Geothermics, Vol. 11, No. 4, pp. 259 – 268, 1982.
Printed in Great Britain.

0375 – 6505/82/040259 – 10 $03.00/0
Pergamon Press Ltd.

GEOTHERMAL EXPLORATION AND RESERVOIR MONITORING USING EARTHQUAKES AND THE PASSIVE SEISMIC METHOD

G. FOULGER

Science Institute, University of Iceland, Dunhagi 3, Reykjavík, Iceland

(*Received* 1 *December* 1981; *accepted for publication* 27 *May* 1982)

Abstract—This paper reviews the use of earthquake studies in the field of geothermal exploration. Local, regional and teleseismic events can all provide useful information about a geothermal area on various scales. It is imperative that data collection is conducted in properly designed, realistic experiments. Ground noise is still of limited usefulness as a prospecting tool. The utility of the method cannot yet be assessed because of its undeveloped methodology and the paucity of case histories.

INTRODUCTION

Volcanism, tectonic activity and seismicity often occur side by side. This, in turn, results in the commonly observed association of geothermal areas and seismicity. That the seismicity of a geothermal area is often different from that of the surrounding region was first established in Iceland (Ward *et al.*, 1969; Ward and Björnsson, 1971), and its potential as a prospecting tool quickly pointed out (Ward, 1972).

In the intervening years substantial advances have been made in the development of a methodology. In common with all other geophysical exploratory disciplines, the passive seismic method has both advantages and disadvantages. It can supply information that is unobtainable otherwise, and is most useful when results are reviewed in the light of information obtained by other exploratory methods. The fact that the method has received relatively little attention as a prospecting tool has doubtless much to do with the unconstrained and poorly understood nature of the energy source, which, in the case of local earthquakes, is, in part, a measure of our lack of understanding of geothermal areas. However, interest has increased with the greater keenness in exploiting geothermal resources, and at present ideas and theories are rather more numerous than case histories.

This paper aims at presenting a brief review of the main results obtained by application of the passive seismic method to the field of geothermal prospecting, together with illustrative case histories. A minimum of reference will be made to the voluminous speculation which appears in the literature.

SEISMICITY OF GEOTHERMAL AREAS

Many geothermal areas exhibit a higher background seismicity than their surrounding regions (e.g. Ward *et al.*, 1969; Ward and Björnsson, 1971; Marks *et al.*, 1978; Conant, 1972) whereas others have a similar seismicity to the regional (e.g. Evison *et al.*, 1976; Majer, 1978; Hunt and Latter, 1979) or even define a seismic gap (e.g. Steeples and Pitt, 1976; McEvilly *et al.*, 1978b).

Certain aspects of the seismicity of some seismically active geothermal areas have been related to the regional. For example, the South Iceland zone of destructive historic earthquakes extends into the Hengill geothermal area (Björnsson and Einarsson, 1974). Focal mechanism solutions of earthquakes in The Geysers area, California, are consistent with the regional Coastal Range tectonics (Bolt *et al.*, 1968). Majer and McEvilly (1979) concluded that in The Geysers area the direction of failure is controlled by regional stress, whilst the rate of failure is controlled by local stress levels.

It may be concluded, therefore, that in many cases the location of a geothermal area coincides with an area where regional stress is being released at a different rate to the surrounding areas. The differences in the seismicity of different geothermal areas may hence reflect differences in the regional tectonics of the areas in question.

For example, the Krafla geothermal area lies within the Northern Volcanic Zone of Iceland, (Björnsson *et al.*, 1977; Björnsson *et al.*, 1979) whereas the Hengill geothermal area lies at a triple junction where the Reykjanes Peninsula, the Western Volcanic Zone and the South Iceland Seismic Zone meet (Foulger and Einarsson, 1980). There is evidence that the Hengill area has displayed fairly continuous seismicity, with events up to magnitude 6, over the last half century, whereas no such seismicity is recorded from the Krafla area (Tryggvason, 1973; Tryggvason, 1978a; 1978b; Tryggvason, 1979). The two areas hence lie in contrasting regional tectonic settings, and display contrasting seismicity patterns. It may be that the ongoing seismicity of the Hengill area is intimately related to the proximity of the South Iceland zone of destructive historic earthquakes.

Variations in the seismicity within a large geothermal area may correspondingly reflect variations in tectonics within the area. For example, Smith *et al.* (1974) interpreted spatial and temporal variations in seismicity in the Yellowstone geothermal area in terms of contrasting crustal structure inside and outside the caldera.

LOCAL EARTHQUAKES

Spatial distribution

The spatial epicentral distribution of local earthquakes and its relation to surface features has provided insight into the geographical distribution of high temperature areas. The positions of high temperature areas on the Reykjanes Peninsula have been correlated with the points where fissure swarms cross the trace of the plate boundary as defined by earthquake epicenters (Klein *et al.*, 1977).

Marks *et al.* (1978) correlated the spatial distribution of microearthquakes in The Geysers area with two pressure sinks in the steam field. A well-defined westward dipping seismic gap between them was interpreted as an unfractured, impermeable barrier separating the two discrete sinks. Denlinger and Bufe (1980) likened the spatial distribution of The Geysers earthquakes to a two-stemmed mushroom occupying the zone of pore pressure decline, fluid withdrawal and negative reservoir dilation.

Microearthquakes located in the Ahuachapan geothermal area were found to define a plane of activity extending down to 6.5 km depth (Fig. 1). The interpretation that this plane of activity defines a fault that allows hot water to circulate to the surface was supported by evidence from borehole data and surface tectonics (Ward and Jacob, 1971).

Combs and Hadley (1977) likewise inferred the presence of a fault passing through the East Mesa geothermal area, California, on the basis of microearthquake locations.

Variations in the pressure and temperature conditions in the crust can also cause variations in seismicity. Laboratory experiments have demonstrated that a mechanism known as "stick-slip", where motion occurs as a series of discrete, rapid slips, occurs most readily at high pressures and low temperatures. This implies that the maximum depth at which earthquakes occur may be temperature dependent, and thereby of possible use as a geothermometer (Brace and Byerlee, 1966; Brace and Byerlee, 1970). Majer (1978) found that depths of microearthquakes beneath Grass Valley, Nevada, were within the range 0 – 8 km, contrasting

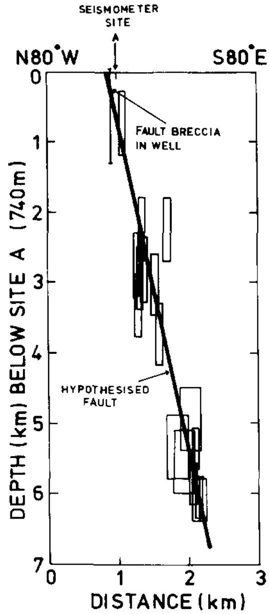

Fig. 1. Earthquake hypocenters projected on a vertical plane striking S80°E through the Ahuachapan geothermal area, 70 km WNW of San Salvador, El Salvador. Rectangles denote approximate precision in location. (Adapted from Ward and Jacob, 1971.)

with the typical 10 – 15 km for the region. This was thought to suggest a broad, shallow high temperature regime. Majer and McEvilly (1979) noted that the maximum depth of earthquakes beneath The Geysers area is 5 km as compared to 11 – 12 km for the surrounding region, and interpreted this as indicating elevated temperatures at 4 – 5 km depth. They also noted that earthquake activity tended to occur at the edges of, and not within, the production zone itself. This was interpreted as indicating relatively high pore pressure gradients at the edges of the field, and low pressure and high temperatures within it. The random distribution of events and lack of event migration was considered to indicate the intensely fractured state of the area and the lack of throughgoing faults.

Walter and Weaver (1980), in a study of the Coso Range, California, found no anomalous hypocentral depth distribution beneath the Coso geothermal field, which they concluded to indicate the absence of liquidus temperatures at shallow depths.

Chronological distribution

Microearthquakes from The Geysers production zones display a relatively high *b* value., i.e. an unusually high preponderance of small magnitude shocks (Marks *et al.*, 1978; Majer and McEvilly, 1979). This implies that the reservoir rocks are in a low stress state. Spectral analysis confirms this. The high *b* value also indicates that the microearthquake activity probably results from crack closing and sliding rather than the propagation of new fractures. It could also be an indication of structural heterogeneity and induced seismicity.

Cavit and Hadley (1980) considered the lack of temporal clustering in events in The Geysers area to indicate that the geothermal wells affected normal seismicity.

Walter and Weaver (1980) found a relatively high *b* value at shallow depths beneath the Coso geothermal area. This was interpreted as indicating short average fault lengths, and less likelihood of a large earthquake occurring within this volume of rock than outside it.

Estimates of Poisson's ratio

Poisson's ratio may be estimated by examining the relative travel times of compressional and shear waves from different microearthquakes to individual stations. Laboratory experiments indicate that dry rocks display a lower Poisson's ratio than wet ones. Somerton (1978) made laboratory measurements of Poisson's ratio in both dry and water-saturated sandstones from the Cerro Prieto (Mexico) geothermal field and obtained values of 0.1 for the dry and 0.3 for the wet samples. Majer and McEvilly (1979) obtained values for Poisson's ratio for The Geysers, by the passive seismic method, and obtained values of 0.15 – 0.2 inside the production zone, and higher values outside it. The low value obtained was thought to reflect steam domination of the reservoir rocks. Low values were also obtained for the Coso Hot Springs area (Combs and Rotstein, 1975) and interpreted as indicating vapour domination.

Studies of water-dominated fields, however, have yielded higher values, typically 0.4, e.g. the East Mesa (California) field (McEvilly *et al.*, 1978b), and the Cerro Prieto (Mexico) field (Majer and McEvilly, 1978a). Estimates of Poisson's ratio, obtained by the passive seismic method, are hence demonstratively useful in indicating degree of water saturation within the reservoir.

Source properties

Studies of the spectra of compressional and shear waves of microearthquakes from The Geysers also yielded exclusive information, especially when supplemented with data obtained from explosion seismology, using the same recording network. An estimate of the volume decrease attributable to seismic failure in the area was found to be much less than that which would be attributable to fluid withdrawal (Majer and McEvilly, 1979). In the absence of aseismic crustal adjustment, the difference must be taken up by re-injection and reservoir recharge.

Evidence of uniform source dimensions (50 m) for these microearthquakes was also obtained, compatible with low pressure and constant permeability and porosity.

Many focal mechanism studies have been done which give information about the strike and dip of active faults, and the direction of movement over the fault plane (e.g. Klein *et al.*, 1973; Majer and McEvilly, 1979).

Seismic attenuation

Information about Q (the quality factor, or the inverse of the attenuation) was also obtained for The Geysers by combining explosion seismological data and data from spectral analyses of microearthquakes. Estimates of Q for both compressional and shear waves were obtained. Variations in Q may be interpreted in terms of variations in the degree of water saturation, pressure, temperature, the presence of gas, partial melting and degree of compositional heterogeneity.

For The Geysers it was found that Q was high at shallow depths within the production zone, and low below it. This is interpreted as a reflection of the low pressure and the low degree of water saturation within the production zone, and higher pressure and degree of saturation below it (Majer and McEvilly, 1979; Kjartansson and Nur, 1981).

At the Cerro Prieto field, spectral analysis of a regional earthquake indicated significant variations in attenuation within the production zone, thought to be associated with subsurface faulting, alteration by precipitation and variations in sediment thickness (Majer and McEvilly, 1978).

At Grass Valley, Nevada, Majer (1978) detected clear high frequency attenuation of both earthquake and explosion generated seismic waves passing through the immediate area of the hot springs. This was attributed to effects within the upper 1 km of the crust beneath the springs.

At Sierra La Primavera, Mexico, McEvilly *et al.* (1978a) observed substantial differences in the attenuation of body waves of regional events, which they attibuted to the effects of the upper few kilometers of crust, possibly a shallow zone of partially molten material.

In Long Valley Caldera (California), Hill (1976) detected clear high frequency attenuation of explosion generated seismic waves passing at shallow depths beneath a region of hydrothermal alteration and hot spring activity. This could be caused by crustal heterogeneity, or indicate that the system reaches near boiling pressure at the depths penetrated by the seismic energy (Kjartansson and Nur, 1981). Unusually high attenuation of explosion generated seismic waves has also been reported from geothermal areas at Wairakei (Modriniak and Studt, 1959), Broadlands (Hochstein and Hunt, 1970) and Kawerau (Studt, 1958), New Zealand; Matsukawa, Japan (Hayakawa, 1970); and Coso Hot Springs, California (Combs and Jarzabeck, 1977).

Monitoring operations

Earthquakes provide one of the few means for continuous monitoring of the whole field, both before and after production has commenced. With the latter aim in mind, it is very important to commence data collection prior to exploitation so that seismicity during exploitation can be compared to pre-exploitation data.

There are many examples of changes in surface geothermal activity consequent to earthquakes. For example, changes in the eruptive cycle of Old Faithful Geyser, and also effects on hot springs in Yellowstone have been correlated to seismic activity (Rinehart, 1969; Rinehart and Murphy, 1969; Marler, 1964). Dramatic changes in the Geysir area in Haukadalur, Iceland, have been noted on several occasions, accompanying large destructive earthquakes in southern Iceland (Einarsson, 1964).

Significant changes in surface thermal activity have also accompanied earthquake swarms, e.g. in Reykjanes, Iceland (Ward *et al.*, 1969; Tryggvason, 1970) and the Krafla area, Iceland (Björnsson *et al.*, 1977, 1979; Einarsson and Brandsdóttir, 1980). Periodic variations in the discharge and temperature of thermal water in the Kagai hot spring area, Japan, has been correlated to changes in seismic rate within the Matsushiro sequence (Kasuga, 1967).

Data from The Geysers area, California, suggest that microearthquake activity has increased by a factor of 10 in the past 7 years. This has been attributed to accelerated fluid withdrawal and re-injection (Marks *et al.*, 1978; Majer and McEvilly, 1979; Denlinger and Bufe, 1980). The apparent increase in the frequency of large events may also be attributable to the re-injection process. Majer and McEvilly (1979) considered the microearthquakes to indicate regions of expansion of the vapour-dominated zone, and the level of activity to be a measure of the degree of departure from a state of equilibrium. Migration of the active zones may thus indicate migration of the boundaries of the vapour-dominated zone.

REGIONAL EARTHQUAKES

Both regional and local events have been used in several instances to define areas of anomalous shear wave attenuation, interpreted as magma chambers in the crust and upper mantle. Some notable examples are studies done in the Katamai volcanic range, Alaska (Kubota and Berg, 1967; Matumoto, 1971); the Sulphur Springs geothermal region, St. Lucia, West Indies (Aspinall *et al.*, 1976); at Socorro, New Mexico (Sanford *et al.*, 1977a,b); and Krafla, Iceland (Einarsson, 1978).

Spectral analysis of seismic waves of regional earthquakes indicated anomalous *P*-wave attenuation in the East Mesa, California, area, and was presented as evidence for a magma chamber (Combs and Jarzabeck, 1978).

TELESEISMS

Teleseismic techniques can be used to obtain precise information from much greater depths than most conventional geophysical methods. A technique which has been used in geothermal exploration with good results involves accumulating records of distant earthquakes on a relatively dense network of stations over the area under investigation. Delays in the relative arrival times of the *P*-waves are then assessed for each station. The data can be used to estimate the size and position of low velocity bodies in the crust and upper mantle. Assuming that such delays are attributable to hot or partially molten rock, rough estimates of the total heat anomaly may be made.

In the Yellowstone region, delays of up to 2 s were recorded on a dense network, and attributed to a zone of partial melting extending to a depth of 200 – 250 km, with a higher proportion of melt at shallow depths. The *P*-wave velocity within this body was estimated to be about 10% lower than that of the surrounding rocks. This would imply average temperatures of the order of 1000°C within the body (Iyer, 1979; Steeples and Iyer, 1976).

Delays of up to 0.3 s were detected at the Long Valley geothermal area. Analysis of these data revealed that the anomalous body is probably shallower than 25 km, with a velocity contrast in the range 5 – 15%. Steeples and Iyer (1976) found that a spherical body 14 km in diameter with its upper surface at 6 km depth was consistent with these and gravity observations (Fig. 2). The order of magnitude of the teleseismic delays indicated that temperatures within the body are such that some degree of partial melt must be present. The fact that teleseismic shear

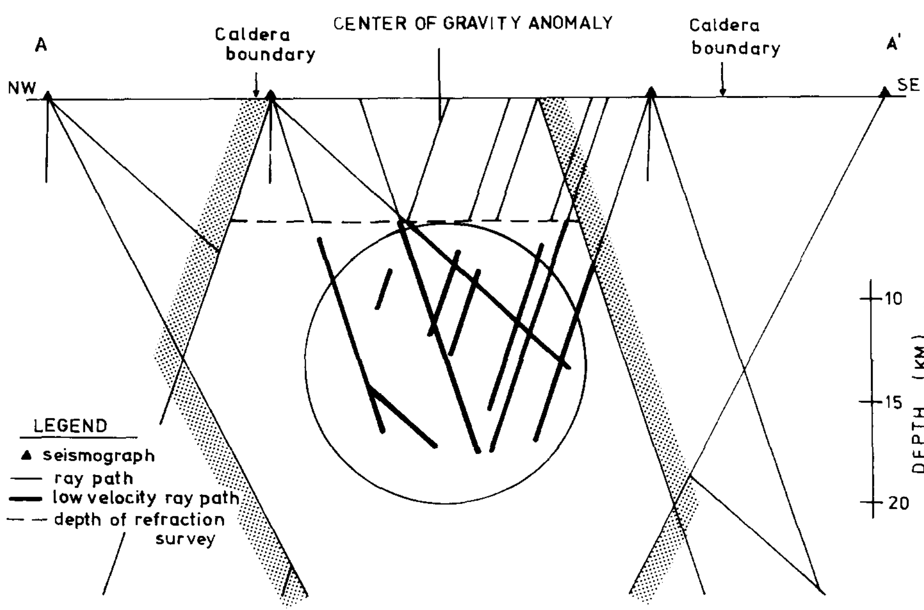

Fig. 2. Sectional view of Long Valley Caldera. Stations within 1 km of section line have been projected on to the section. Model represents velocity contrast of 15% along heavy ray paths. Heavy lines on ray paths are proportional in length to delays detected along respective ray paths. Velocities outside the stippled zone are normal (6.0 km/s) and the effects of the upper 6 km of material (depth penetrated by refraction survey) have been removed. The circle is a sectional view of a sphere 14 km in diameter. A density contrast of 0.18 g/cm^3 would produce a 10 mGal gravity anomaly at the surface directly above the center of the sphere. (Adapted from Steeples and Iyer, 1975.)

waves were transmitted indicated, however, that either the molten portion of the body must exist in pockets less than 4 km in diameter (the approximate wavelength of the shear waves), or the body must have sufficient viscosity to transmit shear energy. *S*-waves from local earthquakes were found not to penetrate deep enough to sample the anomalous body.

P-wave delays of more than 1 s have been observed in The Geysers – Clear Lake area. Iyer *et al.* (1979) postulated the existence of a magma chamber with a core of severely molten rock beneath Mt. Hannah, to the NE of The Geysers area, and partially molten rock underlying The Geysers field itself. Both zones extend down to a depth of 20 km or more.

Young and Ward (1978) analysed *P*-waves from over 60 teleseismic events recorded over the Coso Hot Springs, California, and modelled two-dimensional variations in attenuation. In a later study the model was extended to three dimensions (Young and Ward, 1980). Their method involves dividing the rock volume of interest into cells, and calculating the attenuation of each of these. Based on these data a model for the Coso Hot Springs KGRA is proposed with high attenuation at near surface (0 – 5 km), due to 'lossy' lithology or partial fluid saturation; an intermediate fractured zone (5 – 12 km), through which heat is transferred by fluid movement; and a deep zone (12 – 20 km), containing intrusives or magma. A two-dimensional model has also been calculated for The Geysers – Clear Lake field, which indicates a high attenuation zone in a position corresponding to gravity anomalies and *P*-wave travel time residuals (Ward and Young, 1980).

GROUND NOISE

For completeness, some mention of ground noise should be made. The method commonly consists of sampling the ambient ground noise over the surface of the geothermal area with a dense station network. Two main lines of approach are followed.

Firstly, the theory has been proposed that hydrothermal processes within reservoirs (e.g. a phase change) radiate seismic energy. To test this hypothesis noise levels are contoured on the surface in order to delineate noise sources.

Secondly, consideration is taken of the type of elastic wave propagating, and processing then aimed at inverting the propagation characteristics to obtain information about the distribution of medium properties (e.g. seismic velocity and attenuation), and the location of sources.

High levels of ground noise have been detected at many sites in the vicinity of hot springs and geysers, and some evidence presented for the existence of radiating sources at shallow depths. For example, at Grass Valley, Nevada (Liaw and McEvilly, 1979); Long Valley, California (Iyer and Hitchcock, 1976); and Yellowstone National Park (Iyer and Hitchcock, 1974). At Wairakei and Waiotapu, New Zealand, depths of less than 200 m were attributed to noise sources beneath sites of surface activity (Whiteford, 1975). There is also some evidence that geothermal areas with no visible surface expression may not be noise emitters, e.g. East Mesa, California (Iyer and Hitchcock, 1975) and Kurobe, Japan (Ehara and Yuhara, 1978). As yet, no deep subsurface emitter has been located.

Formidable difficulties are associated with the application of this method to geothermal prospecting. The greatest of these are high levels of cultural and meteorological noise interference, laterally variable and unassessable ground amplification (due to variable geology), and poorly constrained crustal models. The lack of station correlation prohibits the location of sources, and powerful surface sources such as hot springs mask weaker seismic waves from buried emitters, so amplitude mapping will only yield information about shallow sources. In order to demonstrate the existence of body waves radiated by deep emitters within geothermal systems, large station arrays and sophisticated processing methods would be necessary.

Analysis of ground noise remains, as yet, a tool of very limited applicability in geothermal prospecting.

FURTHER DEVELOPMENTS

The main points emerging from a review of the use of the passive seismic method as a prospecting tool at present is that there is a paucity of case histories, and no well-developed systematic methodology for the delineation of anomalous portions of the crust, with the possible exception of the treatment of teleseisms. Indeed, results may be contradictory, or ambiguous in the light of parallels drawn with laboratory experiments, or at best difficult to interpret. On the positive side, however, seismic monitoring is cheap, has yielded valuable information about reservoirs in many cases, and can be used to 'see' much deeper than conventional geophysical methods.

When exploration of a prospect commences there is a strong case for conducting a preliminary monitoring experiment using a small number of stations, over as long a period as possible, in order reliably to assess the level, and temporal nature, of local and regional activity. Only then can a more sophisticated experiment be designed on a realistic basis. In areas which have a high seismicity prior to exploitation there is an especially strong case for commencing monitoring early, particularly if injection wells are likely to be used.

To the author's knowledge, The Geysers is the only exploited geothermal area where continuous seismic monitoring and spectral analysis was, and is, being done with a view to modelling the reservoir.

Many questions remain to be answered before the potential utility of the method can be fully assessed, and many more case histories with intensive studies are necessary. The possibility of developing methods to map attenuating areas should be investigated. Spectral analysis, and the calculation of source characteristics could provide an answer to whether the phenomenon of a 'geothermal' earthquake exists. Changes in the rate of seismicity due to production activities should be evaluated, and efforts made to find discriminants between natural and induced earthquakes. Finally, more laboratory work needs to be done to enable the results of seismic studies to be interpreted with a minimum of ambiguity, in terms of reservoir characteristics.

Acknowledgements—My grateful thanks are extended to the following people for reading the manuscript and making many valuable suggestions for its improvement: Sveinbjörn Björnsson, Axel Björnsson, Páll Einarsson, Gudmundur Pálmason and Ernie Majer.

REFERENCES

Aspinall, W. P., Michael, M. O. and Tomblin, J. (1976) Evidence for fluid bodies beneath the Sulphur Springs geothermal region, St. Lucia, West Indies. *Geophys. Res. Lett.* **3**, 87.

Björnsson, S. and Einarsson, P. (1974) Seismicity of Iceland. In *Geodynamics of Iceland and the North Atlantic Area* (Edited by Kristjánsson, L., D.) pp. 225–239. Reidel, Dordrecht, Holland.

Björnsson, A., Johnsen, G., Sigurdsson, S., Thorbergsson, G. and Tryggvason, E. (1979) Rifting of the plate boundary in North Iceland. *J. geophys. Res.* **84**, 3029–3038.

Björnsson, A., Saemundsson, K., Einarsson, P., Tryggvason, E. and Grönvold, K. (1977) Current rifting episode in North Iceland. *Nature, Lond.* **266**, 318–323.

Bolt, B. A., Lomnitz, C. and McEvilly, T.V. (1968) Seismological evidence on the tectonics of central and northern California and the Mendocino escarpment. *Bull. seism. Soc. Am.* **58**, 1725–1767.

Brace, W. F. and Byerlee, J. D. (1966) Stick-slip as a mechanism for earthquakes. *Science* **153**, 990–992.

Brace, W. F. and Byerlee, J. D. (1970) California earthquakes: why only shallow focus? *Science* **168**, 1573–1575.

Cavit, D. S. and Hadley, D. M. (1980) Geothermal earthquakes: temporal and spatial characteristics. *EOS* **61**, 1051.

Combs, J. and Hadley, D. (1977) Microearthquake investigation of the Mesa geothermal anomaly, Imperial Valley, California. *Geophysics* **42**, 17–33.

Combs, J. and Jarzabeck, D. (1977) Preliminary seismic wave attenuation and Poisson's ratio studies at the Coso geothermal area, California: The Coso Geothermal Project. Technical Report, No. 5, Center for Energy Studies, U.T. at Dallas, Richardson, Texas.

Combs, J. and Jarzabeck, D. (1978) Seismic wave attenuation anomalies in the East Mesa geothermal field, Imperial Valley, California: preliminary results. *Geoth. Res. Council, Trans.* **2**, 109–112.

Combs, J. and Rotstein, Y. (1975) Microearthquake studies at the Coso geothermal area, China Lake, California. *Proc. Second U.N. Symp. Develop. Use Geoth. Res., San Francisco, U.S.A.* Library of Congress Catalog Card No. 75-32682, pp. 909–916.

Contant, D. A. (1972) A microearthquake survey of geothermal areas in Iceland, 1970. *Earthq. Notes* **43,** 19–32.

Denlinger, R. P. and Bufe, C. A. (1980) Seismicity induced by steam production at The Geysers steam field in Northern California. *EOS* **61,** 1051.

Ehara, S. and Yuhara, K. (1978) Seismic noise measurements of some geothermal areas in Japan. *Geotherm. Res. Council, Trans.* **2,** 171–172.

Einarsson, P. (1978) *S*-wave shadows in the Krafla caldera in NE-Iceland, evidence for a magma chamber in the crust. *Bull. volcan.* **43,** 1–9.

Einarsson, P. and Brandsdóttir, B. (1980) Seismological evidence for lateral magma intrusion during the July 1978 deflation of the Krafla volcano in NE-Iceland. *J. Geophys.* **47,** 160–165.

Einarsson, T. (1964) *Geysir í Haukadal. (The Great Geysir in Haukadalur).* The Geysir Committee, Reykjavík, Iceland.

Evison, F. F., Robinson, R. and Arabasz, W. J. (1976) Microearthquakes, geothermal activity and structure, Central North Island, New Zealand. *N.Z. Jl. Geol. Geophys.* **19,** 625–637.

Foulger, G. and Einarsson, P. (1980) Recent earthquakes in the Hengill-Hellisheidi area in SW-Iceland. *J. Geophys.* **47,** 171–175.

Hayakawa, M. (1970) The study of underground structure and geophysical state in geothermal areas by seismic exploration. *Geothermics Special Issue* 2, 347–357.

Hill, D. P. (1976) Structure of Long Valley Caldera, California, from a seismic refraction experiment. *J. geophys. Res.* **81,** 745–753.

Hochstein, M. P. and Hunt, T. M. (1970) Seismic, gravity and magnetic studies, Broadlands geothermal field, New Zealand. *Geothermics Special Issue* 2, 333–346.

Hunt, T. and Latter, J. H. (1979) Seismic activity near Wairakei geothermal field. *Proc. N.Z. Geotherm. Workshop 1979,* Part 1, 14–19.

Iyer, H. M. (1975) Short note: search for geothermal seismic noise in the East Mesa area, Imperial Valley, California. *Geophysics* **40,** 1066–1072.

Iyer, H. M. (1979) Deep structure under Yellowstone National Park, U.S.A.: a continental 'hot spot'. *Tectonophysics* **56,** 165–197.

Iyer, H. M. and Hitchcock, T. (1974) Seismic noise measurements in Yellowstone National Park. *Geophysics* **39,** 389–400.

Iyer, H. M. and Hitchcock, T. (1975) Seismic noise as a geothermal exploration tool: techniques and results. *Proc. 2nd U.N. Symp. Develop. Use Geotherm. Res., San Francisco, U.S.A.* Library of Congress Catalog Card No. 75-32682, pp. 1075–1083.

Iyer, H. M. and Hitchcock, T. (1976) Seismic noise survey in Long Valley, California. *J. geophys. Res.* **81,** 821–840.

Iyer, H. M., Oppenheimer, D. H. and Hitchcock, T. (1979) Abnormal *P*-wave delays in The Geysers–Clear Lake geothermal area, California. *Science* **204,** 495–497.

Kasuga, I. (1967) Aspect on the relation of thermal water and Matsushiro earthquakes in Kagai hotspring area, Nagano prefecture. *J. Geol. (Tokyo)* **76,** 16–26.

Kjartansson, E. and Nur, A. (1981) Attenuation due to thermal relaxation in porous rocks. *Geophysics,* in press.

Klein, F. W., Einarsson, P. and Wyss, M. (1973) Microearthquakes on the Mid-Atlantic plate boundary on the Reykjanes Peninsula in Iceland. *J. geophys. Res.* **78,** 5084–5099.

Klein, F. W., Einarsson, P. and Wyss, M. (1977) The Reykjanes Peninsula, Iceland, earthquake swarm of September 1972 and its tectonic significance. *J. geophys. Res.* **82,** 865–888.

Kubota, S. and Berg, E. (1967) Evidence for magma in the Katmai volcanic range. *Bull. volcan.* **31,** 175–214.

Liaw, A. L. and McEvilly, T. V. (1978) Microseisms in geothermal exploration—studies in Grass Valley, Nevada. *Geophysics* **44,** 1097–1115.

Lofgren, B. (1978) Monitoring crustal deformation in The Geysers geothermal area, California. U.S.G.S. Open-file Report 78-597, 19 pp.

McEvilly, T. V., Mahood, G. A., Majer, E. L., Schechter, B. and Truesdell, A. H. (1978a) Seismological/geological field study of the Sierra La Primavera geothermal system. Preliminary Report, Department of Geology and Geophysics, Berkeley, California, 13 pp.

McEvilly, T. V., Schechter, B. and Majer, E. L. (1978b) East Mesa seismic study. Annual Report, Earth Sciences Division, Lawrence Berkeley Laboratory, University of California, pp. 26–28.

Majer, E. L. (1978) Seismological investigations in geothermal regions. Ph.D. thesis, University of California, Berkeley.

Majer, E. L. and McEvilly, T. V. (1978) Seismological studies at Cerro Prieto. Annual Report, Earth Sciences Division, Lawrence Berkeley Laboratory, University of California, pp. 29–33.

Majer, E. L. and McEvilly, T. V. (1979) Seismological investigations at The Geysers geothermal field. *Geophysics* **44,** 246–269.

Marks, S. M., Ludwin, R. S., Louie, K. B. and Bufe, C. G. (1978) Seismic monitoring at The Geysers geothermal field, California. U.S.G.S. Open-file report 78-798.

Marler, G. D. (1964) Effects of the Hebgen Lake earthquake of August 17, 1959, on the hot springs of the Firehole Geyser Basin, Yellowstone National Park. U.S.G.S. Professional Paper 435, pp. 185–197.

Matumoto, T. (1971) Seismic body waves observed in the vicinity of Mount Katmai, Alaska, and evidence for the existence of molten chambers. *Geol. Soc. Am. Bull.* **82**, 2905 – 2920.

Modriniak, N. and Studt, F. E. (1959) Geological structure and volcanism of the Taup-Tarawera district. *N.Z. Jl. Geol. Geophys.* **2**, 654.

Rinehart, J. S. (1969) Old Faithful performance, 1890 through 1966. *Bull. volcan.* **33**, 153 – 163.

Rinehart, J. S. and Murphy, A. (1969) Observations of pre- and post-earthquake performance of Old Faithful Geyser. *J. geophys. Res.* **74**, 574 – 575.

Sanford, A. R., Mott, R. P., Shuleski, P. J., Rinehart, E. J., Caravella, F. J., Ward, R. M. and Wallace, T. C. (1977a) Geophysical evidence for a magma body in the crust in the vicinity of Socorro, New Mexico. *Am. Geophys. Union Monog.* **20**, 385 – 403.

Sanford, A. R., Rinehart, E. J., Shuleski, P. J. and Johnston, J. A. (1977b) Evidence from microearthquake studies for small magma bodies in the upper crust of the Rio Grande Rift near Socorro, New Mexico. *Trans. Am. Geophys. Union* **58**, 1188 (Abstr.).

Smith, R. B., Shuey, R. T., Freidline, R. O., Otis, R. M. and Alley, L. B. (1974) Yellowstone hotspot: new magnetic and seismic evidence. *Geology* **2**, 451 – 455.

Somerton, W. (1978) Some physical properties of Cerro Prieto cores. *Proc. First Symp. Cerro Prieto Geothermal Field, Baja California, Mexico.* Berkeley, Lawrence Berkeley Laboratory, LBL-7098.

Steeples, D. W. and Iyer, H. M. (1975) Teleseismic *P*-wave delays in geothermal exploration. *Proc. Second U.N. Symp. Develop. Use Geotherm. Res., San Francisco, U.S.A.* Library of Congress Catalog Card No. 75-32682, pp. 1199 – 1205.

Steeples, D. W. and Iyer, H. M. (1976) Low velocity zone under Long Valley as determined from teleseismic events. *J. geophys. Res.* **81**, 849 – 860.

Steeples, D. W. and Pitt, A. M. (1976) Microearthquakes in and near Long Valley, California. *J. geophys. Res.* **81**, 849 – 860.

Studt, F. E. (1958) Geophysical reconnaissance at Kawerau, New Zealand. *N.Z. Jl. Geol. Geophys.* **1**, 219 – 246.

Tryggvason, E. (1970) Surface deformation and fault displacement associated with an earthquake swarm in Iceland. *J. geophys. Res.*, **75**, 4407 – 4422.

Tryggvason, E. (1973) Seismicity, earthquake swarms and plate boundaries in the Iceland region. *Bull. seismol. Soc. Am.* **63**, 1327 – 1348.

Tryggvason, E. (1978a) Jarðskjálftar á Íslandi 1930 – 1939. (The seismicity of Iceland 1930 – 1939.) Science Institute, Reykjavík, Report RH-78-21, 92 pp.

Tryggvason, E. (1978b) Jarðskjálftar á Íslandi 1940 – 1949. (The seismicity of Iceland 1940 – 1949.) Science Institute, Reykjavík, Report RH-78-22, 51 pp.

Tryggvason, E. (1979) Jarðskjálftar á Íslandi 1950 – 1959. (The seismicity of Iceland 1950 – 1959.) Science Institute, Reykjavík, Report RH-79-06, 90 pp.

Walter, A. W. and Weaver, C. S. (1980) Seismicity of the Coso Range, California. *J. geophys. Res.* **85**, 2441 – 2458.

Ward, P. L. (1972) Microearthquakes: prospecting tool and possible hazard in the development of geothermal resources. *Geothermics* **1**, 3 – 12.

Ward, P. L. and Björnsson, S. (1971) Microearthquakes, swarms and the geothermal areas of Iceland. *J. geophys. Res.* **76**, 3953 – 3982.

Ward, P. L. and Jacob, K. H. (1971) Microearthquakes in the Ahuachapan geothermal field, El Salvador, Central America. *Science* **173**, 328 – 330.

Ward, P. L., Pálmason, G. and Drake, C. L. (1969) Microearthquakes and the Mid-Atlantic Ridge in Iceland. *J. geophys. Res.* **74**, 665 – 684.

Ward, R. W. and Young, C-Y (1980) Mapping seismic attenuation within geothermal systems using teleseisms with application to The Geysers – Clear Lake region. *J. geophys. Res.* **85**, 5227 – 5236.

Whiteford, P. C. (1975) Studies of the propagation and source location of geothermal seismic noise. *Proc. Second U.N. Symp. Develop. Use Geotherm. Res., San Francisco, U.S.A.* Library of Congress Catalog Card No. 75-32682, pp. 1263 – 1271.

Young, C-Y. and Ward, R. W. (1978) 2-*D* inversion of seismic attenuation observations in Coso Hot Springs, KGRA. *Geotherm. Res. Council, Trans.* **2**, 743 – 746.

Young, C-Y. and Ward, R. W. (1980) Three-dimensional Q^{-1} model of the Coso Hot Springs known geothermal resource. *J. geophys. Res.* **85**, 2459 – 2470.