Geothermal Resources Council, TRANSACTIONS, Vol. 2 July 1978

2-D INVERSION OF SEISMIC ATTENUATION OBSERVATIONS IN COSO HOT SPRINGS, KGRA

Chi-Yuh Young and Ronald W. Ward

CENTER FOR ENERGY STUDIES
UNIVERSITY OF TEXAS AT DALLAS
RICHARDSON, TEXAS 75080

ABSTRACT

Seismograms from over sixty teleseismic events recorded by a sixteen element vertical component telemetered seismograph array installed by the U. S. Geological Survey in the Coso Hot Springs, Known Geothermal Resource Area (KGRA) have been analyzed to determine the lateral variation of seismic attenuation using the reduced spectral ratio technique. Seismograms of events from the SE and NW azimuths observed at a five station linear array have been used to infer a 2-D Q^{-1} model across the center of the geothermal system. A constrained generalized linear inversion algorithm was used to infer the Q^{-1} model. The model contains 15 cells with boundaries at depth of 5, 20, and 33 km and Q variation from 32 to 890. The zone of high attenuation dips from the surface between Coso Hot Springs and Airport Lake toward the northwest beneath Devil's Kitchen.

Introduction

The thermal state and concentration of partial melting present within a geothermal system are known to contribute to high attenuation of seismic waves. The relationship between observed seismic attenuation and temperature within a geothermal system are not well understood, though there are several proposed mechanisms to explain the observations (Jackson and Anderson 1970). Partial melting (Walsh, 1968, 1969), grain boundary relaxation, "high temperature internal friction background" (Jackson and Anderson, 1970) and vapor domination of porous rocks (White, 1975) are all mechanisms that can be acting within a geothermal system to produce the observed seismic attenuation. Mapping the lateral variation of seismic attenuation has proved to be an effective tool in locating and estimating the potential of a geothermal system. This study is oriented toward the development of quantitative inversion techniques for the observed lateral variation of δt^*.

The area of study is the Coso Hot Springs, Known Geothermal Resource Area (KGRA) located in the southwestern part of the Basin and Range Province of the western U. S. The Basin and Range Province exhibits high heat flow (Blackwell, 1970). The Coso Hot Springs KGRA is underlain principally by Mesozoic granitic and metamorphic units that are covered by upper Cenozoic volcanic flows including basaltic flows, dacitic flows, tuff, and ryholitic domes. A Pleistocene age for most of the basaltic and rhyolitic flows has been determined by K-Ar dating. Fumaroles and hot springs are associated with these flows.

The geothermal system is bounded by a N-NW trending normal fault that borders the eastern scarp of the Sierra Nevada Mountains. On the south a W-NW trending Wilson Canyon fault is the terminus of the geothermal activity. Arcuate ring faults center on the geothermal system.

Extensive geophysical surveys have been conducted across the Coso KGRA including seismic background noise survey, resistivity, magnetotelluric, gravity, aeromagnetic, heat flow, microearthquake, teleseismic P-wave traveltime delays, and seismic refraction. The focal depth of the microearthquakes range from 4 to 8 km with depth increasing toward the northwest. The teleseismic traveltime delays observed at the Coso Hot Springs KGRA are much smaller than those observed above other geothermal systems such as Long Valley, Yellowstone, or Geysers-Clear Lake KGRA's. They are generally less than 0.4 sec. Significant seismic attenuation has been reported by one group based on seismic refraction data (Combs and Jarzabeck, 1977) though a second group found a much lower level of attenuation (Weaver, personal communication).

Data

In September 1975, the U. S. Geological Survey installed a sixteen element vertical component telemetered seismograph array in the Coso Hot Springs, KGRA. About one hundred events were judged suitable for attenuation analysis and were digitized using the facilities of the U. S. Geological Survey in Menlo Park, California. The majority of the events occurred in three different azimuths from the array, SW (Tonga, Fiji, Kermadec), SE (Central and South America), and NW (Aleutians, Kurile Islands, Japan). This study will be limited to those events occurring in NW and SE azimuths from the array. The lateral variation of the seismic attenuation along a NW-SE trending cross section will be used to infer a 2-D Q^{-1} model as illustrated in Fig. 1.

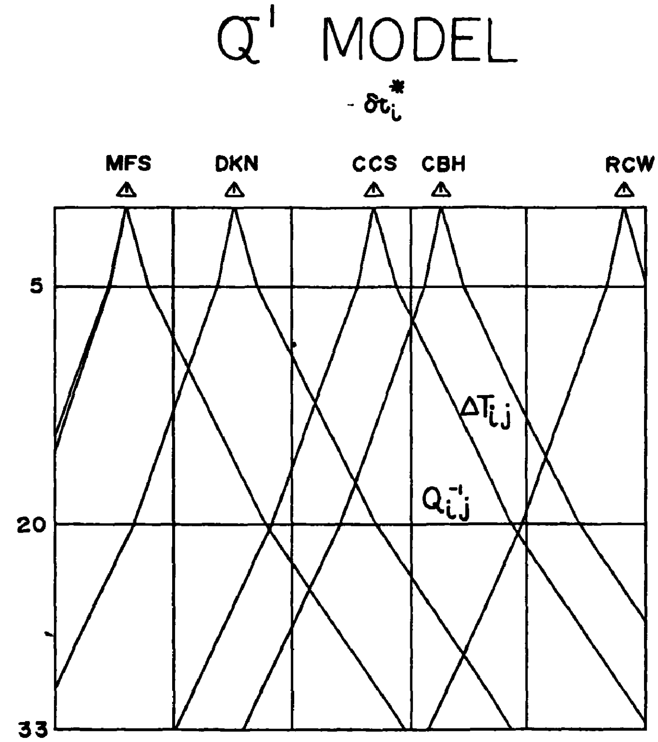

Fig. 1. Ray tracing to 5 stations through a discrete 2-D Q^{-1} model. Depth is in km. ΔT_{ij} is the traveltime of the ray from each event to a station through each cell. Q_{ij}^{-1} will be obtained from generalized linear inversion.

Differential Attenuation

Lateral variation of seismic attenuation exhibits itself in the amplitude, frequency content, and the coda duration of a seismic wave. This information is extracted by applying the reduced spectral ratio technique over a local area (Young and Ward, 1978) such as the Coso Hot Springs, KGRA. The amplitude spectrum of a seismic P-wave is estimated using the Maximum Entropy Method (Burg, 1967; Ulrych and Bishop, 1974).

It has been argued elsewhere (Young and Ward, 1978) that the ratio of the amplitude spectra of teleseismic event observed at two stations can be related to the differential attenuation δt^* by

$$R_{ij}(f) = A_i(f)/A_j(f)$$

$$= C_{ij}\exp\{-f(\delta t_i^* - \delta t_j^*)\}$$

We assume δt^* for the reference stations is zero, which gives

$$\ln R_{ij} = \ln C_{ij} - f\delta t_i^*$$

The δt_i^* is calculated for each station and each of these events. The events are separated into two groups NW and SE, which approach from opposite azimuths.

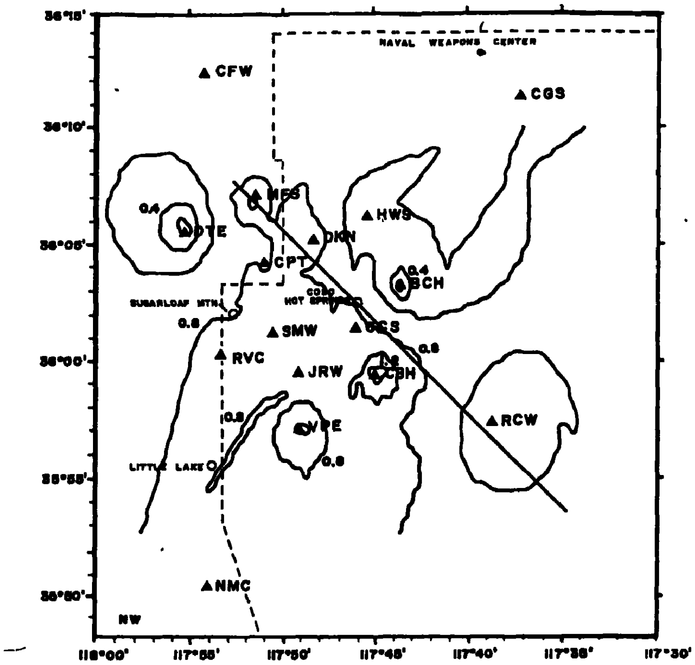

Fig. 2a

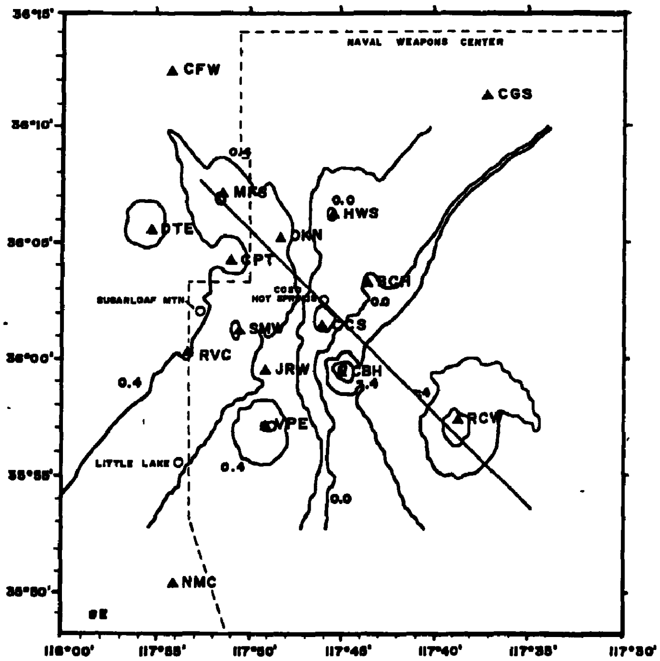

Fig. 2b

Fig. 2. 2-D contours of the averaged differential attenuation factor δt^* for events from a) NW and b) SE

As a first step toward inferring a 2-D Q^{-1} model, we compute the average δt^* of the events from the SE and NW for each station. Since the P-waves from most events are incident within 15° of the vertical, the observed δt^* between events from the same azimuth should not vary greatly. A contour map of δt^* values appears in Fig. 2a for events from the NW and Fig. 2b for events from the SE.

The attenuation anomalies for events from the NW are greatest near stations VPW (Volcano Peak) and CBH (in the north end of the Coso Basin). The anomalies in the Coso Basin shifts toward the northwest for events from the SE indicating that it is buried at depth. The anomaly above the Volcano Peak is independent of the azimuth of the events indicating a very shallow anomaly.

The qualitative interpretation of these δt^* observations must be improved if a quantiative Q^{-1} model is to be constructed, which gives the value of Q, the depth, and horizontal position within the geothermal system of the zones of low Q. In the following section we outline the mathematics for inverting this data to obtain a discrete Q^{-1} model.

Inversion of δt^* Data

A spherical layer earth model is assumed for the velocity variation. The velocity is a function of the radius only, and varies according to $v=ar^b$ (Julian and Anderson, 1968; Bullen, 1965), where a and b are constant for each layer. The velocity model of Herrin et al (1968) is used. The traveltime of the ray through each cell of a 2-D Q^{-1} model (Fig. 1) is calculated as outlined by Julian and Anderson (1968). The differential attenuation factor observed at a station for each event can be written

$$\delta t_j^* = \pi \sum_{ray} \Delta T_{ij} \, (Q^{-1})_i \qquad 1$$

where the summation is taken over the cells through which the ray passes and ΔT_{ij} is the traveltime to the jth station through the ith cell.

A generalized linear inversion algorithm is developed for discrete cells based on equation 1, and Wiggins (1972) and Jackson (1972). The inversion formulation assumes that Q^{-1} is an unknown parameter in each cell of the model. The first order perturbation of the unknown parameter is given by:

$$AX = \pi \begin{bmatrix} \Delta T_{11} & \cdots & \Delta T_{n1} \\ \cdots & & \cdots \\ \Delta T_{1m} & \cdots & \Delta T_{nm} \end{bmatrix} \begin{bmatrix} \Delta(Q_1^{-1}) \\ \vdots \\ \Delta(Q_n^{-1}) \end{bmatrix} =$$

$$= \begin{bmatrix} \Delta\delta t_1^* \\ \vdots \\ \Delta\delta t_m^* \end{bmatrix} = Y \qquad 2$$

The original traveltime matrix, A, is weighted by the observational errors, such that $A = S^{-\frac{1}{2}} A W^{\frac{1}{2}}$, where W is a constant weighting factor. The goal of the inversion iteration is to minimize $X^T W^{-1} X$. The generalized inverse matrix, H, the resolution matrix, R, and information density matrix S are given by:

$$H = V\Lambda^{-1}U^T \qquad 3$$

$$R = HA = VV^T \qquad 4$$

$$S = AH = UU^T \qquad 5$$

where U contains k data eigenvectors of length

m associated with the observation

δt_j^*,

Λ is a diagonal eigenvalue matrix, and

V contains k model eigenvectors of length n associated with the unknown parameters $\Delta(Q_i^{-1})$.

If eigenvalues fall below a threshold of 1% of the maximum they are omitted from the calculation.

This formalism permits the inversion of δt^* observations to obtain a Q^{-1} model for any geometry. In the present case a 2-D Q^{-1} model is inferred.

Q^{-1} Model

Seismogram from sixteen events occurring in NW and SE azimuths and observed at a NW-SE trending array of five stations (MFS, DKN, CCS, CBH, and RCW) were analyzed to obtain δt^* assuming a constant Q^{-1}. These observations formed the input data to a generalized inversion algorithm to determine a 2-D discrete Q^{-1} model for the Coso Hot Springs, KGRA. Layer boundaries were at depths of 5, 20 and 33 km for this preliminary model. A total of 15 cells of constant Q^{-1} were used. The model eigenvectors indicated that the Q^{-1} value was resolved accurately in every cell except the deepest cell beneath station RCW.

The 2-D Q^{-1} model derived (see Fig. 3) from teleseismic P-waves observed at the Coso Hot Springs, KGRA has a very low Q of 34 (high attenuation) in the near surface layer (upper 5 km) extending from the Coso Hot Springs toward the SE into the Airport Lake. This area is covered by Quaternary volcanics which are quite "lossy". The attenuation effect of surficial layers must be separated from the effect of partially molten zones. Both effects are probably operating near the surface. The zone of high attenuation dips toward the NW beneath Devil's Kitchen and then reverses itself moving back toward the SE. The attenuation decreases with depth.

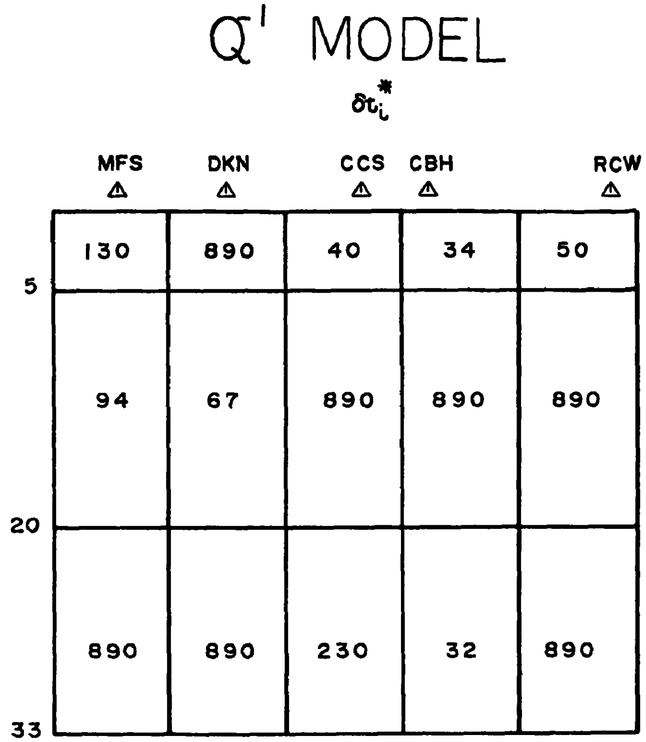

Fig. 3. 2-D Q^{-1} model derived from generalized linear inversion of δt* observations from sixteen events observed at 5 stations. Numbers in the cells are the value of Q.

In future studies these techniques will be extended to 3-D models using additional events and stations from the δt* observations at the Coso Hot Springs, KGRA. The resolution of these observations will be investigated by determining Q^{-1} models with different layer boundaries and many times the number of cells.

Research supported by the National Science Foundation, NSF Grant AER75-23619 and by the U. S. Geological Survey, Department of the Interior, under the USGS Grant No. 14-08-001-G-426.

References

Blackwell, D. D., (1970), The thermal structure of the continental crust, AGU, Geophysical Monograph, v. 14, p. 169-184.

Bullen, K. E., (1965), An introduction to the theory of seismology, 3rd ed., Cambridge Univ. Press, London

Burg, J. P., (1967), Maximum entropy spectral analysis: Paper presented at the 17th Annual Internation Meeting, Soc. of Exploration Geophysicists, Oklahoma City, Oklahoma, October 31, 1967.

Combs, J. and D. Jarzabeck, (1977), Preliminary seismic wave attenuation and Poisson's ratio studies at the Coso Geothermal Area, Calif.: The Coso Geothermal Project. Technical Report #5, Center for Energy Studies, U. T. at Dallas, Richardson, Texas.

Jackson, D. D., (1972), Interpretation of inaccurate, insufficient and inconsistent data, Geophys. J. R. astr. Soc., v. 28, p. 97-109

Jackson, D. D. and D. L. Anderson, (1970), Physical mechanism of seismic wave attenuation, Review of Geophy. and Space Physics v. 8, p. 1-63.

Julian, B. R. and D. L. Anderson, (1968), Travel-times, apparent velocities and amplitudes of body waves, Bull. Seism. Soc. Am., v. 58, p. 339-366.

Ulrych, T. J. and T. N. Bishop, (1975), Maximum entropy spectral analysis and autoregressive decomposition, Review of Geophys. and Space Physics, v. 13, p. 183-200.

Walsh, J. B., (1968), Attenuation in partially melted material, J. Geophys. Res., v. 73, p. 2209-2216.

Walsh, J. B., (1969), A new analysis of attenuation in partially melted rock, J. Geophys. Res., v. 74, p. 4333-4337.

White, J. E., (1975), Computed seismic speeds and attenuation in rocks with partial gas saturation: Geophysics, v. 40, p. 224-232.

Wiggins, R. A., (1972), The general linear inverse problem: Implication of surface waves and free oscillations for earth structure, Review of Geophys. and Space Physics, v. 10, p. 251-285.

Young, Chi-Yuh and Ward, R. W., (1978), Attenuation of teleseismic P-waves in Geysers-Clear Lake Region, California, to be published in U. S. Geol. Survey Prof. Paper.