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ABSTRACT

Single-phase and two-phase geothermal reservoirs are currently being exploited for power production in Italy, Mexico, New Zealand, the U.S. and elsewhere. Vertical ground displacements of upto 4.5 m and horizontal ground displacements of up to 0.5 m have been observed at Wairakei, New Zealand that are clearly attributable to the resource exploitation. Similarly, vertical displacements of about 0.13 m have been recorded at The Geysers, California. No significant ground displacements that are attributable to large-scale fluid production have been observed at Larderello, Italy and Cerro Prieto, Mexico. Observations show that subsidence due to geothermal fluid production is characterized by such features as an offset of the subsidence bowl from the main area of production, time-lag between production and subsidence and non-linear stress-strain relationships. Several plausible conceptual models, of varying degrees of sophistication, have been proposed to explain the observed features. At present, relatively more is known about the physical mechanisms that govern subsidence than the relevant thermal mechanisms. Although attempts have been made to simulate observed geothermal subsidence, the modeling efforts have been seriously limited by a lack of relevant field data needed to sufficiently characterize the complex field system.

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CONTENTS

INTRODUCTION

In many parts of the world geothermal energy is being actively exploited for power generation. Compared to oil and coal, the energy content of a unit mass of geothermal water is relatively small. Hence, power production from geothermal reservoirs, especially those dominated by liquid water, entails the extraction of large volumes of the fluids, leading invariably to the mining of these fluids. This depletion of stored fluid volume is compensated largely by a reduction in the bulk volume of the reservoir with associated reservoir deformation. Abundant field evidence exists to show that the effects of reservoir deformation often propagate to the land surface to be manifested as vertical and horizontal ground displacements. Although the term "subsidence" is suggestive of vertical downward movement of the ground surface, we shall, in this paper, use the term in a more general context to include both horizontal and vertical displacements.

Additionally, the deformations accompanying reservoir depletion may also lead to the activation of movements along preexisting faults, leading to seismic events. The ground displacements which may often attain magnitudes of several meters, can lead to significant environmental consequences in some areas. For example, vertical movements of only a few feet in some coastal areas such as in Texas can lead to flooding and loss of valuable urban or agricultural lands. Abrupt spatial changes in the magnitude of subsidence, on the other hand, can lead to the rupturing of irrigation canals or pipelines.

There exists, therefore, a practical desire to exploit the geothermal resource in such a fashion that the deleterious effects of land subsidence are minimal and acceptable. To achieve this end, a proper understanding of the subsidence mechanism is essential so that the consequences of specific exploitation strategies can be foreseen and appropriate ameliorative measures taken.

The purpose of this paper is to assess our current status of knowledge related to subsidence caused by the removal of geothermal fluids. In particular, we shall address the following questions: What are the patterns and magnitudes of subsidence that have been observed in different parts of the world? What are the physical bases that relate fluid withdrawal and ground displacements? What is our current ability to predict land subsidence with the help of mathematical models? And finally, what are the key questions that need to be answered in order to increase our ability to predict subsidence?

We shall begin the paper with a description of case histories relating to geothermal systems from around the world. Following this, we shall describe the physical mechanisms that govern subsidence and examine how these physical mechanisms may be quantitatively analyzed using mathematical models. We shall close the paper with a discussion of the current status of knowledge and the identification of key issues requiring resolution.

FIELD OBSERVATIONS

In general, geothermal systems can be classified into five categories: normal gradient-, radiogenic-, high heat flow-, geopressured-and hydrothermal systems (DiPippo, 1980). Normal gradient systems are systems in which the temperature gradient in the earth's crust averages about 30°C/km. Exploitation of such a system would require one to drill deep in the earth's crust, rendering this resource to be uneconomical at present.

Geothermal energy produced by the radioactive decay of uranium, thorium, and potassium in the earth's crust forms a radiogenic system. Radioactive decay of 1 kg granite can release about one-billionth of a watt of heat. Thus, a fairly large amount of heat energy can be obtained by tapping radiogenic resource of the earth's crust as a whole. However, this energy is quite diffused and a suitable medium may not be readily available to permit its large scale extraction.

Subsurface temperatures are principally controlled by conductive flow of heat through solid rocks, by convecting flow in circulating fluids, or by mass transfer in magmas. The conduction dominated, high heat flow areas may be associated with regions in which the crust is abnormally thin, thus allowing the mantle to come into closer proximity to the surface, or in which a large, deep seated magma chamber is enclosed within the earth's crust. Such areas are often found to have large thermal gradients, sometimes as large as 2 to 4 times the normal gradient as found in the Hungarian Basin (Boldizsar, 1970) where temperature gradients of 40 to 75°C/km are known. These regions are expected to yield high temperatures at shallow depths. However, such areas may not prove feasible for power production because of the diffused nature of energy contained in them.

The fourth type of geothermal system, the geopressured system, is found in regions where fluid pressures exist in excess of hydrostatic pressure gradient of 9.8 KPa/m (0.433 psi/ft). It is believed that any or all of the following processes are responsible for the existence of a geopressured system: Rapid burial of saturated sediments, with rates of loading exceeding rates of water expulsion; development of osmotic pressure across clay beds; and liberation of water through diagenetic alteration of montmorillonite to illite between temperatures of 80°C to 120°C (Jones, 1969, 1975). Such fields are found along the northern coast of Gulf of Mexico, and in many other parts of the world. These fields do not have high temperature gradients but considerable temperatures are encountered due to great depths (≈6 km) involved. Such systems are of economic importance as they are capable of delivering mechanical energy, thermal energy and large supplies of methane gas. The Gulf coast of Texas and Louisiana is currently being explored with deep wells to harness this resource.

The last geothermal resource, the hydrothermal type, has been extensively exploited and used for power production, space heating and other applications throughout the world because of its proximity to the earth's surface and its amenability to energy extraction. The driving heat energy for such systems is supplied at the base of the convection loop. Hydrothermal systems may be subclassified into two types: vapor dominated and liquid dominated systems, which differ in the physical state of the dominant pressure controlling phase. In vapor dominated systems, pressure is controlled by the steam phase while in the liquid dominated systems, it is controlled by liquid water. Among the geothermal systems discovered to date, hot water systems are perhaps twenty times as common as vapor dominated systems (Muffler and White, 1972). Among the liquid dominated systems, Wairakei in New Zealand and Cerro Prieto in Mexico are currently producing 140 MW and 180 MW of electric power respectively. Electric power is also being produced from the vapor dominated systems such as The Geysers in California U.S.A. (900 MW) and Larderello in Italy (380 MW).

Of the five categories of geothermal systems, only the geopressured and the hydrothermal systems are presently viable for economic power production. Therefore, we shall limit our discussion of subsidence to hydrothermal and geopressured systems.

In the following section field observations are presented from several geothermal sites. Attempt is made to emphasize the important features relevant to subsidence. A total of six case histories are discussed. These include: the liquid-dominated systems at Wairakei and Broadlands in New Zealand and at Cerro Prieto in Mexico; the vapor-dominated systems at The Geysers in California and at Larderello in Italy; and the Geopressured system at Chocolate Bayou in Texas.

Wairakei, New Zealand

Wairakei is located on the North Island of New Zealand. It is situated on the west bank of the Waikato River and lies 8 km north of Lake Taupo (Figure 1). This liquid-dominated field occupies an area of 15 km^2 (Grindley, 1965), and extends about 5 km westward from the river over a relatively flat valley underlain by Taupo pumice alluvium. On the west, it is bordered by hills of Wairakei Breccia that rise 90-150 meters above the valleys and serve as a groundwater recharge area. No boundaries have been indicated towards north and south as evidenced by the behavior of the wells. The structure of this field is controlled by numerous fractures associated with the Wairakei, Kaiapo, and Upper Waiora faults (Grimsrud et al., 1978). The geology of the Wairakei field is described in Grindley (1965), Healy (1965), and Grange (1937). The reservoir engineering data have been compiled by Pritchett et al. (1978) and subsidence related studies are reported in Grimsrud et al. (1978), Viets et al. (1979) and Miller et al. (1980a, 1980b). A mixture of steam and water, in a ratio of about 1 to 4 by weight, is yielded by the Waiora Formation which is considered to be the main geothermal reservoir. Above the Waiora lies a relatively impermeable Huka mudstone. The Wairakei ignimbrites, considered to be practically impermeable, underlie the Waiora. The thickness of the Waiora Formation varies from about 366 m (1200 feet) in the west to more than 793 m (2600 ft.) in the east. The Huka Falls Formation, a relatively fine grained lacustrine rock, is less than 100 m (300 ft) towards southwest of the main production area and thickens to about 310 m (1000 ft) towards northwest and southeast.

Geothermal fluid production at Wairakei started in early 1950. The production increased significantly in 1958 with the commencement of power generation. A total of 141 wells were drilled in the field up to 1968 when drilling activity completely ceased. Of these, 65 bores account for about 95 percent of the total fluid produced from the entire field. It is believed that the reservoir was originally filled with hot water to the base of the Huka Falls formation before production started. Based on the early exploration measurements, initial temperatures and pressures at the sea level were about 250°C and 3965 kPag (575 psig). Data presented by Pritchett et al. (1978) indicate that initial temperatures in the upper part of the reservoir may have been 10°–40°C lower than in the deeper parts. Presumably, the hottest fluids were found in areas close to faults and fissures. In the early years of production (1958–1962), recharge to the reservoir was about 10 percent of the fluid produced. This inadequate recharge led to large pressure drops in the reservoir. For example, pressure drops of the order of 1725 kPag (250 psig) were observed in the western production area and over 2070 kPag (300 psig) in the eastern production area. However, recharge rose to about 90 percent of the fluid produced in the following period, leveling off at a pressure drop rate of less than 69 kPag (10 psig) per year. A total of about 1.05 trillion kg (2.33 trillion pounds) of fluid had been produced from the Wairakei-Tauhara region as of December 31, 1976. This large scale extraction of fluids has led to significant ground deformations in and around the Wairakei field.

Initial surface subsidence measurements were made in 1956 on the basis of bench marks established in 1950. Periodic measurements since then have shown that the area affected by subsidence exceeds 3 km^2. Subsidence at Wairakei has been reported by Hatton (1970), Stilwell et al. (1975), and subsequently, thoroughly reviewed by Pritchett et al. (1978). Observed vertical subsidence and horizontal ground movements are shown in Figures 2 and 3. As seen in Figure 2, the area of maximum subsidence occurs east of the main production zone and the maximum subsidence was of the order of 4.5 meters between 1964 and 1975. The horizontal movements, accompanying vertical subsidence, are represented by vectors in Figure 3. These vectors point toward the area of maximum subsidence. Also the observed horizontal deformations increase with increasing vertical settlement. A horizontal movement of about 0.5 meters can be observed near the zone of maximum subsidence in Figure 3. A plot of reservoir pressure drop versus subsidence at bench mark A-97 is shown in Figure 4. This figure shows that subsidence at Wairakei is characterized by a) an off-set subsidence bowl, b) a linear relation between reservoir pressure and subsidence up to 1963 and c) a non-linear relation after 1963.

As discussed elsewhere in this paper, surface and subsurface deformations may be expected to enhance the fault activity and the seismicity of the area. In a recent study Evison et al. (1976) found that both micro earthquakes as well as macro earthquakes were many times more frequent in the Taupo fault belt than either in the adjoining basins or in the Kaingaroa Plateau to the east. Nevertheless, to our knowledge, no such study exists which specifically relates subsidence with seismicity in the Wairakei area. It is difficult to assert at this time that increased seismicity in the Wairakei field is due to increased subsidence.

At Wairakei spent geothermal fluids with approximately 4400 ppm of dissolved solids are discharged directly into the Waikato river (Defferding and Walter, 1978). Since 1968, no new wells have been drilled in the field and about 140 MW of power is being steadily produced since then.

The surface deformations in the field have disrupted pipelines carrying steam, cracked drainage canals and caused the main road to sink by 2-meters (Viets et al., 1979). The recurring cost of repair of steam lines might range from $2000 to $10,000 per year. Fixing of drainage canal cost about $ 250,000.

Broadlands Geothermal Field, New Zealand

The Broadlands geothermal field, located about 28 km northeast of Wairakei is another liquid-dominated geothermal system in New Zealand (Figure 1). Its behaviour, however, appears to be considerably different from that at Wairakei, largely due to the presence of significant amounts of carbon dioxide gas. The New Zealand Electricity Department is expecting to produce 150 MW of electricity from this field by mid 1980's. The first 50-MW unit may be in operation by late 1983. The exploration in this area began in early 1960's and drilling activity started in 1965. As of 1977, a total of 32 wells had been drilled, of which only 16 are considered to be good producers (DiPippo, 1980). These are wells BR2, 3, 8, 9, 11, 13, 17 to 23, 25, 27 and 28 (Figure 5). The depth of the wells in this field vary between 760 and

1400 meters with one well (BR15) reaching down to 2418 meters. Over the field three thermal anomalies have been recognized (DiPippo, 1980). First is the Broadlands thermal anomaly covering an area of roughly 365 m radius centered around well BR7; second is the Ohaki thermal anomaly with an area of about 550-m radius is centered on well BR9 and third is an elongated area between wells BR6 and BR13 which extends about 1220 meters in north south direction. Of these, the Ohaki anomaly is associated with best production.

The geology of the Broadland's area has been extensively studied and reported in Grindley (1970), Browne (1970), Hochstein and Hunt (1970), and Grindley and Browne (1975). The subsurface formations in the descending order include: Recent Pumice alluvium, Huka Falls formation, Ohaki Rhyolite, the Waiora Formation, Broadlands Rhyolite, Rantawiri Breccia, Rangitaiki Ignimbrites, Waikora Formation, the Ohakuri group, and the Graywacke basement. The thickness of these formations is spatially variable. The Waiora Formation and Rantawiri Breccia are the two main aquifers which provide most of the fluid produced. The formations below the lower aquifer are quite dense and almost impermeable. The Huka Falls Formation and the Ohaki Rhyolite provide confinement to the Waiora aquifer while the Broadlands Rhyolite apparently acts as a boundary separating the two aquifers. The local disruption of alternating permeable and impermeable formations by faults and dikes provides steep channels, for fluid flow (Browne, 1970). The Ohaki and the Broadlands faults lie in the respective thermal areas. The lateral extent of the field, as determined from the resistivity surveys (Risk, 1975) is also shown in Figure 5. The resistivity of the region, enclosed by bars, is less than 5 ohm-m and the resistivity anomaly encloses an area of about 10 square kilometers. The boundary between hot and cold ground is essentially vertical down to a depth of at least 3 km.

During a period of five years between 1966 and 1971, a total discharge of about 34 billion kg (74 billion pounds) of fluid and 4.4 x 10^{16} joules (42 trillion BTU) of heat had been extracted from the Broadlands field. The entire field was almost shut down for over 3 years between August 1971 and December 1974. Initial temperatures of about 260°C and 300°C existed in the upper and the lower reservoirs before exploitation.

The response of the Broadlands geothermal field is noticeably different from that of a conventional liquid dominated system due to the presence of non condensible gases, mainly CO_2. The partial pressure of gases reduces the boiling point of water by 3°C at 300°C and by about 1°C at a temperature of 260°C (Macdonald, 1975). Thus, a two-phase region is expected to exist in the reservoir within a depth of about 2 km during the preproduction state. Standard hydrostatic pressures, as defined by Hitchcock and Bixley (1975), existed in the aquifers of the Broadlands system prior to production. Since the commencement of production in early 1966, the reservoir pressures have continued to decline until 1971 when the production ceased almost completely. During exploitation it was found that in the Ohaki area to the north the reservoir behaves as a single, interconnected unit, whereas Broadland area to the south is characterized by considerably lower permeability and contains several isolated pockets tapped by individual wells. According to Grant (1977), the communication between the Ohaki bores is also not perfect, as it takes about a year for pressure transients to propagate across the Ohaki region. This behaviour is quite different from that observed in the Wairakei field where the communication between wells is very good (Pritchett et al., 1978). Another remarkable difference observed between these two fields is in the size of the pressure drop. A pressure drop of as much as 1400 kPa (14 bars) was observed in some wells when exploitation ceased in 1971 in the

Broadlands field (Hitchcock and Bixley, 1975). For a comparable amount of fluid withdrawn, the pressure drop observed in the Wairakei field was quite small. Apparently, the presence of small quantities of carbon dioxide has played a major role in determining the response of the Broadlands field. The lower effective permeability as evidenced at Broadlands can be attributed to the presence of the two fluid phases, each of which impedes the flow of the other (Grant, 1977). According to this same author, the pressure drop in the reservoir is mainly the drop in the partial pressure of the gas, with a little drop in the steam phase pressure. After the 1971 shut in, pressures in the field started to build up with a recovery rate of about 173 kPa/year (25 psi/year) over a three-year period (Hitchcock and Bixley, 1975). Grant (1977) believes that this pressure recovery is primarily the recovery of gas pressure and the amount of pressure increase is the measure of the total amount of CO_2 recharge to the system.

Under these conditions it is reasonable to infer that in the absence of CO_2, the pressure behavior should have been similar to that of Wairakei. Following the pattern of drawdown and recovery, ground subsidence and rebound are also observed at the Broadlands field as described below.

An extensive precise level network was established in the Broadlands area during 1967-69 period. Approximately 500 bench marks were installed over an area of 65 square kilometers, covering a route distance of about 78 km. Initially a precise level survey was carried out in May 1968 which was then resurveyed in September 1969 and in March 1974. Local subsidence surveys were conducted in September 1969, December 1969, June 1970, January 1971, January 1972, February 1975, and February 1976. Total vertical subsidence observed between May 1968 and March 1974 is shown in Figure 6 which also includes the recovery in the ground levels during the shut-down period between 1971 and

1974. Thus, the subsidence shown in Figure 6 is in fact smaller than the maximum magnitudes observed up to August 1971. For example, a maximum vertical displacement of 220 mm was observed over the period February 1969 to January 1972 compared to 190 mm shown in Figure 6, indicating clearly that a rebound has occurred. It may also be noted by comparing Figures 2 and 6 that unlike at Wairakei, the subsidence bowl in the Broadlands occurs directly over the region of maximum discharge. By using the pressure data from Hitchcock and Bixley (1975) and subsidence data from the Ministry of Works and Development (1977), a plot of reservoir pressure drop versus subsidence has been prepared for bench mark H468A in the vicinity of well BR9 and is shown in Figure 7 for the period September 1969 to October 1973. For this figure September 1969 is taken as the datum, at which time a certain amount of subsidence and a pressure drop of about 600 kpa (6 bars) was already existing at well BR9 (Hitchcock and Bixley, 1975). A total subsidence of 6 mm took place at H468A between January 1972 and February 1975 (Ministry of Works and Development, 1977). Assuming a linear relation, we have calculated the subsidence for October 1973. Note from Figure 7 that the slope of the curve tends to change between January 1971 and January 1972 and drastically changes beyond January 1972. This is due to the rebound of the ground surface associated with rising reservoir pressures caused by the shut in of the field in August 1971. Horizontal displacements associated with subsidence have also been observed in the Broadlands field and are shown in Figure 8. As seen in this figure, the maximum movement over a six year period (1968-1974) is about 120 mm. A reinjection plan is underway in the Broadlands area to minimize subsidence effects. Reinjection tests were conducted on wells BR7, BR13, BR23, BR33 and BR34 for periods varying from a few weeks to 3 years (Bixley and Grant, 1979). In all cases, water, super-saturated with silica, was injected and it was found that the permeability of the injected formation either increased or remained the same. Thus, reinjection tests are quite encouraging at the Broadlands field and a full scale reinjection scheme might be coming forth in the future. The potential problems related to subsidence at this site include flooding by Wairakei rivers and the role of ground deformation in siting power houses and steam lines.

Broadlands geothermal field, which lies in the Taupo-Reparoa basin, has very little micro earthquake activity as compared with that in the Taupo fault belt where the activity is about two orders of magnitude higher (Evison et al., 1976). It appears that neither geothermal fluid production nor subsidence has any effect on the microseismicity of the Broadlands area.

Cerro Prieto, Mexico

Cerro Prieto geothermal field is the first liquid dominated system in North America to be exploited for electric power generation. It is located in the Mexicali valley in the Colorado River delta and is located about 30 km south of the border of Mexico and the United States (Figure 9). It occupies a relatively flat area of about 30 square kilometers and exhibits some surface geothermal manifestations such as mud volcanoes (5 cm to 2 m high), steam and gas vents, hot springs, boiling mud ponds and a 200-m high black volcanic cone known as Cerro Prieto, after which the geothermal field is named.

Geologically the Cerro Prieto field is underlain by deltaic sediments which are classified into two units, Unit A and Unit B. Unit A has a thickness of 600 to 2500 meters and contains nonconsolidated and semi-consolidated sediments of clay silts, sands, and gravels. Unit B consists of layered consolidated sediment shales and sandstone and is more than 2 km thick. The depths to the producing layers vary over the field between 600 to 900 meters and 1300 to 1600 meters to the west of the railroad track and between 1800 to 2000 meters and 2200 meters to 2500 meters to the east.

The structure of the Cerro Prieto field is controlled by numerous faults related to the San Andreas fault system. The locations of some important faults are shown in Figure 10. Hydrologically, these faults may or may not act as conduits for the influx of fluids from the basement. For example, the Cerro Prieto fault is believed to act as a western hydrologic boundary to the field while the Morelia fault acts as a leaky boundary to the north. The Delta-, the Patzcuaro- and the Hidalgo faults appear to act as conduits to fluid flow. The eastern boundary of the field is not yet well established. Based on geophysical data and interference tests it has been inferred that both Cerro Prieto I and Cerro Prieto II areas, lying on the west and east of the railroad track, respectively, are hydrologically interconnected. According to Mercado (1975), hot water in the eastern part of the field rises up and flows towards west.

Electric power generation in the Cerro Prieto field began in 1973. In April 1979 the capacity of the plant was doubled to 150 MW, as two new units came into operation (Lippmann and Goyal, 1980). Consequently, fluid production rate has increased from about 2.8×10^6 to 4.2×10^6 kg (2800 to 4200 tonnes per hour. Total heat and mass produced as of November, 1980 has been estimated to be 6×10^{13} kcal (2.4×10^{14} BTU) and 1.9×10^{11} kg (1.9×10^8 tonnes), respectively (Goyal et al., 1981). Figure 10 shows the location of over 60 deep wells that have been completed in the field. These wells produce a water-steam mixture, the weight ratio of which varies from well to well from 0.5:1 to 4:1. Under natural conditions, the waters in the producing strata are believed to have existed at or below the boiling point (Truesdell, A.H., 1980, personal communication). This view is supported by temperature logs which show that the highest temperature of the water has been equal to the saturation temperature corresponding to the hydrostatic pressure of the hot saline water. Enthalpies of the produced fluid were found to vary from well to well from about 200 kcal/kg to 450 kcal/kg. Temperatures of about 300°-310°C and pressure about 10,000 kPa (100 bars) are believed to exist in the field at a depth of about 1300 meters (Lippmann and Mañón, 1980). Some approximate locations of isotherms, expected to exist at different depths, are also shown in Figure 10. These profiles are likely to have changed because fluid production has been in progress since 1973. Pressure drops and temperature drops of about 500 to 2000 kPa (5 to 20 bars) and 10-15°C have been observed in some wells.

Subsidence associated with this large scale fluid extraction was anticipated to occur in the Cerro Prieto field. Therefore, the Direccion General de Estudios del Territorio Nacional (DETENAL) and the U.S. Geological Survey, jointly laid out the first network to measure horizontal and vertical deformations in the Mexicali Valley in 1977. The second survey conducted in 1978 and reported by Garcia (1980) show both uplift (max. 33 mm) and subsidence (max 28 mm) over an area extending from the U.S.A-Mexico border to the south of the field. According to these results, subsidence in the producing area was very small. Nevertheless, it must be noted that the interpretation of these results will depend upon the location of the chosen datum of zero subsidence. Horizontal contraction of about 31 μ strain/year in NW-SE direction and extension of 0.7 μ strain/year in NE-SW was also observed in the field during the second survey of 1978 (Massey, 1980). It is likely that the Cerro Prieto area might also be undergoing some tectonic deformations similar to those observed in the Imperial Valley. Zelwer and Grannell (1982) provide gravimetric evidence to the effect that between 1977 and 1981 approximately

45 cm of vertical subsidence has occurred east of the power plant, caused by the 6.1 magnitude earthquake of June 8, 1980. Their evidence also suggests that almost all the fluid withdrawn is replenished by recharge.

Increased fluid withdrawal at Cerro Prieto may cause changes in pore pressure, temperature gradients, volume and stress patterns which in turn may influence the seismicity of the area (Majer et al., 1980). Seismic studies have been conducted at the Cerro Prieto field since 1971 and have been reported by Albores (1980) and Majer et al. (1980). It was found that microearthquake activity in the production area was lower compared to that in the surrounding region. One explanation for this may be that the effective stress in the production zone is increased due to fluid exploitation. This leads to an increase in the effective strength of the rock against slippage, which in turn reduces the seismic activity in the production zone. The regional seismicity may be due to tectonic stresses rather than the geothermal activity.

As of August 8, 1979 reinjection was underway into well M9 with untreated water separated from well M29. Response of well M9 and that of peripheral wells is being monitored over a period of time. About 40,000 kg/hr (40 tonnes/hr) of approximately 165°C fluid is injected into an aquifer located between 721 and 864 m depth. The injection rate had decreased to about 25,000 kg/hr (25 tonnes/hr) by December 1979 (Alonso, et al., 1979). No subsidence-related damages have been reported from the Cerro Prieto geothermal field so far.

The Geysers, California, U.S.A.

The Geysers is a vapor dominated geothermal system and is the largest producer of geothermal electric power in the world. As of early 1982, Pacific Gas and Electric Company is generating approximately 960 MW of electricity from steam supplied by Union Oil of California, Magma Power Company, Aminoil U.S.A.

and Thermogenics, Inc. The Geysers field is located about 120 km north of San Francisco in the northern coast ranges of California (Fig. 11). Electric power production at The Geysers began in 1960 when a 12.5 MW generating plant went on line using about 114,000 kg (250,000 pounds) per hour of steam supplied by four wells. Since then power production at The Geysers has been steadily increasing through the addition of more wells to the production line. The field is being exploited by private companies and much of the reservoir performance data is not in the public domain. Good reviews of subsidence related literature of the Geysers are contained in Grimsrud et al. (1978) and Miller et al. (1980a,b).

The Geysers field, a tectonically active area, can be characterized by a series of generally northwest-trending fault blocks and thrust plates. It is underlain by four major geologic units; the Franciscan assemblage, the Ophiolite, the Great Valley sequence and the Clear Lake volcanics. The Franciscan graywacke, that has undergone slight to moderate metamorphism, constitutes the reservoir rock. These sandstones are very dense and have low permeabilities (≈1 md) and porosities (≈10 percent). The steam is thus expected to be confined to open fractures and fault zones, the presence of which have been confirmed by drillers' logs and tested cores. Two reservoirs are believed to exist at the Geysers: a small, shallow reservoir and a deep extensive one. The depth to the shallow reservoir, which has produced about 50 billion kg (110 billion pounds) of steam (Garrison, 1972), is about 640 m (2100 feet). The main deep reservoir is located between 760 to 1520 m (2500 to 5000 feet). These two reservoirs are in communication with each other at some locations while at others they are not. The vertical extent of the reservoir is estimated to be greater than 3050 m (10,040 feet), (Lipman et al,

1977) and the lateral extent is believed to be 4580 m by 4580 m (15,000 feet by 15,000 feet). The top of the reservoir is estimated to be near sea level in elevation.

Prior to 1968 the shallow reservoir was the source of steam for power production. But by the early 1970's most of the fluid was being produced from the deep reservoir that is believed to have an initial temperature and pressure of about 240°C and 3545 kPa (514 psi), respectively. It is speculated that a deep boiling water table may exist at a depth of about 4580 to 6100 m (15,000 to 20,000 feet) which supplies the steam to the producing wells. Noncondensable gases up to 2 percent by weight are also produced at The Geysers along with the steam. By 1975 there were 110 wells in the field, providing about 3.65 million kg (8 million pounds) of steam per hour to generate about 500 MW of electricity. At present, about 900 MW of electricity is being generated. Future plans to increase the capacity are underway. This commercial steam production has reduced reservoir pressures considerably and has caused land deformations. A pressure drop of 1240 kPa (180 psi) was observed in the reservoir between 1969 and 1977. The relative changes in the elevations of the ground surface in The Geysers area are shown in Figure 12. The maximum subsidence to about 13 cm has occurred in the area of maximum fluid withdrawal. It is interesting to note that the vertical changes in the vicinity of the power plants 9-10 are minimal, even though large-scale steam productions from these units since 1972. It was also observed that the reservoir pressure drops and the rates of subsidence were largest soon after the new sources of steam were put on line and they gradually diminished as recharge gradients reached steady state conditions (Grimsrud et al., 1978). The vertical displacements observed during 1973-75 and 1975-77 along section AA" (Figure 12) are shown in Figure 13. Two types of ground movements may be observed here.

First, a downward local tilt of about 3.5 cm towards WNW and second, a substantial subsidence in the areas of steam production overlapped by circles in Figure 12. A maximum subsidence rate of about 4 cm per year during 1973-75 decreased to about 2 cm per year during 1975-77 near power plant 5-6. A nearly uniform uplift from 1975-77 in the ESE area of the main production zone can either be due to the thermal expansions of the overburden in the newly drilled areas or may be attributable to the assumption of zero subsidence at bench work R1243. Reservoir pressure drop and subsidence along section AA' (Figure 12) are shown in Figure 14. As could be seen from this figure, the areas of maximum subsidence are the areas of maximum pressure drop. Horizontal displacement rates were found to vary from 1.5 cm per year in the areas of heaviest fluid withdrawal to 0.4 cm per year in the peripheral areas (Lofgren, 1978).

In an effort to reduce subsidence and extract maximum thermal energy, reinjection of the steam condensate back into the formation began in 1969 at The Geysers. By 1975, six wells were being used to reinject condensate of about 1.78 x 10^4 m^3 (4.7 million gallons) per day, about 25 percent of daily steam output, back into the reservoir between 720 and 2450 m (2364 and 8045 feet) depths. The injected condensate flows down by gravity to depths below adjacent producing wells, in situ steam pressure being less than the hydrostatic pressures. It is believed that the injected fluid can contribute to an increase in the seismicity of the region by inducing slippage along the planes of weaknesses. In addition, microearthquake activity can also be caused by volume changes due to fluid-withdrawal and subsidence (Majer and McEvilly, 1979). A comparison of the seismic activity during pre-production (1962-63) and peak production (1975-77) times showed that the regional seismicity (magnitude $\geq$2) in the area increased to 47 events per year in the latter period as opposed to 25 events per year in the former (Marks et al., 1979). Also the microearthquakes at The Geysers are strongly clustered around the regions of steam production and fluid injection. It appears likely that much of the present seismicity at The Geysers is induced by one or more of the following phenomena: steam withdrawal, injection of condensate and subsidence. The exact relationship between these phenomena is not firmly established, it is suspected that the slippage of fault blocks past one another due to subsidence movements may in part contribute to the microearthquakes. Or it is that subsidence may give rise to the formation and propagation of micro-fissures which may enhance micro-seismic activity. Initiation and propagation of microcracks is also attributed to the thermal stresses produced by circulating geothermal fluids and this phenomenon is termed "thermal stress cracking" (Nelson and Hunsbedt, 1979). The mechanisms, by which this increase in seismicity has occurred warrants further study. No environmental hazards are reported in the Geysers area due to land subsidence.

Larderello Geothermal Field, Italy

Electric power generation at the Larderello field began as early as 1913, making it the first geothermal field in the world producing electric power from geothermal steam (DiPippo, 1980). This field is part of an arch of high heat flow extending along the east coast of the Italian peninsula from Tuscany to Sicily (Mongelli and Laddo, 1975). The Larderello system contains many geothermal anomalies. As of March 1975, the producing anomalies included San Ippolito, Gabbro, Larderello, Serrazzano, Castelnuovo V.C., Sasso Pisano, Lagoni Rossi, Lago, Monterotondo and Molinetto anomalies. Figure 15 shows the location of the field and its various anomalies. The Larderello field extends over a distance of about 20 km from Monterotondo in the south to the Gabbro in the north. The area covered by this field is about 170 km^2 (Ceron et al., 1975). As of March 1975, a total of 511 wells were drilled in the region with an average depth of about 650 meters. Of these, 194 were connected to the production line and 9 wells were used as observation wells for reservoir engineering studies. Installed capacity at this time was about 380 MW with an average fluid production of about 15,000 kg/hr (15 tonnes/hr) per productive well at an operating well-head pressure of 100 to 800 kPa abs (Ceron et al., 1975).

The geologic map of the Larderello field along with a cross section view is shown in Figure 16. From a hydrogeological point of view, the lithology here can be grouped into three main complexes: The first is an impermeable cap rock complex made of outcrops of "Argille Scagliose" comprising shales, limestones, etc., and "Macigno" and "Polychrome Shales" overlain in places by clay, sand and conglomeratic sediments. The second is the Tuscan formation, which constitutes the principal reservoir and forms the circulation region for the endogenous fluid. It comprises radiolarites to evaporite deposits. The third is the basement complex, consisting of phyllitic-quartzitic formations, which is highly impervious where phyllites predominate but may be locally permeable where intercalations of quartzites and crystalline limestones are present. Because the cap rock is not continuous, the Tuscan Formation is exposed at some places and allows the geothermal aquifer to be recharged by rainfall. The steam produced at the Larderello field originates from the meteoric water that may have undergone either a deep regional circulation or a local shallow one (Petracco and Squarci, 1975).

Geophysical studies indicate that the reservoir is characterized by a distinct resistivity high of greater than 100 ohm-meters and is located at depths of less than 1000 m. Thermal gradients of the order of 300°C/km to a maximum of 1000°C/km at some places are found in the area. The accepted normal gradient for the area is about 30°C/km with a heat flow of greater than 3 HFU (heat flow units). At some places heat flows of 5 to 10 HFU's exist. The highest reservoir temperature and pressures encountered in the field are 300°C and 3000 kPa (440 psi), respectively. The geothermal fluid produced consists of dry saturated or slightly superheated steam and some noncondensible gases. The amount of noncondensible gases varies over the field from 1 percent to 20 percent by weight with an average of about 5 percent of which CO_2 is the dominant one. Temperatures, pressures and flow rates vary from well to well and from area to area. There has been a significant decrease in the mass flow rates and reservoir pressures over the decades of production. For example, wells 85 and Fabiani of the Larderello anomaly show a significant decrease in the mass flow rate over a 20 to 30 year period (Figure 17). This figure indicates that flow rates are apparently tending to a steady state. The productive wells in the Larderello field were first shut-in in 1942 but systematic measurements of relative pressures began only in 1955 (Celati et al., 1977). The water table data was used by Celati et al. (1975) to determine formation pressures during this period. It was found that the initial pressures in the Larderello region varied from 1960 to 3920 kPa (284 to 568 psi) and that these values were affected by nearby producing zones as a result of the expansion of the explored area. Pressure depth plots indicated that in different parts of the field both water dominated and vapor dominated systems had existed prior to intensive exploitation began. Reservoir pressure distributions in the Serrazzano area for 1970 are shown in Figure 18. The highest pressure downhole pressure formed boserved among all Italian steam fields was measured in the Travale field where a pressure of about 6000 kPa (about 870 psi) existed in new wells drilled from 1972 onwards. Such high pressures indicate that the wells in this field have reached a water dominated reservoir. Considering that Larderello is in much the same hydrogeological and thermal situation as the Travale field, it can be deduced that Larderello, in undisturbed conditions, might also have had similar pressures (Celati et al., 1975). Continued production from the field has led to a considerable drop in the reservoir pressures and water levels. Since 1963-64, water levels in the western part of the field have dropped by about 100 meters compared to the other parts where the drop is less than 50 meters (Celati et al., 1977). In a related study, Atkinson et al. (1978) calculated an initial pressure of 3920 kPa (570 psi) at the Serrazzano field and steam reserves of about 1.7×10^{11} kg (170 million tonnes).

To monitor the effects of the injection of liquid wastes on surface and ground waters, a reinjection program was started in the Larderello region during the early 1970's. About one fifth of the waste liquid is returned to the reservoir by reinjection through the wells at the periphery of the field and the remaining 80 percent is discharged directly into the local streams (Defferding and Walter, 1978) in spite of high concentrations of boron. Generally reinjection was successful. However, in one case cold reinjected liquids reached a production well.

No subsidence has been reported at the Larderello field, although production has been in progress for 60 years and on a relatively large scale for 30 years (Kruger and Otte, 1976). Microearthquake studies in the Larderello area do not seem to have been repeated. However, a few earthquakes of magnitude 4 in the Larderello area and one earthquake centered about 15 km east of Larderello have been reported. It seems that microseismicity could be recorded at the Larderello field if sensitive instruments are used.

Geopressured Systems

Several reservoirs of brine at high pressures and moderate temperatures are known along the Gulf Coast of Louisiana and Texas. Wallace (1979) estimates 17.1×10^{21} Joules of thermal and methane energy in place in the reservoir fluids underlying a 310,000 km^2 area of the northern Gulf of Mexico basin. Temperatures and pressures of most waters in these areas vary between 163-204°C and 69 to 103 MPa (10,000 to 15,000 psi), respectively, in the depth intervals of above 3.5 to 5 km (Wilson et al., 1974). Temperature gradients of about 20°-40°C/km exist in the Gulf Coast region in the upper 2 km. Geothermal gradients exceeding 100°C/km are found within and immediately below the depth interval where maximum pressure gradient change occurred (Jones, 1970). The depth to the top of the geopressured zone conforms in a general way with a 120°C isotherm which occurs in the depth range of a 2.5 to 5 km below sea level (Jones, 1970). The loss of load bearing strength due to thermal diagenesis which takes place between 80°C to 120°C is considered most responsible for creating the top of the geopressured zone. The location and depth of occurrence of the geopressured zones in the Gulf Coast region are shown in Figure 19.

In this paper we shall confine our attention to one geopressure field which is under exploitation. It is the Chocolate Bayou Field in Texas from which oil and gas have been produced since the early 1940's.

Chocolate Bayou, Texas

Chocolate Bayou is an oil and gas field in Brazoria County, Texas, and is located about 30 miles south of Houston (Figure 20). Oil and gas production from both normally and geopressured zones have been responsible for a land subsidence of about 0.6 m (2 ft) in this area. The Austin Bayou Prospect, a proposed geothermal exploration site, is located 8 km (5 miles) southwest of

Chocolate Bayou and has essentially the same geohydrological conditions. One should expect that the subsidence at the Austin Bayou prospect will not be much different from that at Chocolate Bayou under geothermal fluid production. The Chocolate Bayou field occupies an area of about 69 km^2 (25 mi^2) and has a surface elevation of 3 to 12 m (10 to 40 feet) above sea level with a gentle southeast dip. The formations from surface downwards include Pliocene to Holocene sand and clay beds in the upper 760 m (2500 feet), Miocene and Pliocene sands in the next 1200 m (4000 feet); and Oligocene and Miocene shales in the remaining 610 m (2000 feet) down to a depth of about 2650 m (8700 feet). Underlying these formations are the productive geopressured sediments, which might occur down to depths of about 4900 m (16,000 feet). The producing zones are underlain and overlain by thick shales and vary in thickness from less than 3 to more than 60 m (10 to more than 200 feet). The depth to the top of the geopressured zone varies from 2,440 m to 3,050 m (8000-10,000 feet). Numerous faults with no surface expressions exist in the field. Relative fault movement is believed to be responsible for the generations of abnormal pressure zones by bringing shales into contact with sands and thereby preventing the communication between upper and lower aquifers (Miller et al., 1980b). Faults are also believed to act as complete or partial barriers to the fluid flow (Gustavson and Kreitler, 1976).

Since the early 40's the Chocolate Bayou field has produced more than 4.5×10^{11} cubic meters (16×10^{12} standard cubic feet) of natural gas and 5.6×10^6 m^3 (35 million barrels) of oil, respectively. Gas wells have contributed an additional 6.5×10^6 m^3 (41 million barrels) of liquid hydrocarbons. Annual production of the field is shown in Figure 21. The annual production of brine from the field is not shown in this figure, because these data are not readily available (Grimsrud et al., 1978). The reinjection of brine since 1965, for the Phillips Petroleum Company's wells, is also shown in this figure. This reinjection is believed to be about 10 percent of the brine produced. As may be noted from this figure, brine is the predominant liquid produced at the field since 1965. Current total production of hydrocarbons is less than 8×10^4 m^3 (0.5 million barrels) per year and that of brine is more than 4.8×10^5 m^3 (3 million barrels) per year (Miller et al., 1980b; Grimsrud et al., 1978). Initial conditions of the reservoir vary within the field from one location to another. The producing zones in west Chocolate Bayou area are all normally pressured; in East Chocolate Bayou, both normally pressured and abnormally pressured zones are present; and in South Chocolate Bayou all zones are abnormally pressured. Pressure versus depth relationships in wells located west of the Chocolate Bayou area is shown in Figure 22. Based on well logs, a temperature gradient of about 3°C/100 m seems quite reasonable down to about 5000 m. Data presented by Bebout et al. (1978) for a well from the South Chocolate Bayou field indicates that bottom hole pressures had declined by 55 to 62 MPa (8000 to 9000 psi) during a ten-year period 1964 to 1974, although bottomhole temperatures remained fairly stable at about 162°C (323°F). The reduction in the formation pressures is expected to cause some land subsidence. In fact, bench mark K691 located in the Chocolate Bayou area has subsided .55 m (1.8 feet) since 1943 (Figure 23). Besides oil and gas production, ground water withdrawals and tectonic movements are also considered as potential causes of this subsidence (Grimsrud et al., 1978). Groundwater withdrawals between 1943 and 1974 have caused a subsidence of more than 2.13 m (7 feet) in the Houston-Galveston area and are believed to be responsible for some subsidence in the Chocolate Bayou area (Miller et al., 1980). Some estimates have been made on the component of Chocolate Bayou subsidence attributable to groundwater pumpage. It has been suggested that the subsidence of about 0.3 m (1 foot) out of a total of .55 m (1.8 feet) can be attributed to the groundwater withdrawal (Sandeen and Wesselman, 1973; Grimsrud et al., 1978). It may be noted from Figure 23 that the bench marks P53 and M691 located in the Chocolate Bayou field show a change of slope in late 40's and early 60's. This increased subsidence can be related to increased production of hydrocarbons as shown in Figure 21. Another interesting point may also be noted in Figures 21 and 23 that although the hydrocarbon production at the Chocolate Bayou field has been decreasing since 1964, the average rate of subsidence from 1964 to 1973 was greater than that from 1959 to 1964. This suggests either a lag-time of at least several years between extraction of deep fluids and the appearance of subsidence effects at the surface or a transition of the sediments from a state of over consolidation to that of normal consolidation (Holzer, 1980). No surface effects, such as faulting, ground cracking, disruption of well casings, damages to structures, etc., are reported in the Chocolate Bayou area due to ground subsidence.

Summary of Field Observations

There is clear field evidence from different parts of the world to confirm that geothermal fluid extraction can cause to vertical as well as horizontal displacements at the land surface. These deformations, which can cause significant damage to property, often show specific patterns of variations in space and in time, indicative of complex interaction of physical phenomena.

At Wairakei in New Zealand a well-defined subsidence bowl, with an area of about 1.5 km^2, has developed due to fluid production, with a maximum vertical displacement in excess of 4.5 m. Within this bowl, pronounced horizontal displacements directed towards the center have also been observed with a maximum magnitude of about 0.5 m. Spatially, the subsidence bowl is centered about 1.5 km ENE of the center of the main production area (Figure 2), presumably controlled by peculiarities of local geology. The annual rate of vertical displacement also shows a significant relation to time. An analysis of the subsidence at bench mark A97 in the SSW part of the bowl suggests that the rate was about 4 cm/year between 1953 and 1962. Around 1963 this rate underwent a marked increase and, between 1971 and 1974 had attained a magnitude of about 15 cm/year. An examination of the pressure drop at the same bench mark suggests that the marked change in the rate of subsidence is indicative of a marked increase in the compressibility of the material undergoing subsidence. Although recent studies indicate an increase in the micro- and macroseismic activity in the Taupo fault belt area (which passes to the West of the field in a northerly direction), it is difficult to decide at present whether this increase is to be attributed to the exploitation activity or to natural tectonism.

The Broadlands field, New Zealand, also exhibits conclusive evidence of subsidence associated with geothermal fluid production. Unlike at Wairakei, the subsidence bowl at Broadlands is centered almost directly over the production area, with a maximum displacement of the order of 0.2 m since the mid 1960's. The field was almost completely shut down during August 1971 and again during September 1974. Both these events were followed by a marked decrease or even a reversal in the direction of ground displacement. Between 1969 and 1972 the average rate of subsidence was of the order of 5 cm/year. As at Wairakei, the horizontal movements at Broadlands are also directed towards the center of the bowl, with maximum displacements of the order of 10 cm. Recent investigation indicate no noticeable seismic activities attributable to geothermal exploitations at Broadlands.

The liquid-dominated geothermal field at Cerro Prieto, Mexico, has been producing since 1973. Observations to date indicate that small vertical displacements, of the order of a few centimeters, have occurred over the field since 1977. However, these displacements cannot be conclusively attributed to fluid production. It is probable that the ground deformations and seismicity that have been measured recently in Cerro Prieto could be attributed by tectonic activity in this structurally active part of the earth's crust.

The Geysers field in Lake County, California constitutes an excellent example of a vapor-dominated system. Relative to a 1973 datum, vertical movements of as much as 13 cm have been observed in The Geysers up until 1977. The maximum subsidence rate was about 4 cm/year between 1973-75 and declined to about 2 cm/year near Power Plant 5-6. At the same time, a nearly uniform uplift of about 2 cm occurred during 1975-77 was also observed in the ESE part of the production area. There is evidence that the areas of maximum subsidence over the field are correlated with areas of maximum pressure drop. Since 1969, spent condensate fluids have been reinjected into the formation and provide pressure support to the reservoir. A comparison of the preproduction seismicity to that of the peak production period, 1975-77, shows that regional seismicity (magnitude ≥ 2) had increased from 25 to 47 events/year. In addition, micro-earthquake activity is strongly clustered in regions of production and injection.

Another well known vapor dominated system is the one in and around Larderello in Italy. Although this field has been under production since early 1940's, no noticeable subsidence has apparently occurred. A few earthquakes of approximate magnitude 4 have been reported. No data are available on the micro-earthquake activity in this region.

The geopressured geothermal systems of the Gulf Coast in Texas and Louisiana are currently under exploration. As such, no case histories are now available to evaluate their deformation behavior in response to fluid withdrawal. Some clues to their possible behavior, however, can be obtained by studying the deep oil and gas fields that have been exploited near the exploration sites. In the Chocolate Bayou oil fields of Texas producing oil and gas horizons are known between depths of 2650 and 4900 m (8700 to 16,000 feet). These zones may be either normally pressured or may be geopressured. Within the Chocolate Bayou area maximum subsidence of up to 0.55 m subsidence attributable to oil and gas production between 1944 and 1972 has been measured. It is believed that part of this subsidence could be due to shallow groundwater development.

PHYSICAL BASIS

As evidenced by field observations the important questions requiring consideration in analyzing subsidence due to geothermal fluid withdrawal are as follows:

1. The nature of the size and the shape of the subsidence bowl. Distributions of horizontal and vertical displacements.
2. The location of the subsidence bowl in relation to the location of the area of fluid production.
3. The variations in the time-rate of subsidence as a function of time.
4. Differential subsidence.
5. Fault movement and induced seismicity, if any.

The fundamental sequence of events leading to land subsidence is as follows: (1) fluid withdrawal causes reduction in fluid pressure; (2) fluid pressure reduction causes an increase in stresses on the rock matrix, accompanied by a reduction in the reservoir bulk volume; (3) the reduction of reservoir volume leads to the generation of a three dimensional displacement field within the reservoir and some deformation may also be induced by contractions associated with temperature declines; and (4) the reservoir displacements propagate to the land surface to cause horizontal and vertical ground displacements.

For purposes of analysis, it is convenient to distinguish between the reservoir proper where the displacements originate, and the overburden through which the reservoir displacements are merely transmitted. We shall define the "reservoir" to include those portions of the system from which geothermal fluids are drained (released from storage) to compensate for the fluids removed at the wells. Specifically, the reservoir includes the highly permeable horizons (the aquifers) as well as the slowly draining formations (aquitards or caprocks). The overburden, on the other hand, has little hydraulic continuity with the reservoir. Thus, there is no drainage of fluid from the overburden to make up for the geothermal fluids exploited.

The reservoir and the overburden basically differ in the manner in which they are subject to loading and deformation. The reservoir is subjected to loads originating from within the pores (<u>endogeneous loading</u>) whereas the overburden is subject to stresses or displacements imposed on its boundary (<u>exogeneous loading</u>). Stated differently, the reservoir is subject to drained loading while the overburden is subjected to undrained loading.

In practice, it may be hard to define the exact location of the reservoir-overburden interface. This boundary will obviously change its disposition with time, unless there exists a sharp impermeable contact between the two. Nevertheless, there is reason to suspect that results of overall analysis may not be very sensitive to the uncertainties inherent in locating this contact.

Deformation of the Reservoir

In a geothermal reservoir deformation may occur due to mechanical as well as thermal causes. Although a reasonable theoretical basis is currently available to discuss mechanically induced deformations, much remains to be done to properly explain the complex interactions existing between thermal and mechanical deformations that occur within a geothermal reservoir under exploitation.

In the following discussions on reservoir deformation, we shall restrict ourselves to subsidence caused by fluid withdrawal (that is, endogeneous loading). The duration of reservoir exploitation is considered to be much smaller than that over which tectonic stresses change and hence, we will treat the total stresses on the system to remain unchanged in time.

Mechanical Deformation: This phenomenon can perhaps be best explained by considering an elemental volume of the reservoir and its response to an imposed change of fluid mass at constant temperature. Such a change in fluid mass is indeed induced when geothermal fluid is mined from the reservoir. The mass of fluid contained in an arbitrary volume element is given by $M_f = V_v \rho_f S_f$, where M_f is mass of fluid, V_v is volume of voids, ρ_f is fluid density and S_f is fluid saturation. Consequently,

$$\Delta M_f = \rho_f S_f \Delta V_v + V_v S_f \Delta \rho_f + V_v \rho_f \Delta S_f \tag{1}$$

Of the three quantities on the right hand side of (1) the first, which denotes the component of ΔM_f arising due to pore volume change, is the phenomenon which is directly determines the magnitude of subsidence. The remaining two terms, which govern the dynamics of pore-fluid pressure change, indirectly contribute to subsidence. Therefore, in so far as the mechanism of pore volume change is concerned, we may focus attention on the first term on the right hand side of (1). Thus, only that term needs further consideration. As already stated, ΔV_v is caused by a reduction in the pore fluid pressure following removal of fluid mass. In addition to the change in void volume one has also to consider the change in the volume of the solid grains in order to evaluate the change in the bulk volume of the element. Note that it is the change in the bulk volume that controls subsidence. The basic ideas of combining ΔV_v and ΔV_s (where V_s is volume of solids) to compute ΔV has been discussed in detail by Skempton (1961). In accordance with Skempton's development, we can show that when the total stress is constant, that is, $\Delta\sigma = 0$,

$$\frac{\Delta V_v}{V} = -\,c\Delta p \qquad (2)$$

where c is the compressibility under drained flow conditions in which external stress is increased with no change in pore pressure (p). And:

$$\frac{\Delta V_s}{V} = c_s\Delta p \qquad (3)$$

where c_s is the bulk compressiblity of the solids. As justified experimentally (Skempton, 1961), the assumption inherent in (3) is that a medium subjected to the same magnitude of internal fluid pressure and external stress will behave as if the entire medium was made up of the solids. Based on the theory of elasticity, Nur and Byerlee (1971) provide a theoretical justification for this. In (2) and (3) a reduction in volume is associated with a positive sign. Combining (2) and (3):

$$\frac{\Delta V}{V} = - c\alpha\Delta p \tag{4}$$

where $\alpha = (1-c_s/c)$. Equation (2) pertains to volume change solely due to grain-grain slippage and (3) relates to volume change due to elastic compression of the grains. Physically, (4) implies that falling pore pressure will be accompanied by compaction due to grain-grain slippage and a dilation due to the expansion of the solids. Usually $c > c_s$ and hence grain-grain slippage will dominate the deformation process.

Furthermore (2) implies that for void volume change and bulk volume change, respectively, the following constitutive relations between effective stress, σ', and p are valid:

For void volume change:

$$\Delta\sigma' = - \Delta p \tag{5}$$

For bulk volume change:

$$\Delta\sigma' = - \left(1 - \frac{c_s}{c}\right) \Delta p = - \alpha\Delta p \tag{6}$$

When more than one fluid phase is present in a geothermal system (e.g., steam and water), the relation between change is fluid pressure and change skeletal stresses becomes more complex than (6). No published work, to our knowledge, addresses this relationship for a two-phase, stream-water system. The somewhat analogous problem of deformation of soils partially saturated by water and partially by air has been addressed experimentally by soil mechanicians. The discussions immediately below outline their findings.

Bishop (1955) suggested that (2) would need to be modified. Skempton (1961) generalized Bishop's ideas and proposed, for a porous medium with a water phase and an air phase,

$$\sigma' = \sigma - S_\chi p_w \tag{7}$$

where

$$S_\chi = 1 + (1 - \chi) \frac{p_a - p_w}{p_w} \tag{8}$$

in which p_w is the pressure in the water phase, p_a is the pressure in the air phase and χ is a parameter dependent on the capillary pressure $(p_a - p_w)$. If one assumes σ to be constant, (4) implies that (5) should be modified to:

$$\Delta\sigma' = - [S_\chi \Delta p_w + p_w \Delta S_\chi] \tag{9}$$

and (6) should be modified to

$$\Delta\sigma' = - \alpha [S_\chi \Delta p_w + p_w \Delta S_\chi] \tag{10}$$

More recent work by Fredlund and Morgenstern (1976) suggests that χ may be dependent on σ in addition to $p_a - p_w$. It is important to take note of the fact that the change in pore pressure in response to fluid drained will be governed by (2). The Δp controlled by (2) will then govern the solid volume change as in (3) or the bulk volume change as in (4).

In view of the foregoing, we may express the rate of change of void volume with reference to fluid pressure by the relation,

$$\frac{\Delta V_v}{\Delta p_w} = \frac{\Delta V_v}{\Delta \sigma'} \quad \frac{\Delta \sigma'}{\Delta p_w} = - S_\chi' \frac{\Delta V_v}{\Delta \sigma'} \tag{11}$$

where $S_\chi' = \left[S_\chi + p_w \frac{\Delta S_\chi}{\Delta p_w} \right]$. Noting that $c = -(\Delta V_v/V)(1/\delta\sigma')$, we may replace (11) by

$$\frac{\Delta V_v}{\Delta p_w} = S_\chi' cV \tag{12}$$

Skempton (1961) presented data indicating that c_s/c is very nearly zero in almost all unconsolidated sediments. But in rocks such as granite and quartzite it may attain values of 0.7 or more. Also, for fully saturated materials, $S_\chi = S_\chi' = 1.0$.

Very little is known about the nature of the χ and S_χ functions for steam water systems at elevated temperatures, although one would suspect that a steam-water system would obey an expression similar to (12).

Insofar as the phenomenon of fluid flow is concerned, it is the void volume-change, a scalar quantity, that is of critical importance. However, for purposes of subsidence analysis it is necessary to be able to evaluate the vertical and lateral displacements that accompany bulk volume change. How much of the horizontal displacement seen at the land surface is directly related to horizontal displacements in the reservoir is one of the important questions that needs resolution at the present time.

One of the simplest methods of converting volume change to displacement is to assume that due to the large lateral dimensions of geologic systems, horizontal strains are essentially negligible and that all volume change is caused by vertical displacements. If so, the change in the vertical dimension of a regular prism of crossectional area A will be given by, $\Delta h = -(c\alpha\Delta p)h$. Where h is the height of the prism. This is the assumption of one-dimensional consolidation theory, which has proven engineering validity in many field situations. In this case, c is usually obtained by testing samples in an oedometer or a device in which lateral strains are prevented.

For those field situations in which lateral strains may not be neglected, the general three dimensional deformation field accompanying volume change has to be considered. The boundary conditions obtaining in such systems cannot generally be duplicated in the laboratory and hence it is not possible to know c a priori. In this case, change in bulk volume is a function of linear displacements in different directions. For simplicity, if we consider a system with elastic, isotropic materials undergoing small strains, the change in bulk volume may be related to directional displacements by the relation,

$$\epsilon_v = \epsilon_x + \epsilon_y + \epsilon_z \qquad (13)$$

where ϵ_v is the volumetric strain $\Delta V/V$ and ϵ_x, ϵ_y and ϵ_z are linear strains in the direction of the coordinate axes. The task here is to evaluate ϵ_x, ϵ_y and ϵ_z based on the linear moduli of the material and the boundary conditions. In addition there also arise distortions of the elemental volume in addition to displacements. These distortions, caused by shear forces, do not contribute to volume change. If we assume the porous medium to be an elastic solid obeying Hooke's Law, then, three dimensional strain components, ϵ_{ij}, i = 1,2,3 can be related to the three-dimensional stress components σ'_{ij} through three material properties, Young's modulus, E, shear modulus G and Poisson's ratio, ν (see for example, Popov, 1968). For purposes of analyzing three dimensional deformation, the effective stress relations (5) and (6) may be generalized to (Garg and Nur, 1973):

$$\sigma'_{ij} = \sigma_{ij} - S_X \delta_{ij} p_w \tag{14}$$

for void volume change and

$$\sigma'_{ij} = \sigma_{ij} - \alpha S_X \delta_{ij} p_w \tag{15}$$

for bulk volume change.

It is widely known from experience that the parameters governing volume change, c and E, are often strong functions of effective stress, in addition to being dependent on the direction of the loading path and having a memory of past maximum loads. It is practically most expedient to treat these complexities of behavior by imposing the elasticity assumption over small ranges of stress increment. Accordingly we will restrict ourselves to the assumption of linear elasticity.

Having considered the phenomenon of deformation of our elemental volume, we now proceed to consider the forcing function that causes volume change. In the present case, the forcing function is the withdrawal of fluid from a geothermal field and the consequent reduction in fluid storage. The task then is to relate (12) to the dynamics of transient fluid flow in a deformable porous medium.

<u>The single-equation approach</u>: The simplest way to couple fluid flow and deformation is to assume that the boundary condition controlling deformation in the field can be reproduced in the laboratory (e.g., oedometer test) and

that the compressibility of the porous medium is known. In this case, the entire problem may be represented by a single governing equation,

$$\rho_w G + \nabla \cdot \frac{\rho_w k}{\mu} (\rho_w g \nabla z + \nabla p_w) = m_c^* \frac{\partial p_w}{\partial t} \tag{16}$$

where G is the source/sink term, k is absolute permeability, μ is viscosity, ρ_w is fluid density, z is elevation above datum, g is acceleration due to gravity, p_w is fluid pressure, and m_c^* is a generalized storage coefficient,

$$m_c^* = \rho_w [nS_w c_w + S_w S_x' c + n\, dS_w/dp_w] \tag{17}$$

where n is porosity and c_w is compressibility of water. When two phases are present, an equation similar to (16) has to be set up for the second phase. The solution of the above equation(s) merely leads to the evaluation of the bulk volume change. If one makes an assumption of the pattern of displacement (e.g., zero horizontal displacement) one can easily compute vertical displacements in the reservoir. Essentially such an assumption was used by Helm (1975) in his one-dimensional simulation of land subsidence. Narasimhan and Witherspoon (1978) combined this one-dimensional deformation assumption in conjunction with a three dimensional fluid flow field. This assumption is perhaps realized within the low-permeability, high-storage aquitards that may exist in geothermal systems. If this assumption is used, the horizontal displacements observed at the land surface will have to be explained solely in terms of the deformation of the overburden. Another limitation of the simple equation approach is that the stress field evaluation is treated in a perfunctory manner; that is, only the change in mean principal stress or the vertical stress is accounted for.

<u>The Coupled-Equation Approach</u>. A more comprehensive solution of the reservoir deformation problem requires that the physics of the problem be described in terms of two coupled equations: one for fluid flow and the other for porous medium deformation. These equations will have to be reinforced by an energy transport equation in geothermal systems, as we shall see later.

The coupled approach was originally propounded by Biot (1941) and later revised by him in 1955. Biot's approach has since been applied extensively in the fields of soil mechanics and rock mechanics (e.g., Sandhu and Wilson (1969), In the general three dimensional situation we need to separate out the change in void volume and the expansion of water in (16) and then rewrite equation as,

$$\rho_w G + \nabla \cdot \rho_w \frac{k}{\mu}(\rho_w g \nabla z + \nabla p_w) = \rho_w \frac{\partial \delta_{ij} \epsilon_{ij}}{\partial t} + \rho_w n \beta_w \frac{\partial p_w}{\partial t} \tag{18}$$

where ϵ_{ij} are the elements of the strain tensor and δ_{ij} is Kronecker Delta. Also, $\delta_{ij}\epsilon_{ij} = \epsilon_v$, the volumetric strain. Because the ϵ_{ij}'s are unknown in (18), we need to solve a second equilibrium equation which balances total loads. That is,

$$\frac{\partial \sigma_{ij}}{\partial x_j} + F_i = 0 \tag{19}$$

where σ_{ij} denotes the total stress tensor and F_i denotes body forces. In order to couple (18) and (19), we may express σ_{ij} in (19) in terms of ϵ_{ij} and p_w using appropriate consitutive, stress-strain laws. For an elastic isotropic material, σ_{ij} may be expressed as (Biot, 1941):

$$\sigma_{ij} = \frac{E}{1+\nu}\,\epsilon_{ij} + \frac{E\nu}{(1+\nu)(1-2\nu)}\,\delta_{ij}\delta_{kl}\epsilon_{kl} + \delta_{ij}p_w \qquad (20)$$

where E is Youngs modulus and ν is Poisson's ratio. The physical basis for fluid flow in a deformable porous medium is provided by (18) and (19) modified by (20). These equations are subject to appropriate initial conditions, boundary conditions and sources.

Thermal Deformation: Although the phenomenon of thermal expansion of solid materials is extremely well known, the role of thermal expansion in relation to fluid flow in deformable media has so far been treated only in a simplistic manner (see for example, Golder Associates, 1980). In this approach, the volume change due to temperature change is explicitly added to the volume change due to pore pressure change. In particular, an increase in temperature leads to an increase in the volume of the solid. Intuitively one would expect that part of the solid volume expansion will tend to decrease the pore volume while part of it will contribute to an increase in the bulk volume. Whether the ensuing decrease in pore volume will generate a pore pressure component or not will have to depend on the relative thermal expansivities of the solid and water. If we neglect these questions, the expansion of the solids due to an increase in temperature will affect bulk volume change in the same sense as a decrease in pore-fluid pressure. Thus, if β_s is the coefficient of thermal expansion of the solids defined as $\beta_s = 1/V_s\,(dV_s/dT)$, then the bulk volume change due to a simultaneous change of Δp_w and ΔT is (under conditions of constant σ):

$$\frac{\Delta V}{V} = -\,c\left(1 - \frac{c_s}{c}\right)\Delta p_w - (1-n)\,\beta_s\Delta T \qquad (21)$$

It is obvious that ΔT has to be obtained by solving a separate energy balance equation in addition to the fluid flow equation (19) and the force balance equation (20). In recent years many papers have appeared in the literature on the formulation of the energy balance equation in single phase and two-phase system. Notable among these are: Coats (1977), Garg et al. (1975), Mercer and Faust (1979), Pinder (1979), and Witherspoon et al. (1977).

In summary then, a physical description of the subsidence phenomenon accompanying geothermal fluid exploitation involves the simultaneous consideration of three coupled equations: one for the conservation of fluid mass; one for the maintenance of force balance and one for the maintenance of energy balance. These equations would need to be supplemented by appropriate relations between pore pressure on the one hand and skeletal stresses on the other as well as information on the compressibilities of the bulk medium and the solids and the thermal expansivity of the solids.

OVERBURDEN DEFORMATION

The overall effect of the reservoir deformation is that its interface with the overburden is deflected downward. In addition, points on this interface may also be subjected to some horizontal displacements. In response to these displacements on its bottom boundary, the overburden itself deforms, leading to vertical as well as horizontal displacements at the land surface.

It is clear at the outset that the overburden deforms essentially in an undrained fashion. The deformation of the overburden is therefore governed by a force balance equation such as (19), subject to a prescribed displacement boundary condition at the bottom and a zero stress, free surface boundary condition at the top. Additionally, the overburden may be constrained by other lateral boundaries such as faults and basin margins. The response of the overburden is largely governed by the ratio of its thickness to the radius of the deformed region as well as the properties of the materials constituting it. Where the thickness of the overburden is relatively small compared to the areal extent of deformation, the land subsidence observed will almost be a replica of the deformation pattern at the reservoir overburden interface. However, as the thickness increases, the displacements at the reservoir-overburden interface may be modified and attenuated before reaching the land surface.

One of the intriguing questions still unanswered concerns the nature of the mechanisms that cause horizontal displacements at the land surface; does the horizontal displacement at the surface definitely imply significant horizontal displacement in the reservoir? In principle it is conceivable that horizontal displacements may be caused at the land surface simply because of the elastic overburden system responds to the curvature of the underlying subsidence bowl. Such horizontal displacements could be accounted for by means of the force balance equation (19) already mentioned. However, Helm (1982) has been investigating the possible importance of horizontal deformations originating within the reservoir by treating the solid grains as constituting a viscous fluid. The chief difficulty in answering this question is that no field data are presently available on the variation of horizontal displacements with depth within the overburden. Indeed, a recent report by O'Rourke and Ranson (1979) indicates that not only are instruments non-existent at present to measure horizontal displacements along a vertical profile but that such instruments may not be available in the foreseeable future. Until sufficient field data is forthcoming, all the hypotheses that attempt to explain horizontal displacements will remain largely untested.

ROLE OF FRACTURES

Our discussion of physical bases has so far centered exclusively on porous materials. Yet there is reason to believe that many geothermal reservoirs (e.g., Wairakei, New Zealand, The Geysers, U.S.) are dominated by fractures. Additionally, the overburden may also be traversed by individual faults or a system of faults. Despite these physical realities, incorporating fractures into the physical basis is not easy. Fractures may control subsidence both at microscopic level, in terms of microfracturing and development of secondary porosity (Noble and Vonder Haar, 1980) or on a macroscopic scale in terms of differential subsidence across major faults as has been noted at Long Beach, California. In fractured reservoirs, fractures appear more to constitute highly permeable channels of fluid flow rather than constituting storage. Fractures may indirectly govern deformation by influencing the rate of fluid transmission and discrete discontinuities may contribute to deformation by acting as failure planes. However, it is doubtful whether the deformation of fractures themselves will contribute greatly to bulk volume changes in the reservoir. Apart from this qualitative reasoning, very little quantitative information is currently available in the literature to evaluate the role of fractures in the subsidence process.

RANGE OF VALUES OF PARAMETERS

Because of the difficulties associated with the collection of undisturbed samples from geothermal reservoirs and the difficulties associated with measuring physical properties of rocks under simulated in situ conditions, reliable data on physical properties relevant to geothermal subsidence are very limited in extent. Within the last decade, the U.S. Department of Energy has funded a few projects aimed at understanding the mechanism of subsidence in hydrothermal systems as well as geopressureed geothermal systems. As part of this effort physical properties of core samples from the Wairakei reservoir in New Zealand, East Mesa reservoir of the Imperial Valley in California; the Cerro Prieto geothermal system in Mexico and the Pleasant Bayou exploratory geopressured well in Texas have been measured. Even among these, only the East Mesa samples and the Cerro Prieto samples were subjected to elevated temperatures and pressures. The ranges of values, as evidenced by these studies, are as follows.

Wairakei, New Zealand: The producing formation at Wairakei is the Waiora Formation, which is a volcanic tuff. This is overlain by the Huka Falls Formation, a mudstone of lacustrine origin which in turn is overlain by the Pumice zone. Both permeability and mechanical properties of cores from these formations were measured at room temperature by Hendrickson (1976). The effective porosities of the aforesaid formations were as follows: Waiora, 35.6 to 41.6 percent, Huka Falls, 39 to 41 percent and Pumice, 48.8 percent.

The permeability of the Waiora Formation was found to be in the micro-darcy range and was found to be distinctly sensitive to effective stress. In the effective stress range of 5 to 15 MPa, absolute permeability of the Waiora was found to decrease from 50 microdarcies ($4.93 \times 10^{-17} m^2$) to about 10 microdarcies ($9.86 \times 10^{-18} m^2$). On unloading, the permeability was distinctly lower than it was during loading. A sample of the Huka Falls formation indicated a permeability of 63 microdarcies ($6.22 \times 10^{-17} m^2$).

The bulk compressibility of the Waiora formation was found to vary from 3.5×10^{-10} to $2.44 \times 10^{-9} Pa^{-1}$, with compressiblity decreasing with increasing confining pressure. The compressibility of Huka Falls mudstone varied from 4.5×10^{-10} to $1.2 \times 10^{-9} Pa^{-1}$. The Pumice was found to be far more compressible than the other two rock types with compressibility varying from 3.45×10^{-9} to $3.13 \times 10^{-8} Pa^{-1}$. The Waiora rock indicated a linear thermal expansion of $8.2 \times 10^{-6} m/m°K$ and specific heat of about 0.18 cal/g°C.

East Mesa, California and Cerro Prieto, Mexico: Recently, Schatz (1982) studied the physical properties of cores from East Mesa and from Cerro Prieto under elevated conditions of temperature and pressure and in the presence of fluids similar in chemical composition to the reservoir fluids. In addition, permeability and compressibility, Schatz also studied the creep behavior of the samples under elevated temperatures and pressures.

The observations, in regard to mechanical properties, are summarized in Tables 1 and 2. In all the tests, the samples were first carefully subjected to confining pressures and pore fluid pressures expected at the appropriate depth in the reservoir. Following this, the confining pressure was maintained constant and the pore pressure was dropped by 6.9MPa (1000 psi) to simulated pressure drop due to fluid production. The accompanying instantaneous strains were then measured. The compressibilities given in Tables 1 and 2 are the ratios of observed strains to the change in pore pressure. Both the East Mesa samples and the Cerro Prieto samples clearly exhibited increased deformation, when loaded beyond the preconsolidation stress level. The rebound compressibility varied between 50 to 75 percent of the virgin compressibility. Porosities of the rocks varied between 15 and 20 percent, these being functions of lithology as well as depth. The ranges in the compressibilities of rocks from both reservoirs are remarkably similar, varying between 6×10^{-11} to 3×10^{-10} Pa^{-1}, depending on lithology and conditions of testing. The uniaxial compressibility is generally higher than the corresponding isotropic case.

In order to verify the assumption of the Terzaghi effective stress concept, Schatz also conducted tests in which the confining pressure was increased by 6.9 MPa (1000 psi) rather than dropping the pore pressure by that amount, after attaining simulated reservoir conditions. He found that within statistical limits the strains observed in either case were approximately equal, suggesting that the concept of effective stress is indeed useful for geothermal reservoirs.

Considering the fact that hydrothermal systems are very active physically and chemically, one should expect that the deformation of the system to any imposed load would require time to equilibrate. Thus, compaction due to creep or, in other words, the dependence of strains on time at constant loads could be very important. Schatz (1982) addressed this issue experimentally as well as theoretically using cores from East Mesa and Cerro Prieto. He also monitored the chemical composition of the waters expelled during the creep tests to decipher the physico-chemical mechanisms accompanying creep.

The duration of the creep experiments varied from about a day to a maximum of about 9 days. The average long-term creep rate (over 4 days or more) was of the order of $1\text{x}10^{-9}\text{sec}^{-1}$ for East Mesa and $0.3\text{x}10^{-9}\text{sec}^{-1}$ for Cerro Prieto. Because of experimental considerations, longer term creep tests were not feasible. The results suggest that creep rate tended to decrease with increasing grain size and decrease with decreasing porosity. Under elevated temperatures and pressures, pressure solution effects exert significant influence in creep. As a result, less altered materials which are not yet fully in equilibrium with existing physico-chemical conditions are likely to creep more than already hydrothermally altered materials. For the 6.9 MPa (1,000 psi) pore pressure reduction imposed in the experiments the observed instantaneous compressibility was about $1.5\text{x}10^{-10}\text{Pa}^{-1}$ ($1\text{x}10^{-6}\text{psi}^{-1}$) for the East Mesa and Cerro Prieto rocks. Schatz (1982) estimates that assuming a long-term creep rate of $1\text{x}10^{-9}\text{sec}^{-1}$, the bulk strain could increase by a factor of two in a day and by a factor of ten in about three months over that suggested by the instantaneous value.

During the creep tests, permeability measurements were also made on the Cerro Prieto cores. The permeability at room temperature varied from 0.5 to 14.5 millidarcies (4.9×10^{-16} to $1.4\times10^{-14} m^2$) with an average of about 4 millidarcies ($4\times10^{-15} m^2$). Measured data indicate that temperature increase from room temperature to 150°C as well as creep cause permeability to change by a statistically significant 40 percent. However, permeability reduction merely due to pore pressure reduction was not significant. There is a possibility that pressure solution accompanying creep could selectively close throats connecting individual pores, thereby leading to permeability reduction.

In regard to permeability values it must be emphasized that effective field permeabilities in the field are likely to be larger due to the presence of interconnected macroscopic fractures and lithology changes, too large to be manifest in the small cores tested in the laboratory.

Pleasant Bayou, Texas: The U.S. Department of Energy has drilled two exploratory wells at Pleasant Bayou, Brazoria County, Texas. These wells reached down to a depth of 4775 meters. Core samples from these wells have been studied by Gray et al. (1979) in respect of instantaneous compressiblity and permeability and by Thomson et al. (1979) for creep behavior. All these studies have been carried out under room temperatures. Little is known about the possible mechanical behavior of these materials under the observed reservoir temperatures of about 160°C (320°F).

Data from 25 different depth intervals between 4478 and 4775 m indicate bulk compressibilities with pore pressure set up to atmosphereic varied from 4.4×10^{-11} to $5.1\times10^{-10} Pa^{-1}$ (0.3×10^{-6} to $3.5\times10^{-6} psi^{-1}$). Under conditions of elevated pore pressures, however, bulk compressibilities were somewhat lower, varying from 2.9×10^{-11} to $3.6 \times 10^{-10} Pa^{-1}$ (0.2×10^{-6} to $2.5 \times 10^{-6} psi^{-1}$). Gray et al. (1979) take this to be definitely indicative that the compressibility of grains cannot be ignored. The uniaxial compressibility for these samples varied from 5.1×10^{-11} to $2.2 \times 10^{-10} Pa^{-1}$ (0.35×10^{-6} to $1.5 \times 10^{-6} psi^{-1}$) and the porosities varied from 2 to 20 percent. Gray et al. (1979) report that all samples showed that deformation was stress-path dependent and that the materials stiffened noticeably with increased effective stress. Thompson et al. (1979) studied the creep behavior of some of the Pleasant Bayou core samples by holding the stresses constant for up to a third of a day and observing the dependence of volumetric and distortional strains as a function of time. They found that the volumetric behavior could be treated as that of a modified Kelvin body and that distortional behavior could be treated as that of a Maxwell material. The implication in relation to subsidence is that the effective long-term compressibilities will be much larger than the instantaneous, elastic values, leading to slower pressure declines and larger subsidence potential. The permeabilities of the sandstones from the geopressured horizons varied from 2 to 100 millidarcies (2×10^{-15} to $1 \times 10^{-13} m^2$).

In summary, the compressibilities of geothermal rocks are of the order of $1.5 \times 10^{-6} psi^{-1}$, which is about one-third the compressibility of water. Non-elastic and time-dependent deformation (creep) is a rule than the exception. There is reason to believe that pressure solution and reprecipitation effects may be significant in controlling creep behavior of geothermal rocks. The intergranular permeabilities of these rocks are in 1 to 100 millidarcy range and are significantly modified by temperature changes and porosity changes. The effect of temperature on absolute permeability of unconsolidated Ottawa silica sand was investigated by Sageev et al. (1980). They found that up to 300°F, absolute permeability did not depend on temperature.

DATA SYNTHESIS AND PREDICTION

The ultimate objective of geothermal subsidence analysis is to forsee the pattern and magnitude of subsidence for a given production strategy and then to modify production strategies suitably or prepare for alternate ameliorative measures to minimize adverse consequencies of subsidence. Predictive models used for this purpose vary widely in sophistication. Most of these models have been developed in the fields of petroleum reservoir engineering and hydrogeology. The simplest of these models is motivated by a need for engineering solutions in the absence of even a minimum amount of required field data. At the other extreme, highly sophisticated models have been developed within the past decade based on detailed theoretical considerations. Presumably, these models can make detailed predictions; but their important data requirements far exceed our ability to collect appropriate field data.

As already discussed under Physical Basis, the task of prediction entails two aspects: the deformation of the reservoir and the deformation of the overburden. Predictive algorithms could therefore be developed either separately for each aspect or a single generalized algorithm could be developed to handle both in a single frame work. Some of the algorithms that have appeared in the literature under each of these categories are briefly discussed below.

Reservoir Deformation Models

The simplest reservoir deformation models involves the direct application of Terzaghi's one dimensional consideration theory, neglecting the effects of thermal contraction. The key assumption here is that the reservoir is compressed vertically over a wide area so that lateral strains are negligible. Such an assumption is likely to be realistic in those situations where a) pore pressures decline over a large area in a well-field involving several production wells or b) when water drains vertically from a low permeability high storage aquitard to the aquifer characterized by high permeability and low storage over a large area. The assumption is likely to be unrealistic where pressure drawdowns are highly localized and strong spatial gradients in pressure drawdown exist. For the one-dimensional approximation, the vertical deformation of a prism of the reservoir material can be computed by

$$\delta H = - c_m H \delta_p \qquad (22)$$

where δH is the change in the prism height, c_m is the uniaxial compressibility, H is the prism height and δp is the change in pore-fluid pressure. The simplicity of the expression enables it to be used either for computing the ultimate deformation where δp is the ultimate pressure change or for computing in a more general fashion, the time-dependent variation of δH in a transient system. In situations where pore pressures may rise and fall, one could account for non-recoverable or non-elastic compaction by using either the normal consolidation value or the rebound value for c_m. Helm (1975) used this model with success to simulate the observed subsidence history near Pixley in the central valley of California over a 11-year period. His model has subsequently been used by the U.S. Geological Survey to analyze land subsidence due to groundwater withdrawal in other parts of the United States. Helm's approach consisted in modeling only the aquitard material as a one-dimensional, doubly-draining column, subject to prescribed time-dependent variations of fluid potentials at the boundaries.

The one-dimensional approximation of deformation has been used in a more general context by other workers. Following Narasimhan (1975), Lippmann et al. (1977) developed an algorithm in which the one-dimensional deformation approximation is used in conjunction with a general three-dimensional field of non-isothermal fluid flow. For computing fluid pressure changes, this algorithm accounts for the temperature dependencies of fluid density and viscosity and allows for variations of permeability in space or due to stress changes. To the extent that the assumptions are appropriate, this algorithm has the advantage of avoiding the need to solve the force balance equation.

Many numerical models have been proposed in the literature to simulate geothermal reservoirs, neglecting a detailed consideration of pore volume change in response to fluid withdrawal. The primary goal of these models is to follow the evolution in time of the fluid pressure and temperature fields (or equivalently, fluid density and internal energy fields) over the reservoir. Among these simulators one should include the following: the two-dimensional, areal, finite element model of Mercer et al. (1975); the three-dimensional finite difference model of Pritchett et al. (1975); the vertically integrated, two dimensional, finite difference model of Faust and Mercer (1979); and the three-dimensional, integrated finite difference model of Coats (1977) and Pruess and Schroeder (1977).

Although these models do not in themselves handle matrix deformation, they could be readily coupled with an algorithm designed to solve the stress-strains equation. A good example of such a coupling is the work of Garg et al. (1976) in which they coupled their finite difference heat-mass transfer model with a two-dimensional finite element model for static stress-strain analysis.

Of the algorithms mentioned above, those of Mercer et al. (1975) and Lippmann et al.(1977) are for single-phase liquid-dominated systems and the rest are for two phase, water-steam systems.

Overburden Deformation Models

The goal of overburden deformation models is to ignore the presence of fluid in the system and to solve the stress equilibrium equation over the system. The solution itself may be carried out with analytical techniques or through the use of more general numerical models.

Among the analytical techniques, two deserve special mention. One of these is the nucleus of strains method, variants of which have been developed by Gambolati (1972), Geertsma (1973) and others. Essentially these models involve the superposition of the fundamental exact solution of a uniform pressure drop within a spherical region in an isotropic, homogeneous, elastic half space. The superposition enables the handling of irregularly shaped regions. The second analytical approach involves the use of the more recent Boundary Integral Equation Method approach. In this approach, the required solution for the steady state problem is found by integrating the product of the boundary values and the normal derivative of the appropriate Green's function over the surface bounding the domain of interest, using numerical techniques.

Since the mid-1960's numerical methods, notably the finite element method, have been successfully used to solve the problem of static equilibrium in a loaded linear elastic continuum (Desai and Abel, 1972). Non-linear material properties can be included in such models by an iterative process using effective elastic modules. Pritchett et al. (1975) adapted such a model to simulate reservoir deformation and overburden response in a geothermal reservoir. Although not developed specifically for geothermal reservoirs, several two- and three-dimensional finite element deformations models are known from the soil mechanics and the rock mechanics literatures (e.g., Sinha, 1979). These could be easily applied to simulate overburden deformation in geothermal systems, provided that sufficient data is available to characterize the subsidence.

Coupled Reservoir-Overburden Models

The most sophisticated of all the approaches to modeling is, undoubtedly, the fully coupled approach in which the fluid flow equation, the energy transport equation and the force-balance equation are all simultaneously solved over the entire region extending from the land surface down to the base of the reservoir. Although no such model is available for geothermal systems (nor is one apparently warranted due to lack of data), algorithms do exist for isothermal systems in which the fluid flow equation and the force balance equation are simultaneously solved. Perhaps the earliest such model is that of Sandhu and Wilson (1969), which considers an elastic, fluid-saturated medium in the light of Biot's (1941, 1955) theory. A more recent example of a fully coupled model is that of Lewis and Schrefler (1978).

Comparison of Geothermal Subsidence Models

Recently, Miller et al. (1980b) carried out an in-depth comparison of a set of typical models available for simulating geothermal subsidence. Their study included reservoir models, deformation models as well as coupled models. In addition, to comparing the conceptual contents of the models, they also solved typical problems with different models.

One of their major conclusions is that the lack of suitable data precludes the need for highly sophisticated, fully coupled models. Indeed, in many cases full coupling may increase cost more than it does accuracy. Depending on the stage of activity, exploration, drilling, testing and development, model sophistication should be commensurate with quality of field data. Currently available reservoir and deformation models are conceptually adequate to handle field problems in relation to the inaccuracies introduced by lack of data. Miller et al. (1980) found that there is a greater need to have the algorithms in readily usable forms than to develop newer and more sophisticated models.

Some Simulation Results

In their model calculation studies Miller et al. (1980b) applied models of varying sophistication to simulate the observed subsidence at The Geysers in California and at Wairakei in New Zealand.

For The Geyser simulation, they assumed the following parameters: α, coefficient. of linear thermal expansion = 10^{-5} ft/°C; K, bulk modulus of the reservoir rock = 1.44×10^{-8} psf and a temperature versus pressure relation, $\Delta T = 9.31 \times 10^{-4} \Delta p$ where T is in degrees C and p is in psf. At the outset, they found that the thermal contraction at the Geysers may be over four times as large as contraction due to decreasing pore pressure. A comparative study of the Boundary Integral Element Method, the Nucleus of Strains Method, and simple, back-of-the envelope type of calculations, indicated that a good match with field observation could be obtained with any of the methods using appropriate assumption. However, if one wanted to increase model certainty, a major program of field investigations would be essential.

The Miller et al. simulation of Wairakei subsidence was restricted purely to deformation modeling; fluid flow was not considered. The required pressure-change profiles were obtained from Pritchett et al. (1978). The methods used for simulation included one-dimensional hand calculations, two-dimensional finite element calculations and three-dimensional nucleus of strain calculations. Although the simulations yielded overall similarities with several simulations, they could not match the pronounced localization of the subsidence bowl at Wairakei. Miller et al. concluded that Wairakei subsidence was dominated by inhomogeneity of pressure drops, strong variabilities in the thickness of beds or pronounced variations in material compressibility and that data was grossly inadequate to model these phenomena accurately.

It is pertinent here to cite two recent attempts at simulating Wairakei subsidence. Pritchett et al. (1980) have recently carried out detailed one and two-dimensional simulations of the Wairakei geothermal field, using a two-phase non-isothermal numerical model. In their approach, the authors start with the premise that 90 percent of the total reservoir compaction occurs within the permeable Waiora formation. In addition, they also assumed that a) the Waiora formation thickens towards the region of maximum subsidence and b) the late-time subsidence of the Waiora's is about 15 times larger than that at early times, apparently due to preconsolidation effects. Based on their simulations they conclude that the subsidence bowl lies close to the margin of the geothermal field and that local phenomena such as a seismic slippage along preexisting faults control the offset location of the subsidence bowl from the main production area. Based on available data and parametric studies, Pritchett et al. (1980) feel that pore collapse cannot adequately explain the peculiarities of Wairakei subsidence.

Narasimhan and Goyal (1979) carried out a preliminary three dimensional analysis of a Wairakei-type idealized system to investigate whether the offset of the subsidence bowl and the plastic deformation noticed in Wairakei could be explained in terms of a leaky-aquifer-type situation with the Huka Falls mudstone acting as an aquitard. By assuming variable aquitard thickness and suitable preconsolidation stresses, they were able to show that the observed patterns could indeed be simulated in space and time. As pointed out by Pritchett et al., the compressibility values used by Narasimhan and Goyal were effectively 6 to 12 times higher than what had been measured on a few core samples from Wairakei. Although Narasimhan and Goyal did not carry out a detailed analysis of the field data, their results did indicate that pore collapse, in the context of heterogeneities and variation of material compressibility does have a reasonable chance of explaining a major portion of Wairakei subsidence. As pointed out by Pritchett et al. (1980), very little subsurface data is available from the area of the subsidence bowl to resolve this question satisfactorily.

CONCLUDING REMARKS

That deformations of the land surface (vertical displacements, horizontal displacements, differential subsidence) may accompany geothermal fluid production under favorable hychogeological and exploitation conditions is very well documented. Such deformations are induced by volume changes in the geothermal reservoir caused by depletion of fluid storage as well as thermal contraction. Conceptual models do exist at the present time to explain the phenomena that are involved in the subsidence mechanisms. Compared to our ability to collect field data to characterize the subsidence history as well as the geologic system itself with adequate resolution, our conceptual models are exceedingly sophisticated. Even with the present technological revolution, the cost of collecting imput data to do justice to the resolution of sophisticated mathematical models appears to be excessive. It is also doubtful whether certain kinds of data such as the depth-wise change of stresses and horizontal displacements within the geologic system will ever become available in sufficient detail in the near future.

Under the circumstances, the best course of action appears to be to establish an adequate surface and subsurface, deformation monitoring system, so that measurements are made continuously as the system evolves in time from the exploration through the exploitation phase. There is a need to develop improved, economic measuring devices to measure deformations, especially as a function of depth. Without this valuable data, most of our sophisticated mathematical models will be practically useless, since they can never be validated in a credible fashion.

Mathematical, predictive models have an important role to play. During the early stages of development, when data is scarce, simple models can be used to predict a range of consequences and help decide whether a particular field could be developed in environmentally and economically acceptable fashion. Models should grow with a field as the field evolves in time and more and more data are accumulated. The status of modeling at present is such that even with adequate data-base, only short-term predictions (over a period of a few years) can be attempted.

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Table 1
Mechanical Properties of Reservoir Rocks from the East Mesa Geothermal Reservoir, California (from Schatz, 1982).
(Note: In all the tests external stresses were maintained constant and pore pressure discussed by Δp).

Rock Type	Depth m	Porosity %	Type of Test	Initial Values: Mean Effective Stress, MPa	Initial Values: Deviator Stress MPa	Pore Pressure MPa	Δp MPa	Compressibility Pa^{-1} Isotropic: Normal Compaction	Compressibility Pa^{-1} Isotropic: Rebound	Compressibility Pa^{-1}: Uniaxial	Temp °C
Med. Grain Gray Siltstone	1000	19.3	Isotropic Compression	13.3	0	9.5	6.9	1.6×10^{-10}	1.0×10^{-10}		150
		20.0	Isotropic Compression	10.7	3.8	9.5	6.9	2.8×10^{-10}	1.6×10^{-10}		150
		19.7	Isotropic Compression	8.2	7.6	9.5	6.9	3.0×10^{-10}	2.0×10^{-10}		150
		17.7	Uniaxial	10.6	3.8	9.5	6.9			4.4×10^{-10}	150
Fine-Grained Sandstone	1676	18.5	Isotropic Compression	22.8	0	15.9	6.9	7.3×10^{-11}	5.8×10^{-11}		150
		19.9	Isotropic Compression	18.4	6.5	15.9	6.9	2.2×10^{-10}	1.2×10^{-10}		150
		20.3	Isotropic Compression	14	13	15.9	6.9	3.4×10^{-10}	2.3×10^{-10}		150
		19.9	Uniaxial	18.4	6.5	15.9	6.9			1.3×10^{-10}	150
Very Fine-Grained Sandstone	2180	16.9	Isotropic Compression	29.3	0	20.1	6.9	1.2×10^{-10}	9×10^{-11}		150
		15	Isotropic Compression	23.8	8.4	20.1	6.9	1.0×10^{-10}	7.3×10^{-11}		150
		15.2	Isotropic Compression	18.2	16.8	20.1	6.9	1.5×10^{-10}	8.7×10^{-11}		150
		15.9	Uniaxial	23.8	8.4	20.1	6.9			1.7×10^{-10}	150

Table 2
Mechanical Properties of Rocks from the Cerro Prieto Geothermal Reservoir, Mexico (from Schatz, 1982).
(Note: In all the tests external stresses were maintained constant and pore pressure discussed by Δp).

Rock Type	Depth M	Porosity %	Type of Test	Initial Values: Mean Effective Stress, MPa	Initial Values: Deviator Stress MPa	Pore Pressure MPa	Δp MPa	Compressibility Pa^{-1} Isotropic: Normal Compaction	Compressibility Pa^{-1} Isotropic: Rebound	Uniaxial	Temp °C
Coarse-Grained Siltstone	2115	18.9	Isotropic Compression	26.5	0	21.4	6.9	8.7×10^{-11}	7.3×10^{-11}		150
		18.96	Isotropic Compression	15.9	15.9	21.4	6.9	1.0×10^{-10}	7.3×10^{-11}		150
		18.4	Uniaxial	15.9	15.9	21.4	6.9			1.3×10^{-10}	150
Fine-Grained Sandstone	2175	19.6	Isotropic Compression	27.2	0	22.1	6.9	1.2×10^{-10}	6×10^{-11}		150
		19.4	Isotropic Compression	16.2	16.4	22.1	6.9	1.3×10^{-10}	1.0×10^{-10}		150
		20.4	Uniaxial	16.2	16.4	22.1	6.9			1×10^{-10}	150
Very Fine-Grained Sandstone	2440	19.6	Isotropic Compression	30.4	0	24.8	6.9	8.7×10^{-11}	4.4×10^{-11}		150
		19.3	Isotropic Compression	18.1	18.4	24.8	6.9	7.3×10^{-11}	4.4×10^{-11}		150
		19.3	Uniaxial	18.1	18.4	24.8	6.9			1.2×10^{-10}	150

Figure 15. Location map of the Larderello geothermal field, Italy (from Atkinson et al., 1977).

Figure 16. Geologic map and cross section of the Larderello geothermal field (from ENEL, 1976).

Figure 17. Production history of well 85 and the Fabiani well, Larderello, Italy (from ENEL, 1976).

Figure 18. Spatial pressure distribution, Serrazano area, Larderello, Italy during 1970. Contour values are in kg/cm^2 (from Celati et al., 1976).

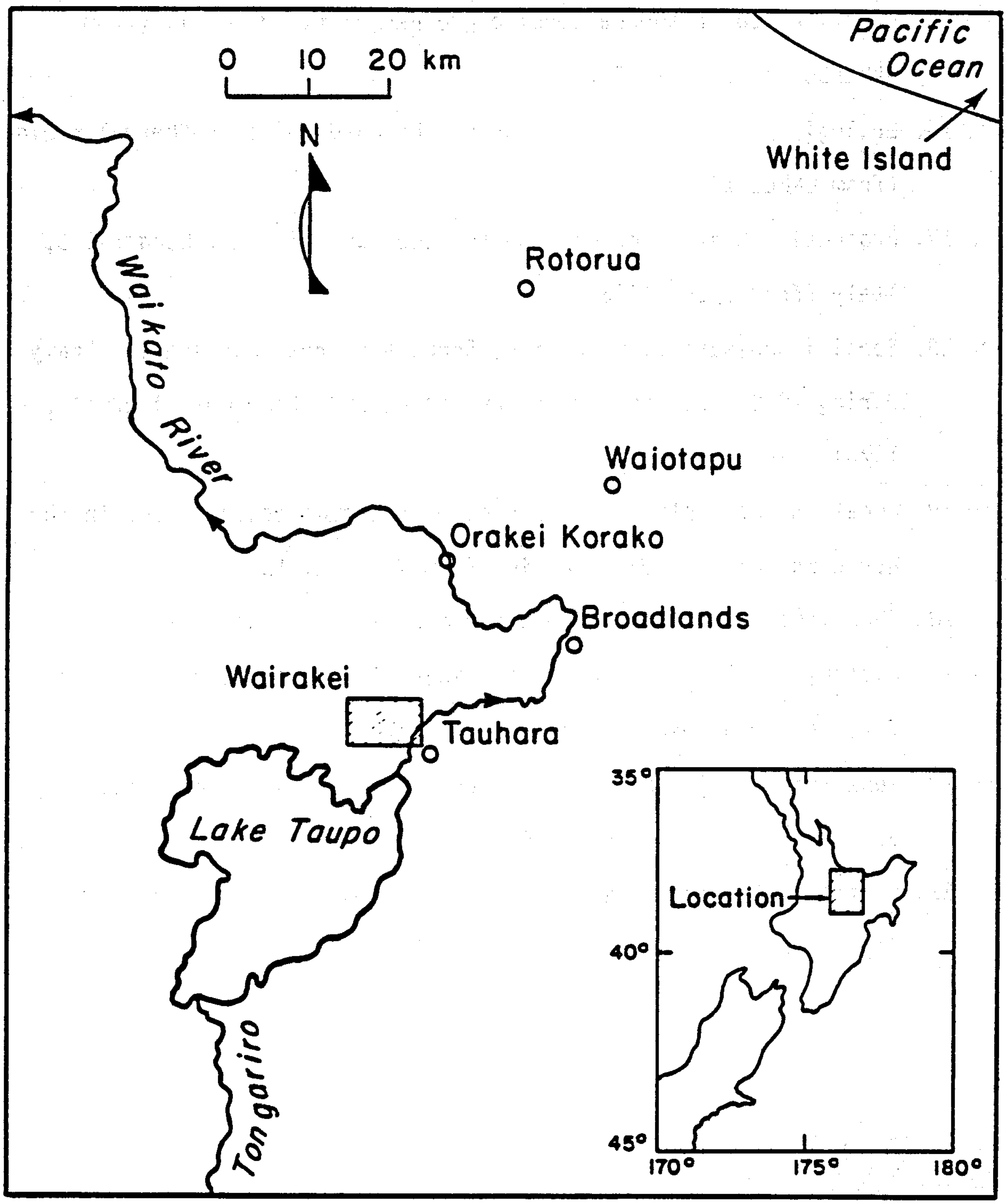
Figure 19. Location and depth of occurrence of the geopressured zone in the northern Gulf of Mexico basin (from Wallace, 1979).

Figure 20. Chocolate Bayou location map (from Grimsrud et al., 1978).

Figure 21. History of fluid production and injection, Chocolate Bayou Area, Texas (from Grimsrud et al., 1978).

Figure 22. subsurface fluid pressure gradients from wells in the Austin Bayou area, 8 km SW of Chocolate Bayou area (from Bebout et al., 1978).

Figure 23. Vertical displacements individual bench marks in the Chocolate Bayou area (from Grimsrud et al., 1978).



XBL 7912-13630

Figure 1. Location of the Wairakei geothermal field (from Mercer et al., 1975).

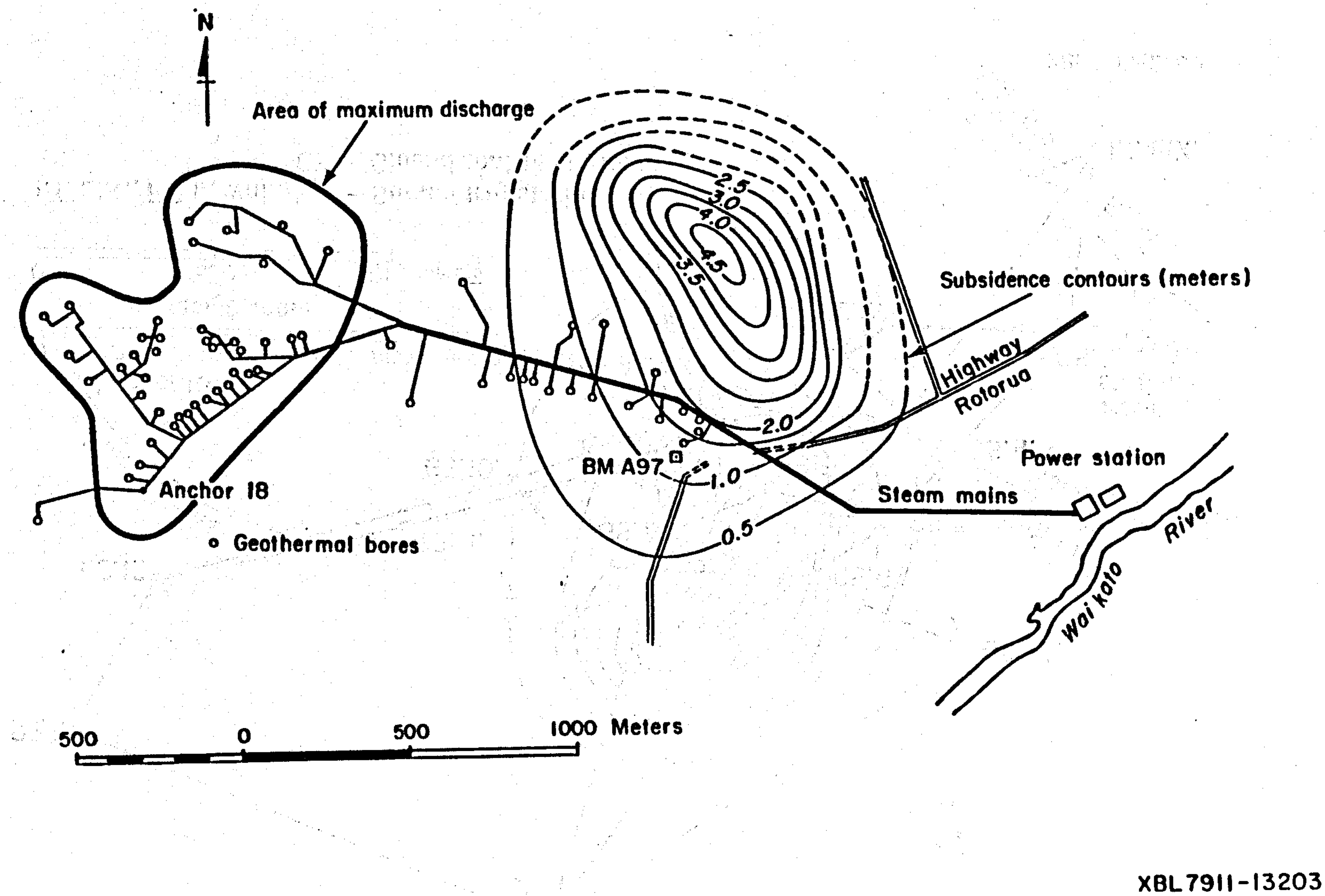


Figure 2. Vertical displacements due to geothermal fluid withdrawal at Wairakei, New Zealand, 1964-1975 (from Stillwell et al., 1975).

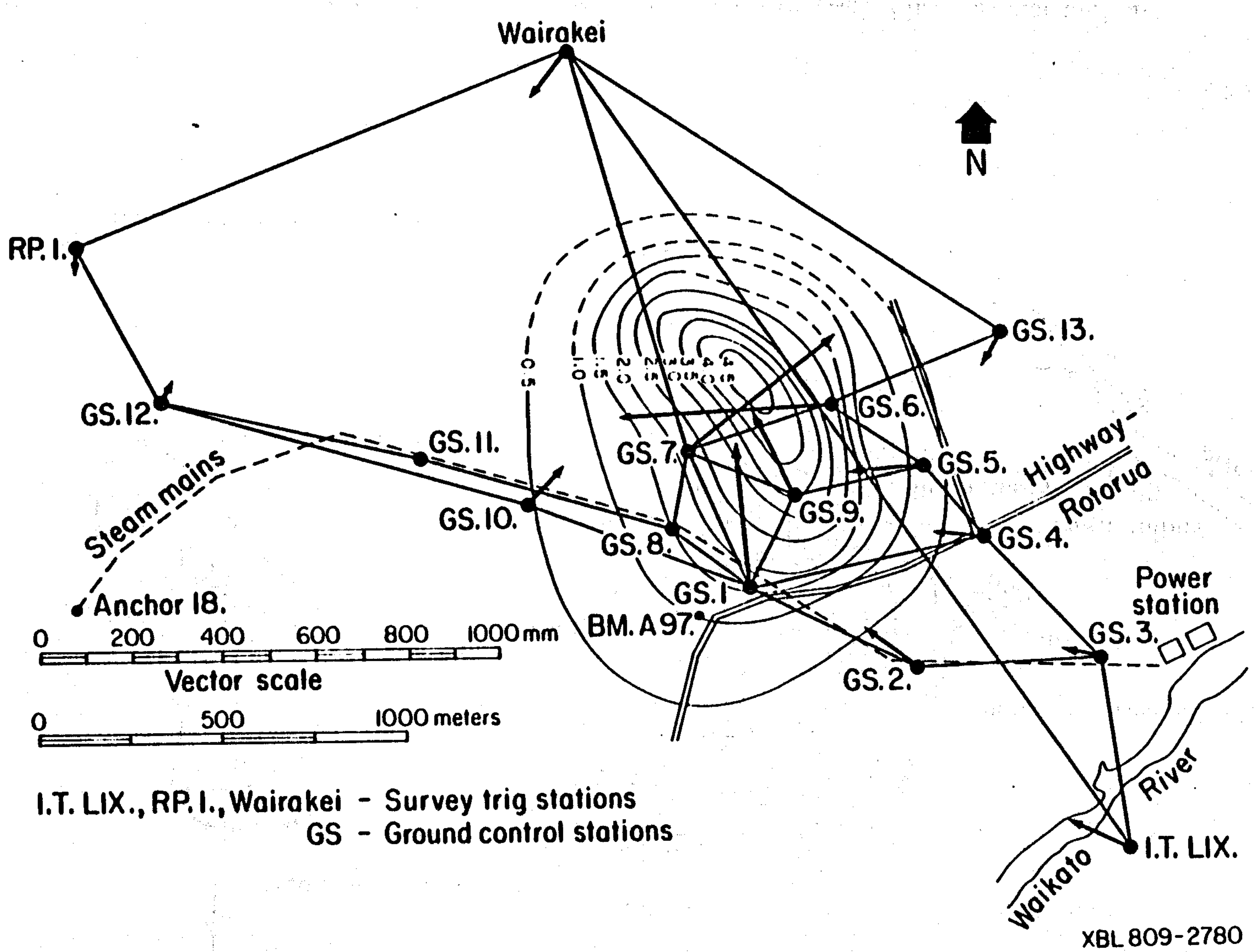


Figure 3. Horizontal displacements due to geothermal fluid withdrawal at Wairakei, New Zealand, 1964-1975 (after Stillwell et al., 1975).

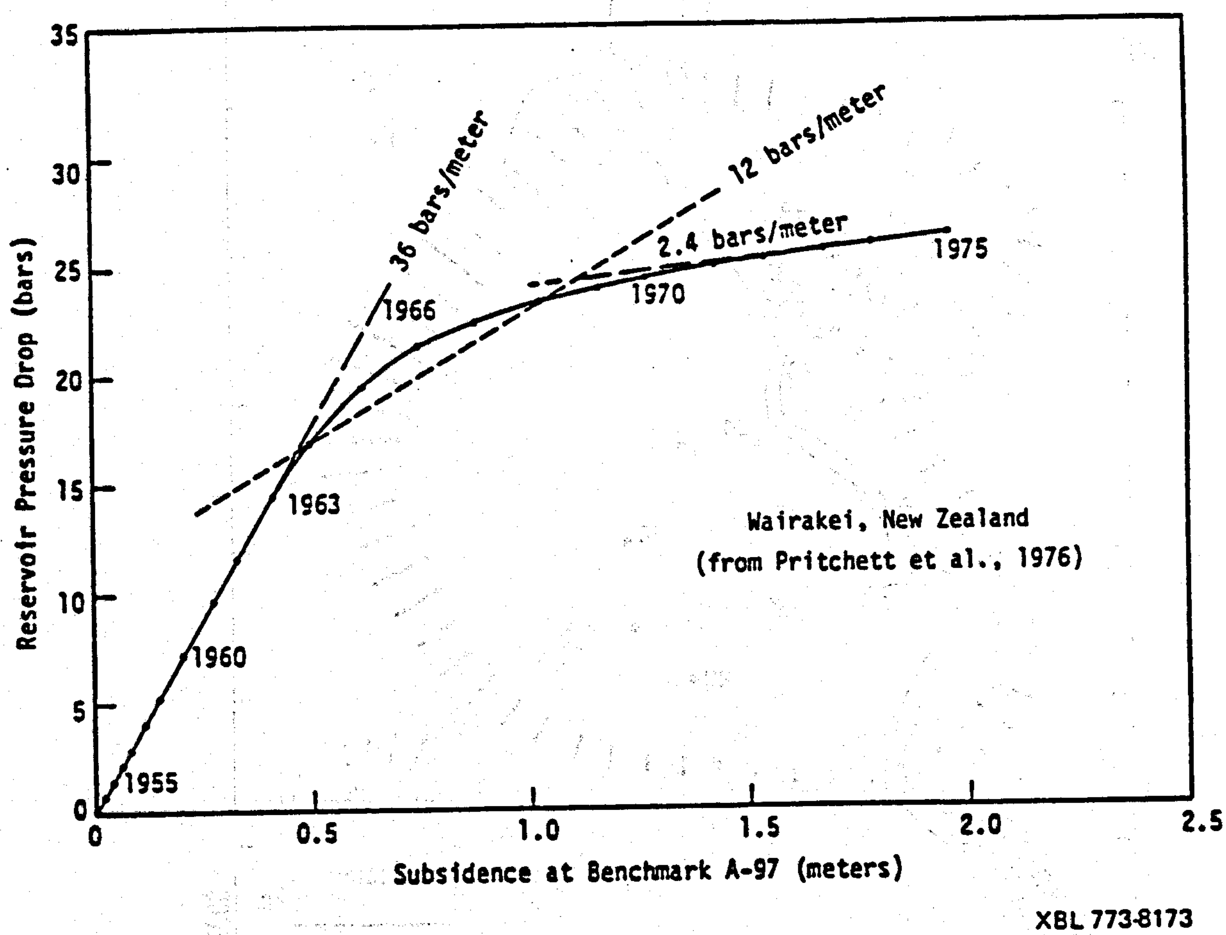
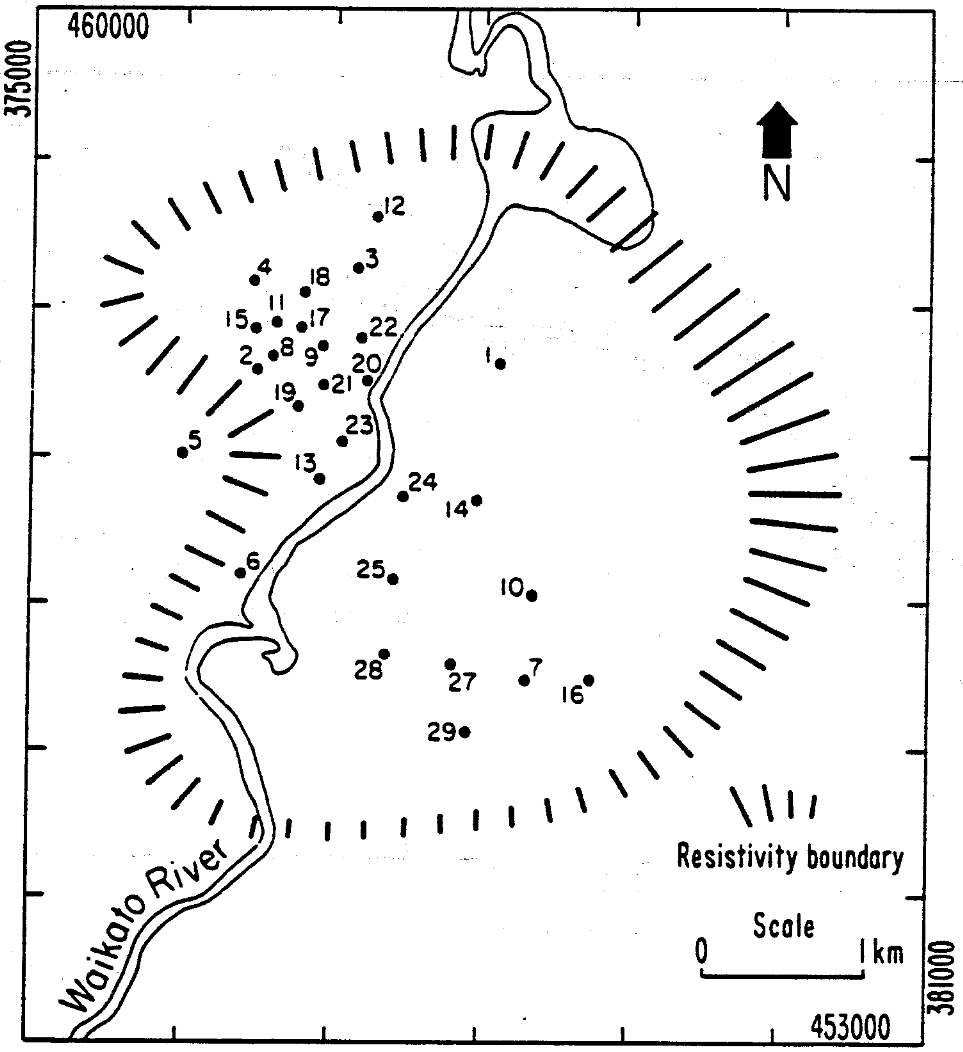


Figure 4. Relation between reservoir pressure drop and subsidence at Wairakei, New Zealand (from Pritchett et al., 1976).



XBL 809-2779

Figure 5. Locations of wells and resistivity boundary at the geothermal field at Broadlands, New Zealand.

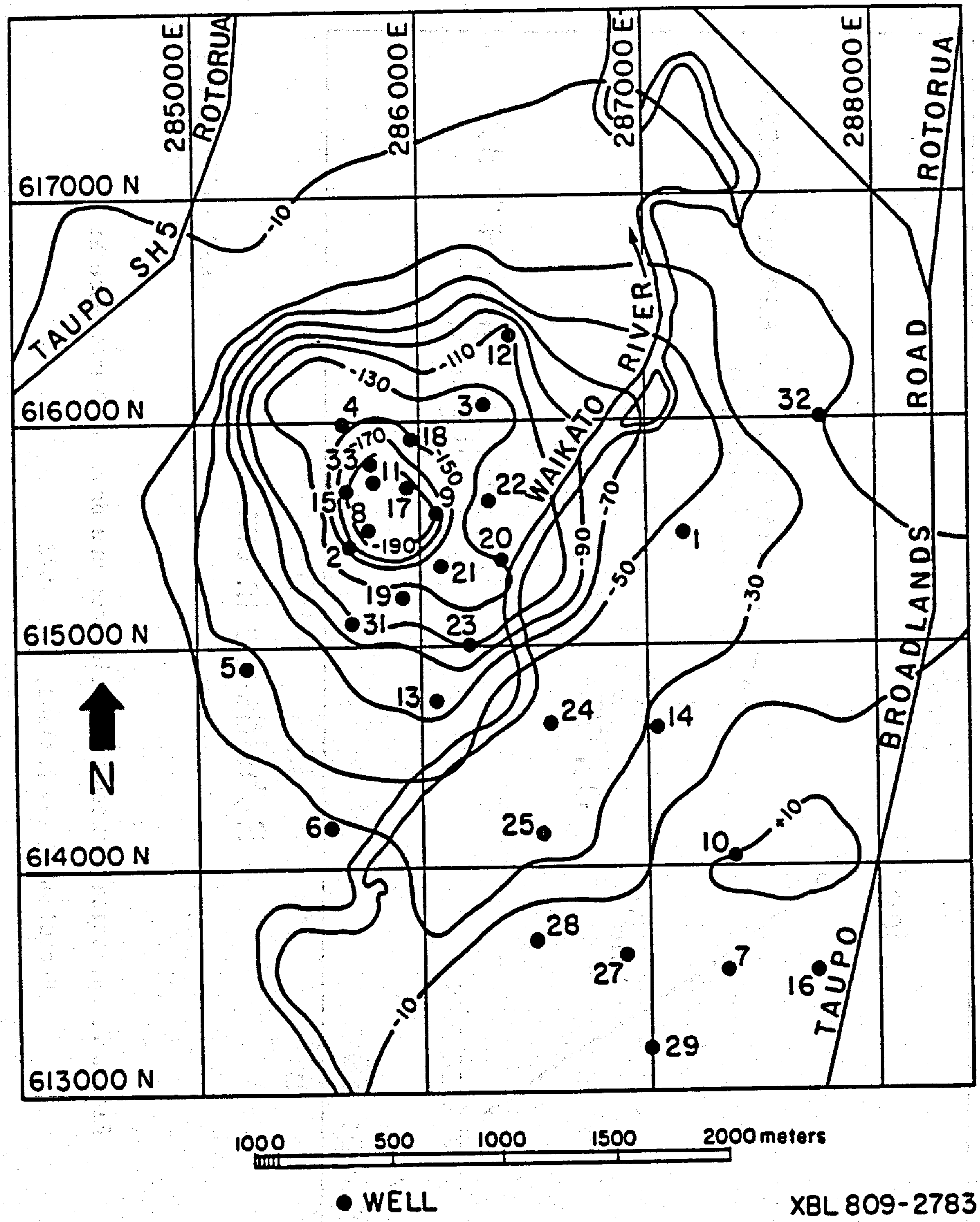
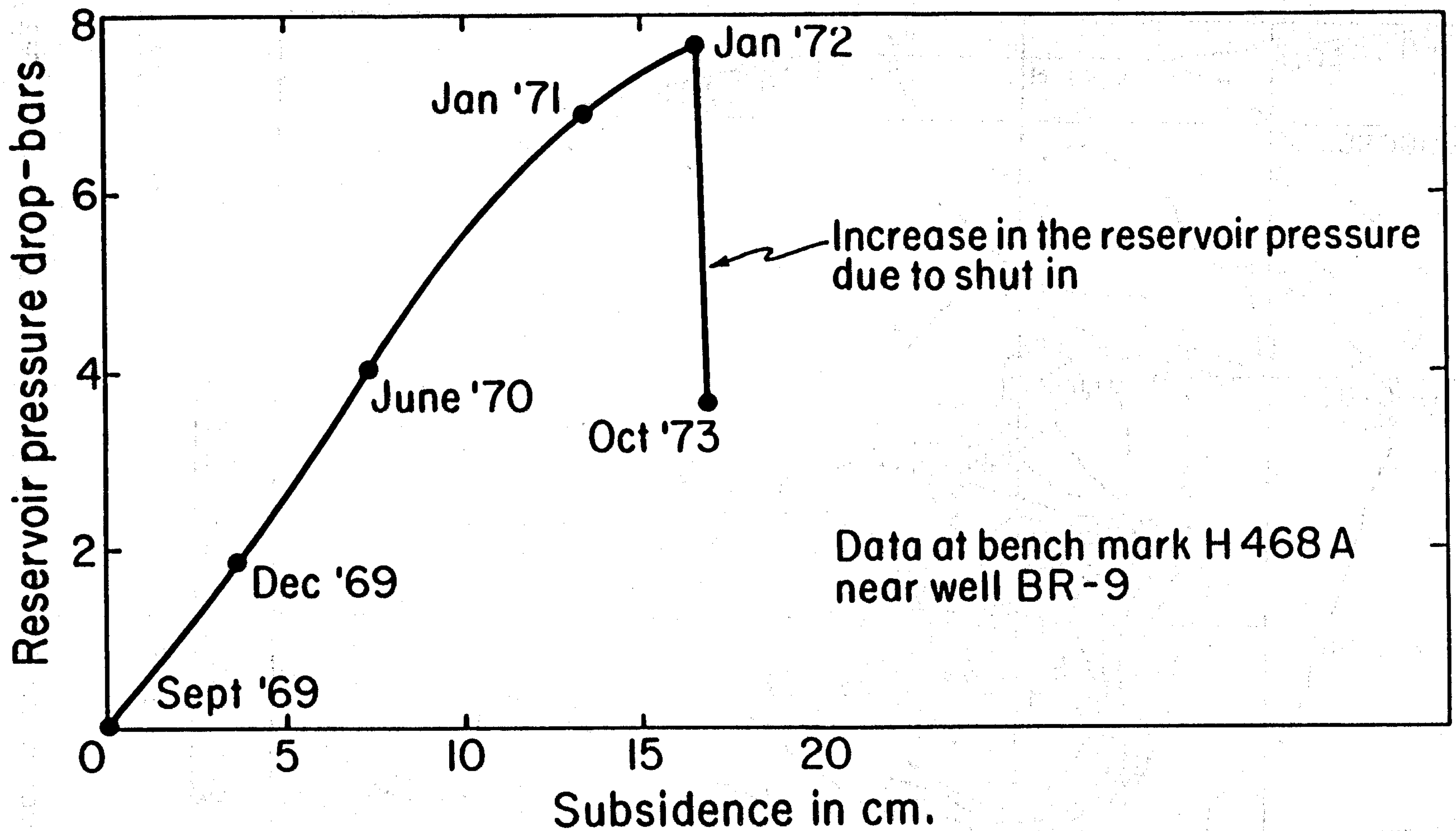


Figure 6. Subsidence at the Broadlands geothermal field between May 1968 and March 1974. Contour values are in mm. (from Ministry of Works and Development, 1977).



XBL 809-2782

Figure 7. Relation between reservoir pressure drop and subsidence at Broadlands, New Zealand, 1969-1973.

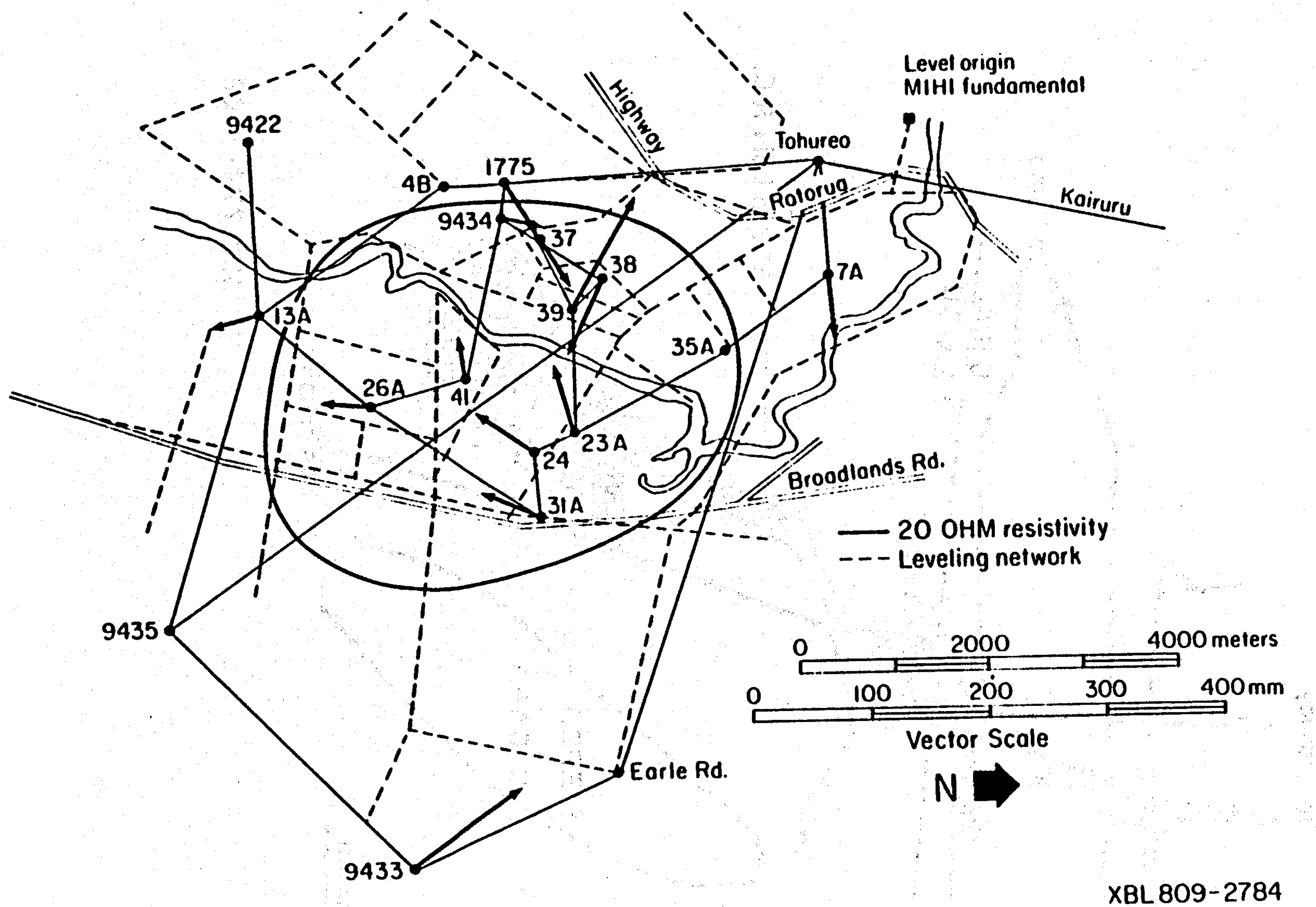


Figure 8. Observed horizontal displacements at the Broadlands geothermal field, New Zealand, 1968-1974 (from Stilwell et al., 1975).

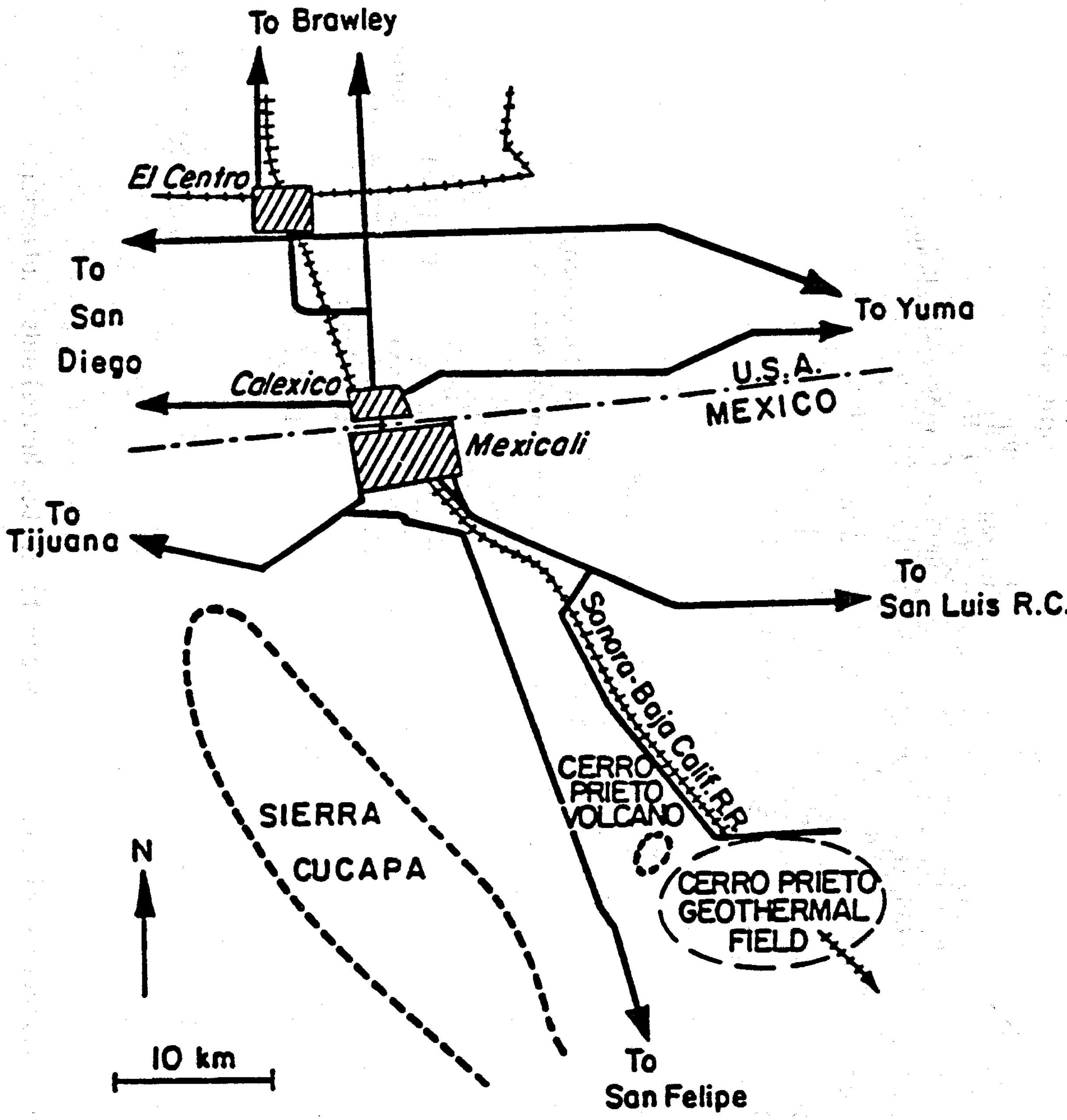


Figure 9. Location of the Cerro Prieto geothermal field, Mexico.

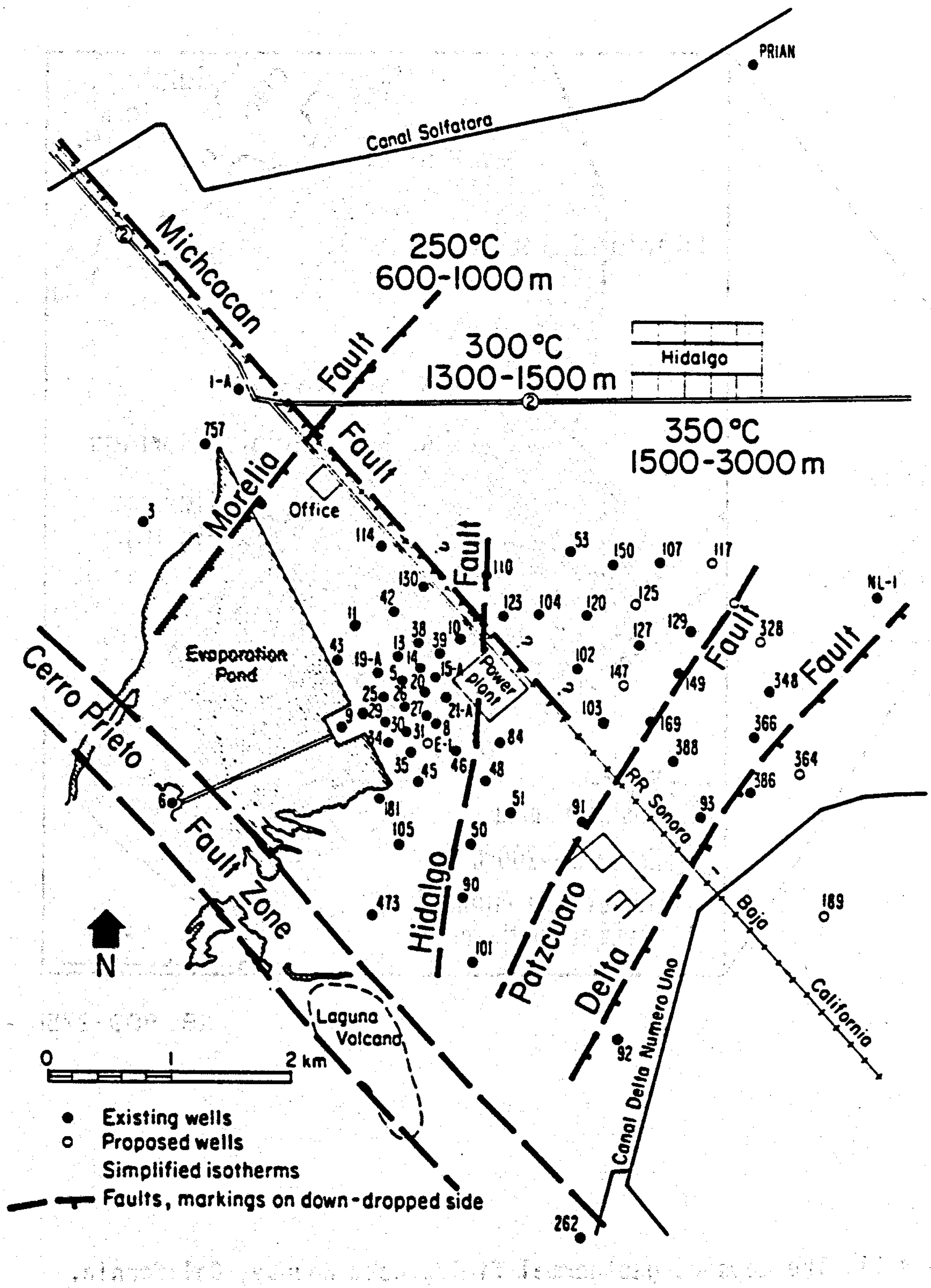


Figure 10. Location of major faults and wells, Cerro Prieto geothermal field, Mexico (from Lippmann and Manon, 1980).

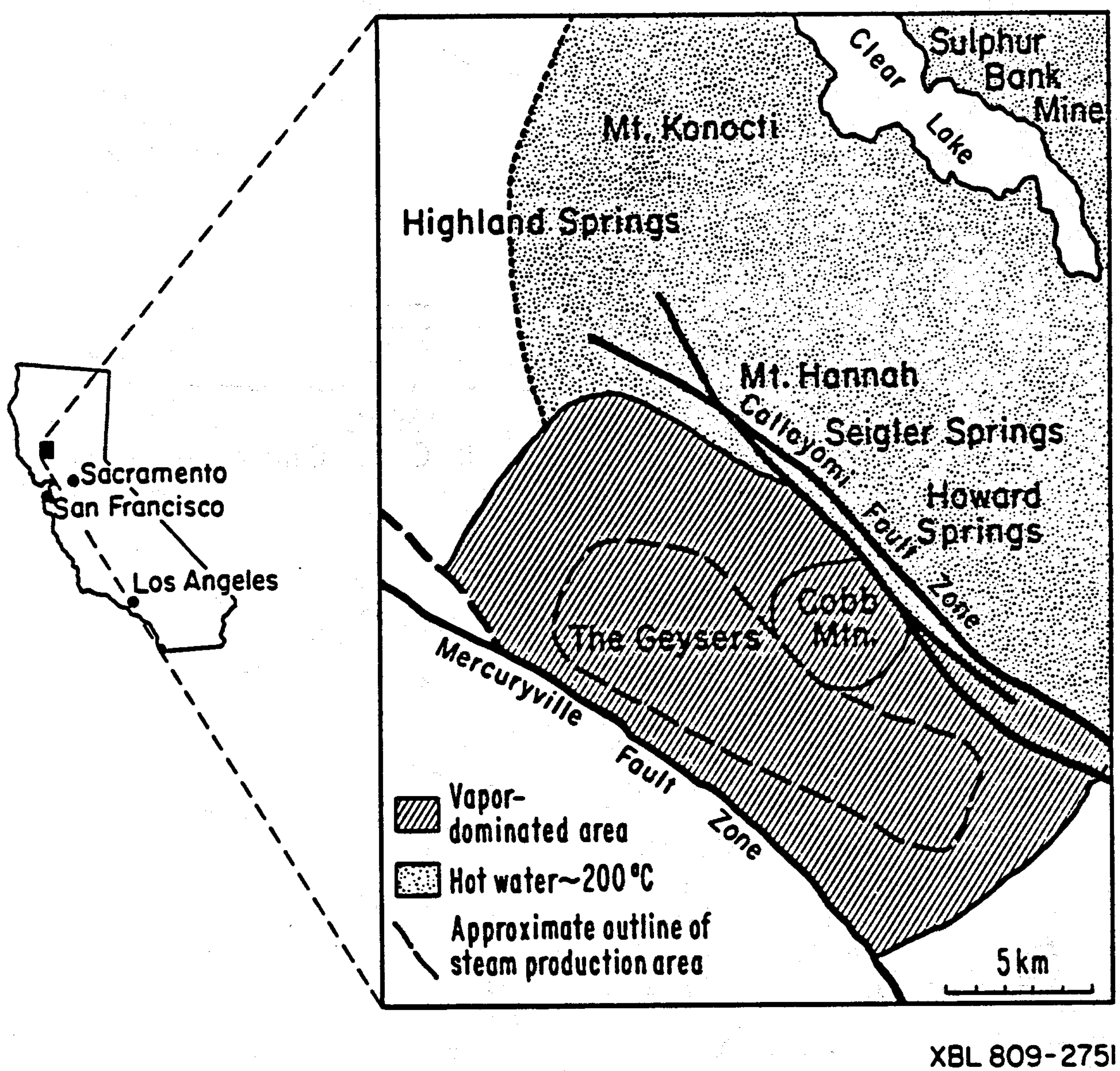
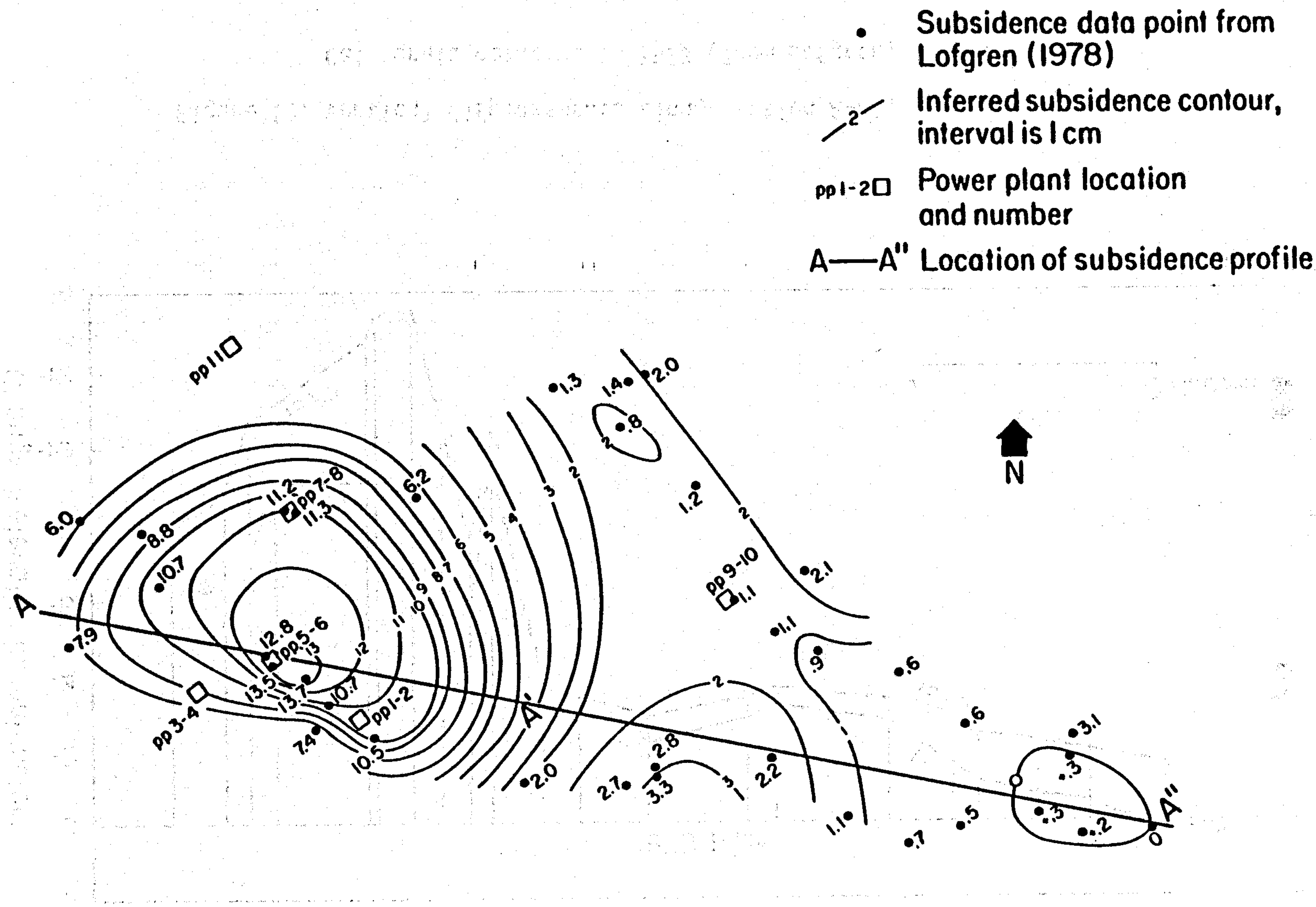


Figure 11. The Geysers geothermal field, Lake County, California.



XBL 809-2781

Figure 12. Vertical displacement field, 1973 to 1977, The Geysers, California. Circles denote areas of steam production supporting the power plant at the center (from Grimsrud et al., 1978).

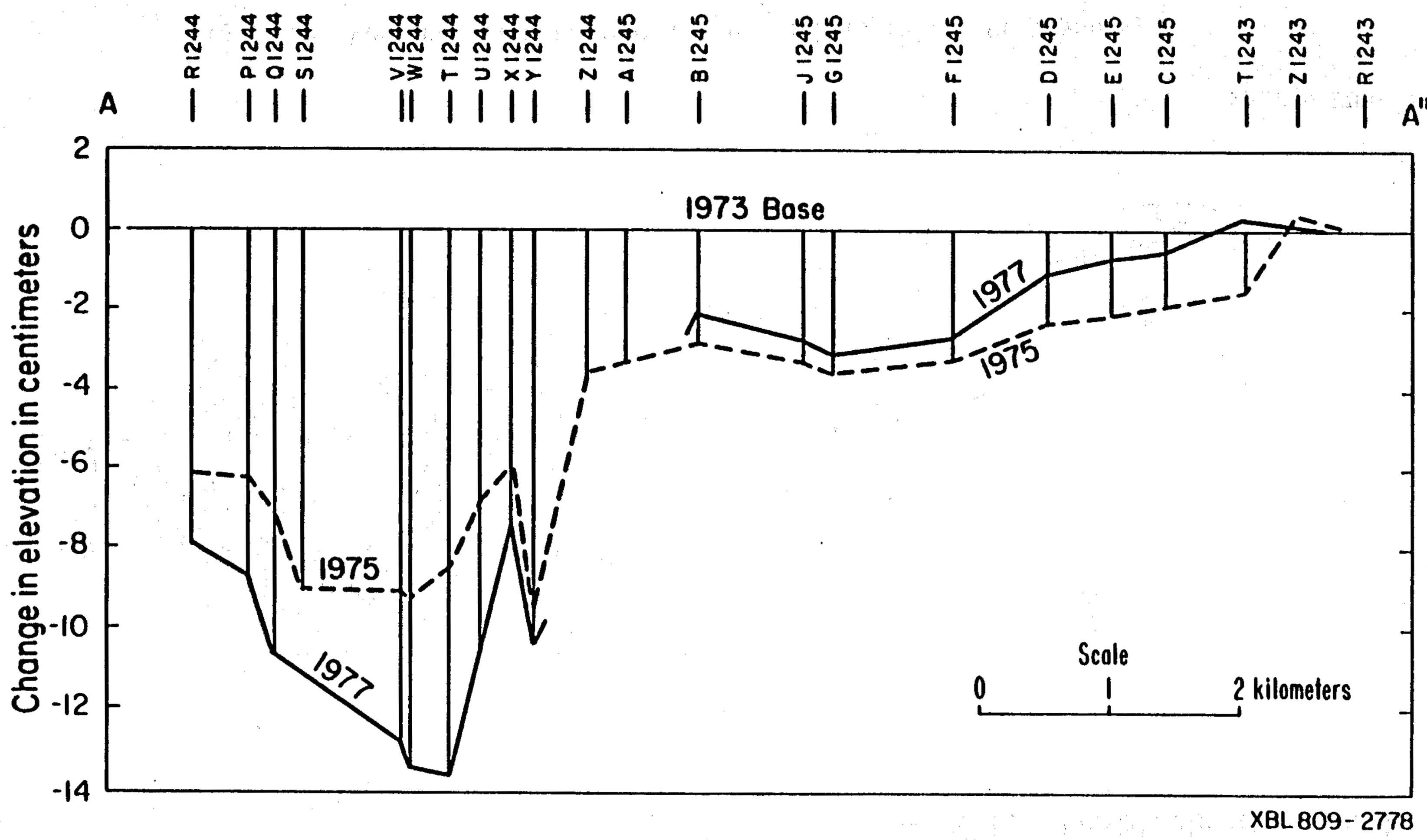


Figure 13. Vertical displacements along section A-A', the Geyser area, California relative to 1973 (from Lofgren, 1978).

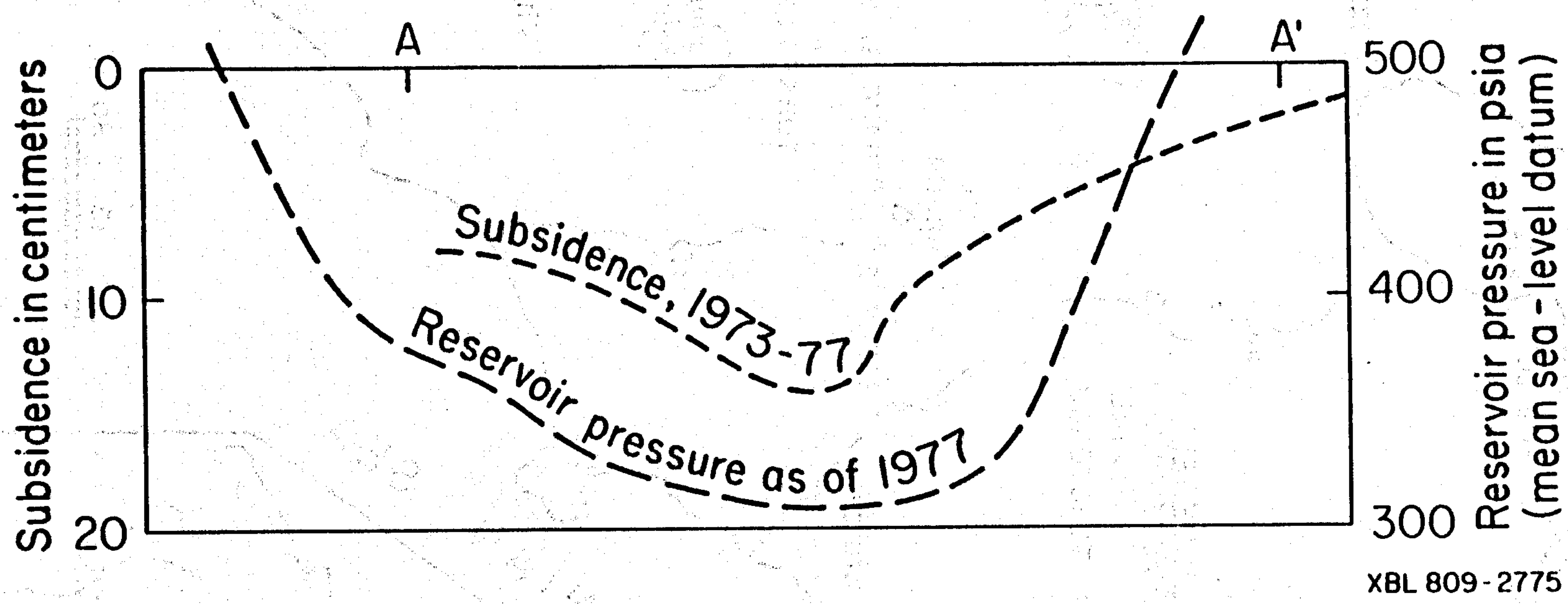
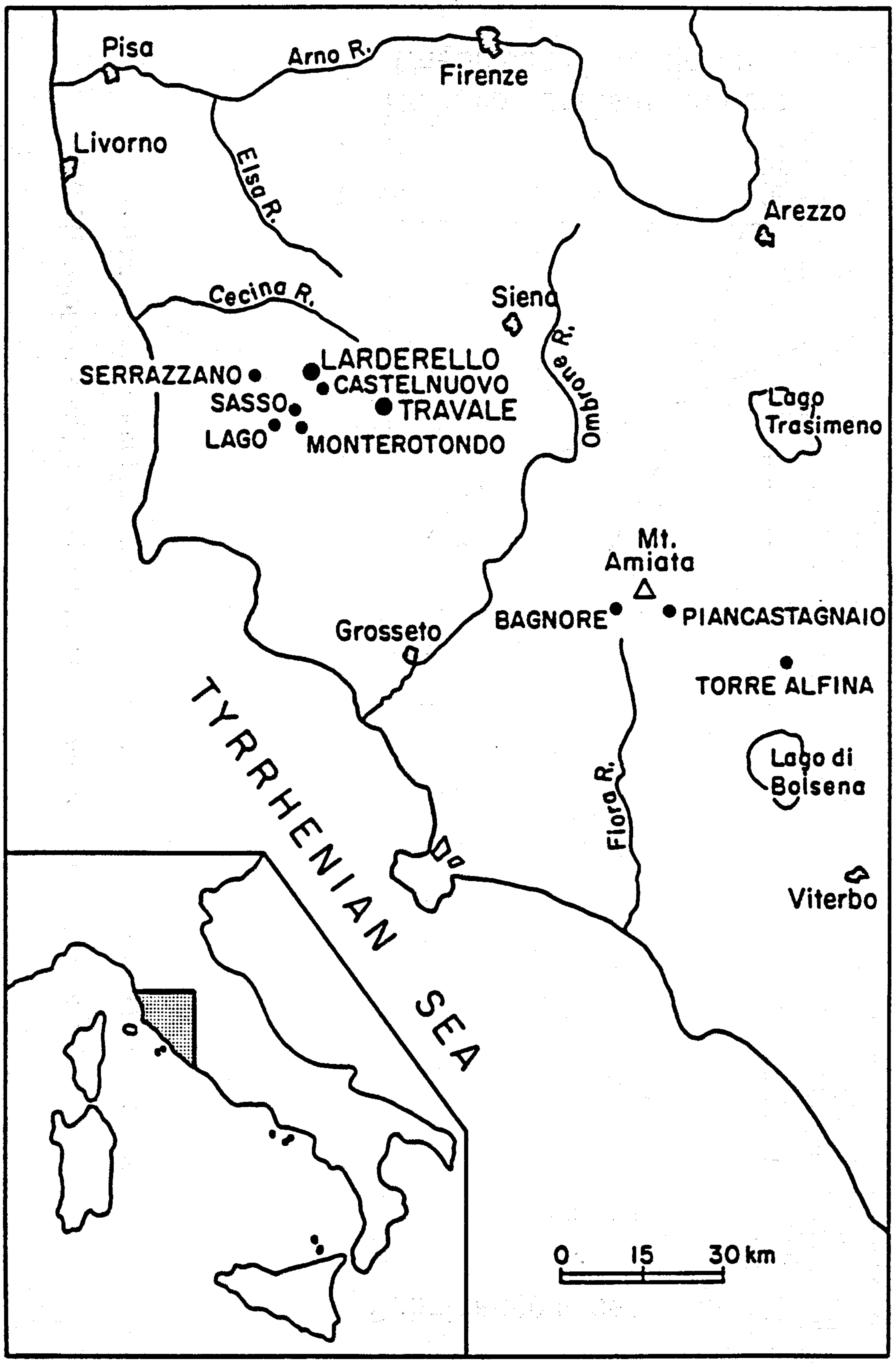
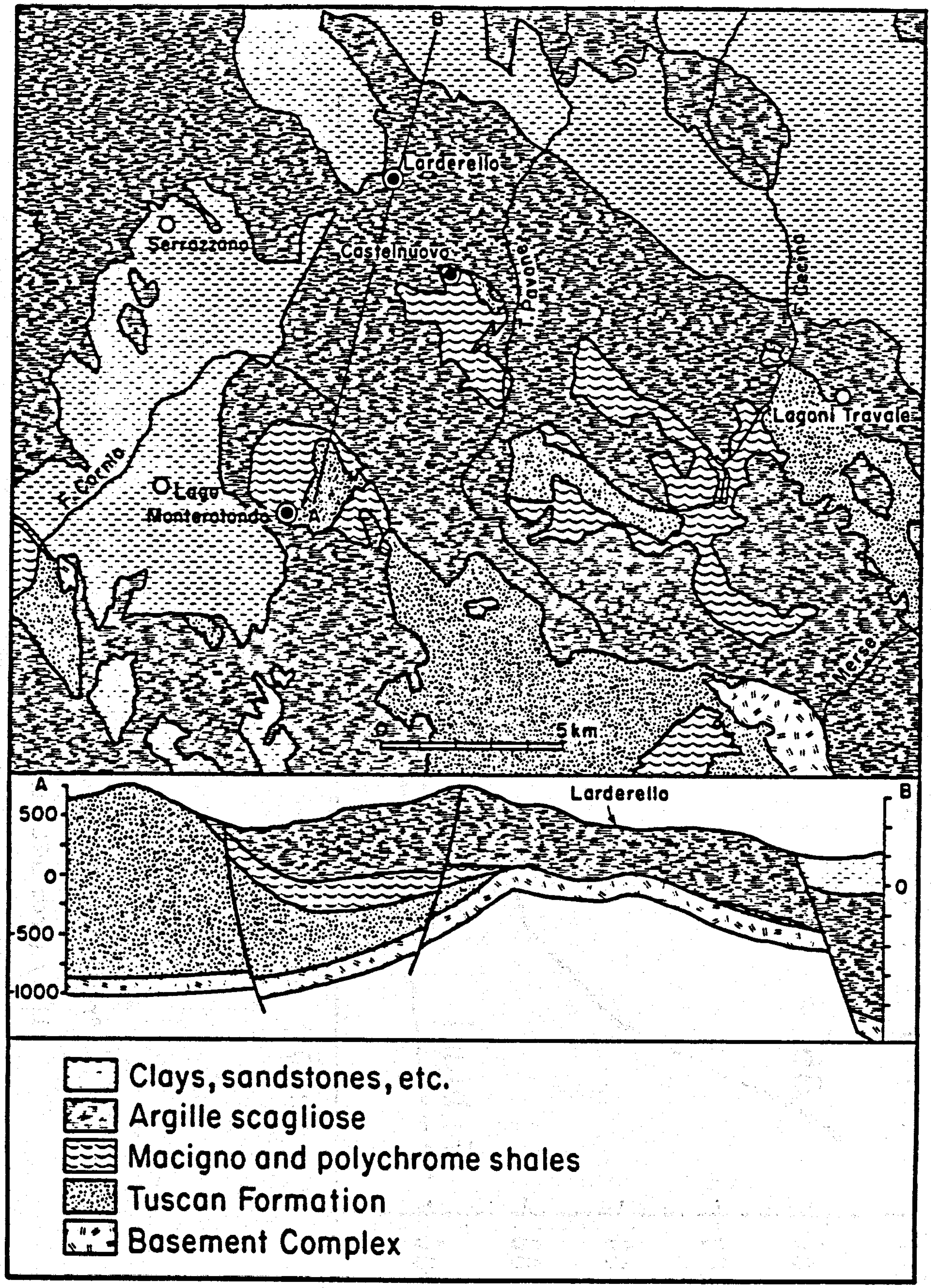


Figure 14. Profiles of reservoir pressure drop and subsidence along section AA" of Figure 12, The Geyser area, California (from Lofgren, 1978).



XBL 809-2798

Figure 15. Location map of the Larderello geothermal field, Italy (from Atkinson et al., 1977).



XBL809-2799

Figure 16. Geologic map and cross section of the Larderello geothermal field (from ENEL, 1976).

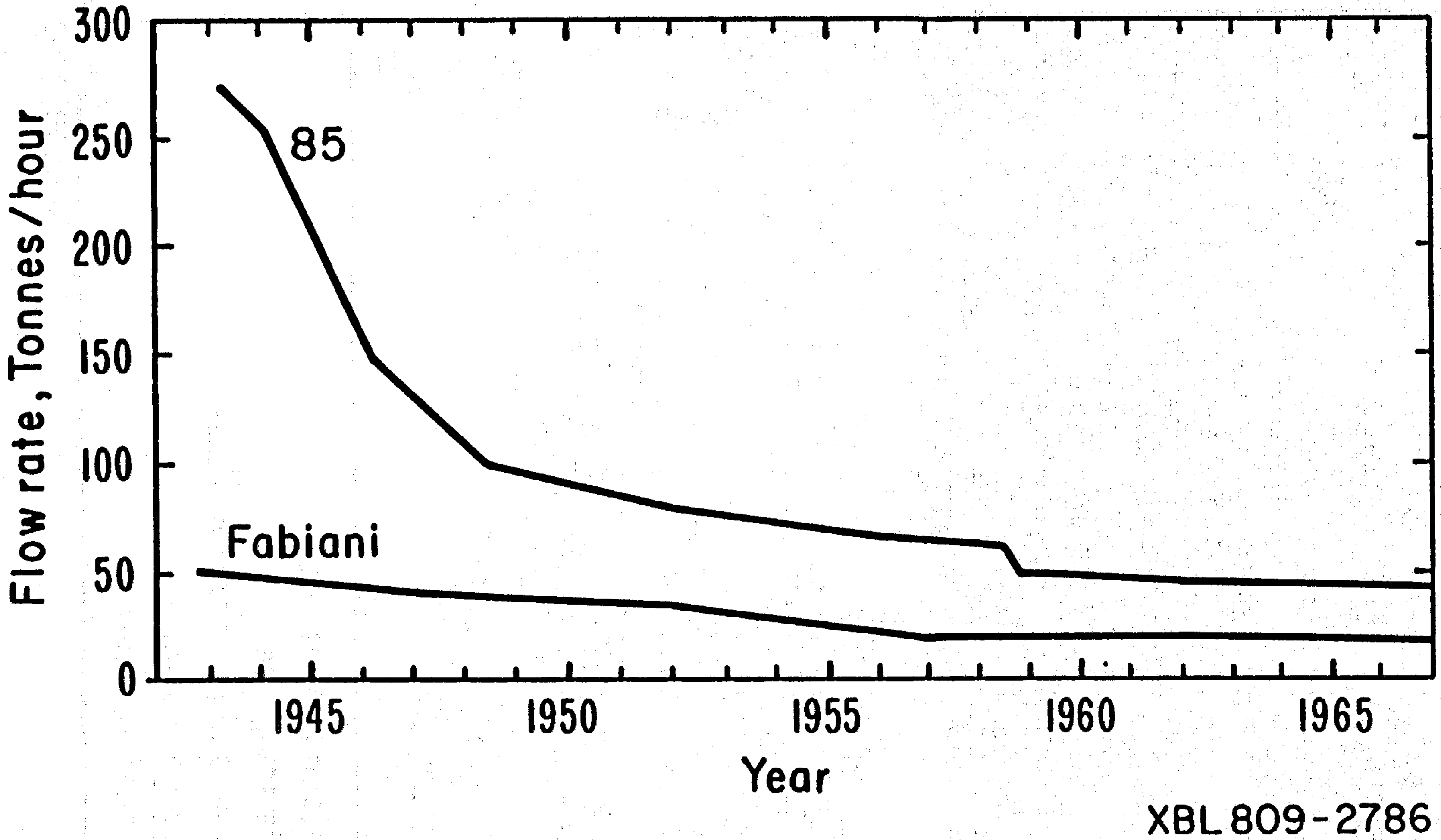


Figure 17. Production history of well 85 and the Fabiani well, Larderello, Italy (from ENEL, 1976).

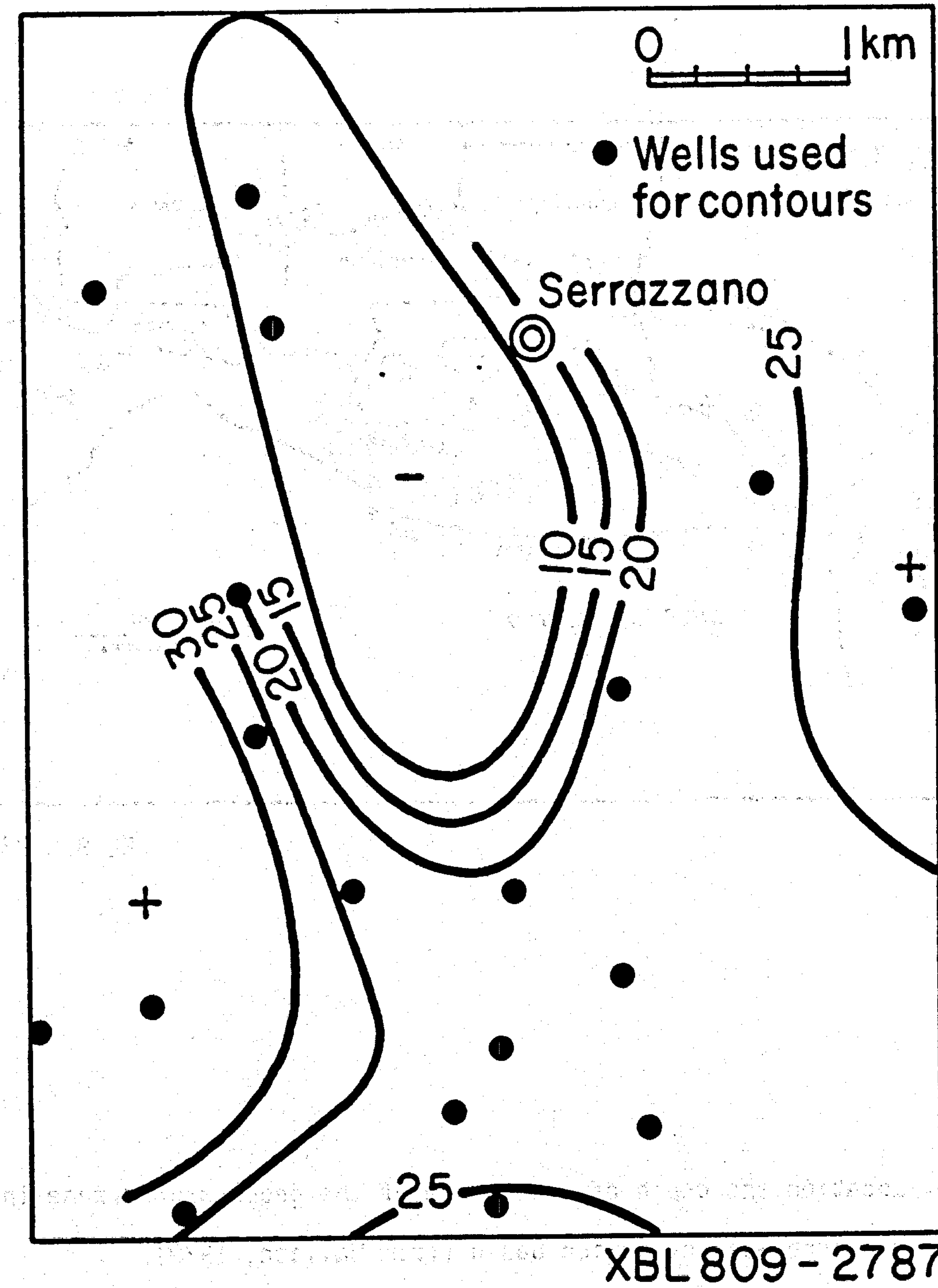
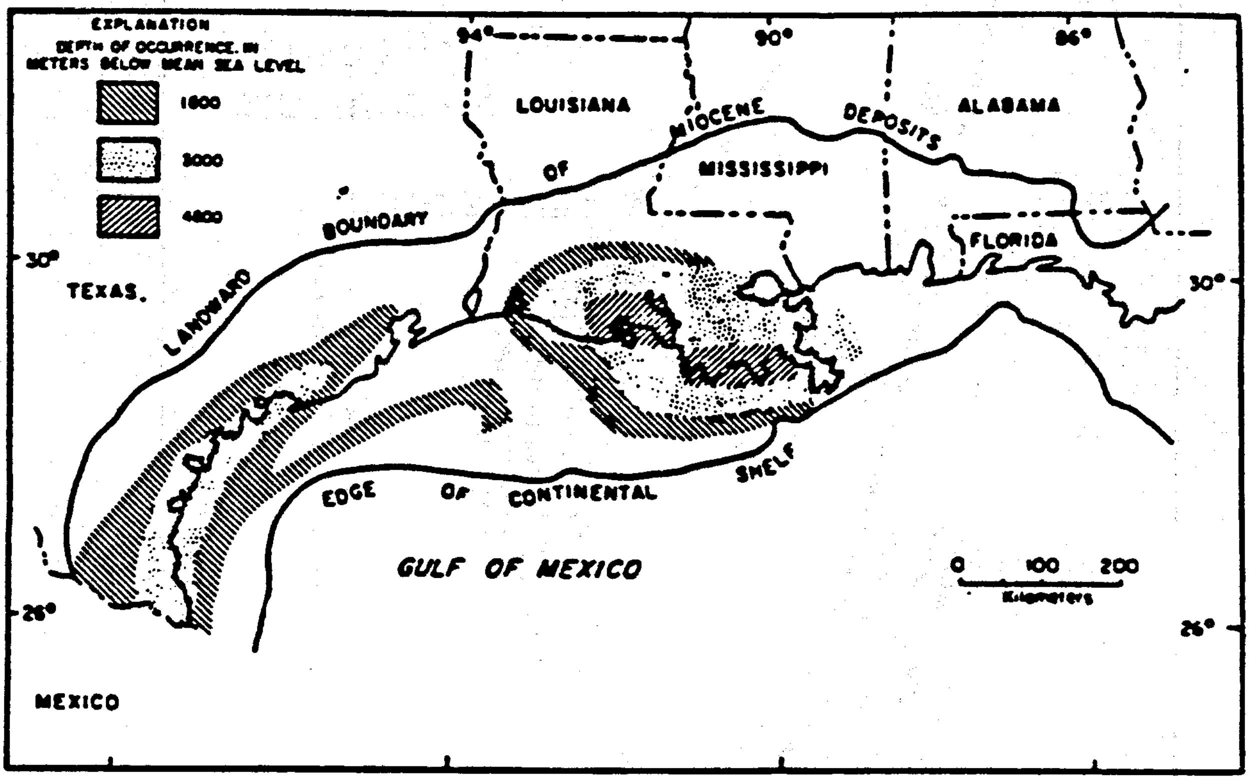
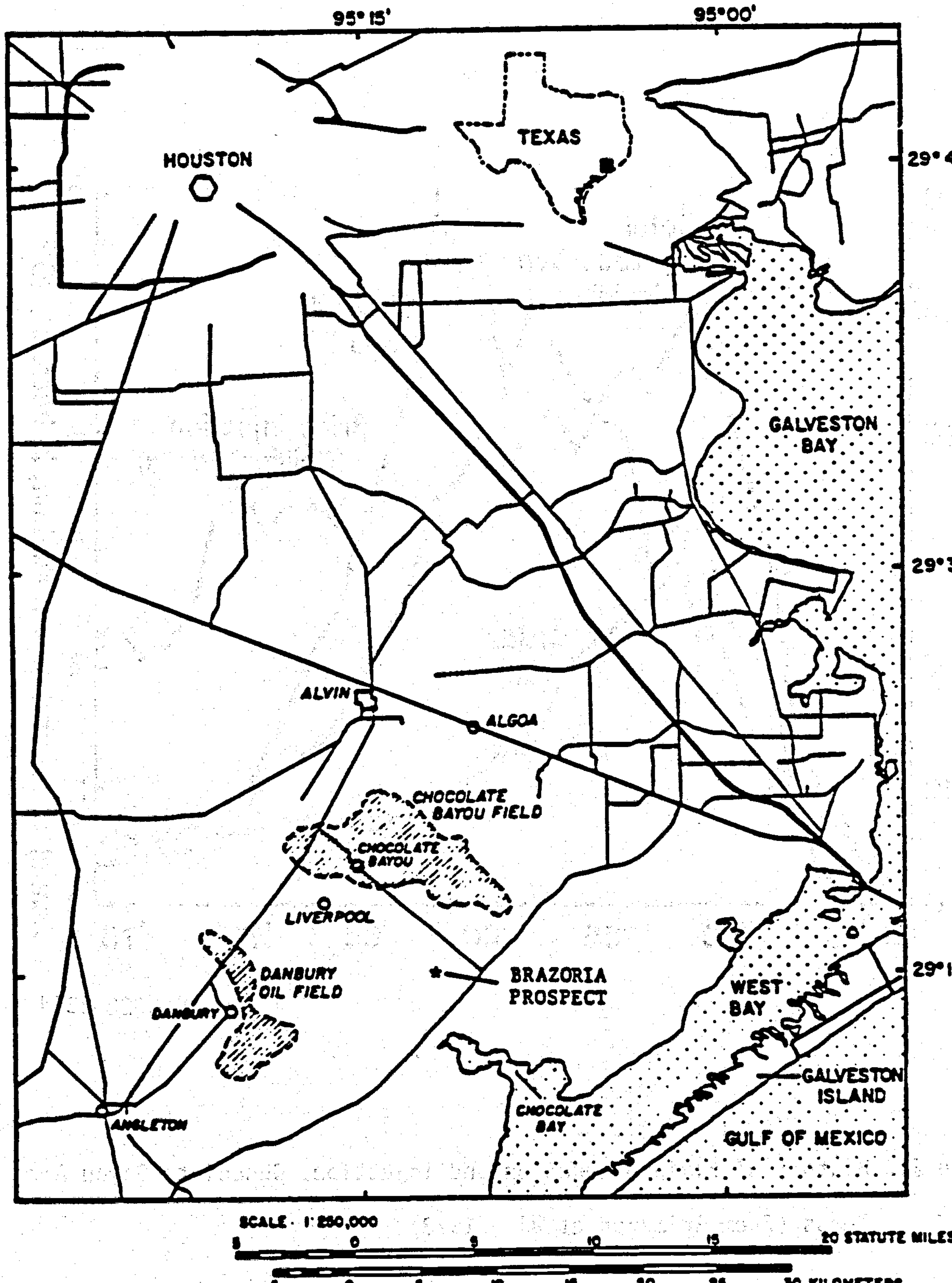


Figure 18. Spatial pressure distribution, Serrazano area, Larderello, Italy during 1970. Contour values are in kg/cm^2 (from Celati et al., 1976).



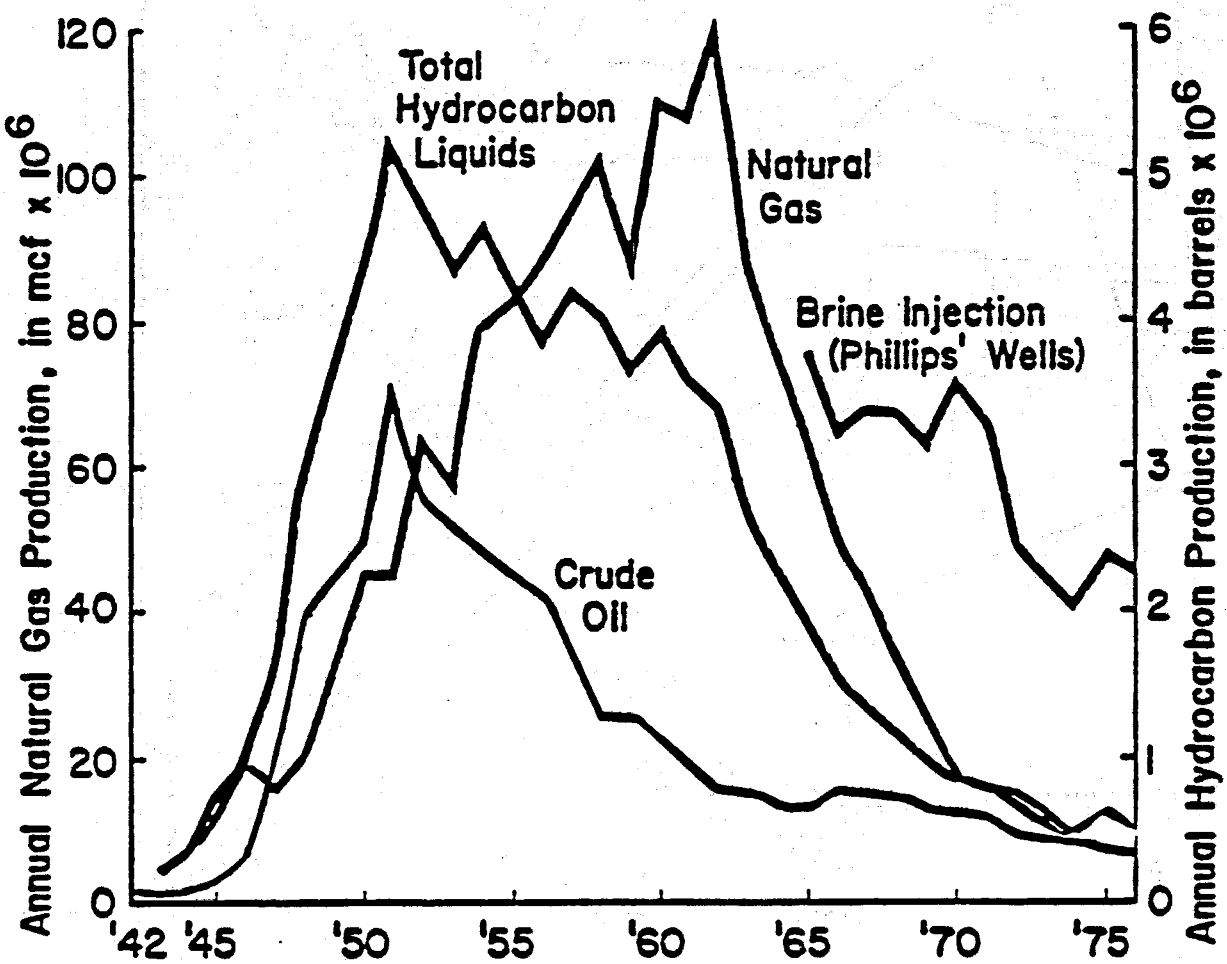
XBL 828-2371

Figure 19. Location and depth of occurrence of the geopressured zone in the northern Gulf of Mexico basin (from Wallace, 1979).



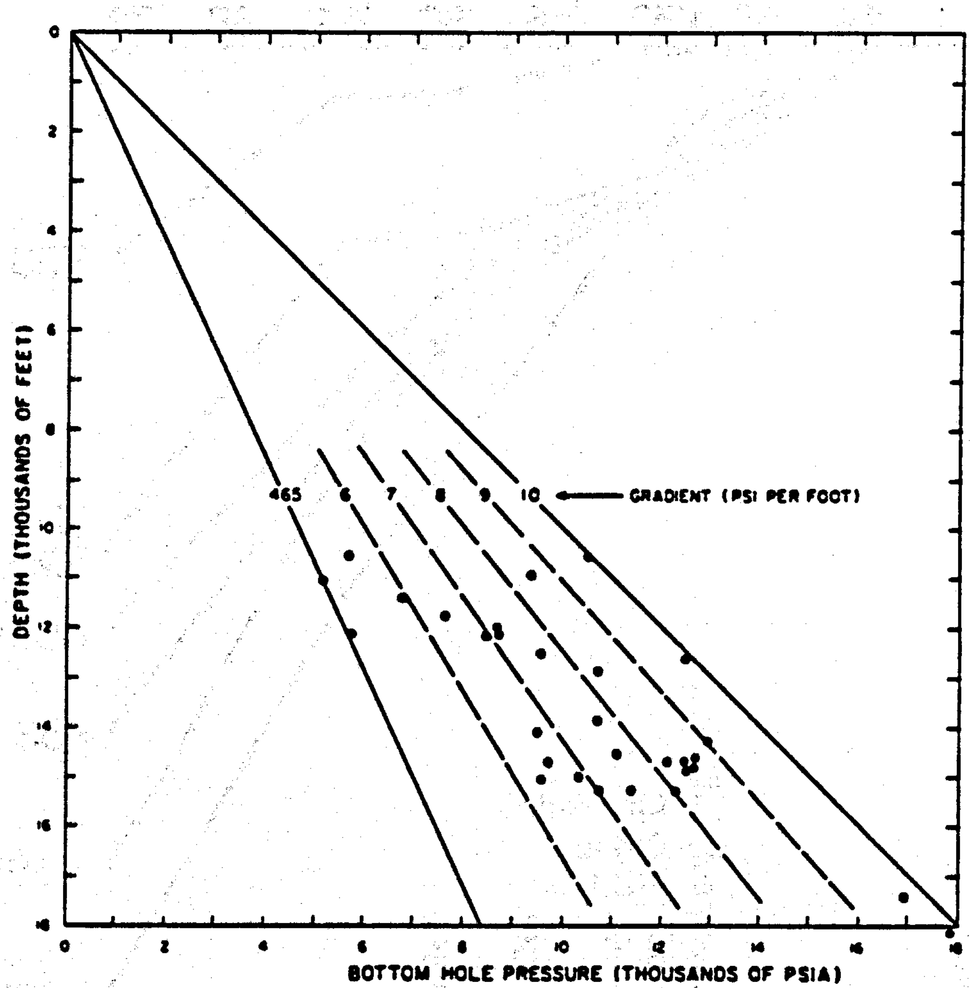
XBL 828-2373

Figure 20. Chocolate Bayou location map (from Grimsrud et al., 1978).



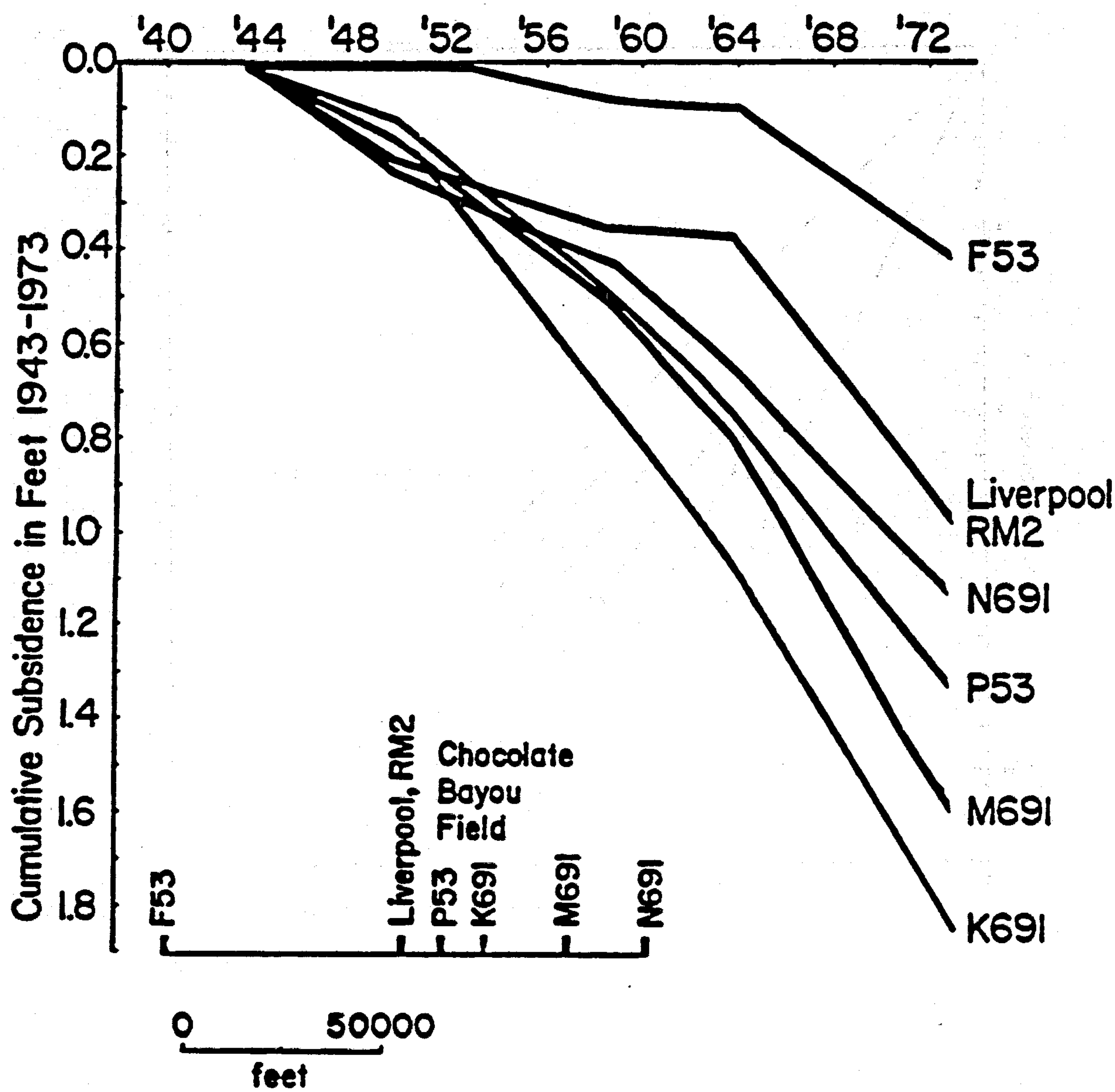
XBL 828-2374

Figure 21. History of fluid production and injection, Chocolate Bayou Area, Texas (from Grimsrud et al., 1978).



XBL 828-2375

Figure 22. subsurface fluid pressure gradients from wells in the Austin Bayou area, 8 km SW of Chocolate Bayou area (from Bebout et al., 1978).



XBL 828-2372

Figure 23. Vertical displacements individual bench marks in the Chocolate Bayou area (from Grimsrud et al., 1978).