

Maximizing Earthquake Hypocenter Location Accuracies in EGS Stimulations

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ABSTRACT

Uncertainties in local models of seismic wave speed cause two kinds of errors in estimated microearthquake locations. Errors that differ between nearby events, and which degrade the resolution of faults and similar structures, can now be greatly reduced by relative hypocenter-location methods. Errors in the absolute locations of events, on the other hand, can be reduced using three-dimensional tomographic models of the wave speeds, but wave-propagation effects inherently limit the accuracy of tomographic models and therefore the effectiveness of this tactic in contexts such as EGS projects.

In typical EGS experiments, fluid is injected one to three km below the surface and monitored by seismometer networks a few kilometers in diameter. Errors in crustal structure produce absolute hypocenter-location errors of the order of a few hundred m. Such errors may preclude use of earthquake locations as targets for production wells. For that purpose, hypocentral accuracies of the order of a few meters are desirable. Adequate accuracies cannot normally be achieved simply by improving the crustal model.

A better approach is to measure directly the travel times between the monitoring stations and the stimulation zone, by either (a) firing a shot in an EGS well at the stimulation depth, or (b) deploying a seismometer in an EGS well at the stimulation depth, and firing a shot at each monitoring station. To our knowledge, no such an experiment has yet been performed. Approach (a), firing a shot down-well, risks damaging the well, and furthermore requires explosives that are stable at high temperature. Approach (b) requires sensors and cables that can withstand high temperatures. Nevertheless, the heat tolerance of sensors is improving, and such experiments may now be possible in some cases.

We model the likely improvements in earthquake locations that these techniques could provide. We simulate a cluster of earthquakes induced by a theoretical EGS injection, and generate synthetic arrival times using a typical volcanic crustal model including a stochastic component. We locate the earthquakes using both an average one-dimensional crustal model and a three-dimensional model of the kind that local-earthquake tomography produces. Finally, we simulate ray-path travel time corrections of the kind expected from a calibration experiment, and compare the locations obtained using those corrections with those obtained using only the three-dimensional tomographic model.

1. INTRODUCTION

The locations of microearthquake hypocenters, as inferred from the arrival times of seismic waves, provide valuable information about processes within geothermal systems. Relative hypocenter locations now have precisions of a few meters (Waldhauser & Ellsworth, 2000), and yield detailed images of the sizes, shapes, and orientations of zones of failure. Unfortunately, *absolute* hypocenter locations cannot be determined with comparable accuracy, placing severe limitations on the use of hypocenter locations for purposes such as guiding drilling in EGS projects.

The fundamental cause of our inability to locate hypocenters accurately is our ignorance about crustal wave-speed structure. Arrival times of seismic waves at short distances can be measured to within a few milliseconds, which would correspond to hypocentral uncertainties of about ten meters if the wave-speed structure were known accurately. In fact, however, seismic wave speeds vary from place to place by 10% or more, effectively introducing uncertainties of tens of milliseconds into arrival times, and of hundreds of meters into inferred hypocenter locations.

In this paper, we investigate hypocentral-location errors by numerical experimentation. We generate theoretical seismic arrival-time data for a suite of realistic cases, including examples based upon both one-dimensional and three-dimensional compressional-wave (*P*-phase) models, and simulate the errors in location that these data predict. Finally, we propose a practical method for reducing hypocenter location errors to approach the theoretical limit imposed by observational timing accuracy. This method involves (a) directly measuring travel times from a potential induced-earthquake hypocentral region to a network of seismometers, (b) calculating the travel-time delays that are unaccounted for by the crustal model, and (c) applying these delays to the real arrival-time measurements used to locate the induced earthquakes.

2. NUMERICAL EXPERIMENTS WITH SYNTHETIC SEISMIC DATA

To investigate the likely effects of crustal heterogeneity, we compute theoretical seismic-wave arrival times on the basis of assumed hypocentral locations and wave-speed models, and use these data to compute hypocentral locations using different models. Comparisons of computed hypocenters with the hypothetical (true) ones upon which the data are based illustrate the types and magnitudes of errors that are expected from using different (wrong), but typical, models.

To make our simulations realistic, we use the geometry of a high-quality existing seismometer network, operated by the U.S. Navy at the Coso geothermal area in southeastern California. The network contains 13 seismometers, some on the surface and some in shallow boreholes, and has an aperture of about 10 km (Figure 1). Station elevations vary by about 0.5 km. We use true elevations in our computations.

In each simulation, we place artificial earthquakes at pseudo-random positions on a rectangular fault plane having a specified location, size, shape, and orientation. For each event, we then compute travel times to seismometers using a hypothetical crustal model. We use these times to determine the hypocenter locations using the conventional iterative Newton-Geiger least-squares method (*e.g.* Stein & Wysession, 2003, Section 7.2) and a crustal model that differs from the assumed one.

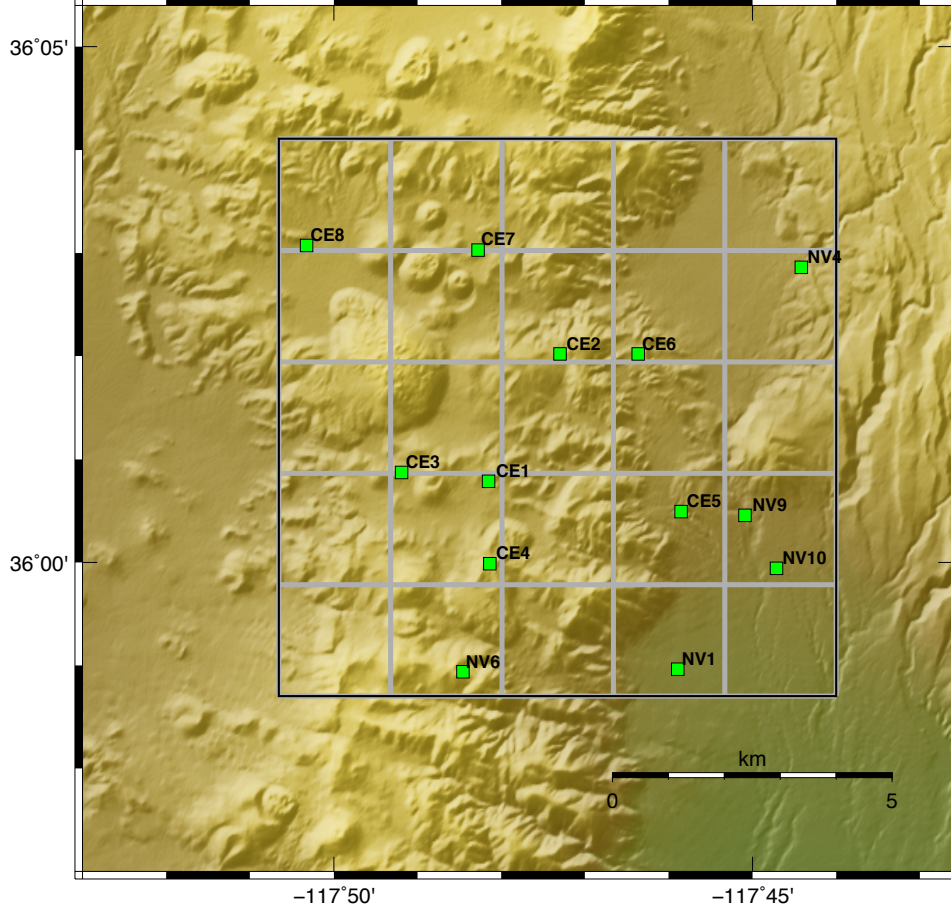


Figure 1: Map of the Coso geothermal area, California. Green squares: The thirteen stations of the U. S. Navy’s permanent seismometer network that are used as the basis for numerical simulations in this paper. Three-dimensional crustal models used in this paper are specified in terms of wave speeds at the nodes of the grid shown in gray, which has nodes spaced by 2 km horizontally and 1 km vertically.

2.1 One-dimensional models

Although the internal structure of the Earth varies in all three directions, it is common, because of lack of better information, to use one-dimensional models in seismological computations. Figure 2 shows two such models that assume homogeneous plane layers, for the Coso area. The first is a model used by the U.S. Navy in routine seismological monitoring. The second is a model derived by inverting a large set of earthquake arrival-time data using the computer program VELEST (Kissling, 1995), which calculates the best-fit crustal model for the entire data set. The upper 3 km of these models (extending from the general surface level at ~ 1 km above sea level, down to about 2 km below sea level) are similar. Below this depth, the models differ significantly, with the VELEST model having compressional-wave speed V_p about 4% higher than the Navy model. We use these models because the difference between them is likely to be fairly typical of the uncertainties in the models used in projects such as this.

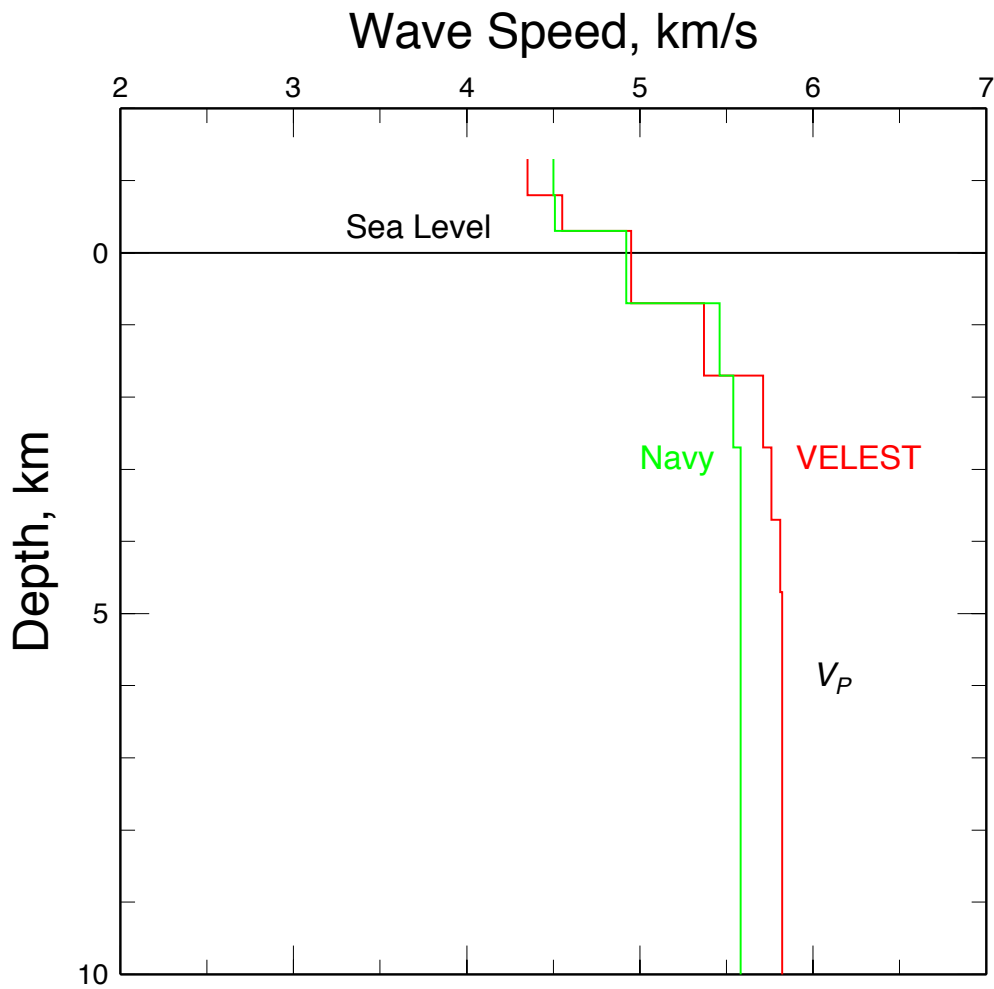


Figure 2: One-dimensional models of seismic wave speeds in the crust at the Coso geothermal area. Green: Model used by the U. S. Navy in routine microearthquake monitoring; red: A revised model derived by inverting a large set of seismic-wave arrival times. The surface at Coso is about 1.3 km above sea level.

Using the "VELEST" model, we generate seismic-wave arrival times for a set of theoretical hypocenters located at pseudo-random positions on a 200 m x 200 m square fault striking north and dipping at 30° to the east. We then re-calculate the locations of the hypocenters using the same arrival times but replacing the "VELEST" crustal model with the U.S. Navy's standard model (Figure 2). The re-computed hypocenters are systematically about 100 m too deep (Figure 3).

In reality, earthquake location program output would provide no clues to the magnitude of such variations in location. By moving the locations, the RMS residual of the arrival-time misfits is minimized. Indeed, this principle underpins the working of conventional earthquake location programs. Examination of the residuals gives an overly optimistic impression of the quality of the location (Figure 4). In this example, the small residuals show that the data are well fit by the hypocenters calculated, and give no indication of the approximately 100-m systematic offset from the "true" locations.

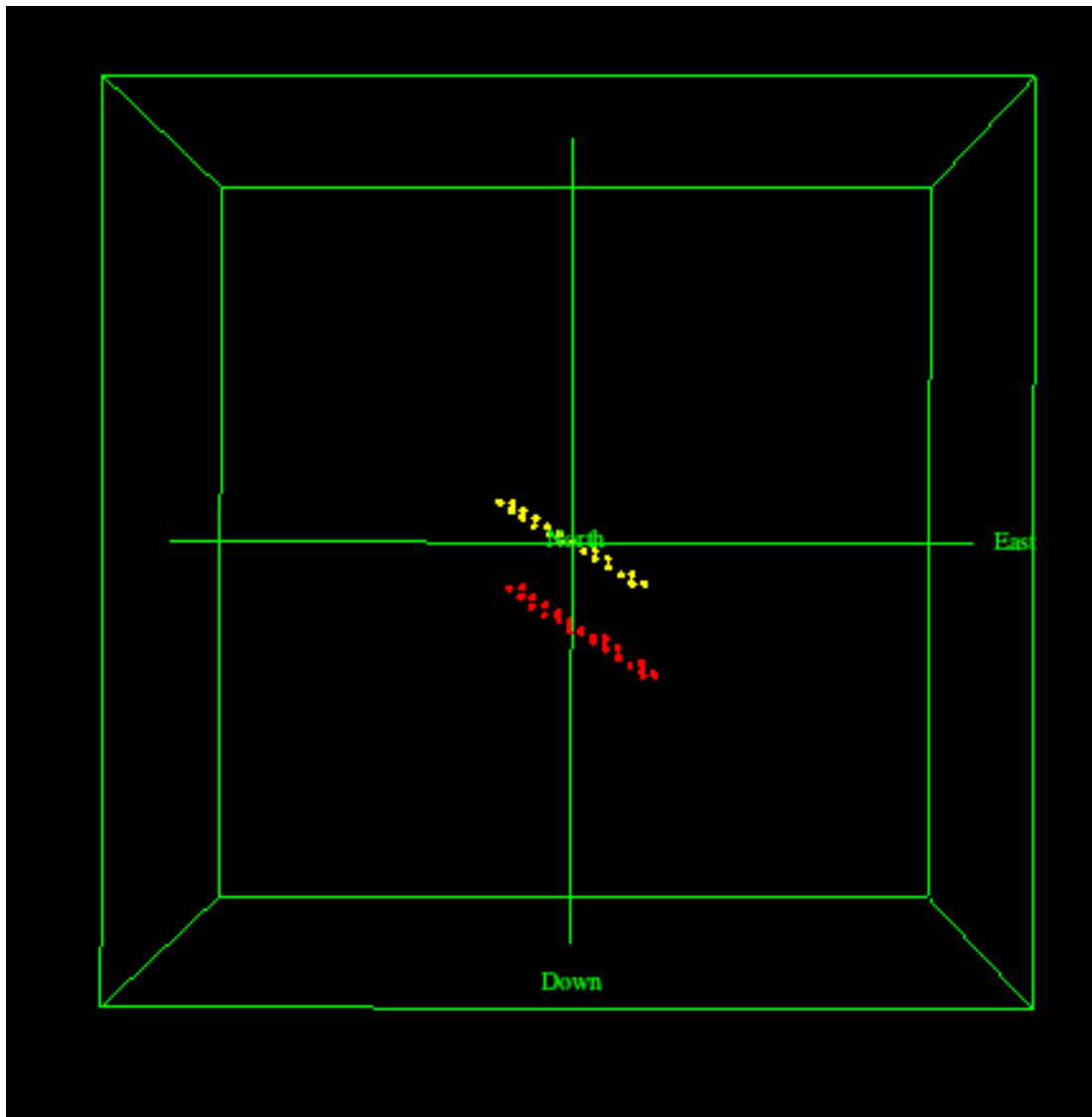


Figure 3: Perspective view of hypothetical and computed hypocenters. The assumed hypocenters (yellow) are placed at pseudo-random positions on a 200 m by 200 m square fault on a plane striking to the north and dipping 30° to the east. The computed hypocenters (red) are determined using standard seismological methods to invert theoretical arrival times at the 13 stations shown in Figure 1. These theoretical times are computed using the assumed locations and the "VELEST" model of Figure 2, but are inverted using the "Navy" model of Figure 2. The computed hypocenters are about 100 m too deep. The green cube is 1 km on a side. The line of view is horizontal and directed northward.

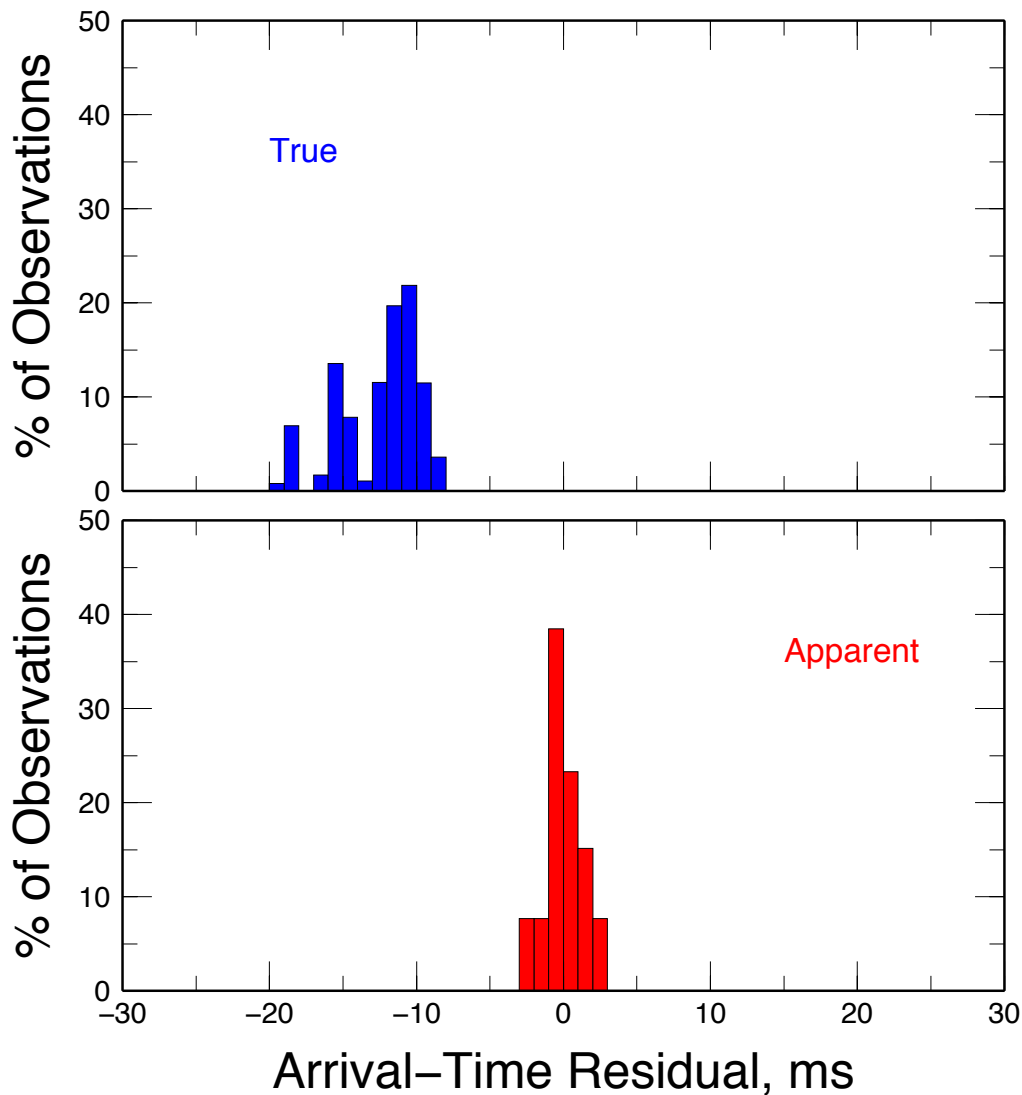


Figure 4: Histograms of P -phase arrival-time residuals computed for 100 hypothetical events on a 200 m x 200 m square fault centered 4 km below seal level near the center of the seismometer network shown in Figure 1. Arrival times are computed using an assumed hypocenter and origin time and the “VELEST” crustal model (Figure 2). The “true” residuals (top) are computed using the same assumed hypocenter and origin time and the model used by the U.S. Navy for routine seismological monitoring (Figure 2). In reality the true residuals would be unknown. Instead, the much smaller “apparent” residuals (bottom) would be obtained after locating the hypocenter using standard methods and the assumed model. The new hypocenter would be significantly different from the original (“true”) one.

2.2 Three-dimensional models

To accurately simulate the problems of hypocenter location, we must use a three-dimensional crustal model. To do this, we generate pseudo-random three-dimensional models, an example of which is shown in Figure 5.

This model consists of two additive two components. The first is a one-dimensional (depth-dependent) “deterministic” component and the second is a three-dimensional pseudo-random component. For the one-dimensional component we use the “Navy” model (Figure 2). For the three-dimensional component we specify values at the nodes of a three-dimensional grid, with node spacing of 2 km horizontally and 1 km vertically (Figure 1). The nodal values are pseudo-random multivariate normal deviates, generated using the algorithm given in Section 7.4 of Press *et al.* (2007) and have a correlation function that decays exponentially with distance.

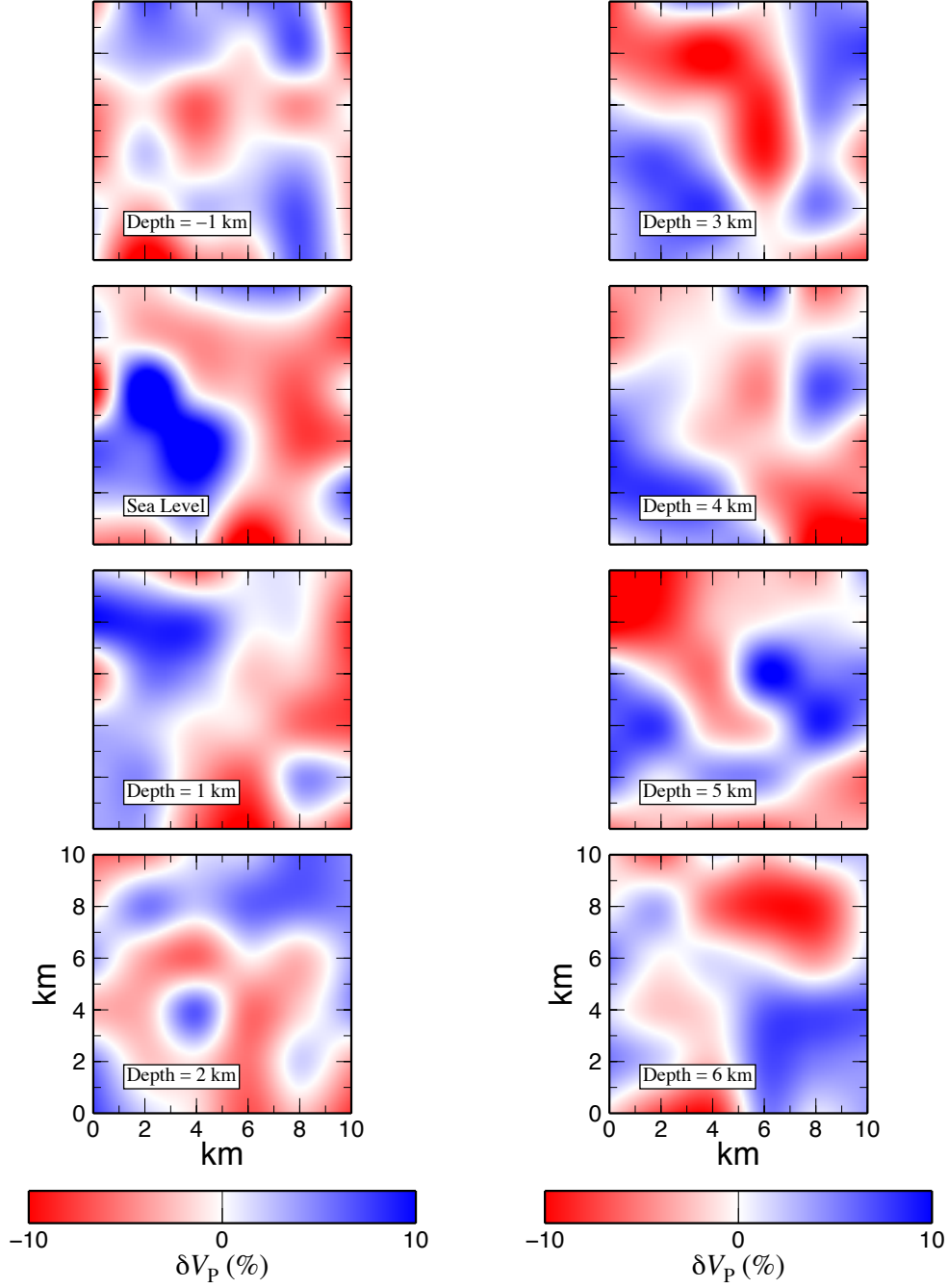


Figure 5: An example of a pseudo-random three-dimensional model of compressional-wave speed V_p . Each square shows the wave speed at a particular depth, and corresponds to the square in Figure 1. Wave speeds at the grid nodes have pseudo-random values with a covariance that decreases exponentially with distance, as described in the text.

Figure 6 shows histograms of arrival-time residuals similar those shown in Figure 4 for the one-dimensional crustal-model case. In constructing Figure 6, however, the three-dimensional model shown in Figure 5 is used to generate the original arrival times. The residuals calculated using the "Navy" model, and holding the original locations fixed, are much larger in absolute value and have a much larger spread (Figure 6, upper panel). Even allowing the hypocenter to move to its optimum position does not reduce the residuals as much as for the one-dimensional case. The spread of the residuals remains large (Figure 6, lower panel).

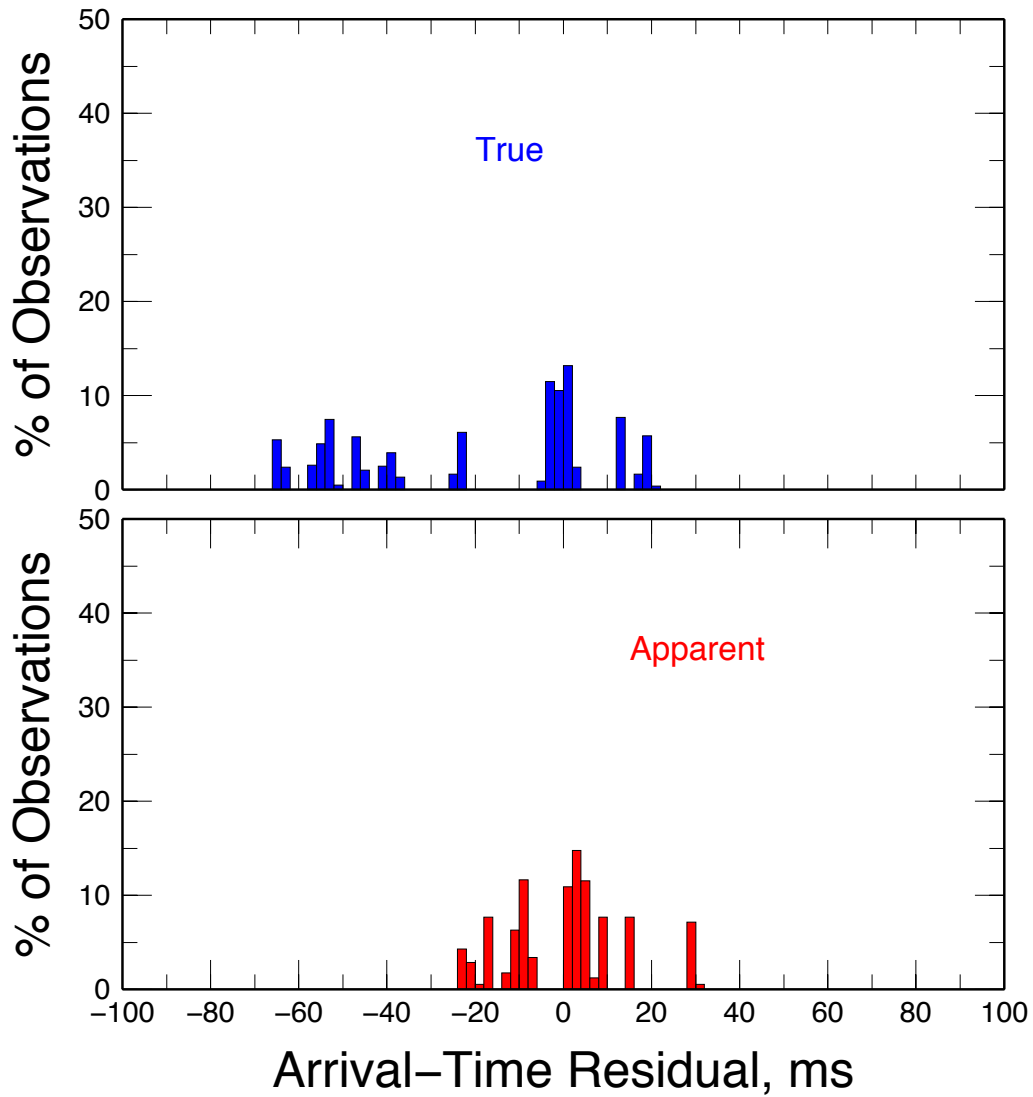


Figure 6: Histograms of theoretical arrival-time residuals, computed for the same hypothetical hypocenter locations as used for Figure 3, but using travel times for the pseudo-random three-dimensional model shown in Figure 5. Note the different scale used for the abscissa compared with Figure 4.

The hypocentral mis-locations are also much larger than in the one-dimensional case. The computed hypocenters are located about 300 m too far west, slightly too far south, and up to 100 m too deep, compared with the “true” locations (Figure 7). Furthermore, the entire fault plane is distorted. The size of the inferred fault is exaggerated, its dip is increased, and it has significant curvature. The details of such distortions will depend on the nature of the true three-dimensionality of the crust. This cannot be known except by performing experiments to determine the three-dimensionality in the volume of interest, and even where detailed results of this sort are available, they are no better than the scale of nodal spacing of the grid used. In the case of the present example, this was 2 km horizontally and 1 km vertically. The effects of heterogeneities on a smaller scale than this will thus remain unknown as will the mis-locations and structural distortions they produce.

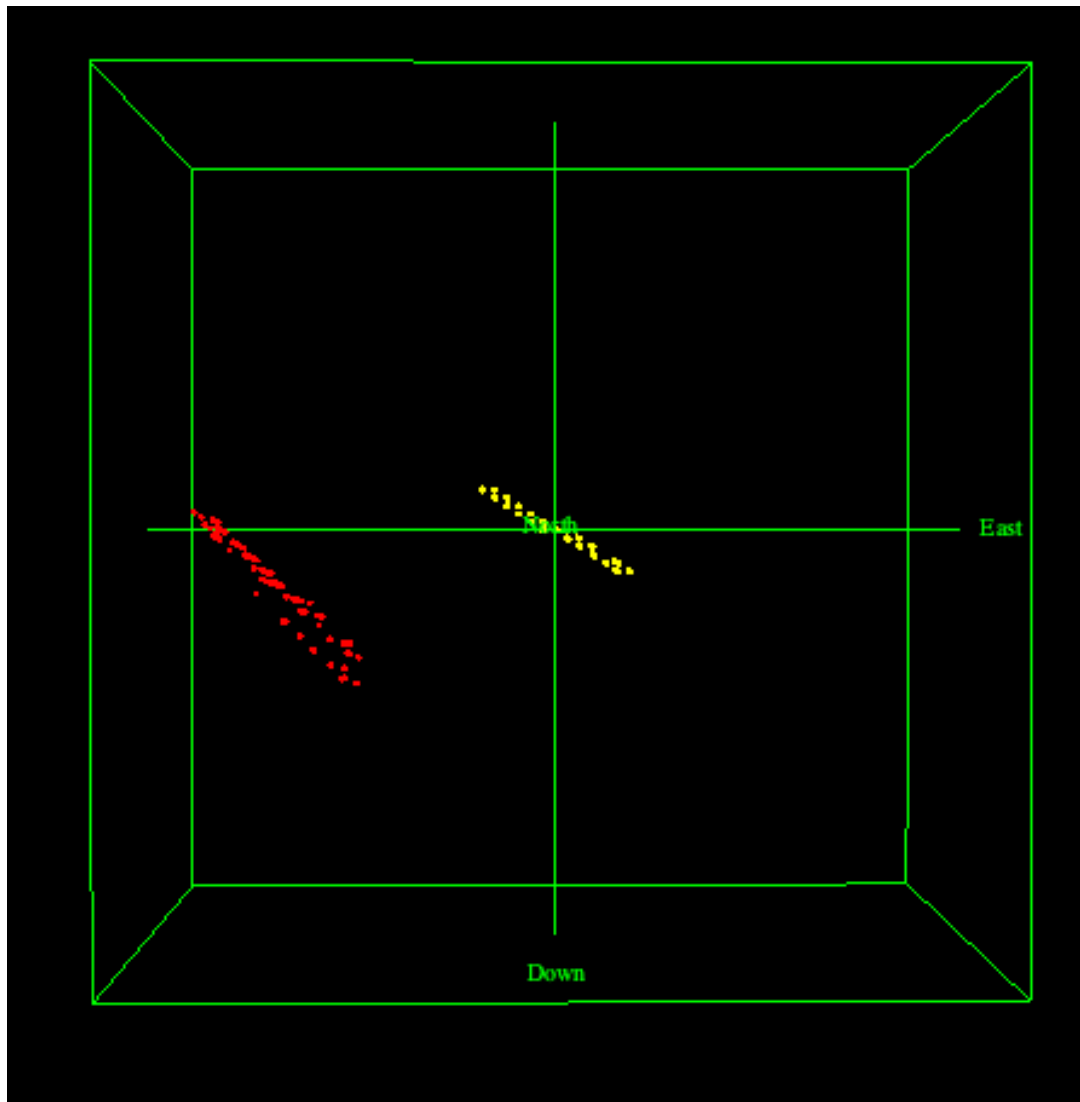


Figure 7: Perspective view of hypothetical and computed hypocenters based upon the three-dimensional hypothetical crustal model shown in Figure 5. The conventions are the same as in Figure 4. The events are mis-located by about 300 m and the orientation, size, and shape of the fault are distorted.

3. A DIFFERENT APPROACH – CREATING AN ARTIFICIAL "MASTER EVENT"

It is not currently possible, as a practical matter, to know Earth structure in sufficient detail to reduce the resultant location uncertainties to a level comparable to errors introduced by effects such as errors in arrival-time measurements and station locations. A different approach is thus required. One such approach is to perform a suitable calibration experiment to obtain explicit corrections for the ray-paths of interest. This approach amounts, conceptually, to creating an artificial "master event" whose location is known almost perfectly. The experimental setup is illustrated schematically in Figure 8.

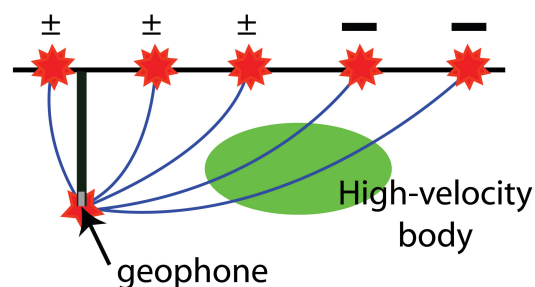


Figure 8: Schematic figure showing a calibration experiment that could provide the information necessary to greatly reduce hypocentral location errors arising from unknown crustal structure. See text for details.

The experiment comprises deploying a geophone in the EGS stimulation well at the depth where fluid injection is planned. Timed explosions are then detonated at each seismometer station of the monitoring network, and recorded on the geophone. Measurements can thus be made of the travel times of seismic waves between the surface explosions (the seismometer stations) and the anticipated hypocentral region of earthquakes generated in the forthcoming hydrofracturing.

The actual travel time measurements are then compared with theoretical travel times for the same ray paths calculated using the location of the geophone and the crustal model that will be used to locate the future induced earthquakes. The differences between the measured and calculated travel times then comprise corrections, applicable to those particular ray paths and the particular crustal model. By applying these corrections to the arrival times measured for the future earthquakes, the effect of unknown crustal structure heterogeneity is effectively removed.

An alternative approach would be to fire a shot in the EGS well at the planned stimulation depth, and to record it on the seismic stations already in place. That approach might be precluded because of the risk of damaging the well and also faces the challenge of finding explosive material that is stable at the high temperatures deep in EGS wells. In deploying a geophone in the well, the problem must be overcome of finding a sensor and cables that have sufficiently high temperature tolerance. Both approaches thus have their challenges, but both yield the same information because of reciprocity.

3.1 Numerical experiment to illustrate the efficacy of the approach

We simulate data to explore the efficacy of the above experiment. We assume a geophone placed in an EGS stimulation well at a depth of 4 km, and well-recorded explosions fired at the 13 stations of the Coso network. We calculate the travel times of seismic waves through the three-dimensional structure shown in Figure 5, to the geophone in the well. We then compare these travel times with travel times calculated for an event at the same location as the geophone assuming the "Navy" one-dimensional crustal model of Figure 2. Differencing these two sets of travel times yields our simulated corrections.

We then assume a set of earthquakes occurring on the same 30°-dipping fault as shown in Figures 3 and 7. The errors in location, when using the one-dimensional crustal model only, are shown in Figure 7. We relocated the events, again using the same one-dimensional model, but this time applying also the simulated corrections. The resulting arrival-time residuals (Figure 9) are reduced essentially to zero, and the locations (Figure 10) (describe, blah blah).

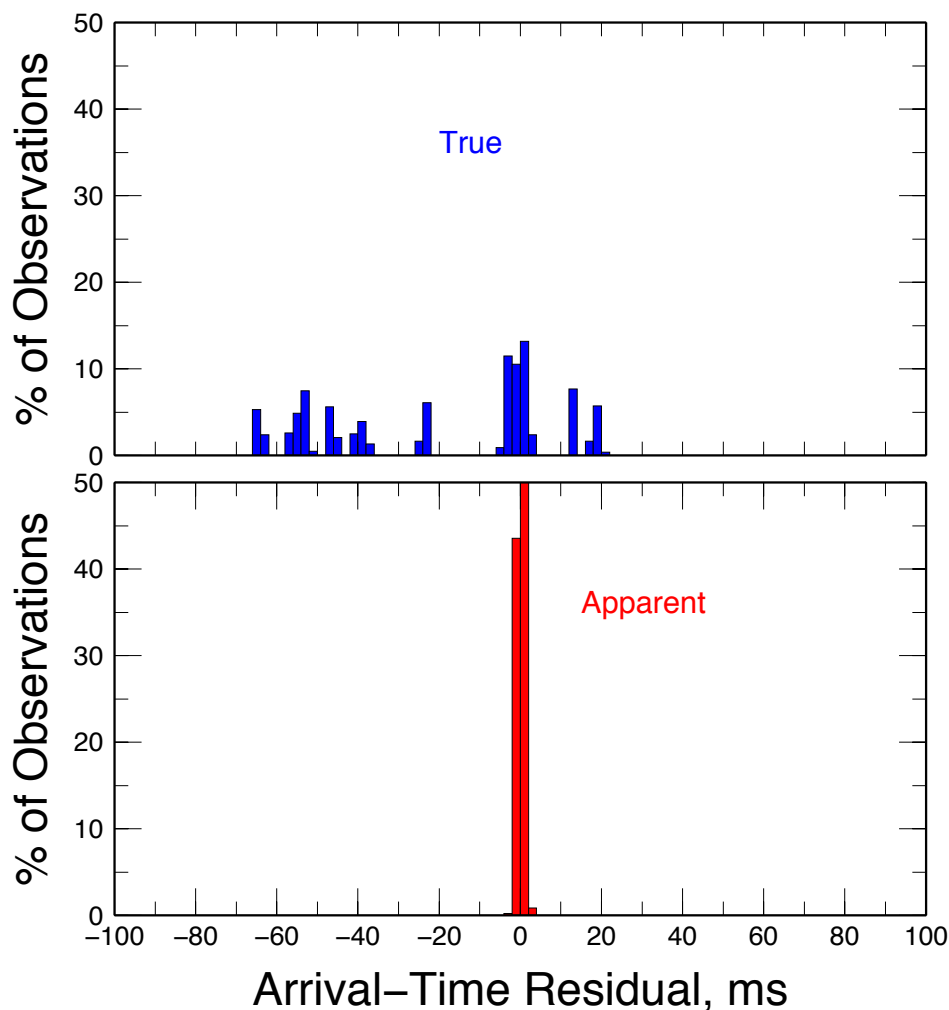


Figure 9: Histograms of theoretical arrival-time residuals, computed as for Figure 6, but applying travel-time corrections determined using calibration explosions, as illustrated in Figure 8. The residuals are reduced essentially to zero.

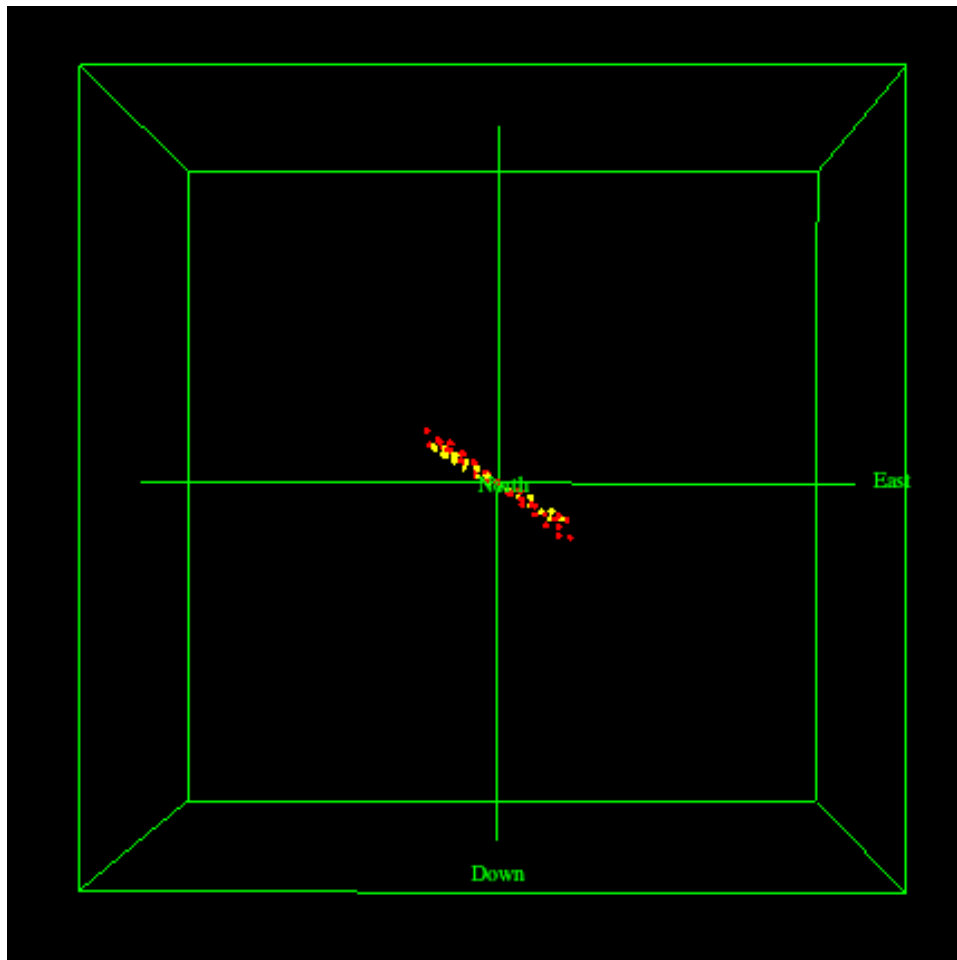


Figure 10: Perspective view of hypothetical and computed hypocenters based upon the three-dimensional hypothetical crustal model shown in Figure 5 and travel-time corrections determined using calibration explosions. The conventions are the same as in Figure 4. The center of the fault is located nearly perfectly.

It can be seen from Figure 10 that applying the corrections has removed almost all of the mislocation bias arising from crustal heterogeneity. Some distortion still remains, because the earthquakes are distributed over a fault 200 m x 200 m in area—they are not all at exactly the same location as the geophone and thus their ray paths are not exactly the same as the explosion-to-geophone ray paths.

4. CONCLUSIONS

Imperfect knowledge of crustal structure dominates absolute earthquake location errors in typical EGS and other geothermal operations. Although the errors can be reduced using various techniques to improve our knowledge of crustal structure, these improvements are not generally great enough to prevent the issue of structure from dominating the error budget. The problem can, however, be effectively solved by calibrating the experiment using a downhole geophone and surface shots, or by performing the reciprocal experiment of firing a downhole shot that is recorded at the stations of the monitoring network. A numerical study of such an experiment confirms that it is a promising approach and can reduce errors in hypocentral locations down to the meter level in typical geothermal situations. Such hypocentral accuracies are sufficiently good to enable targeted drilling to penetrate the hypocentral zones. This approach could thus raise earthquake location studies from the level of providing only general background information of interest, to providing information of direct, first-order use in geothermal production operations.

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