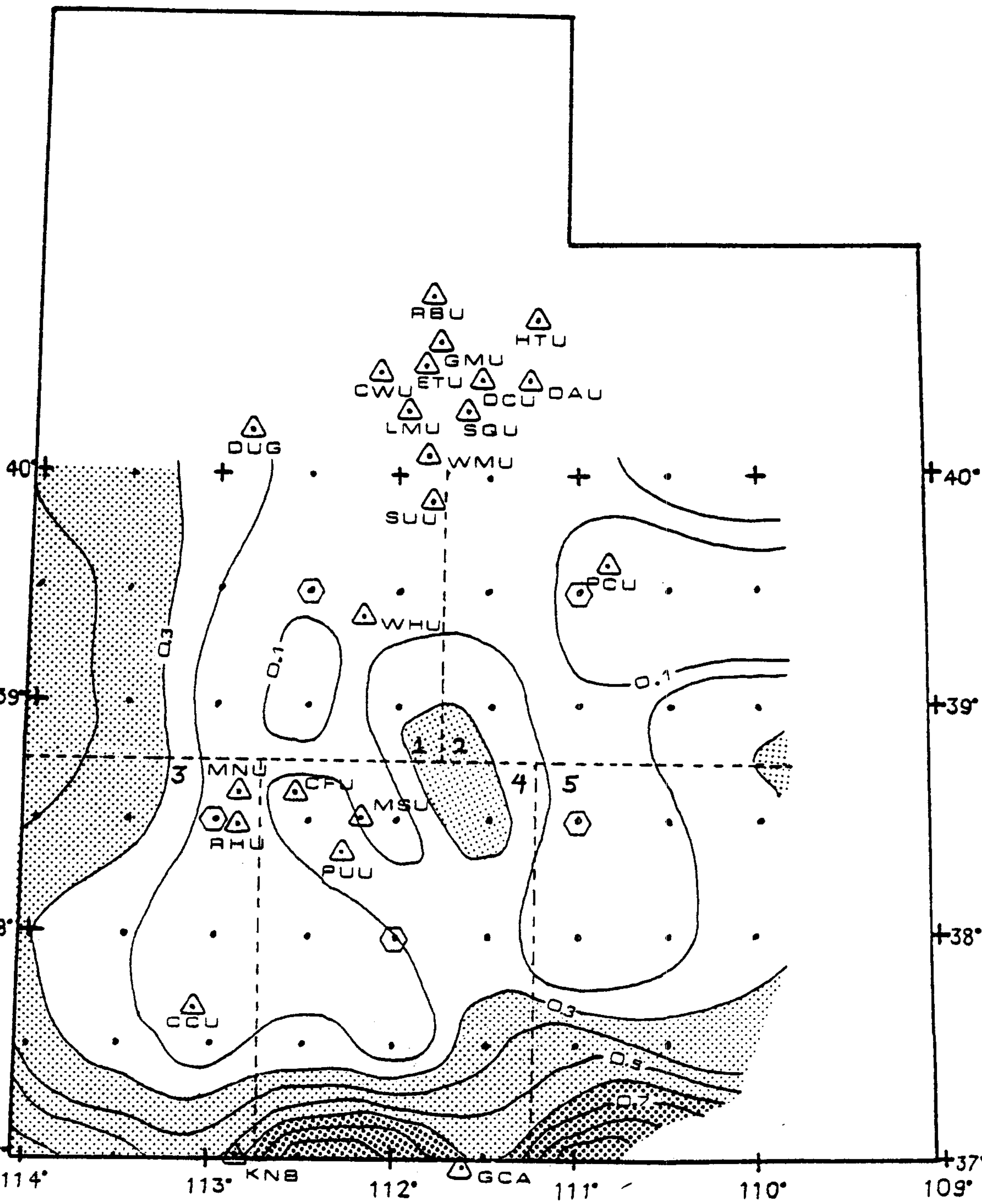
Figure 3.3



SINGLE EVENT: HYPOELLIPSE

STATIONS REPRESENT AFTER 10-01-74

Figure 3.4



JOINT EPICENTER DETERMINATION: JHD77

STATIONS REPRESENT AFTER 10-01-74

REFERENCE EVENT

BOUNDARY OF GROUP LOCATED WITH PARTICULAR REFERENCE EVENT

Figure 3.5

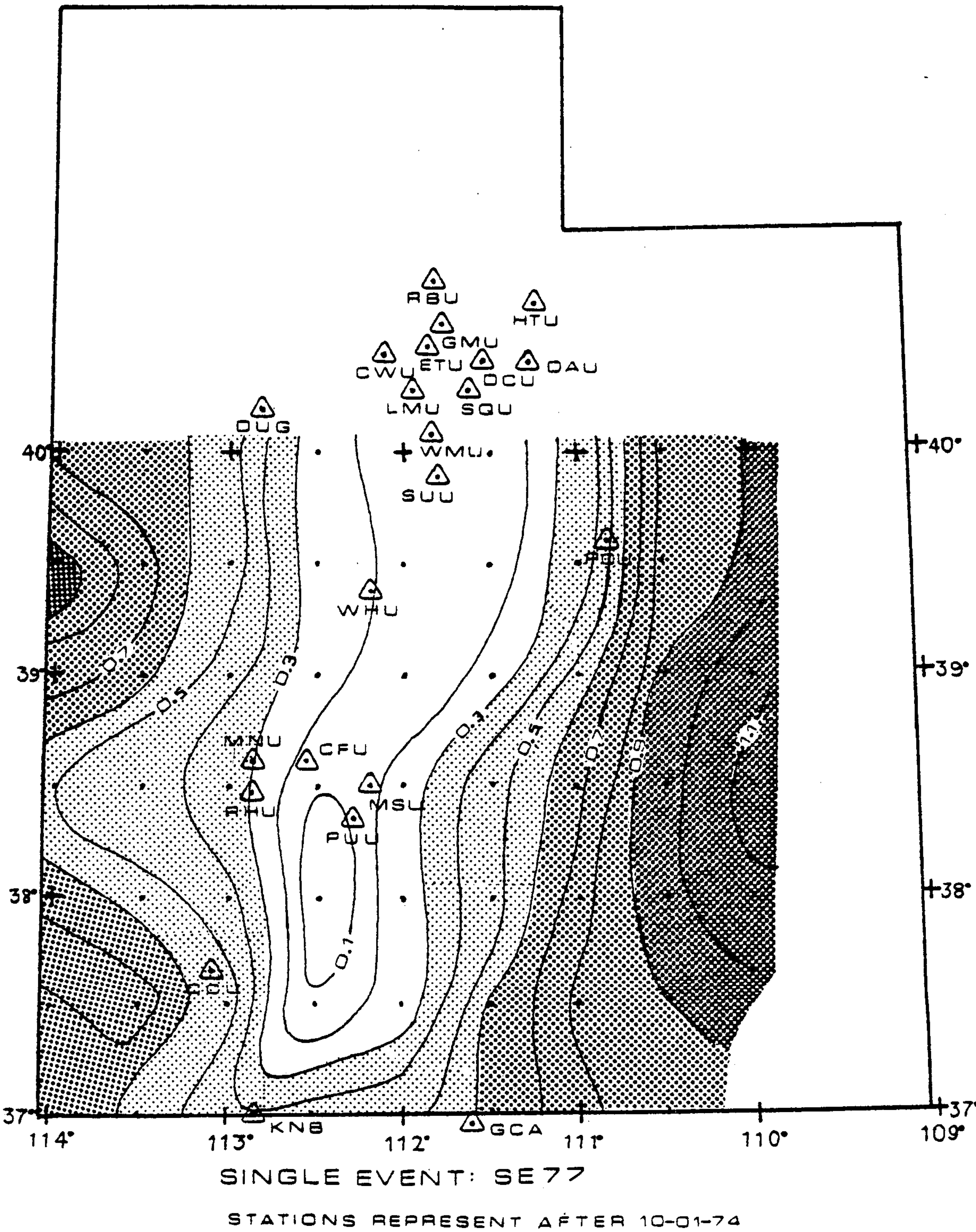

RBU
HTU
GMU
CWU
ETU
DAU
DCU
LMU
SQU
DUG
WMU
SUU
WHU
MMU
CFU
MSU
PUU
KNB
GCA
0.1
0.3
0.5
0.7
0.9
40°
39°
38°
37°
114°
113°
112°
111°
110°
109°
SINGLE EVENT: SE77
STATIONS REPRESENT AFTER 10-01-74

Figure 3.6

contain events with consistent deviations from the true location, since they were to be located together. However, at the same time it was not advantageous to have the group composed entirely of events that tended to be grossly mislocated, because the results would not minimize the differences in deviations over all of southern Utah, which was the other consideration. Therefore areas were not defined strictly on the basis of boundaries suggested by the contours of figures 3.2 and 3.4, but defined to slightly overlap these.

Reference events for each area were chosen in positions either central to the area, to represent most events, or in optimum positions with respect to the recording array geometry. In area 5 (Figure 3.2b) two tests were made, varying the position of the reference event in order to observe possible effects of utilizing a different event. Area 5 was used for this particular test because it exhibited some of the largest deviations from the true epicenters due to the distance between the events and the southern Utah network. The test runs (results shown in Figure 3.7) indicated that the choice of a different reference event, at least in this case, results in very little change in the magnitudes of the mislocations, but may effect the direction of the misplacement. This is especially pronounced for events near the reference event.

The mislocation of the synthetic events in these plots results only from the station geometry. Besides a concept of how well the particular programs can locate events in various places, the plots can also indicate locations of possible false concentrations of events in southern Utah that may be only due to the location capabilities of the

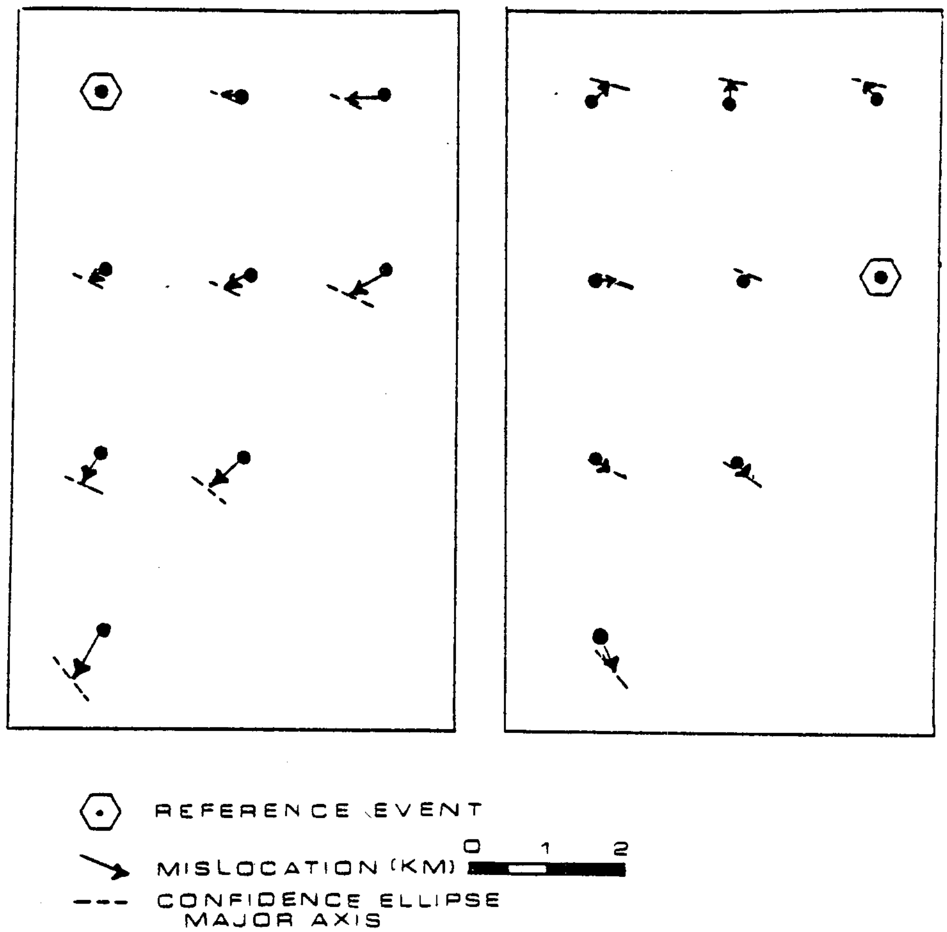


Figure 3.7. Magnitudes and directions of mislocation of synthetic events using two different reference events. Grid points mark true location of event. Note that orientation of major axis of confidence ellipse remains the same, but orientation of vector indicating direction and magnitude of mislocation varies.

array. This is not to say that one can predict in what direction or by how much events will be mislocated by the programs. The contour maps only suggest that, in different regions of Utah with similar numbers of events per region, the station coverage is such that more events will be located in one region than in another. In other words, events may be detected in a region that is poorly covered by the array, but they would be discarded from a regional compilation of epicenters because their locations would not satisfy minimal conditions for determination of the solution.

Comparison of the joint location technique (JED) with the single event HYPOELLIPSE program in Figures 3.2 through 3.5 indicates that, as in previous analyses, JED smooths out the variations in magnitudes of mislocation and, if the reference event is correct, the absolute accuracy will be improved. Because these plots include effects of differences in the programs themselves rather than just the techniques, Figure 3.6 was added to show that comparison with the single event technique as programmed by Dewey results in similar conclusions. It appears that JED can diminish adverse effects of array geometry. This would be anticipated by the theory, since events located within one region will generally show a common bias in mislocations due to their position with respect to the recording stations. Locating them jointly should bring out this effect in the form of station adjustments, provided that the reference event is not subject to this same mislocation to an equivalent degree. This in general cannot be assumed to be valid for most real situations, but the smoothing of variations in displacements should remain a characteristic for real data. Positions of events relative to each other will thus be more accurate. Unless the reference event location has been determined by some other means than just the the same station geometry, the same mislocation error will be largely fixed within the data set when the position of the reference event is fixed. The information gained from plotting the synthetic data can be qualitatively employed in analysis of any epicenter map. Knowing the relative accuracies possible in any given region will aid in critically assessing the importance of various trends in the data. Comparison of the gross characteristics of Figures 3.2 through 3.6 shows many similarities in both time blocks. But since there are fewer stations for the earlier period the elimination of even one will have a stronger adverse effect than would be the case for the later time period. Also, for the group after October 1, 1974 (Figures 3.4 through 3.6), there are a number of stations which were not used for the test cases, but which add density and redundancy to the recording network that is important in a real recording situation. Details of these plots will also be altered by addition and/or deletion of S-arrivals at various strategic stations.

Precision and Accuracy of Results

In an actual situation, of course, the deviations from the true epicenter are unknown. So another criterion that must be evaluated in comparing the technique and programs is what the computer program results tell the user about the precision and accuracy of the estimates. In figure 3.8 three statistical parameters are plotted for the same twelve events discussed previously (Figure 3.1). These parameters, as discussed in Chapter 2, are generated during the computations of locations and are commonly used to judge the quality of the solutions. Figures 3.8a and b are contour maps of the RMS (root mean square) residual as computed by HYPOELLIPSE and the standard error as computed by the JED program respectively. The values of these parameters, when compared to Figure 3.1, do not appear to have any direct correspondence to the magnitudes of the actual mislocations. The RMS residual and standard error (SE) are governed by the station distribution also, but are only indicative of the precision of the solution, not the accuracy. In other words, for a given station distribution and velocity model, the solution corresponding to the minimum in the statistical least squares sense may or may not be the true solution.

In addition, the minimum attainable RMS and SE value varies over the array and is intimately related to the number of station phases used to compute the solution. A minimum can always be found outside the array. Given that the event is within the network and that the velocity model used is correct (which it is for the test cases) a minimum also exists at the true epicenter. However, whether this is reached depends upon when the program stops iterating. Although the process of the computation is based on least squares minimization, iteration is usually terminated based not on the value of a minimized parameter but rather on some other criterion such as the size of the last step in parameter space or simply on a maximum number of iterations.

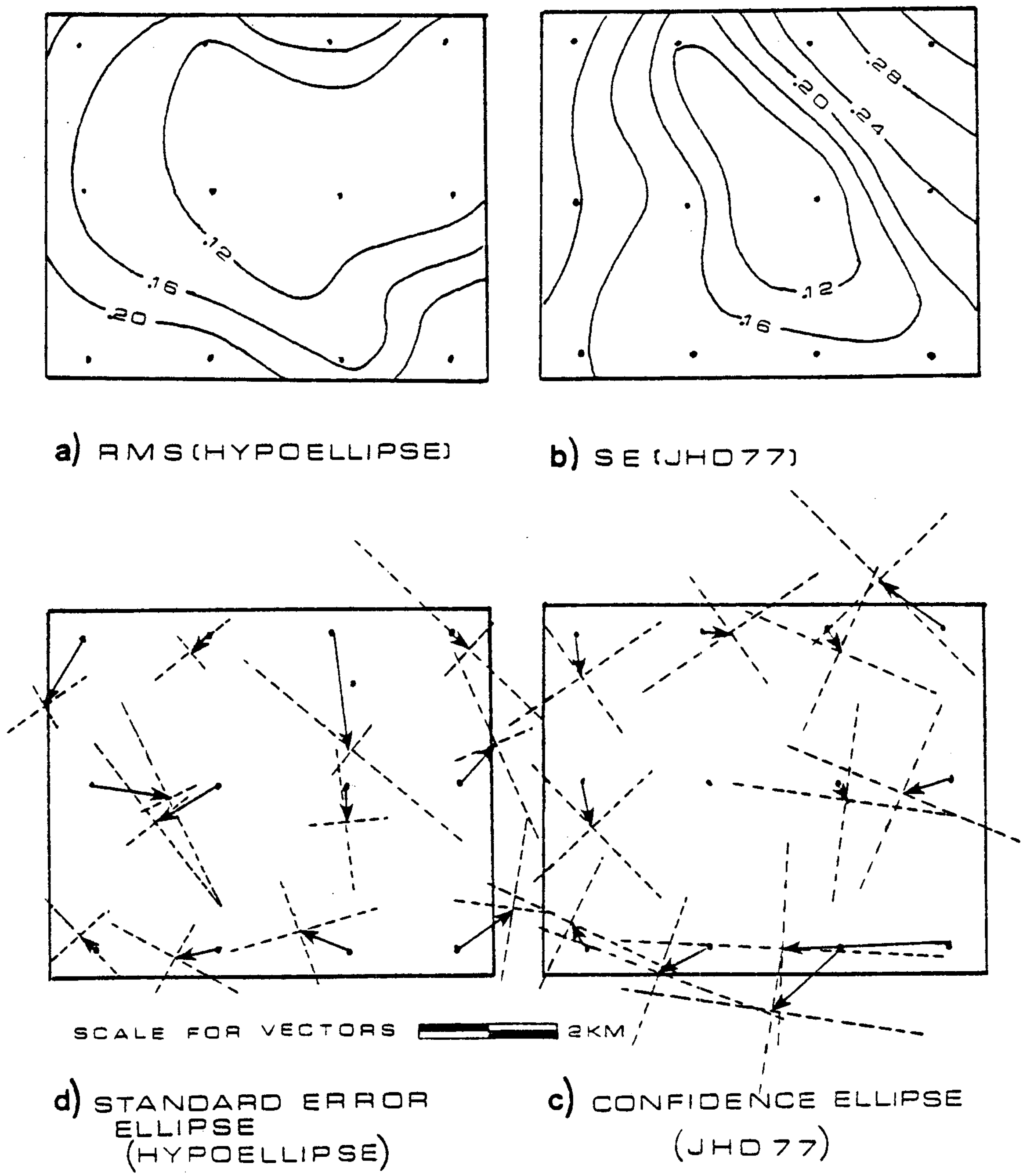


Figure 3.8. Statistical results for synthetic events of Figure 3.1. a) Root mean squared residuals computed by single-event program HYPOELLIPSE. b) Standard error computed by joint-determination program JHD77. c) Standard error ellipses and misplacement vectors. d) Confidence ellipses and misplacement vectors.

Figures 3.8c and d are plots of the same twelve events showing the directions of the mislocations of the estimated solutions. Misplacements are indicated by the arrows, scale in kilometers; also superimposed on these are the orientations of the major semi-axes of the joint confidence ellipses (JED) and the standard error ellipses (HYPOELLIPSE). For the JED solutions (Figure 3.8d) the ellipses satisfy the specified conditions, i.e., at least 90% of the 90% confidence ellipses cover the true location. Only one does not, and for this one the true location lies just on the edge of the ellipse. For HYPOELLIPSE the orientations of the axes are similar, which is encouraging since both use the same computations and these are functions only of the station distribution and geometry. However, contrary to the program description (Lahr, 1979), the ellipses computed by HYPOELLIPSE are not joint confidence regions. Less than 50% of these ellipses (5 events out of the total twelve) cover the true epicenter; two more touch the true location. To make these into confidence regions would require a few additional computations.

From Figures 3.8c and 3.8d it is apparent that there is no consistent correlation between either magnitude or orientation of the mislocation vector and the size and orientation of the confidence or standard error ellipses. This follows from the discussion of confidence ellipses in Chapter 2. However, in the situation of synthetic test cases with no noise, the deviation of the estimated solution from the actual solution is a function of only the station distribution. Since confidence ellipses are also functions of the station distribution, one might expect some sort of correlation.

A correlation was discovered in the analysis of perfect synthetic data (no noise). Figures 3.9 and 3.10 are similar to Figures 3.8c and 3.8d, and represent, respectively, HYPOELLIPSE and JHD77 mislocations for a station distribution common for events in Utah before October, 1974. Because of the number of stations involved, similar figures were not completed for the period after October 1, 1974. The main characteristics would be similar, however. Some of the differences in magnitude between HYPOELIPSE and JHD77 mislocations are the result of the fact that computations for the generation of synthetic data are the same as those used in the JED program, but differ from the forward routine in HYPOELLIPSE. But the major conclusion is that, for confidence ellipses with large major to minor semi-axis ratios (>2) indicating a large directional bias in the ability of the array to place an event at that point, the direction of displacement generally corresponds to the orientation of the major semi-axis. For synthetic data, larger confidence ellipses also correspond to larger deviations of the estimate from the true solution.

Thus a useful result is that for a specific region with less station coverage solutions may show an artifical linear trend or clumping. The possibility of this occurring can be anticipated by studying the ratios and orientations of the semi-axes of the confidence ellipses. The success of predicting such a phenomenon is lessened by the fact that, for noisy data, the results are valid only if the effects of poor station distribution are greater than the noise, as was observed previously in Figure 3.8. A further complication exists when using the JED technique. It was stated that

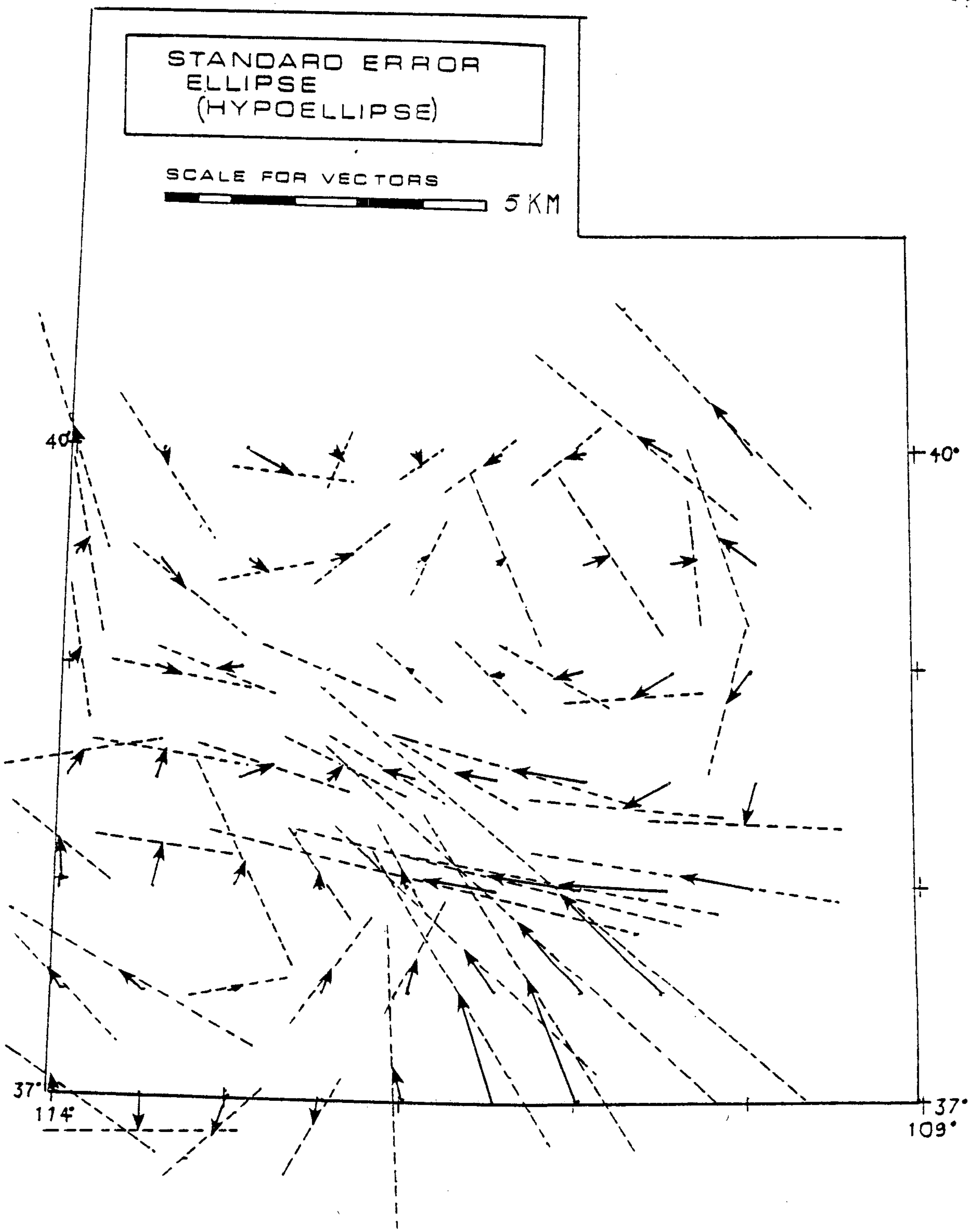


Figure 3.9. Misplacement (km) of synthetic events of Figure 3.2. True locations are grid points; tips of arrows are computed locations. Dashed lines are major axes of standard error ellipses.

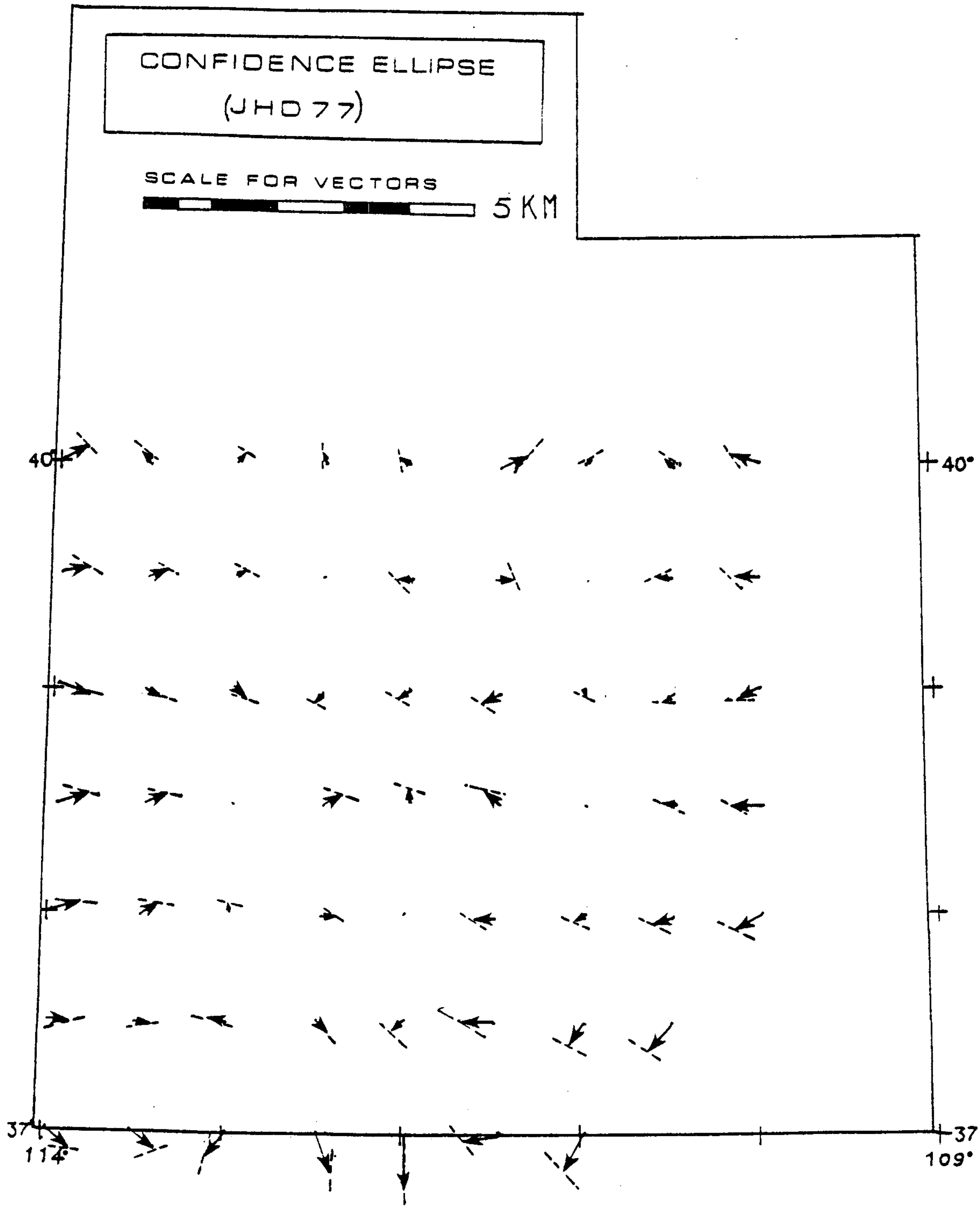


Figure 3.10. Misplacement (km) of synthetic events of Figure 3.3 . True locations are grid points; tips of arrows are computed locations. Dashed lines are major axes of confidence ellipses.

the orientation of the deviations from the true location were dependent in part on the choice of reference event. Confidence ellipses are of course the same regardless of the reference event. Figure 3.7 shows some of the effects of the reference event on the displacement and the relationships to the orientations of the confidence ellipses.

Analysis on a Local Scale

Seismic event location on the scale of a microearthquake survey is different in some respects from location with a regional network of permanent seismograph stations. The area under consideration is usually covered by more stations (usually more than 10), and events generally lie within the array. Station distances range to about 60 or 70 kilometers. This means that the possibility of resolving the focal depths exists in many cases. In a local survey, sources of random noise such as are contributed by different types of stations and different analysts picking and timing are probably reduced.

The hypocenter programs were run on a number of test cases using stations in a microearthquake survey set up in the Roosevelt Hot Springs and Cove Fort areas during the summers of 1974 and 1975 (Olson, 1976). The synthetic data were generated in a manner similar to that for the regional test data.

The first objective was to determine the differences between the two joint location programs. Some aspects of Dewey's program JHD77 (Dewey, 1978) have already been discussed under the regional discussion, where it was used as a joint epicenter determination (JED) program. The other program, GHYP1 (Bolt et al., 1978) was written by R. Urhammer to invert on earthquake locations, station adjustments, and P and S velocities for a simple model composed of a single half-space. Because of the half-space model it is applicable only to a local network with event-to-station distances not exceeding about sixty kilometers.

A variety of different test runs were performed to experiment with a number of variables such as station distribution, number of P and S arrivals, sensitivity to the depth parameter, and initial starting location and origin time. Figure 3.11 shows the results of a test run on noisy data (standard deviation of noise = 0.10 sec). For this case the data were generated for different depths, but in the solutions the depths were fixed at one intermediate depth (7.0 kilometers). Stations used were different for each event; the number of recordings was intended to be typical for actual events located at similar positions. True locations are at the center of each plot; the points represent the deviations of the JHD epicenters from the true positions.

Both programs failed to locate accurately one event that was purposely created with few arrival times and at a position at the edge of the recording array. GHYP1 mislocated two other events that were also lacking close (within 20 km) stations. However, for the most part the results for the two programs were comparable in accuracy. JHD77 is the more versatile and general in format; GHYP1 works as well for a situation in which little or nothing is known about the velocity model.

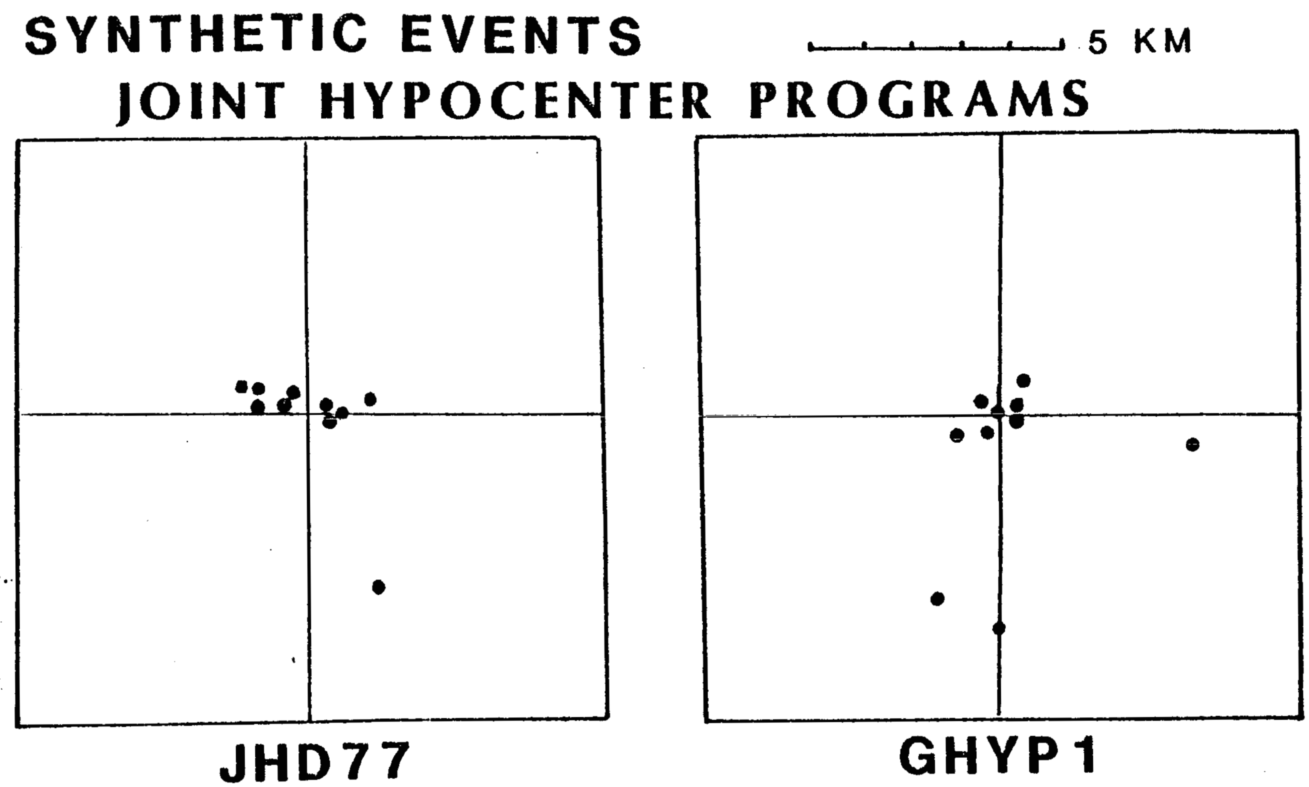


Figure 3.11. Comparison of two joint hypocenter determination programs. Nine synthetic events; random noise added with standard deviation of 0.10 sec. Depths varied from 0 to 15 km, but were fixed in computations at 7 km. Points indicate positions of computed location relative to true locations plotted at the center of each square.

This introduces the question of how sensitive the computed locations are to varying the velocity model. At first it seemed rather surprising that GHYP1 had comparable results to JHD77 since the velocity model was known and used in JHD77 whereas an average velocity, purposely computed to be slightly too high, was used in GHYP1. But the results emphasize one advantage in using a joint hypocenter determination (JHD) method that simultaneously computes station adjustments. To a large degree, any inadequacies or errors in the velocity model are incorporated into the station adjustments. In both single event programs considered, HYPOELLIPSE and SE77, the option exists to essentially improve upon known velocity models by adding appropriate station delays, or adjustments. The problem still remains concerning how to compute these adjustments initially.

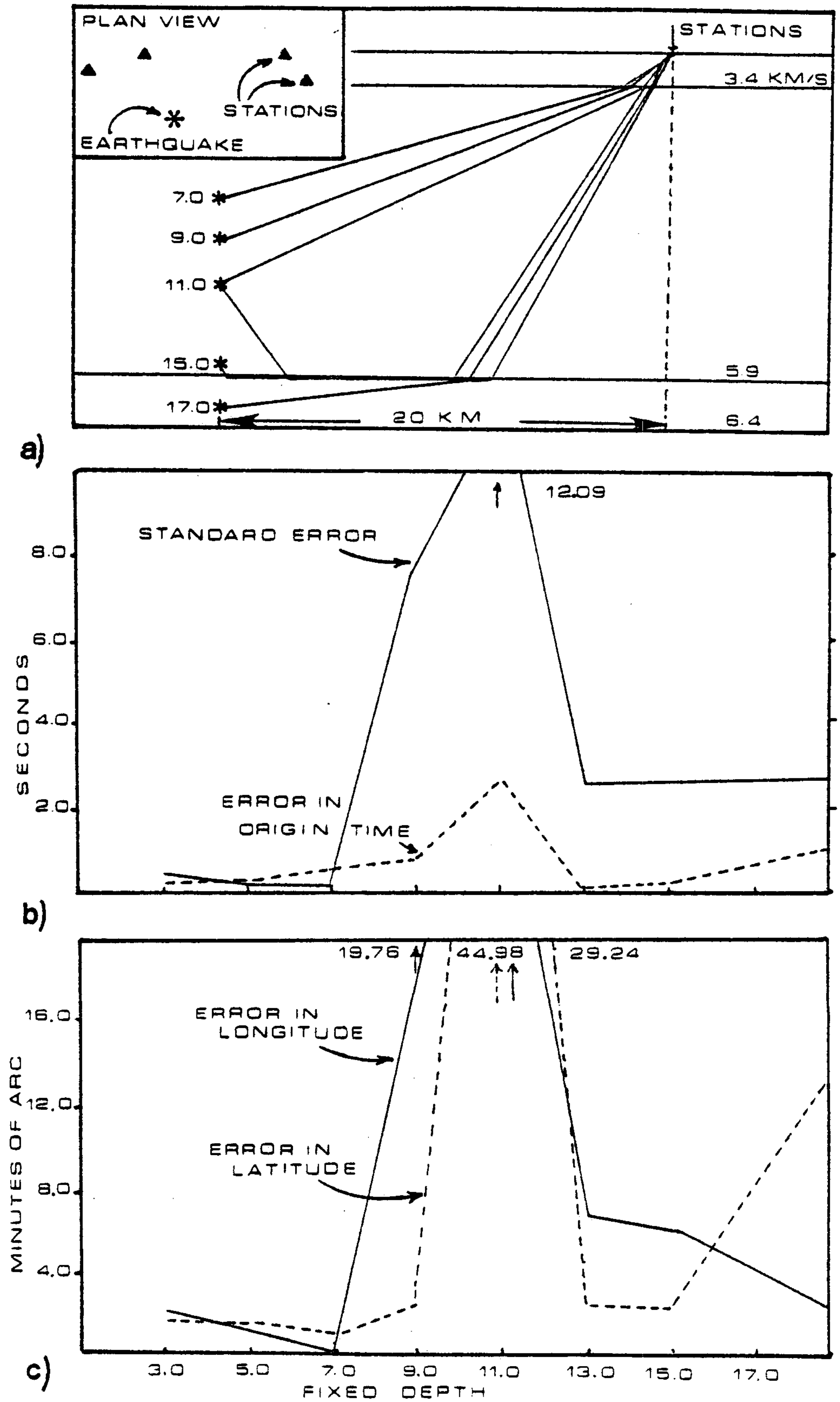
There are some features in the program GHYP1 that are helpful in diagnosing some characteristics of hypocenter solutions, although they may not be necessary for routine location work. These additional features provided some valuable information during the test runs. One feature was the printout of the parameter correlation matrix for each event. For reasons discussed in Chapter 2, the depth and origin time parameters were highly correlated for all events (correlation coefficients greater than 0.80). Closer scrutiny of the degree of correlation with respect to various station distributions revealed that as distances from event to station decreased the correlation coefficient between depth and origin time increased. Similar observations were made for greater azimuthal control, also often accompanied by an increase in the correlation coefficient. Initially this seems contrary to what might be expected, but it can be qualitatively explained by noting that, for a very poor distribution of stations in either distance or azimuth, control over the focal depth parameter is essentially lost, i.e., it is not correlated to any measured quantity. As station distribution improves, the program regains some control over the depth and tries to resolve it; then it is more highly correlated to the origin time parameter.

A second aspect of the program GHYP1 that contributes some additional information is the method used by Urhammer (Bolt et al., 1978) to terminate the iteration procedure. It is similar to HYPOELLIPSE in that it is based upon the step size in parameter space. In the test runs the convergence criterion was rarely satisified if the focal depth calculation was included, so the program was modified to include only the epicentral coordinates in this particular computation. Using a convergence criterion instead of simply specifying the number of iterations, as Dewey (1978) does, gives the user more control over the uniformity of the results. Of course a similar effect can be accomplished by experimenting with the number of iterations and studying the output to see if the results satisfy some criterion, but it is less time-consuming to let the computer do this. The disadvantage is that every one of the group of events must satisfy the specified value for convergence in GHYP1, otherwise the program will stop short of completing the statistical analysis.

During the initial tests of JHD77 the number of iterations was set too low (3 iterations), and an interesting result was noted which contributed some insight into the stability of solutions and the practice of fixing depths. Figure 3.12 shows the position of a particular event with respect to the recording array geometry. Theoretical travel times for this synthetic event were computed for a depth of 15 km, but in the program the depth was set first at 7 km, then at 15 km. The epicentral solution for the correct depth was not as good as it was for the incorrect depth at 7 km. Seven more runs were then carried out with depths fixed at several values, and the results are seen in Figures 3.12b and 3.12c. It can be seen that a minimum is reached at a fixed depth of 7 km for the epicentral mislocation and for the standard error. The explanation I propose for this is that for each particular combination of hypocenter, velocity model, and station distribution the path through parameter space is not smooth. This is an example of a false minimum in the solution space that could have been avoided if the program was allowed more iterations (subsequently verified in further test runs). From tests such as this with poor data where instabilities are likely to occur, the conclusion was that at least five to six iterations are necessary when running JHD77, although with better data the solutions did not change significantly after the third or fourth iteration. This problem would not occur in programs such as GHYP1 and HYPOELLIPSE, because testing the size of the previous step in parameter space would have forced the computations to continue for at least one more iteration.

The preceding discussion is applicable to situations where conditions are such that the data are almost sufficient to resolve the depth. In most cases when the focal depth is fixed there is no

Figure 3.12. Illustration of a false minimum in the solution space for a particular event located with three iterations of JHD77.a)Location of event, velocity section, and some possible ray paths to stations. The true focal depth was 15 km. Several different depths were fixed in computations; after three iterations the errors in solutions were as shown in b) and c). The preferred depth was 7 km. Six iterations resulted in smaller errors and a preference for the true focal depth of 15 km.


PLAN VIEW
EARTHQUAKE
STATIONS
STATIONS
3.4 KM/S
5.9
6.4
7.0
9.0
11.0
15.0
17.0
20 KM
a)
12.09
STANDARD ERROR
ERROR IN ORIGIN TIME
SECONDS
8.0
6.0
4.0
2.0
b)
19.76
44.98
29.24
ERROR IN LONGITUDE
ERROR IN LATITUDE
MINUTES OF ARC
16.0
12.0
8.0
4.0
3.0
5.0
7.0
9.0
11.0
13.0
15.0
17.0
FIXED DEPTH
c)

significant change in the epicentral location for different values of the fixed depth. This indicates that the underlying assumption for fixing the depth is valid; that is, the fixed parameter no longer has any influence on the solution because its changes are negligible compared to the other parameters.

Further tests were conducted using JHD to determine whether the error in depth, when included as a parameter in the solution (not fixed), was more dependent on the distance to the closest station, the size of the largest gap between event-to-station azimuths, or the distance between the event under consideration and the reference event. Plots of the error in focal depth versus each of these influences separately and in combination revealed no simple empirical relationship for a single cause or combination of these influences.

Tests were also conducted to determine the sensitivity of the hypocenter location and station adjustments to varying numbers of stations, azimuthal control, and to characteristics of the reference event. Details of the tests and results are not essential to this discussion, but several conclusions were reached which are applicable to all of the programs used. Conclusions regarding the reference event apply equally to the master event and JHD techniques.

The standard error (in JHD77) and the RMS residual (in HYPOELLIPSE) can decrease with decreasing station density and poorer distribution of stations with respect to distance and azimuth. This can be accompanied by no difference in the location of the event, indicating that the distribution was good enough with fewer stations and removing some of the stations was equivalent to removing random noise. It can also occur if the distribution becomes so poor that a number of different locations will satisfy the least squares procedure equally well.

Selecting the wrong focal depth for the reference event appears to effect the accuracy of the other locations very little, but station adjustments may absorb the error introduced into the travel-time computations, differing by several tenths of a second in some cases. Setting the wrong origin time for the reference event will bias all origin times and station adjustments by similar amounts. Setting the wrong location for the reference event will bias all other locations and station adjustments by similar amounts; the direction of the mislocations will also be similar. Stations not recording the reference event do not show significantly larger magnitude station adjustments or residuals for the JHD technique, but are variable for the master event technique.

As was previously mentioned, results for Dewey's JHD77 and Urhammer's GHYP1 joint location programs are comparable. JHD77 is more general and more applicable to situations where the velocity model is not completely unknown. For these reasons it was decided that further work would be done only with JHD77 as representative of the joint hypocenter determination (JHD) method.

The next step was to compare three techniques: the JHD, master event, and single event methods. Figures 3.13a, b, c, and d show the deviations of 14 estimated epicenters from the known synthetic event locations at the center of each plot. The data for this comparison was generated without any noise added. All of the stations recording

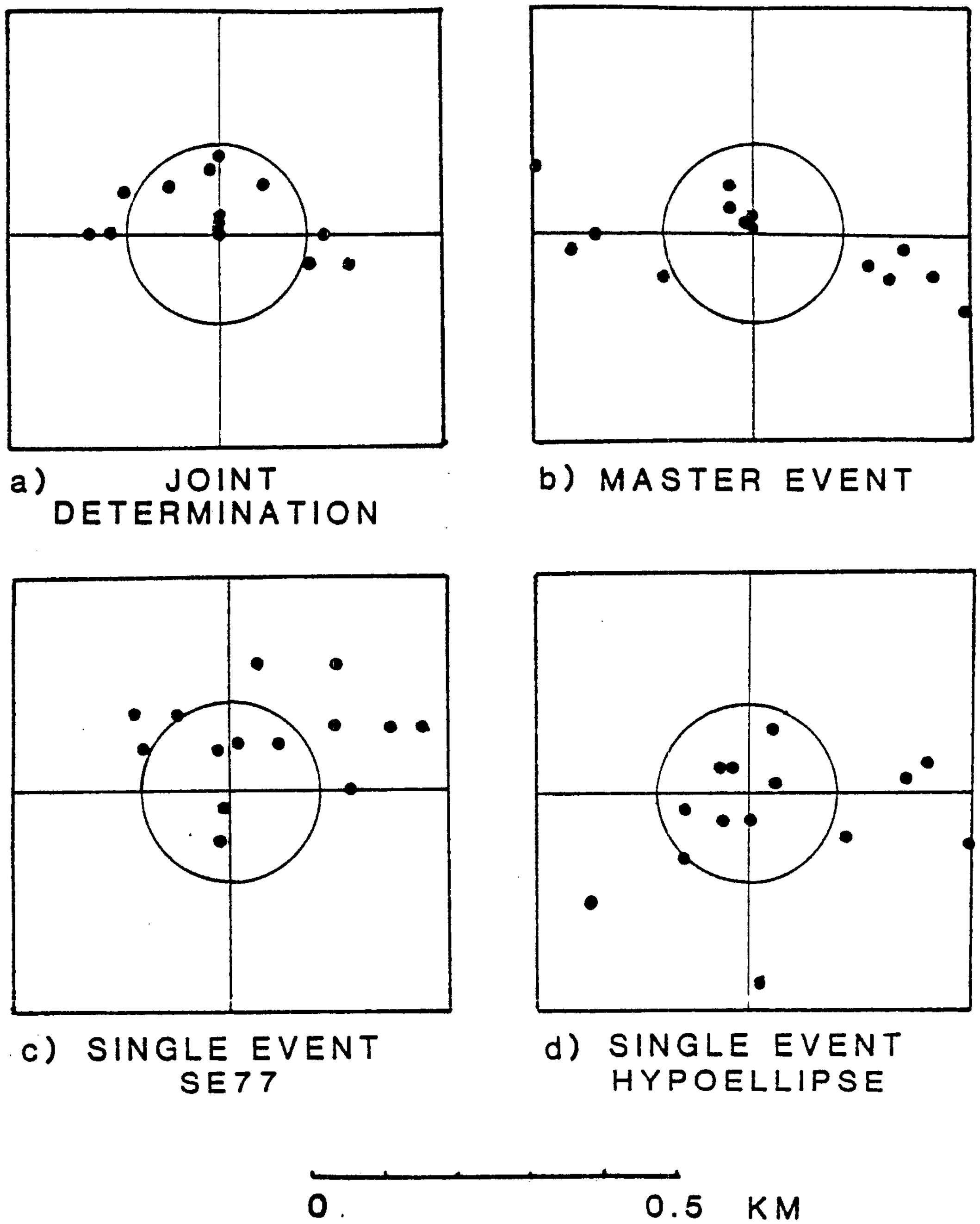


Figure 3.13. Deviations of 14 estimated epicenters from true locations of synthetic events. No noise added to arrival times, but events in different locations used different subsets of stations and different numbers of S-wave arrivals. Circle of diameter 1.25 km is for reference.

real events in the region of the synthetic event location were used, but S-arrivals were computed only for the nearest station. Note that there is a difference in scale between these plots and those in Figure 3.11. This indicates some of the difference that, regardless of the location method used, can be expected when random noise is added to the data. In Figure 3.11 mislocations were on the order of 0.5 to 1.5 kilometers, whereas in Figure 3.13a the mislocations were all within 0.25 kilometers.

The results in Figure 3.13 show clearly that JHD and master event techniques are better than the two single event programs at locating most events correctly with respect to the reference event. The scatter is reduced in Figures 3.13a and 3.13b, and there are more events closer to the correct locations at the center. For the synthetic cases the location of the reference event is known perfectly, so the plots show the absolute accuracy of each synthetic event.

Comparison of Figures 3.13a and 3.13b reveals a feature of the master event technique that can lead to a poorer location than with a single event method. This can occur if the station adjustments computed relative to the master event are incorrectly applied to another event, i.e., the reference event is not adequately representative of all events it was used to help locate. The events which fall outside the circle of radius 0.25 km in Figure 3.13b are probably in this category, and all of them are events farthest from the reference event or approaching the edges of the array coverage.

This leads to one of the significant advantages of the JHD technique. Because the reference event in JHD is introduced to contribute additional stability to the equations of condition and not only to compute station adjustments, the effect discussed in the previous paragraph less pronounced. Adjustments are computed relative to a group of up to fifteen events, and no station should have an adjustment that is unreasonable for some of the events. If these adjustments are then used to relocate more events there is a better chance that they will be representative of the whole region in question, rather than just a small area surrounding the one master event.

The fact that JHD must take into account all station-event residuals in computing station adjustments is usually advantageous, as just discussed. But if one or more events are considerably different from the rest of the group their influence can reduce the effectiveness of the technique. This was observed in one case where the whole group of JHD-located events was within a smaller radius of the true locations than the same group located with the master event technique. However, there were a few events that located better with the master event method. Thus, for JHD, the group of events as a whole will have better locations than with single event or master event methods, but the accuracy of the best events may suffer slightly to compensate for the changes in station adjustments due to improvements in the locations of the poorest events.

A summary of the major comparisons and conclusions discussed in this chapter was compiled into Plate 1. Specific problems and situations are listed, and the applicability and/or limitations of each program studied are presented for a list of specific problems and situations. Information derived from my experience with the methods (applications from Chapters 4 and 5) was also incorporated into Plate I.

CHAPTER 4

PATTERNS OF EARTHQUAKE DISTRIBUTION IN SOUTHERN UTAH

The primary result of tests on the various epicenter location techniques was that, for detailed analyses of earthquake occurrence, the joint epicenter determination (JED) method provides better solutions for relative earthquake locations. In this chapter interpretations of earthquake patterns are presented for relocations of epicenters in southern Utah using the JED technique. Throughout the following discussions the question will be considered: How does the relocation of epicenters contribute to the understanding of earthquake occurrence, and are the patterns tectonically significant?

Summary of Structural and Tectonic Characteristics

The transition between two major physiographic provinces, the Basin-and-Range and the Colorado Plateau, is expressed in south-central Utah as a 150 km to 200 km -wide zone of changing crustal and probably upper-mantle properties. This zone may have represented a structural and depositional transition zone between cratonic and marginal crust since the end of the Precambrian (Stewart, 1978); hence, the coincidence of many structural features has a long and complex history. A summary of all of the geologic and tectonic characteristics and possible models to explain their occurrence woud be too extensive for the purposes of this study, and can already be found elsewhere (GSA Memoir 152). However, it is appropriate to briefly review some of the geological and geophysical observations that may have a direct relationship with concentrations of earthquakes in southern Utah.

Eaton et al. (1978) compared regional, long wave-length Bouguer gravity data of the wstern U.S. to other geological and geophysical information, including seismicity, heat flow, crustal thickness, surface topography, and Cenozoic volcanism and faulting. They found evidence for a broad crustal upwarp expressed by a regional gravity low (-90 mgals) in the Great Basin. The gravity low coincides with relatively high heat flow ($\geq$1.25 HFU) and low P_n velocities of 7.5 to 7.6 km/s (Smith, 1978). This anomalous region is bounded on the east and west by a decrease in the age of faulting and increase in seismic activity.

Cenozoic faulting is considered to be contemporaneous with bimodal basalt-rhyolite volcanism along much of the western margin of the Colorado Plateau (Stewart, 1978; Christiansen and McKee, 1978). Silicic volcanism, the composition of which is indicative of melting of crustal material (Nash and Hansel, 1975; Lachenbruch and Sass, 1978), could have an origin in crustal partial melt zones. The presence of partial melt zones in the crust is suggested by the abnormally high heat flow (Blackwell, 1978; Lachenbruch and Sass, 1978) and crustal low velocity layers (Smith et al., 1975).

There are other mechanisms that can explain the presence of crustal low velocity layers. The presence of listric faulting, with fault planes that rotate to almost horizontal at depths of 3 to 7 km, is indicated by reflection data in the overthrust belt from Wyoming to southern Utah. These fault planes may converge to a single detachment horizon that may be the locus of high concentrations of pore fluids. One model for the listric faulting and associated décollement zones has been proposed by Davis and Coney (1979). They relate the presence of a sedimentary cover deformed by "growth faults" to the extensional tectonics and concurrent formation of metamorphic core complexes. Prograde metamorphism could then provide a source for excess fluids. Additionally, the décollement surface could be genetically related to regional unconformities or to previous planes of Sevier or Laramide thrusting (Davis and Coney, 1979). The chronological sequence of events has not been resolved, however.

A number of mechanisms have been suggested to account for all or some of the phenomena observed in the region of the Great Basin - Colorado Plateau transition. Some prefer an active spreading mechanism, such as mantle upwelling (e.g. Eaton et al., 1978). Christiansen and McKee (1978) have summarized evidence lending support to a model that explains the properties of the region as the passive behavior of inherently different crustal regions responding to stresses introduced by plate movements. Another theory summarized and favored by Stewart (1978) is the presence of back-arc spreading and associated mantle upwelling above a subduction zone.

General Patterns of Earthquake Occurrence

The character of epicentral trends and patterns in Utah and their relationships to tectonics and structure has been discussed in several recent publications (e.g., Smith, 1979; Smith, 1978; Anderson, 1978). The present study elaborates on some of the ideas suggested by these authors, but the emphasis is on assessing the contribution of accurate and precise earthquake location to the interpretations.

In examining the very general trend of Utah seismicity, it has been observed (Smith and Sbar, 1974) that earthquakes occur in a rather diffuse zone that trends north-south along approximately 112° W longitude. At approximately 38° 30' N latitude the trend of earthquakes changes to a more southwest orientation (Smith, 1979). Figure 4.1 is a map of Utah showing the relocated earthquakes in this zone of seismic activity and their spatial relationships to regional Bouguer gravity contours (Cook et al., 1975) and a line marking the eastern extent of thrust faulting of the Sevier orogeny according to Hintze (1975).

The eastern margin of the gravity low discussed by Eaton et al. (1978) coincides with the zone of earthquakes (Figure 4.1) in Utah, and also with an increase in crustal thickness, as determined from refraction profiling (Smith, 1979). The characteristics of this transition from the Great Basin to the Colorado Plateau have been studied extensively, and sumaries can be found in many papers (e.g. Anderson, 1978; Smith, 1978; Best and Hamblin, 1978). Faulting in south-central Utah (Anderson, 1978; Best and Hamblin, 1978) and focal mechanism solutions complied by Smith and Lindh (1979) suggest that the area is subject to east-west extension.

Some general observations can be made in comparing the relocated earthquakes in Figure 4.1 to previous compilations of epicenters. The

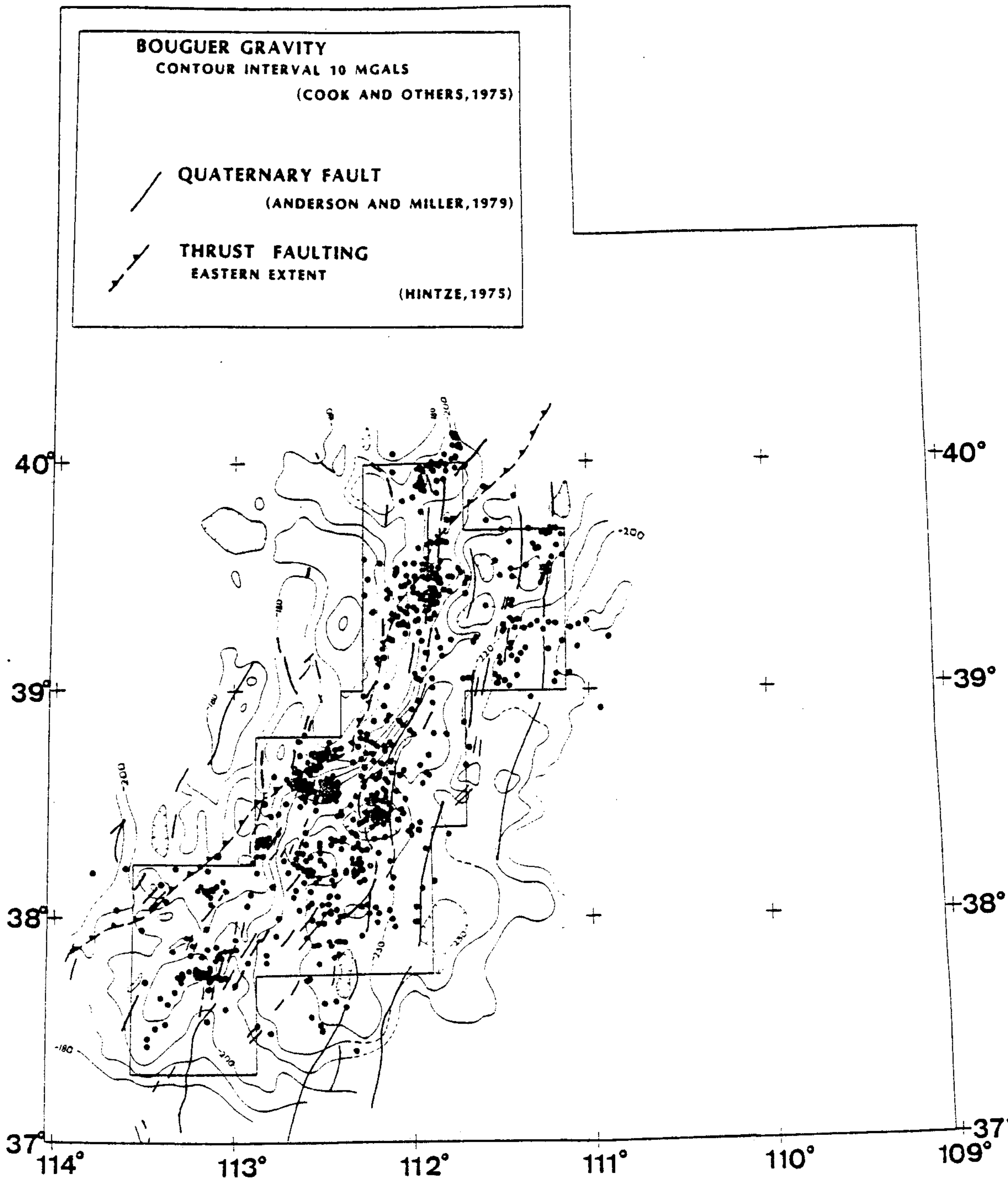


Figure 4.1. Map of relocated epicenters in southern Utah. Area of original earthquake data set is within boxed outline.

relocated earthquakes exhibited a more linear character paralleling the major Wasatch and Joe's Valley fault zones, which will be discussed in more detail pressently. Clusters of epicenters were observed to tighten-up after being relocated, covering less area on the map. Precise locations of earthquakes can thus better define their relationships to regional fault patterns, although the addition of focal depth control is probably essential in resolving the relationship with possible listric faulting.

Activity is also concentrated south of the Elsinore fault, at approximately 38° 15' N latitude. This cluster corresponds spatially to a regional gravity low of about 20 to 30 mgals. The orientation of the earthquake zone in Utah changes in this same area from a north-south to a southwest direction. Possible east-west trends in the regional gravity contours (Figure 4.1) have been recognized by Eaton et al. (1978). They suggest that major east-west extension requires transverse structural features parallel to the direction of extension. Attempts to recognize major east-west trends in the seismicity have been largely qualitative. Precise locations of earthquakes relative to each other would have delineated any major trends. But the spatial association of geothermal areas, Tertiary and Quaternary volcanism, and epicenter clusters in this area of Utah is better resolved by the relocations.

There are other possible causal mechanisms to explain patterns of earthquake occurrence, which have only an indirect relationship to any tectonic framework. Decreases in water tables or aquifers due to abnormal demands close to population centers or heavily irrigated regions could produce local subsidence and differential compaction that might result in small, shallow earthquakes. Any abnormal increase or decrease in the water table could change the effective stress in a particular region by changing the pore pressure (Gretener, 1978). The brittle or ductile behavior of the rock units may also be affected by a change in pore pressure (Brace, 1972). Either of these mechanisms could conceivably be sufficient to produce shallow earthquakes, or at least to localize stress release to enhance the occurrence of events in clusters.

Procedures for Application of Method

For these analyses catalogs and maps in southern Utah from 1962 through 1978 (Arabasz and McKee, 1979) were studied to delineate areas of interest, including areas where (i) changes in directions of regional trends were apparent, (ii) possible correlations with general geologic or structural characteristics were observed (e.g., faulting, recent volcanism), (iii) concentrations of earthquakes in linear trends, zones, or clusters existed. Consideration was also given to the position of areas with respect to the station distribution, for reasons discussed in Chapter 3.

Within each of the six regions delineated in Figures 4.2 and 4.3, a number of events were chosen from the catalog of Utah earthquakes (Arabasz and McKee, 1979) as representative reference events for that particular area. The considerations in effect for the choice of reference events, in order of relative importance, were number of phases recorded for the event, azimuthal gap in station distribution,

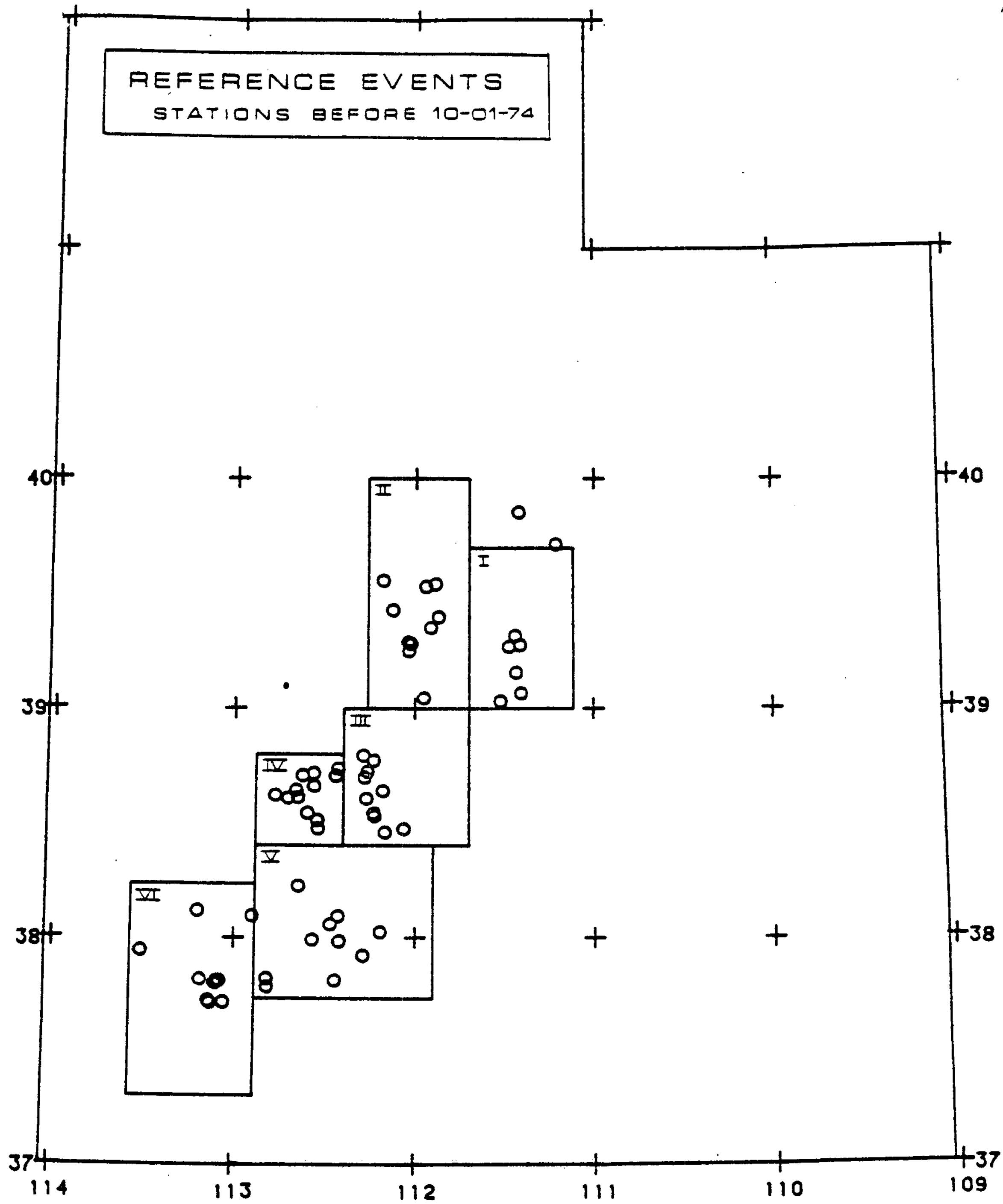


Figure 4.2. Six areas of regional epicenter studies in southern Utah. Circles represent reference events located with the joint hypocenter determination program JHD77. Roman numerals are region numbers referred to in text.

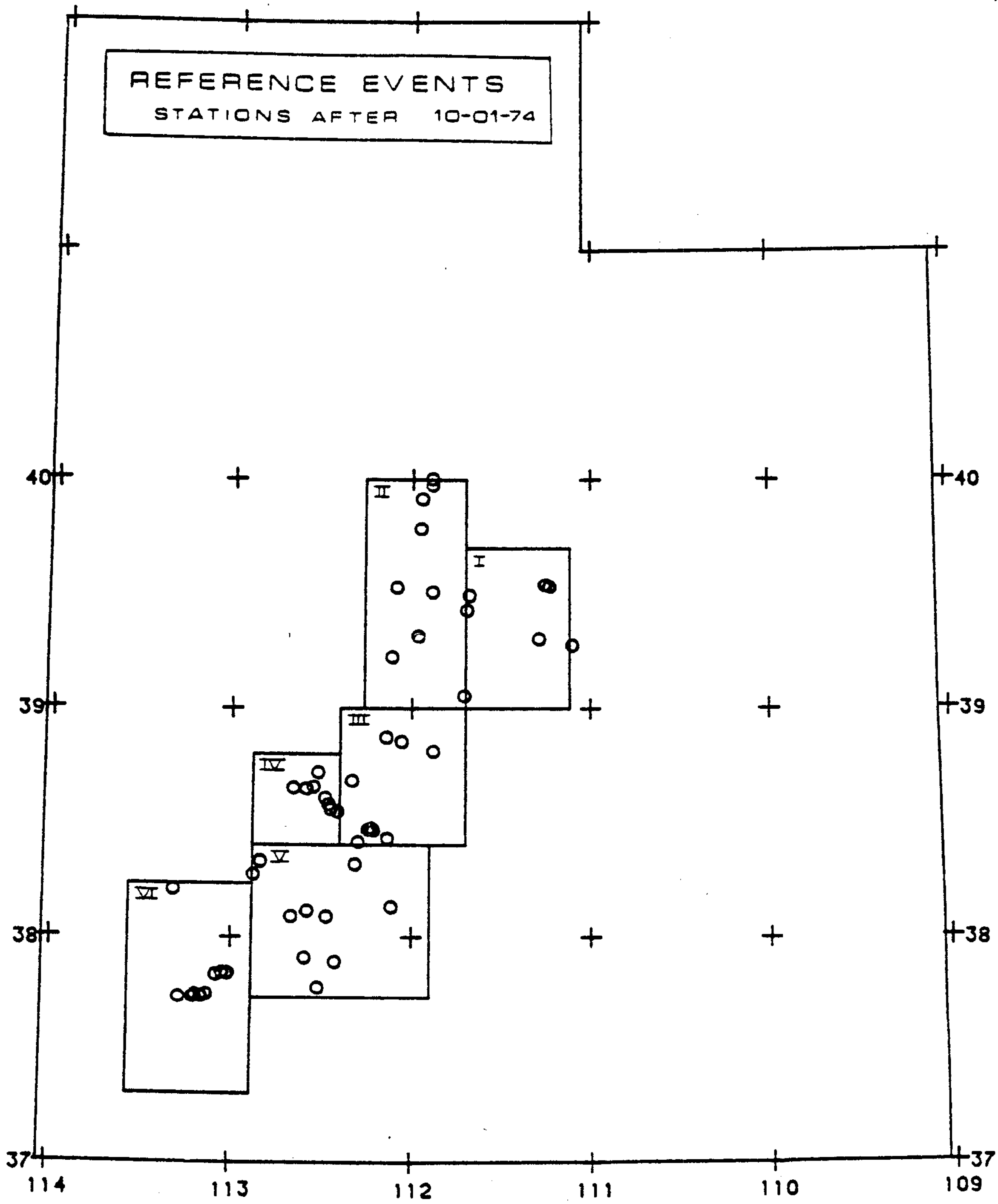


Figure 4.3. Areas same as Figure 4.2, with different reference events for period after October 1, 1974.

magnitude and RMS residual. Reference events had to be chosen for the two time blocks discussed in Chapter 3: before and after October 1, 1974.

The method originally employed for the relocation of this data was an adaptation of the single-event program HYPOELLIPSE (Lahr, 1979) to compute station adjustments. This technique and its limited application was discussed briefly in Chapter 3. The problem is that the single-event locations for the reference events must be used. The residuals are not treated as random variables (Dewey, 1971b) during location of the reference events; therefore, the final station adjustments are based on reference event locations that do not have the benefit of simultaneous travel-time adjustments. It was found, however, that in some cases the relative locations of groups could be improved using this technique. The basis for assessing the improvement was the observation of reduction in the scatter of events, accentuating clusters and trends in the data.

Using the procedure suggested by Dewey (1978) and outlined and tested in Chapter 3, the earthquakes in the areas of interest were relocated using the JED programs. The first step was to locate the reference events and compute station adjustments using the program JHD77 (Dewey, 1978). A total of twelve different sets of station adjustments were obtained in this manner: two for each of the six regions representing the two time blocks studied. These station adjustments were then applied to other events in the same region using the single event program, SE77 (Dewey, 1978). Starting positions for both techniques were the original catalogued single-event

(HYPOELLIPSE) locations. However, one data set was also located using the center of the group of earthquakes as the starting position for the JED events within that region. The solutions for both starting locations were essentially identical, differing by less than one kilometer. A listing of all relocated earthquakes appears in Appendix C.

Maps of relocated earthquakes depict the final epicenter locations using the JED technique. Comparison of the two sets of final locations, JED and adjusted HYPOELLIPSE, resulted in conclusions that were consistent with expectations based on theory and on the previous test cases. The JED solutions showed considerably less scatter than those determined by the single-event HYPOELLIPSE program with station adjustments.

Initially it was thought that it might be possible to partially adjust for mislocations due to the recording station distribution by first creating a set of test events for each region, computing station adjustments with JED, and applying these adjustments to the reference events. This was attempted for all of the regions for the stations in the data set before October 1974. It was found that the adjustments computed for the test cases were, at most, an order of magnitude smaller than the final adjustments computed for the reference events (hundredths compared to tenths of seconds). Any effects on subsequent station adjustments or locations by the addition of these preliminary adjustments were essentially indiscernable. It was concluded that, beyond the qualitative knowledge of the varying location capability across the array obtained from the test cases, the effects of poor station distribution cannot be significantly corrected by any method.

For earthquakes in southern Utah relocated with the regional seismic array two velocity models were used. Seismograph stations were divided into two groups, one comprised of stations located within the eastern Great Basin and the other including stations in the Middle Rocky Mountains and Colorado Plateau (Richins, 1979). All stations west of 111° W longitude, except for the station GCA, were placed in the Great Basin group, for which the velocity model was based on a refraction profile of Keller et al., 1975. It will be referred to as the Wasatch Front model. The second velocity model, referred to as the Colorado Plateau model, was from a reversed profile discussed by Roller (1965). Velocities and depths for these two models are given in Table 4.1.

Interpretations of Revised Earthquake Locations

The general epicentral patterns were evident before the relocations were done, and have not been greatly altered by relocation of the epicenters using the JED technique. However, focusing on specific regions of interest on a local scale emphasizes some more definite results. The maps of Figures 4.4 through 4.9 outline the same regions as in Figures 4.2 and 4.3, but include all of the relocated earthquakes in each region occurring from July 1962 through October 1978. The following discussion examines these regions in more detail, concentrating on spatial and temporal grouping such as clustering of epicenters in swarms or along coherent trends. As was stated previously, recognition or delineation of grouped epicenters is

Table 4.1. Velocity models for regional relocations: southern Utah.

WASATCH FRONT MODEL

P-WAVE VEL (Km/s)	DEPTH TO TOP (Km)	LAYER THICKNESS (Km)
3.4	.00	1.4
5.9	1.40	14.1
6.4	15.50	9.9
7.4	25.40	-

COLORADO PLATEAU MODEL

P-WAVE VEL (Km/s)	DEPTH TO TOP (Km)	LAYER THICKNESS (Km)
3.0	.00	1.5
6.2	1.50	26.0
6.8	27.50	12.5
7.8	40.00	-

Figures 4.4, 4.5, 4.6, 4.7, 4.8, 4.9. Regional maps of relocated earthquakes, June 1962 through August 1978.

EXPLANATION

Earthquake epicenter

Town

Quaternary fault

Lake or reservoir

Swamp

Qb :Quaternary basalt

Tv, Tg :Tertiary volcanics or intrusives

Qa, Ta :Tertiary to Quaternary alluvium, colluvium

pЄ-T :Precambrian to Tertiary formations

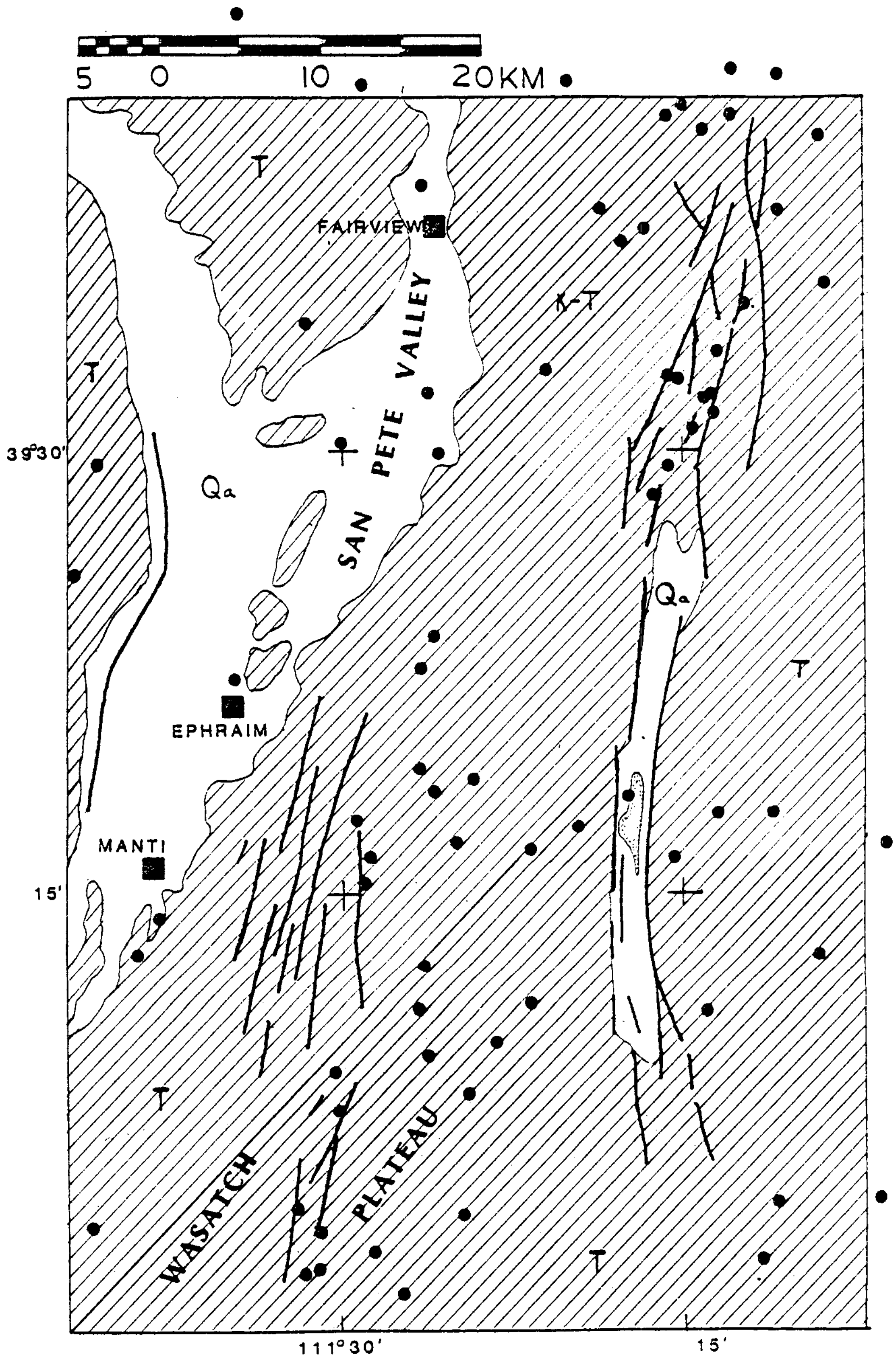


Figure 4.4. Region I.

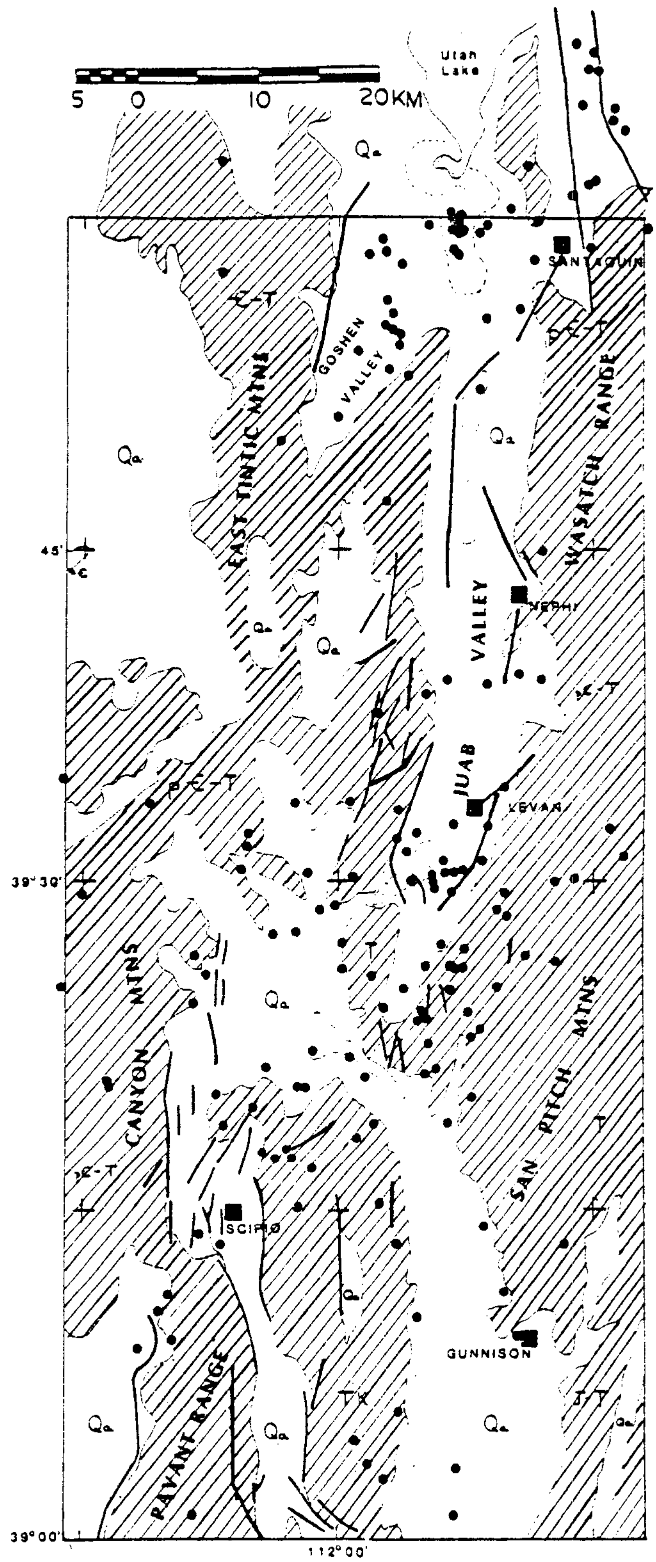


Figure 4.5. Region II.

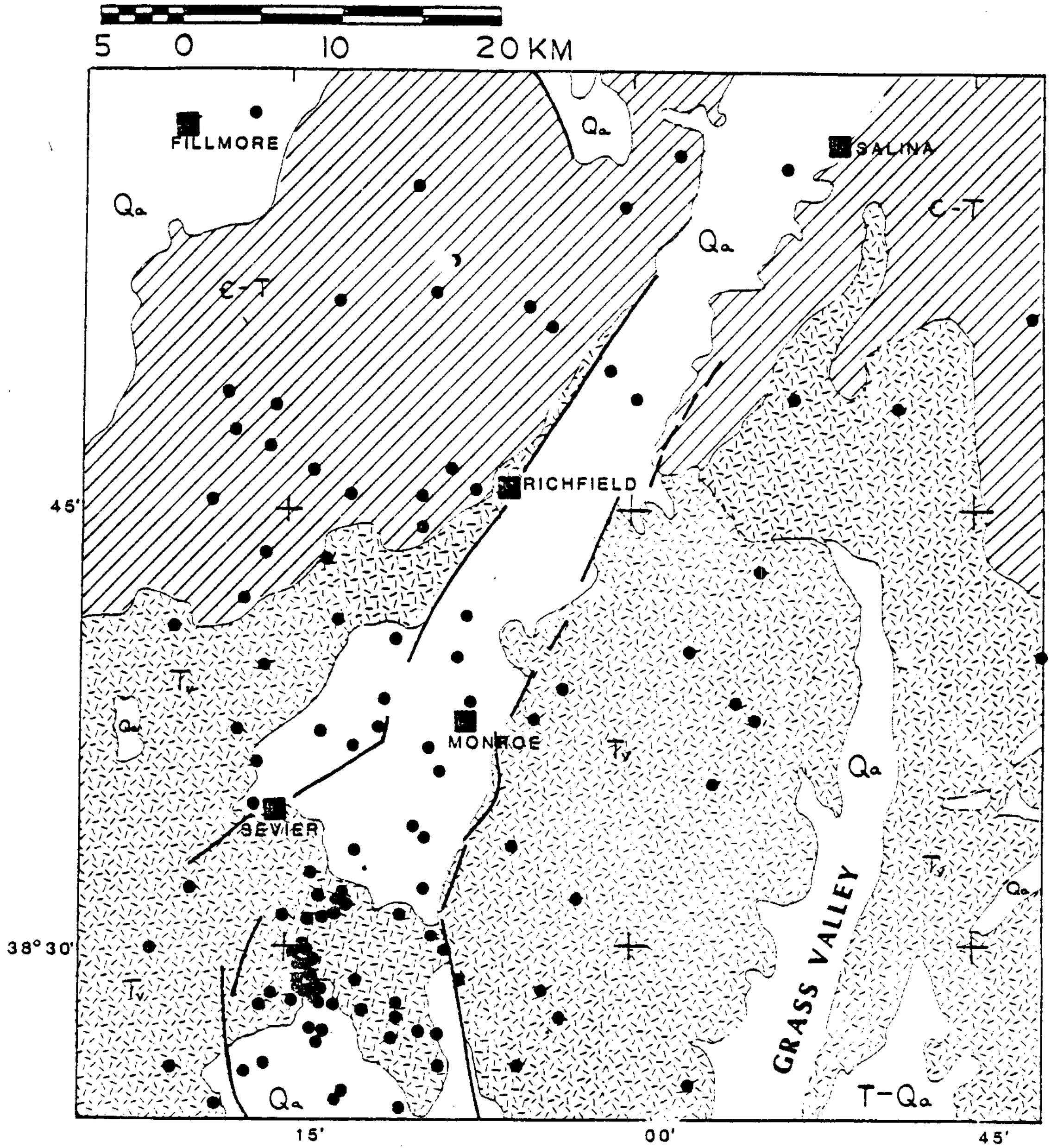


Figure 4.6. Region III.

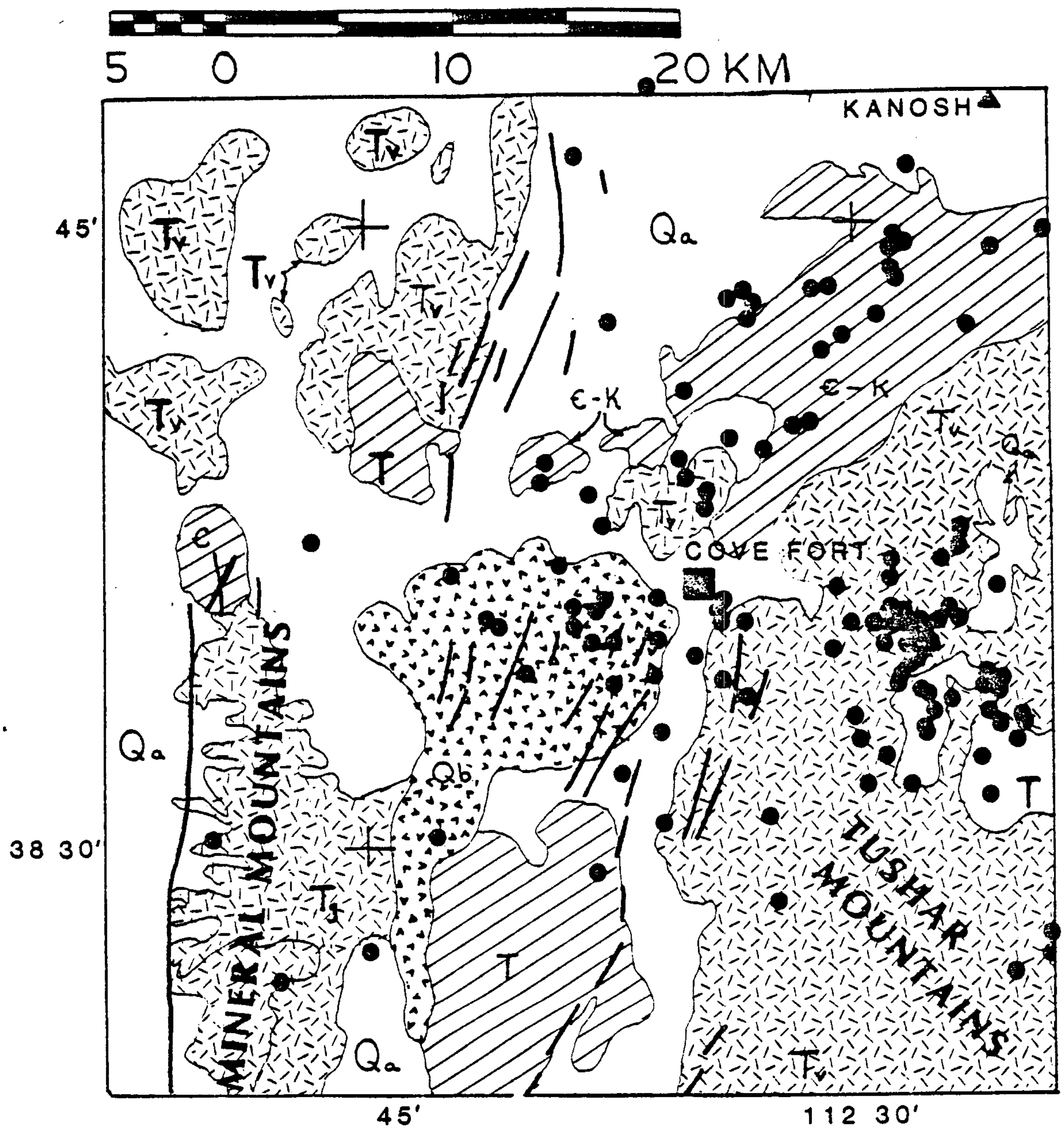


Figure 4.7. Region IV.

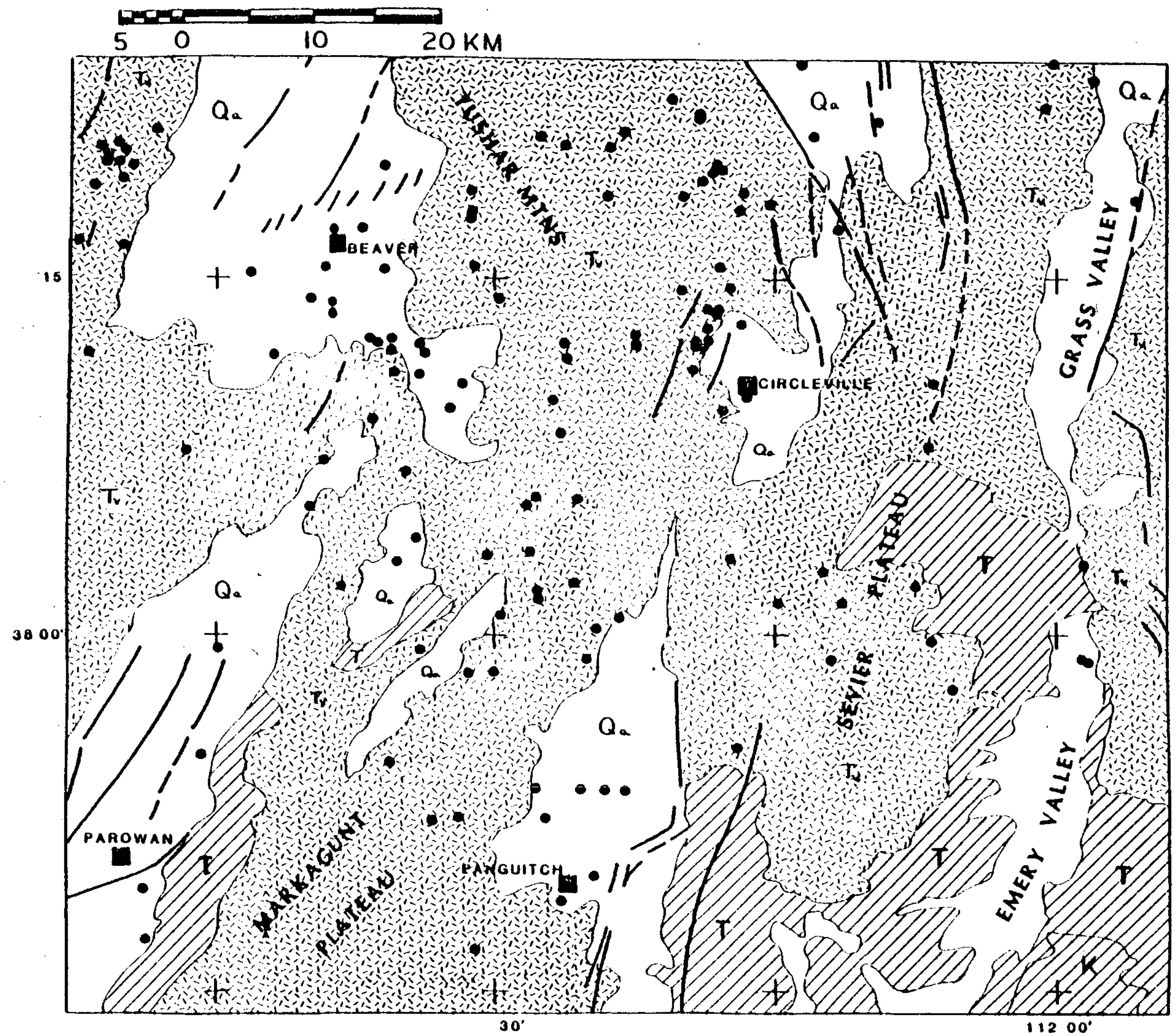


Figure 4.8. Region V.

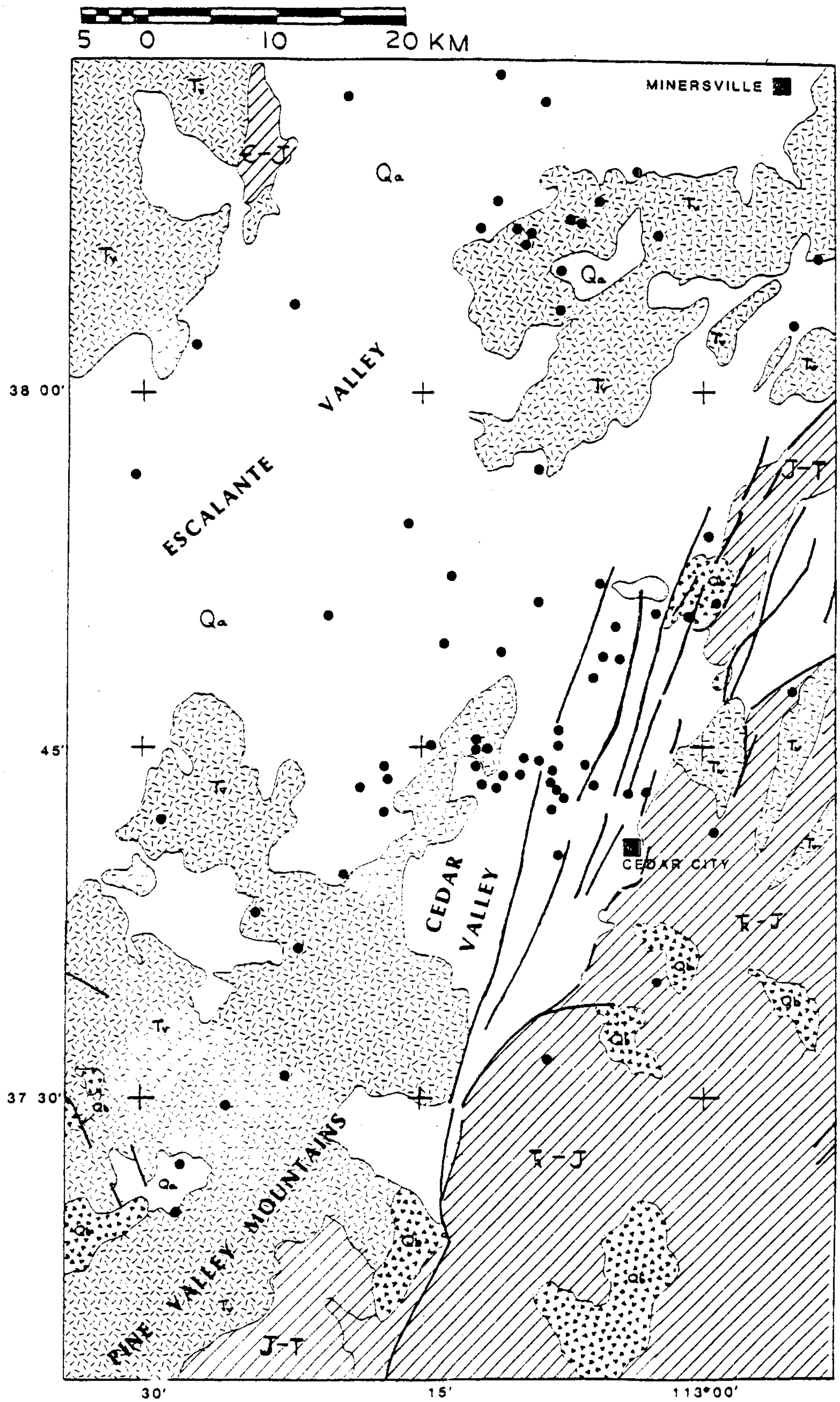


Figure 4.9. Region VI.

enhanced by the use of the JED technique. Additionally, detection of a regional bias in the locations, as indicated by systematic changes in the relocations using JED, is important in interpreting possible regional variations in the velocity model.

The maps in Figures 4.4 through 4.9 show only gross characteristics of geology and structure that appear to have some relationship to the seismicity. Features included, therefore, may not be the same on each map. In the following discussions Regions I through VI refer to the areas outlined in Figures 4.2 and 4.3.

Region I. Figure 4.4 shows the area northwest of the San Rafael Swell, an area of Laramide uplift. It includes the Wasatch Plateau east of the San Pete Valley. In the northeast portion of the map a number of earthquakes occur within the Joe's Valley graben or fault zone. These faults are geologically young, exhibiting displacement as recently as 11,000 years ago (Anderson and Miller, 1979). Recent microearthquake surveys in the same region (W. Arabasz, oral comm.) have shown activity concentrated in generally the same area, which suggests that the regional JED locations are more accurate than the single-event locations. But the most noticeable feature on this map is the diffuse character of the epicenter distribution. Except for the northern part of Joe's Valley, and a few epicenters possibly associated with faulting in the central portion of the region, there appears to be no obvious correlation with any surficial geology or structure.

Earthquakes in the southern portion of this area were systematically relocated approximately 5 to 10 km east of their previous locations. This suggests that there is a bias in the travel-times to stations from these events. This could either be due to a lateral velocity inhomogeneity, or departure from the simple flat-layered velocity model, or it could reflect a constant error in the focal depth (fixed at 7.0 km) for this group of events. Since the station distribution in this area is fairly good, I favor the former possibility because an error in focal depth would result in systematic mislocations only if the azimuthal and distance control were quite poor.

Region II. In Figure 4.5 a zone of rather diffuse seismicity represents the southern continuation of the Wasatch fault. The valleys east of the southern portion of the Wasatch range were areas for which possible correlations between water-level changes and seismicity could be made. At the northern end of the Juab Valley, in the Goshen and Utah Valleys, clusters of epicenters are composed in part of events occurring throughout the recording period (1962-1978); however, there are also several swarms (3 events) during particular years (1973, 1976, 1977, 1978). Water-level change maps have been compiled annually for the Utah and Goshen Valleys, and during these years document fluctuations of -3 to +5 feet west of Santaquin (U.S.Geol. Survey, Cooperative Investigations Reports 13, 15, 16, 18). Records in the Utah and Goshen Valleys are particularly complete, showing changes for water-table aquifers and for various depths in the artesian aquifer. However, changes in water levels in other years, with no swarm activity, are not significantly different. The same is true for possible spatial and temporal association with water-level changes noted for the small cluster of events in the Juab Valley west of the Chicken Creek Reservoir. Lack of complete areal coverage of the water-level records precludes direct correlation with seismicity, but the spatial coincidence should not be ignored in attributing earthquake occurrence entirely to tectonic causal mechanisms. Focal depth information would aid in separation of these events from tectonic-related earthquakes.

Region III. The diffuse earthquake occurrence in the area in Figure 4.6 is even more pronounced in the northern part than it was in Region II. The prominent fault zone includes the northern extension of the Sevier fault, with segments that have been active within the last 3 MY (Anderson, 1978). The Elsinore fault passes through the town of Richfield, and the area was the site of large earthquakes in 1901 (M_L 6.5) and 1921 (M_L 6.0) (Arabasz and Smith, 1979).

South of the town of Sevier is a prominent cluster of epicenters that includes swarms during October 1967 and in late June through August 1978. In this area the Marysvale volcanics, composed of Oligocene to Miocene basalt flows, breccias, and ash flow tuffs, straddle the boundary between the Basin-and-Range and Colorado Plateau provinces (Steven et al., 1978). Extensive Basin-and-Range normal faulting in this area probably began about 21 MY ago, contemporaneous with a change in the overall composition of the Marysvale volcanics from intermediate to silicic-alkalic. About 21 - 17 MY ago, eruptions of ash flow tuff accompanied formation of the Mount Belknap and Red Hills Caldera (Steven et al, 1978). The latter coincides with the present occurrence of the epicenter swarm. An east-west trend in the gravity contours in this region (see Figure 4.3 and Eaton et al., 1978) may be indicative of a causal mechanism on a larger scale that would account for the spatial association of hot springs, Tertiary volcanics, and epicenter swarms.

Also notable in this region is the very sparse seismic activity along the Sevier fault zone near the town of Monroe, the site of the Monroe geothermal area.

<u>Region IV</u>. The Cove Fort area was also part of the Marysvale volcanic system. Earthquakes shown in Figure 4.7 exhibit a clustering in the Three Creeks caldera area, situated in the headwater region of Clear Creek. This was the source of the Oligocene Three Creeks Tuff member of the Bullion Canyon Volcanics (Steven et al., 1978). Epicenters are also concentrated in mapped areas of mafic basalt flows of Pleistocene to Holocene age near Cove Fort. However, studies of Holocene faulting (Clark, 1977) in the area delineated the most active faults, and in general the relocated epicenters do not coincide with these. Eruptions in this area were probably along north-south fracture systems; this and local focal mechanism solutions indicate that the area is extensively fractured and a correlation may exist with some unmapped fault. There is also the consideration of intersection of fault planes with the surface. In the case of basalt, fault planes probably change orientation to conform with the

preferential vertical jointing. At depth the dip of fault planes in the Cove Fort area probably decreases to the regional altitude of 55° - 70° (Clark, 1977), and earthquakes occurring on lower portions of the faults would have epicenters on the surface projected off the fault trace.

Also notable is the lack of seismicity along the Opal Mound fault bounding the Mineral Mountains on the west, in the Roosevelt Hot Springs geothermal area. Thus, three geothermal areas, Monroe, Cove Fort, and Roosevelt, close to each other and presumably in the same general tectonic regime, have very different expressions in terms of the regional seismicity. The Cove Fort and Roosevelt KGRA's will be discussed in more detail in Chapter 5.

Region V. The general fault trends and diffuse seismicity discussed for Region III are continued in the southern Tushar Mountains and the Beaver Valley, Figure 4.8. Several clusters are observed: one northwest of Circleville, probably associated with fractures within the graben in the southern Tushar Mountains, and one at the southern end of the Tertiary granitic intrusive of the Mineral Mountains. Near Beaver a cluster of activity is concentrated in the valley and drainage system of South Creek.

This general area in Utah is where the trend of earthquake occurrence changes to a southwest orientation. JED locations did not exhibit any systematic relocation. In particular, events did not align themselves along any east-west trends.

Region VI. This area (Figure 4.9) is dominated by the Hurricane, Paragonah, and Rush Lake fault zones. Anderson and Mehnert (1979), in a recent study of stratigraphic and structural relationships, suggest that the Hurricane fault is too young to be the physiographic boundary between the Colorado Plateau and the Basin-and-Range provinces in this area. Based on the seismicity, which appears to be more prevalent west of the fault zone (although the data are rather sparse) and the fact that there has been no evidence for Holocene surface faulting on the Hurricane fault (unlike the Wasatch fault), Anderson and Mehnert place the present boundary west of the Hurricane fault.

Spatial and temporal correlations can be postulated for the occurrence of epicenters and water-level changes in the Cedar Valley. In 1971 seven earthquakes of magnitude 3 (M_L) or greater, followed by hundreds of aftershocks detected by portable stations installed after the larger shocks, occurred northwest of Cedar City. Ground-fissuring had been observed prior to the events (Arabasz and Smith, 1979), and problems with subsidence combined with analysis of water-level change maps suggest a spatial relationship at least. But the relationship may be a precusory dilatancy effect rather than any causal mechanism (Arabasz and Smith, 1979). More swarm activity in November of 1977 and the 2-3 months following can also be correlated spatially with areas of large water-level fluctuations, from -3 to -6 feet in early 1977 to +3 to +5 feet in late 1978 (U.S.Geol. Survey, Cooperative Investigations Reports 16, 18).

The distribution of the 1971 epicenters in two distinct zones, 10 km apart, raised questions about the accuracy of the locations. The aftershocks located with the portable stations were about 3 to 5 km east of the northern group of epicenters; this and other information led to a preliminary conclusion that all of the epicenters were actually located in the vicinity of the northern group (Arabasz and Smith, 1979). The stations used in the regional locations varied slightly for each event, and the possible mislocations were attributed to some sort of systematic station bias. Figure 4.10 shows the HYPOELLIPSE single event and the JED locations for the seven events of magnitude greater than 3.0 M_L, with arrows drawn from the single-event to the JED location. The four southern events moved 1-2 km north, but remained clearly separated into two zones. Furthermore, later activity (1978) was concentrated in the vicinity of the southern group. For these reasons it was concluded that there were likely two distinct zones of activity during the 1971 swarms.

Comparisons of the single-event HYPOELLIPSE locations, the HYPOELLIPSE relocations with station adjustments, and the JED relocations reveals evidence that the dispersed character of the epicenters is not due only to "smearing" of the locations as a function of the station distribution; in general, there is no consistent correlation between the direction of movement of the relocations and the orientations of the major axes of confidence ellipses. This factor cannot be ruled out, however. From Figures 3.9 and 3.10 it is evident that the principal effect of the station distribution is to shift computed locations in east-west directions. The direction of the shift will vary depending on which stations actually record a specific event, but is reasonable to assume that the

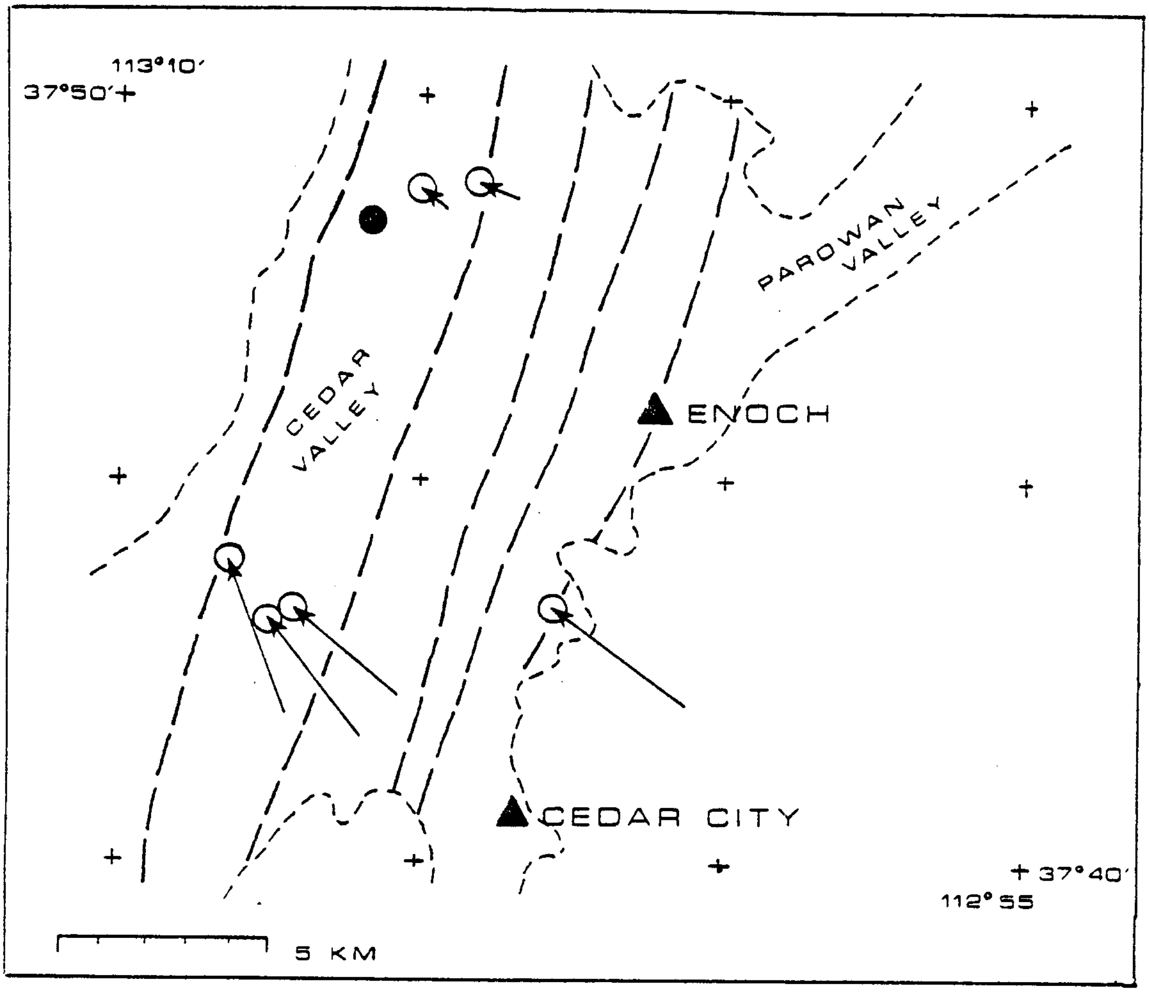


Figure 4.10. Seven events of magnitude greater than 3.0 (M_L) occurring in November 1977 in the Cedar City area. Arrows point from single-event locations to jointly determined locations (open circles). Closed circle is reference event used in JHD77. Heavy dashed lines indicate mapped positions of Quaternary faults. Light dashed line approximates valley margins.

general tendency is to mislocate epicenters in directions perpendicular to the predominant structural trends of the region. The magnitude of this mislocation is dependent on the computed variances, but is probably no more than 10 km for any single event, and usually less than 5 km.

These general results are sufficient to indicate that the diffuse earthquake occurrence in southern Utah is real. A graphical visualization of the nature of earthquake occurrence is depicted in Figure 4.11. Duration magnitudes were used in the relation (R. B. Smith, oral comm., 1978)

$$\log E\ (\text{ergs}) = 9.4 + 2.1\ M_L - 0.02M_L^2$$

to generate histograms of cumulative seismic energy release at 10 km intervals across the transition zone. The dashed line in Figure 4.11 corresponds approximately to the line delineating the east extent of thrust faulting in Figure 4.1, as well as the approximate physiographic boundary between the Colorado Plateau and the Great Basin.

Computations were carried out for a data set larger than that delineated in Figure 4.1. Open bars represent all data from latitudes 37° N to 40° N. Closed (diagonally patterned) bars include data only to 39° N latitude. This was done in order to eliminate the possibly confusing results obtained by including seismic activity in the Price area, much of which is thought to be mining-related. This diagram was generated using a combination of relocated earthquakes and previous single-event locations, but at this scale and due to the averaging effect of 10 km increments over all of southern Utah, the effect of

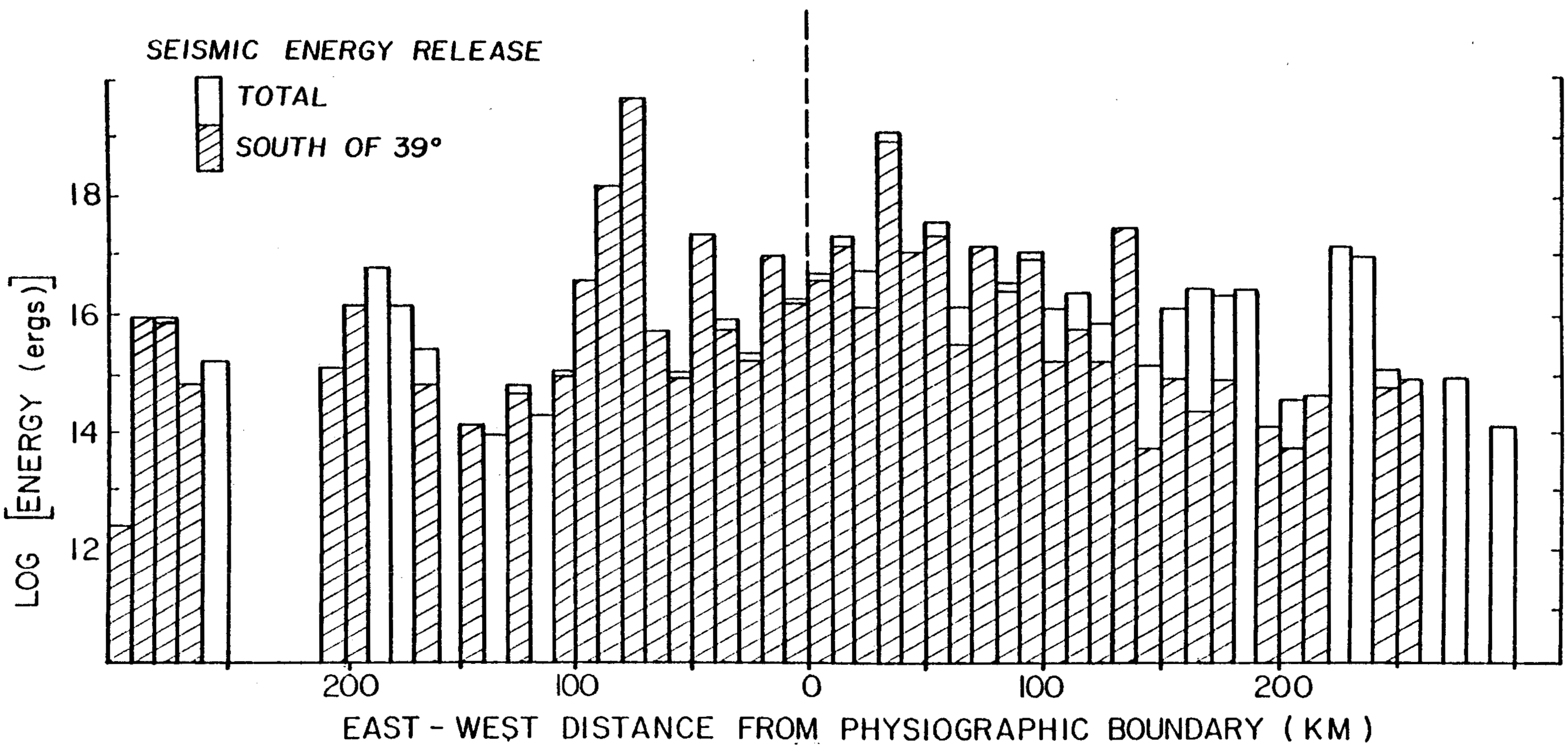


Figure 4.11. Histograms of cumulative seismic energy release along the Colorado Plateau - Basin and Range transition zone. Dashed line represents approximate physiographic boundary.

relocations is negligible. Figure 4.11 emphasizes the point that most of the energy is released over a broad 150 km to 200 km wide zone between the Basin and Range and Colorado Plateau. Previous workers (Anderson, 1978; Smith and Sbar, 1974) have postulated a tectonic mechanism for the loss in coherency of the Intermountain Seismic Belt. They suggest that zones of prominent seismic activity may define intraplate or subplate margins. Epicenters in southern Utah seem to diverge into two zones: one following the boundary of the Colorado Plateau Province and one continuing on in an east-west direction into southern Nevada. If the transition between the Basin-and-Range and the Colorado Plateau is actually represented by a lower crust - upper mantle upwarp, as suggested by refraction, gravity, and P_n-delay data (Eaton et al., 1978; Smith, 1978; Anderson, 1978), then the differing response of upper crustal material to this upwarp could account for the broad zone of activity. The lack of depth control in a regional study such as this prevents much speculation about the occurrence of hypocenters along listric - normal fault planes or zones of decollement, although this could certainly make the epicenters appear randomly scattered on the surface.

Occurrence of swarms (Figures 4.6 -4.8) in the Cove Fort and Sevier areas and southeast of the Mineral Mountains, which cannot be easily attributed to changes in groundwater levels or other mechanisms, may be characteristic of the change in the regional tectonic setting at approximately this latitude. If failure along thrust planes during the Sevier orogeny was enhanced by the presence of abnormal pore pressures, it might be reasonable to assume that localized pockets of anomalous pore pressure still exist in the crust. Alternatively, excess pore fluids could also be present due to metamorphic activity (Davis and Coney, 1979) discussed previously. These could be concentrated in this region due to the change in the lower crustal structure. Some combination of these factors coupled with the possible mantle-upwarp would account for the broad coincidence of geothermal regions, generally diffuse seismicity, and swarm activity in localized areas.

CHAPTER 5

EARTHQUAKE RELOCATIONS AND P-WAVE DELAY DATA FROM BROADSIDE REFRACTION RECORDS: ROOSEVELT HOT SPRINGS AND COVE FORT AREAS

The results of joint hypocenter determination (JHD) of local (epicenter-station distances up to 70 km) earthquakes in the Roosevelt Hot Springs and Cove Fort KGRA's are discussed in this chapter. Interpretations of relocated local and regional earthquakes are then combined to provide a synthesis of earthquake data with information about the velocity structure in the Milford and Cove Fort areas. New information includes an interpretation of local P-delays from a broadside refraction survey. The purpose is to present the seismic data and also to assess the possible value of regional and local earthquake monitoring in geothermal areas in light of the previous discussion of specialized earthquake location techniques.

Earthquake Activity in Geothermal Areas

A number of studies have been carried out in the past several years on the seismicity of geothermal areas (e.g., Ward and Bjornsson, 1971; Hamilton and Muffler, 1972; Hill et al., 1975; Combs and Hadley, 1977). Ward (1972) documented known occurrences of microearthquakes in geothermal areas, and additionally proposed several uses for accurately located microearthquakes. Among them: (i) location of active faults that may act as conduits for hot water, (ii) delineation of the fracture-dominated reservoir, (iii) correlations of earthquakes with aspects of the tectonic setting, to aid in understanding the occurrence of the geothermal system, and (iv) recording of seismic activity before and after fluid withdrawal. Other investigators (e.g. Steeples and Pitt, 1976) have studied geothermal areas in which the lack of seismicity was interpreted as significant. They suggested that absence of seismic activity, combined with the observation of other phenomena such as high heat flow and delayed or attenuated arrivals of P and S waves, might be indicative of a ductile response to stress, due to increased temperatures.

Seismically active geothermal areas may be characterized by continuous or swarm-like activity that has been explained by a number of mechanisms (Ward, 1972; Ward and Jacobs, 1971). Among these are:

(1) High pore pressures changing effective stresses along tectonic planes of weakness (faults). Principal stresses related to potential planes of failure are reduced by increased pore pressure. Depending upon the original value of the deviatoric stress, abnormally high fluid pressures may cause shear failure, shifting the Mohr's circle into the region of instability.

(2) Chemical corrosion of fractures. Weakening of pre-existing fractures may be caused by increased solubility of minerals (e.g., quartz) in solutions under high pressure (Gretener, 1979). This process, however, is dependent also on differential stresses and on temperature conditions.

(3) Weakened rock units due to hydrothermal alteration. This is a complex relationship. Byerlee and Brace (1968) found that some alteration minerals (i.e. calcite and serpentine), favor release of stress by stable sliding rather than stick-slip. Chlorite and micaceous alteration products were found to have varied in responses to applied stress. In some cases these alteration products favored stick-slip. But in other cases the response was stable sliding, thus favoring deformation without brittle failure.

Other localities may have little or no seismicity directly associated with known geothermal reservoirs (e.g., Hamilton and Muffler, 1972; Steeples and Pitt, 1976). Possible causes of this are:

(1) High temperatures favoring failure by stable sliding rather than stick-slip (Brace, 1972). Rocks in higher temperature regimes may not store elastic strain energy to be released by brittle failure, deforming instead by creep mechanisms such as crystal dislocation or diffusional processes (Weertman and Weertman, 1975);

(2) Empirical evidence suggests that porous materials (e.g. tuff) favor failure by stable sliding (Byerlee and Brace, 1968);

(3) Hydrothermal alteration to minerals favoring stable sliding (see #4, above);

(4) Lack of adequate recording spatially or temporally.

Assuming for the present that a possible geothermal system has already been identified, what is the use of microearthquake monitoring to delineate the relationships between earthquake occurrence and the geothermal resource? If the geothermal reservoir is a fracture-dominated system, the chemistry of geothermal brines may reduce permeability and porosity by cementing fractures. It can be postulated that some sort of continuous fracturing must be occurring to maintain an open reservoir. Microearthquakes could be used to help delineate the fault and associated fracture zones or the areal extent of the permeable reservoir. Alternatively, lack of brittle failure may also delineate the extent of a hot region that has a ductile response to stress.

Given that the occurrence of geothermal areas is related to tectonic processes such as are typical along plate margins, in regions of recent volcanic activity, or related to major fault systems, acccurate delineation of microearthquake activity coupled with regional seismicity studies could aid in the interpretation of spatial relationships of geothermal systems with major geologic and/or structural trends. In other words, seismic activity in geothermal areas is probably at least partially the result of aspects of the geothermal environment, such as presence of excess pore fluids contributing to the release of pre-existing tectonic stresses. Finally, earthquakes can be used as sources of energy for studying crustal properties related to geothermal systems from: (i) P and S travel-time delays, (ii) V_p/V_s ratios leading to determination of

Poisson's ratio and (iii) amplitude information for attenuation studies.

It is evident that any conclusions drawn from the above relationships depend upon the precision and accuracy of the earthquake locations. It has already been demonstrated that the precision of earthquake locations, i.e., their locations relative to each other, is substantially improved by the use of joint location techniques. In addition, absolute accuracy can also be improved through the use of a good reference event. If other investigations are being carried out simultaneously, as would probably be the case in a geothermal area, artificial sources (e.g., explosions) could be used as the calibration events.

Relocation of Earthquakes in the Roosevelt Hot Springs and Cove Fort KGRA's

Microearthquakes in the Roosevelt and Cove Fort geothermal areas were originally located by Olson (1976) using the single-event method. The procedures I used for relocation of these events were similar to those used for the regional relocations discussed in Chapter 4, and reference events were chosen using similar criteria. In Olson's (1976) solutions many of the focal depths had to be fixed, especially in the Roosevelt Hot Springs area. Thus an additional specification for the reference events was that the stations recording the reference event had to be sufficient to resolve the focal depth in Olson's solutions.

Figure 5.1 is a map of the 1974-75 microearthquake station array

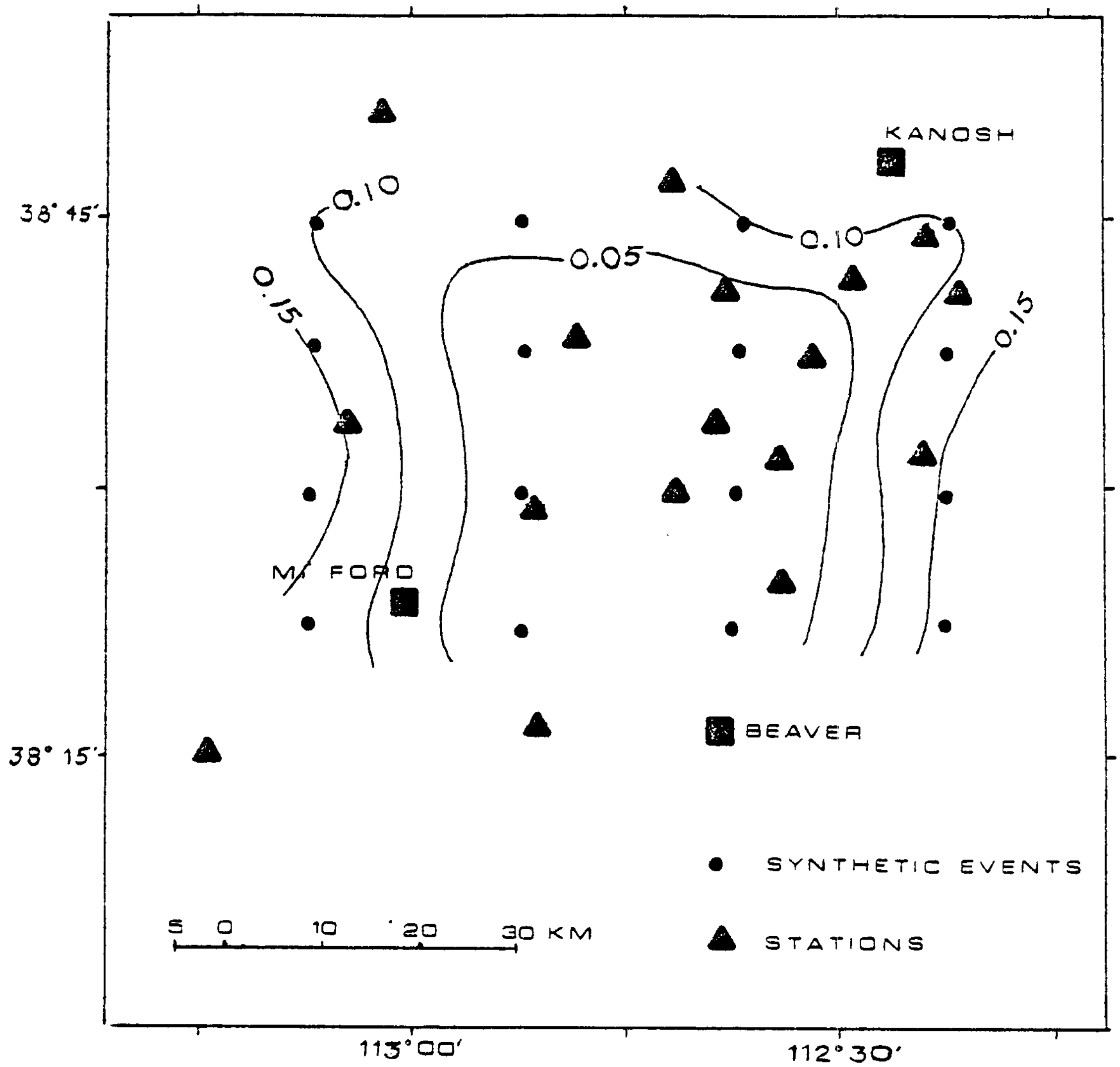


Figure 5.1. Locations of stations from microearthquake survey in the Roosevelt Hot Springs - Cove Fort area (Olson , 1976). Values assigned to gridpoints are magnitudes (km) of mislocation of synthetic events located using joint hypocenter determination. Contour interval is 0.05 km.

(Olson, 1976). Sixteen synthetic events were generated at each grid point in Figure 5.1, using representative subsets of the array. This means that each synthetic event had arrival times at stations that normally recorded eathquakes near that particular grid point. Contours of mislocation of these synthetic events are similar to those produced for the regional test cases in Chapter 3 (Figures 3.2 to 3.6). Figure 5.1 thus gives a generalized indication of the varying location capability of a local microearthquake survey, specifically the portable network deployed by Olson (1976).

Two velocity models were used for these relocations Stations located in the Milford Valley used a model derived from a reversed refraction profile across the Mineral Mountains (Gertson, 1979). Another model was applied to stations in and east of the Mineral Mountains. The latter model was a combination of Gertson's (1979) information and the velocity model for the Colorado Plateau described in Chapter 4. Table 5.1 gives the velocity and depth information for the two models.

Even with the good station distribution of the 1974-75 survey (average station-spacing of 10 km) about 50% of the events were not recorded by a sufficient number of stations to resolve the focal depths, particularly in the Milford region. Epicenters for the reference events were computed twice: once with depths fixed at Olson's (1976) hypocenters and once with depths unconstrained in the final two iterations. When depths were included the systems of equations were generally more unstable. In almost every case when the solution included the computation of the depth parameter, the

Table 5.1. Velocity models for local relocations: Cove Fort - Roosevelt Hot Springs

MILFORD VALLEY MODEL

P-WAVE VEL (Km/s)	DEPTH TO TOP (Km)	LAYER THICKNESS (Km)
1.8	0.0	0.3
4.0	0.3	1.2
6.7	1.5	23.9
7.4	25.4	-

MINERAL MOUNTAINS MODEL

P-WAVE VEL (Km/s)	DEPTH TO TOP (Km)	LAYER THICKNESS (Km)
2.9	0.0	0.2
3.2	0.2	0.3
4.0	0.5	1.1
5.7	1.6	12.7
6.4	14.3	11.7
7.4	26.0	-

JHD-computed focal depths were 1-3 km shallower than Olson's solutions, computed with HYPO71 (Lee and Lahr, 1975). A similar result was noted for synthetic data; i.e., the related program HYPOELLIPSE consistently located events 1-2 km deeper than the true solutions. It is believed, but has not been verified, that HYPOELLIPSE may have a consistent error in focal depth computations.

Since no other major differences were noted in the focal depths computed by HYPO71 and JHD77, and because of the inability to solve for 50% of the focal depths, only epicentral computations were carried out for the remaining (non-reference) events. The focal depths were fixed at those computed by HYPO71 and tabulated by Olson (1976). They are probably correct to within two kilometers, with the error biased towards the deeper values. No further examination of depth was made in the present study, but from Olson (1976) it appears that focal depths are variable throughout the Roosevelt Hot Springs area, occurring mostly within the range of a 0-12 km. A few may be as deep as 14-15 km. In the Cove Fort area the majority of events occur at focal depths less than 8 km.

Interpretation of Relocated Epicenters

Comparison of Olson's (1976) epicenters with the locations computed using the combination JED program and station adjustments in the accompanying single event program showed significant decreases in the sizes of areas covered by swarm activity for the JED events. Figure 5.2 shows the relocated earthquakes relative to concentrations of regional events located with the University of Utah permanent

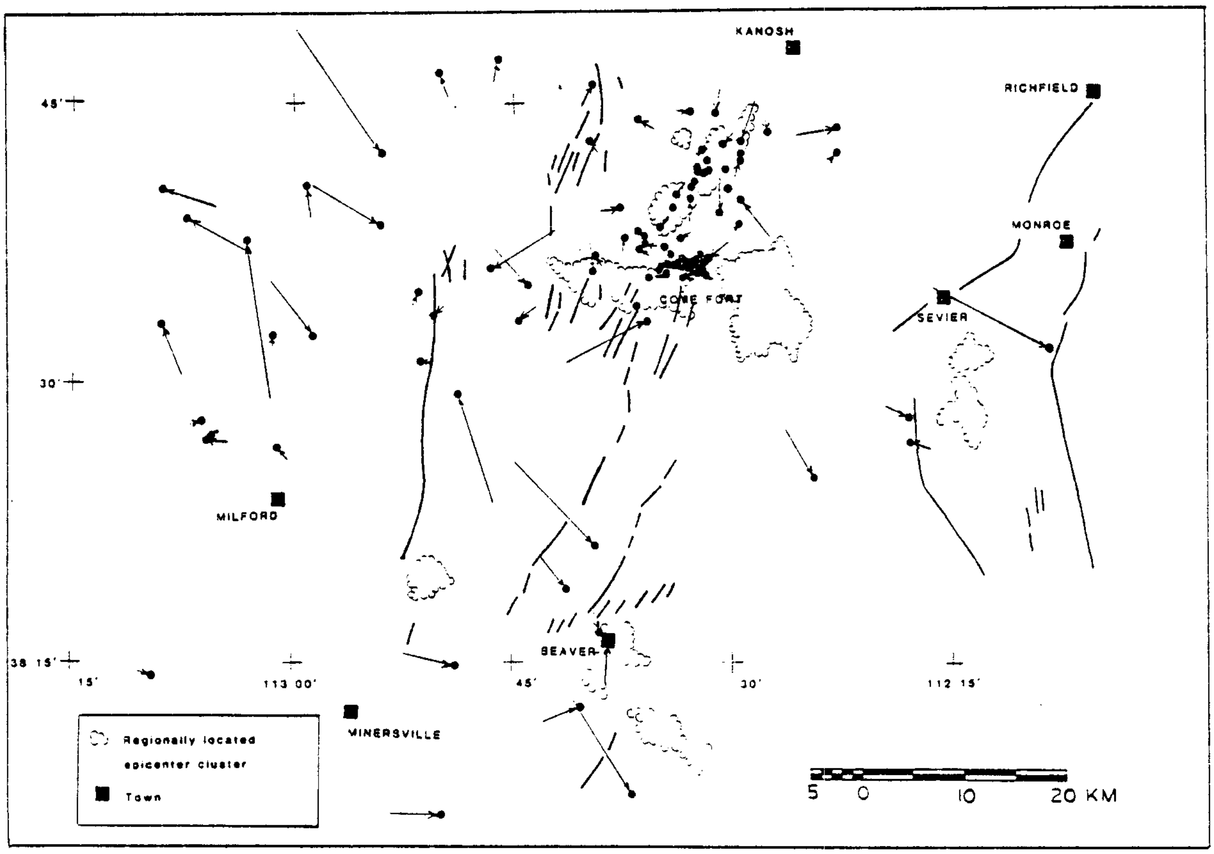


Figure 5.2. Relocated earthquakes in the Cove Fort - Milford area. Arrows point from former single-event locations of Olson (1976) to jointly determined locations (closed circles). Heavy lines are Quaternary faults. Irregular outlines are major concentrations of earthquakes located with the permanent regional seismograph stations from 1962-1978.

network. Arrows are drawn from the former HYPO71 locations to the epicenters computed by the JED program. No systematic shift in regional epicenters is obvious, but north of Milford the directions and magnitudes of some of the large shifts in epicenters (5-10 km) can be attributed in part to the station distribution. Semi-major axes of confidence ellipses are generally oriented northwest in this area, suggesting poorer resolution in that direction.

Some large shifts in locations (greater than ten kilometers) were observed for events with poor station coverage (four recording stations with no S-phase information). Out of an original 163 events, about 10% were eliminated from the relocation procedure because they were recorded by only 3 stations. Fifteen percent of the remaining epicenters showed the shifts of greater than 10 km. Such large differences in two solutions for the same event emphasizes the fact that even in a good microearthquake survey many events are not recorded by enough stations to provide adequate control for their location.

Comparison of the epicenters in Figure 5.2 with the clusters of regional epicenters (also in Figures 4.6, 4.7 and 4.8) shows some similarities in their distributions. About five kilometers east of Cove Fort is a cluster of events that was located with the regional stations. Almost all of this activity occurred in swarms throughout the first six months of 1977. A number of the earthquakes south and east of Cove Fort occurred about this same time (May to mid-June, 1977).

The elongated cluster of events east of Cove Fort in Olson's

1974-75 data set occurs between the two regionally-located clusters. The activity is mostly shallow (< 5 km), and Olson (1976) postulated that the earthquakes occurred along a north-trending fault. However, the cluster of relocated epicenters shows a definite east-west elongation, suggesting an east-trending orientation for the active earthquake zone. Several faults striking east-west along Cove Creek have been mapped by Moore and Samberg (1979). Most of these are concealed by alluvium, and some appear to be truncating, others intersected by, north-south faults that have mapped displacements in the Tertiary volcanics (Moore and Samberg, 1979). Thus at least some of the east-west faults are apparently younger than local Miocene tuff deposits. Based on this information, it is postulated that an active east-west fracture zone may occur along Cove Creek, possibly extending west beneath the alluvium of the Cove Fort graben.

Regional station distribution since 1977 has been fairly good in the Cove Fort - Milford area, including five stations within about 60 km of Cove Fort. However, these five stations (CFU, MNU, MSU, PUU and RHU) were not operating until 1977 (Richins, 1979); hence, none of the earthquakes detected by Olson's (1976) microearthquake array were detected by the regional network.

Considering the different station distributions and the different reference events used for the regional and local earthquakes, it is probable that there is a closer spatial association of the 1974-75 microearthquakes and the post-1977 earthquakes than is obvious from Figure 5.2.

Earthquakes north of Cove Fort define a northeast-southwest trend, more pronounced in the relocations. This swarm coincides with a northeast trending zone of regionally located earthquakes.

Based upon this study of relocated epicenters in the Roosevelt Hot Springs and Cove Fort areas, it is apparent that the more precise locations of earthquakes resulted in finer delineation of seismically active zones. In the Milford Valley there are probably several faults, the presence of which has been detected by local gravity investigations (Crebs, 1976). The occurrence of scattered earthquakes suggests that these may be more active at present than the more prominent faults bounding the Mineral Mountains horst (e.g., the Opal Mound fault). But sparse earthquake activity appears to be characteristic of the Roosevelt Hot Springs area. In contrast, the Cove Fort area shows earthquakes occurring on faults of two distinct orientations: one prominent east-west and several northeast-striking zones, which may reflect active faulting not mapped on the surface. (refer to Figures 5.2 and 4.7).

Velocity Models and P-Wave Delays

A common problem in determining microearthquake locations is inadequate knowledge of the velocity model. In this investigation of earthquakes in the Roosevelt and Cove Fort KGRA's all relocations were done with the two velocity models in Table 5.1, predominately derived from the refraction profiles of Gertson (1979). But Gertson's profiles and local gravity studies (Carter and Cook, 1978; Crebs, 1976) indicate that the flat-layer approximation is not a good model for detailed investigations of this complex area.

Further evidence suggesting lateral inhomogeneity in the velocity models exists in the form of P-wave arrival-time delays measured from refraction blasts. During the 1977 refraction experiment seven portable stations were arranged broadside to the refraction line (Gertson, 1978). Five seismograph stations from the permenanat network also recorded the blasts. Figure 5.3a shows the station locations and shot points. Blast sizes ranged from 1000 to 3000 pounds and produced generally clear, though not always impulsive, P-wave arrivals. These were timed to the nearest 0.02 seconds. A total of 13 S-wave arrivals were also timed but were not sufficient to pursue investigations of V_p/V_s ratios or S-wave attenuation. Three blasts were recorded from shot point 1, two from shot point 9, and one blast from shot points 3, 5, and 7.

Fifteen microearthquakes were also detected during the 2-week period. Initially the earthquakes were to be used to supplement the blast information, but it was determined that the error involved in their location would contribute more to the residuals than the deviations from the velocity models.

Travel-time residuals for P-wave arrivals at each station were computed based upon the flat layer velocity models (Table 5.1.). These residuals are plotted for each shot point in Figures 5.3b, 5.4, and 5.5. The shot point locations are in the center of each plot; each grid point is the mean of all P-wave residuals at that particular azimuth and distance. Positive elements indicate positive delays (late arrivals); negative values correspond to early arrivals.

A systematic examination of the residuals from each shot point a)

b)

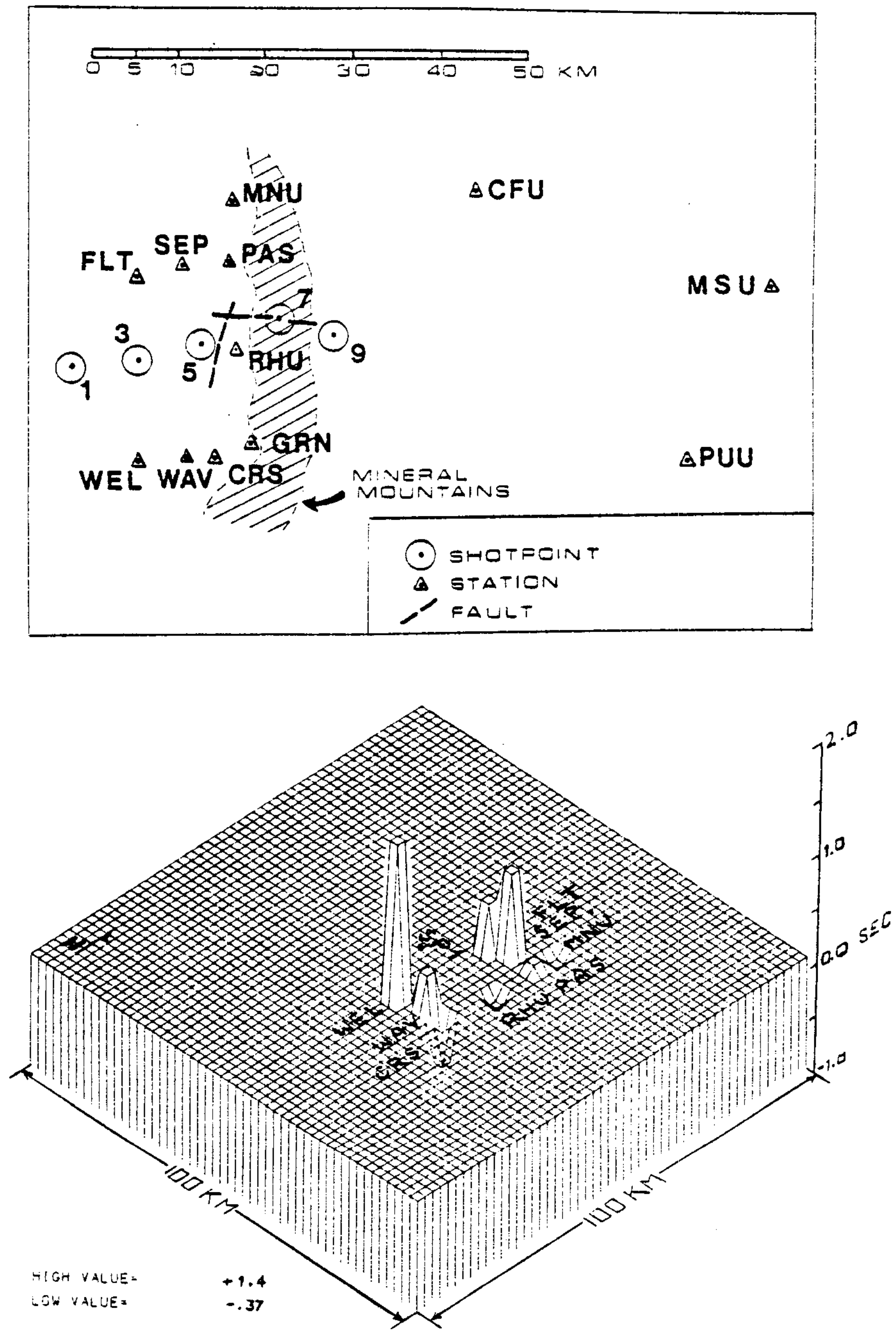
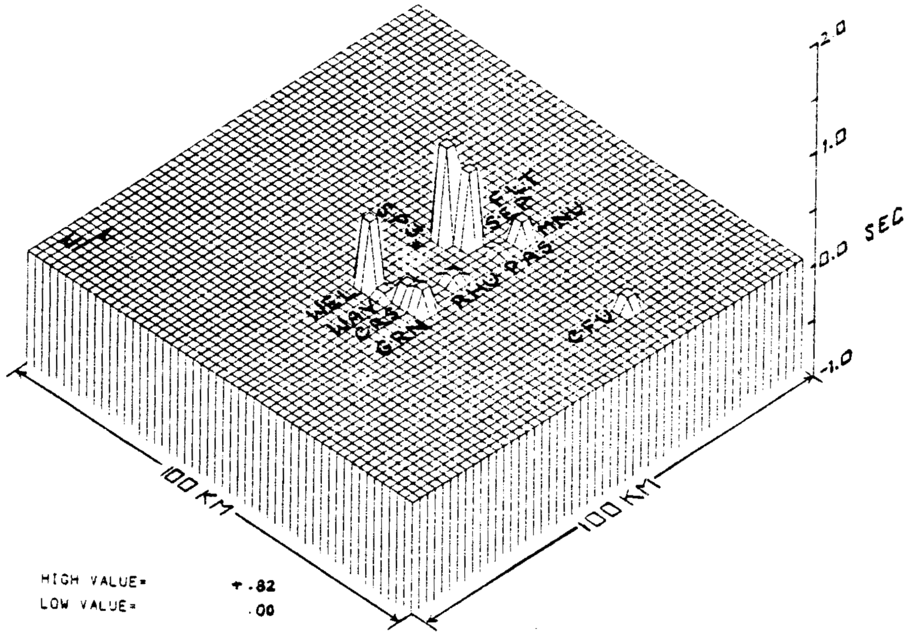


Figure 5.3. a) Location map of broadside refraction array and five permanent seismograph stations relative to shotpoints of 1977 refraction line (Gertson , 1979). b) 3-D plots of travel-time residuals (sec) with distance (km) from shotpoint 1.

a)



b)

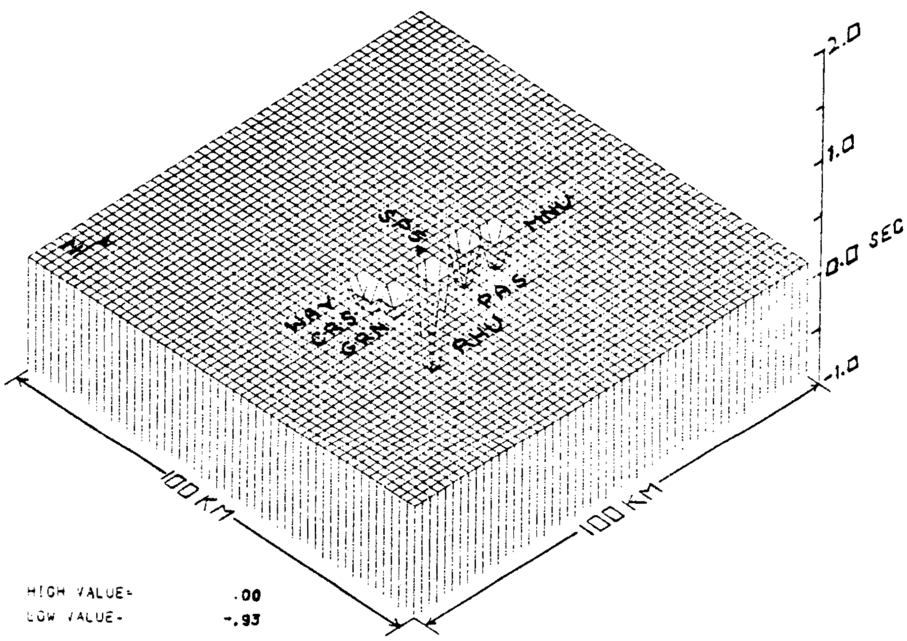


Figure 5.4. a) and b) 3-D plots of travel-time residuals (sec) with distance (km) from shotpoints 3 and 5, respectively.

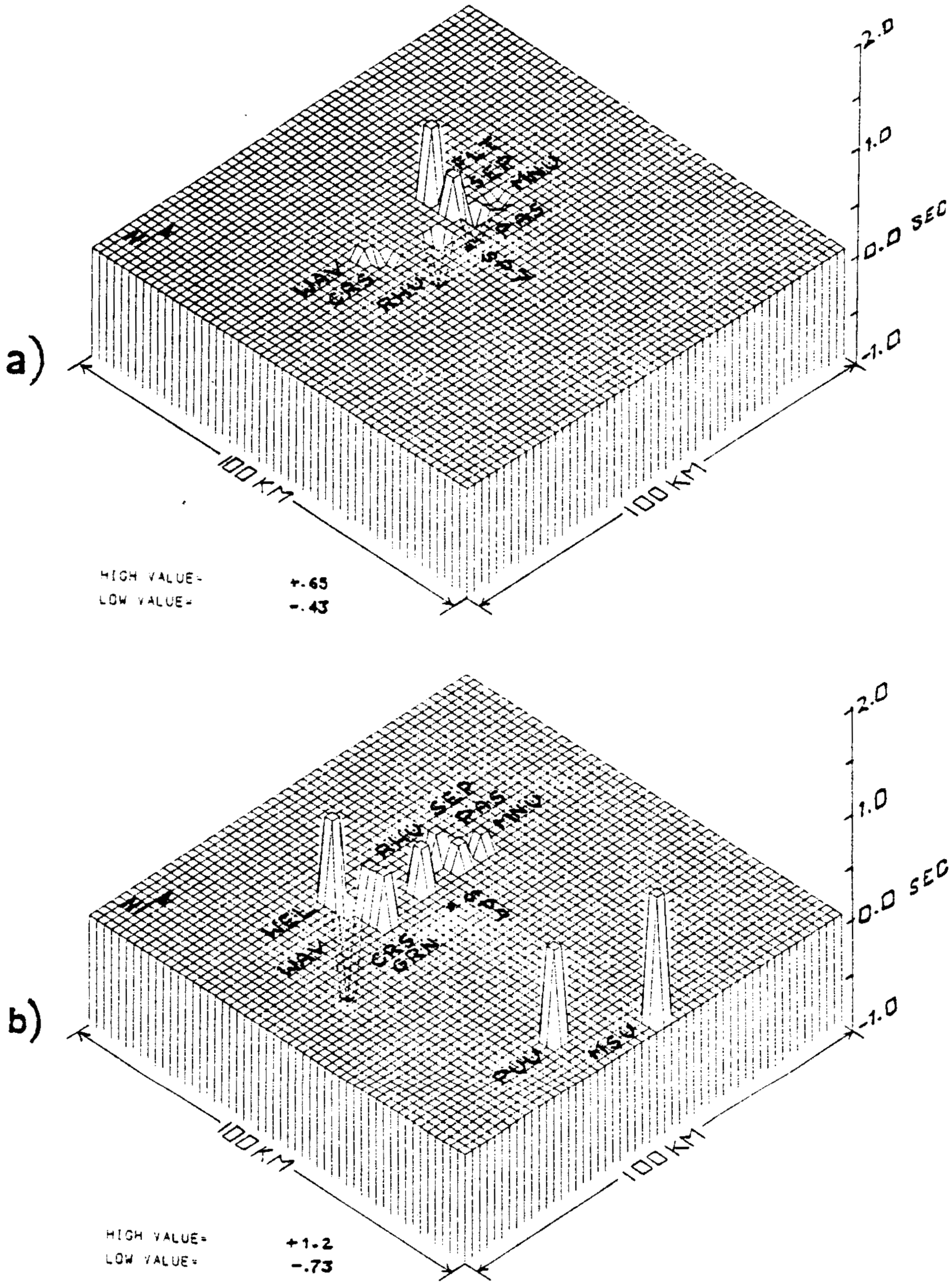


Figure 5.5. a) and b) 3-D plots of travel-time residuals (sec) with distance (km) from shotpoints 7 and 9, respectively.

reveals that most can be attributed to velocity variations in the upper three km of crust below sources and receivers. Stations FLT, SEP, and WEL used the Milford Valley model. All arrivals at these stations were late, with particularly large delays (0.8 to 1.4 sec) recorded for shot points 1 and 3. These delays are observed because the model used was an approximation of the velocity section on the eastern margin of the valley, whereas the depth of the alluvium increases from about 0.3 m to 1.0 km in the center of the valley (Gertson, 1978). Stations recording blasts from shot points 1 and 3 at distances of over 20 km generally show small (less than 0.3 sec) positive and negative residuals. Arrivals at this distance are probably equivalent to P_g, that is, refracted arrivals from the Precambrian crystalline basement, probably banded gneiss mapped in the Mineral Mountains area by Nielson et al. (1978). Gneiss may typically have velocities of about 4.5 to 6.0 km/s (Press, 1966). The value of 4.0 km/s probably represents an adequate average velocity along the ray path. Relative to the velocity model the basement velocities may be higher, but lower velocities would be in effect in the thicker alluvium. Faulting within the Milford Valley and possible velocity anisotropy in the gneiss could account for other small variations (up to 0.2 sec) in the residuals.

Arrivals from shot point 5 are all early (Figure 5.4b), especially at the closest station, RHU. Ray paths are approximately parallel to the Mineral Range front, and the proximity to the Opal Mound fault suggests that they may be traveling through cemented alluvium (Parry et al., 1977), that may account for the higher velocities.

Residuals from the blast at SP7 are generally small (absolute values less than 0.3 sec) and can be attributed to lateral variations such as the presence of faulting, hydrothermal alteration, and dipping layers. Residuals at Stations FLT and SEP is again positive, probably due to inadequate modeling of the thickness of aluvial fill under these stations. The negative RHU residual can be attributed to rays traveling through more granitic material than was modeled for this station.

Except arrivals at stations PUU and MSU in the Cove Fort graben, rays from shot point 9 all traveled through the Mineral Mountains. Large positive delays at PUU and MSU can probably be attributed to the presence of alluvial fill and Pennsylvanian to Permian sediments and Tertiary ash flow tuffs. Arrivals at all other stations except WAV were also late. P-wave arrivals from SP9 were impulsive, including WAV. But one arrival at WAV had a reversed sense of motion (down) while other arrivals were all normal (up). This could indicate that the first arrival was mis-identified because all arrivals from an explosive source should have the same direction of first motion. The polarity was not reversed at station WAV, because arrivals from other shots on the same day were consistent.

A teleseismic P-delay study was recently completed by the USGS for the Roosevelt and Cove Fort region (Robinson and Iyer, 1979). Their main result was the detection of an elongated zone (parallel to the Mineral Mountains range) of low velocity material extending from 5 to 35 km below the Roosevelt Hot Springs. The region ws interpreted as having a maximum velocity decrease of 5 to 7%. This could be associated with a density decrease of about 0.15 g/cm^3. A gravity low of 15-20 mgals (Carter and Cook, 1978) corresponds to this general region (Robinson and Iyer, 1979).

The positive P-delays from SP9 may indicate that rays travelled through the upper part of the same low velocity region detected by the teleseismic ray paths. Robinson and Iyer (1979) attributed the low velocities to a pipe of partially molten material. But the residuals (0.5 - 1.2 sec in the refraction data; 0.1 - 0.3 sec in the teleseismic data) could be associated with the fractured nature and possibly fluid-filled porosity of the western edge of the Mineral Mountains pluton. They could also be due to a compositional change from granite to gneiss. Mean densities for these rocks in the Mineral Mountains area are 2.59 gm/cc and 2.69 gm/cc, respectively (Carter and Cook, 1978). This density contrast of 0.10 gm/cc may be sufficient to produce the observed velocity decrease.

Evidence from the P-delays of the broadside array thus suggest several inadequacies in the simple flat-layered models used to locate in the microearthquake location. Inadequate modeling of the depth of alluvium at some stations, high velocity zones associated with cementation and hydrothermal alteration, low velocity zones associated with fracture systems and lithologic contrasts between granite and gneiss, all contribute to the complexity of the velocity model.

The question becomes how much difference lateral velocity variations make on the hypocenter locations. Based upon tests using Olson's (1976) flat-layered model for all stations and the two models determined from refraction, for single-event locations discrepancies in the two solutions range from 1 to 15 km, with an average difference of 5 km. However, as was previously demonstrated for the test cases in Chapter 3, the JHD (or JED) technique minimizes this difficulty through systematic adjustment of the velocity models through the station corrections.

Based on interpretation of epicenter relocations using the joint hypocenter determination method and on application to regional implications of seismic activity in geothermal systems, several useful results have been obtained. These emphasize the value of joint relocation methods. Analyses of relationships between regional and local seismicity the Roosevelt and Cove Fort KGRA's brought out some practical results of precise and accurate earthquake location. To summarize, tectonic interpretations can be better clarified through the use of precise earthquake location techniques. Models for fracture-dominated geothermal systems rely upon tectonic information which can be easily obtained from regional surveys. But it is clear from the differing characteristics of earthquake activity in the Cove Fort and Roosevelt KGRA's, and also the Monroe geothermal area, that a simple governing model that can be universally applied to help locate geothermal prospects has not been identified. Once the location of a geothermal area has been verified, precise location of microearthquakes can provide low-cost information about the reservoir and/or crustal properties.

microearthquakes can provide low-cost information about the reservoir and/or crustal properties.

PLATE I. Comparison of Hypocenter Location Techniques and Programs

APPLICATION / PROBLEMS	HYPOCENTER LOCATION METHODS: Single Event		Master Event	Joint Determination	
	HYPOELL.	SE77-Single	SE77-Master	JHD77	GHYP1
Routine, Preliminary Event Location	Advantage: Straightforward; ease of data input		Disadvantage: Can be used only if well-located event(s) already exist. Advantage: If known local velocity effects exist at some stations.		
Patterns or Trends in Epicenters	Disadvantage: Relative location accuracy unknown		Advantage: Good relative locations obtained		
Correlation with Known Geology or Geophysics	Disadvantage: Relative location accuracy unknown		Advantage: Resolution of trends and clusters enhanced.		
Velocity Studies:					
a) velocity structure unknown	Disadvantage: Not generally applicable without velocity model		Advantage: Can locate with only preliminary estimate of velocity model		Advantage: can invert for P and S velocities in half-space
b) velocity inhomog. at spec. stations	Disadvantage: Events will be systematically mislocated.		Advantage: Velocities partially corrected by estimates of station adjustments Disadvantage: Adjustments relative to one event	Advantage: adjustments estimated simultaneously relative to: 15 events	10 events

PLATE I. Comparison of Hypocenter Location Techniques and Programs (Cont.)

APPLICATION / PROBLEMS	HYPOCENTER LOCATION METHODS: Single Event — HYPOELL.	Single Event — SE77S	Master Event — SE77M	Joint Determination — JHD77	Joint Determination — GHYP1
c) P or S delay studies	Disadvantage: Station residuals include effects due to structure at source			Advantage: Station adjustments not as much a function of source structure. Residuals indicate source structure	
Control Event			Disadvantage: Master event bias; all stations used must record	Disadvantage: Reference event bias Advantage: Bias not as pronounced; well-recorded, but not necessarily at all stations	
Station Array: a) poor azimuth control	Disadvantage: Any location will solve system			Advantage: Group location better: possible smoothing of location errors	
b) poor distance control	Disadvantage: Velocity model no longer valid		Advantage: Partial correction for inadequate velocity model		Disadvantage: Possible sacrifice good events to locate poor events
Calibration of Regional Stations			Advantage: Use of events with locationgs determined by a local (microearthquake) array. Fix these locations and compute a set of adjustments for regional stations		Advantage: Adjustments can be computed simultaneously

PLATE I. Comparison of Hypocenter Location Techniques and Programs (Cont.)

APPLICATION / PROBLEMS	HYPOCENTER LOCATION METHODS: Single Event — HYPOELL.	SE77S	Master Event — SE77M	Joint Determination — JHD77	GHYP1
Travel Time Bias Included in Variance Calculations	Disadvantage: 1) Incorrect weights applied, leading to incorrect solutions; 2) adverse effect on size and eccentricity of error ellipses			Advantage: Correction of travel-time bias; not included in variance terms	
Inclusion of Events Not Subject to Similar Biases	-			Disadvantage: May affect entire group of locations	
Results for Specific Programs:					
1) Termination of Iteration	Size of last step in parameter space	Number of iterations			Number of iterations; convergence based on size of last step in parameter space
2) Statistical Results	RMS (SE). SE Ellipse; Means of station residuals; Progression of parameter adjustments	Standard error (SE). Confidence ellipse. Station adjustments with variances; means of station residuals; optional: progression of parameter adjustments			

PLATE I. Comparison of Hypocenter Location Techniques and Programs (Cont.)

APPLICATION / PROBLEMS	HYPOCENTER LOCATION METHODS: Single Event		Master Event	Joint Determination	
	HYPOELL.	SE77S	SE77M	JHD77	GHYP1
3) Weighting	Form of Jeffrey's weighting; others	Form of Jeffrey's weighting. No distance weighting. User determines initial weights and variances			Pearson Type VII

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