LA-12778-MS

Thermal Modeling of the Clear Lake Magmatic System, California: Implications for Conventional and Hot Dry Rock Geothermal Development

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Prepared by Lanny Piotrowski, Group EES-1

Cover Photo: View looking north of little Borax Lake, a maar volcano less than 50 thousand years old on Buckingham Peninsula. Mafic volcanism and hot springs in this part of the Clear Lake region are manifestations of relatively shallow magmatism that has enormous geothermal potential.

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UC-251
Issued: June 1997

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ABSTRACT

The combination of recent volcanism, high heat flow (≥4 HFU or 167 mW/m^2), and high conductive geothermal gradient (up to 120°C/km) makes the Clear Lake region of northern California one of the best prospects for hot dry rock (HDR) geothermal development in the U.S. The lack of permeability in exploration wells and lack of evidence for widespread geothermal reservoirs north of the Collayomi fault zone are not reassuring indications for conventional geothermal development. This report summarizes results of thermal modeling of the Clear Lake magmatic system, and discusses implications for HDR site selection in the region. The thermal models incorporate a wide range of constraints including the distribution and nature of volcanism in time and space, water and gas geochemistry, well data, and geophysical surveys. The nature of upper crustal magma bodies at Clear Lake is inferred from studying sequences of related silicic lavas, which tell a story of multistage mixing of silicic and mafic magma in clusters of small upper crustal chambers. Some mafic to intermediate lavas at Clear Lake also contain crustal xenoliths (fragments of rocks foreign to the magma) which provide information about deeper levels of the magma generation. The xenolith suite includes mafic plutonic rocks as well as high-grade metamorphic rocks. The xenolith suite represents fragments of gabbroic intrusions and their contact aureoles. Thermobarometry on metamorphic xenoliths yield temperature and pressure estimates of ~780-900°C and 4-6 kb respectively, indicating that at least a portion of the deep magma system resided at depths from 14 to 21 km (9 to 12 mi).

Thermal modeling based on petrologic and geophysical constraints provides a test of petrologic models, and yields insight into the relationships between observed thermal gradient and magma chamber size, abundance, and emplacement history in the crust. A user-interactive 2-D numerical model was developed which simulates conductive and convective heat transport around magma bodies via a finite-difference numerical approach. The program allows for complex host rocks and multiple emplacements of magma. Conductive models that are broadly consistent with the petrologic history and observed thermal gradients of the Mt. Konocti and Borax Lake areas imply a combination of high background gradients, shallow magma bodies (roofs at 3-4 km (1.9-2.5 mi)), and recent shallow intrusion not represented by eruption. Models that include zones of convective heat transport directly above magma bodies and/or along overlying fault zones allow for deeper magma bodies (roofs at 4-6 km (2.5-3.7 mi)), but do not easily account for the large aerial extent of the thermal anomaly in the Clear Lake region. Consideration of the entire Clear Lake magmatic system, including intrusive equivalents, leads us to conclude that: (1) emplacement of numerous small and shallow silicic magma bodies occurred over essentially the entire region of high heat

flow (about 750 km^2 (290 mi^2)), (2) only a very small fraction (well less than 10%) of the silicic magma emplaced in the upper crust at Clear Lake was erupted, (3) high conductive thermal gradients are enhanced locally by fault-controlled zones of convective heat (geothermal fluid) transport; and (4) except for the Mt. Hannah and possibly the Borax Lake area, most of the silicic magma present in the upper crust has solidified or nearly solidified, and currently bears no clear geophysical signature relative to basement rocks dominated by graywacke (poorly-sorted sandstone of marine origin).

The results of thermal modeling support previous assessments of the high HDR potential of the area, and suggest the possibility that granitic bodies similar to The Geysers felsite may underlie much of the Clear Lake region at depths as little as 3-6 km (1.9-3.7 mi). This is significant because future HDR reservoirs could potentially be sited in relatively shallow granitoid plutons rather than in structurally complex Franciscan basement rocks.

I. INTRODUCTION

Estimates of the heat production of upper-crustal silicic magma bodies provide one measure of conventional and HDR resources in the U.S. Such estimates have traditionally been based on the spatial and temporal distribution of Quaternary volcanic rocks, and simple conductive heat flow calculations (Smith and Shaw, 1975, 1978; Lachenbruch et al., 1976; Kolstad and McGetchin, 1978; Carrier and Chapman, 1981). In this approach, the probable volume of the magmatic system, and the age of its youngest volcanic products are used to constrain the minimum potential heat production from a given system. For example, Smith and Shaw (1975) estimated the potential heat production associated with the Clear Lake volcanic field at 863 x 10^{18} Calories (3610×10^{18} Joules) based on an assumed *active* magma body with a volume of 1500 km^3, an emplacement temperature of 850°C, and a roof at 4 km depth. Although these early estimates demonstrated the huge heat content present in areas of shallow magmatism, they suffered from numerous simplifying assumptions, and a general lack of petrologic and geophysical constraints.

This study employs the same basic approach as pioneering efforts by Smith and Shaw (1975), but incorporates more realistic petrologic and geophysical constraints based on numerous detailed studies of the Clear Lake region (Hearn et al., 1976, 1981, 1988, 1995; Goff et al., 1977, 1993a, b; Lachenbruch and Sass, 1980; Donnelly-Nolan et al., 1981, 1993; Isherwood, 1981; Iyer et al., 1981; Walters and Combs, 1989; Stimac, 1991; Liu and Furlong, 1992; Stimac and Pearce, 1992; Stimac et al., 1992; Griscom et al., 1993; Stimac, 1993; Stimac et al., 1993a, b, c). Advances in numerical modeling and computing easily allow incorporation of these complexities into 2-D conductive and convective heat flow simulations. Thus using a combination of petrologic and geophysical constraints, a variety of models were constructed on the scale of the Clear Lake volcanic field, and on the scale of well studied silicic eruptive centers within the field. Some models assumed all heat transport by conduction, while others allowed for zones of convective heat transport at the roofs of magma bodies or along vertical fault zones. The ultimate goals of these simulations were to: (1) test and refine the magmatic model proposed by Stimac (1991) and Stimac et al. (1992), (2) determine under what range of conditions of magma emplacement the thermal models could reproduce observed heat flow in the region, and (3) determine the implications for conventional and HDR geothermal development. Although the large number of variables insures that no single model can be proven correct, this first attempt to bring petrologic, geophysical, and thermal models into agreement has yielded encouraging results, and further demonstrates the suitability of the Clear Lake region for HDR development (Goff and Decker, 1982; Burns, 1991; Stimac et al., 1992).

This report is organized into three sections. The first section summarizes what we know from previous and ongoing studies of the Clear Lake magmatic-hydrothermal system. The second section describes the thermal modeling technique and the effects of key model variables, and presents model results for the Clear Lake system. The final section discusses modeling results and their implications for HDR development in light of the assumptions and key variables of the system. Abbreviations, units, and acronyms used in this report are summarized in Appendix I.

II. THE CLEAR LAKE MAGMATIC SYSTEM

A. Regional Setting

Clear Lake is located about 135 km north of San Francisco, California, in a broad zone of deformation related to the San Andreas fault system (Fig. 1). The fault system is made up of

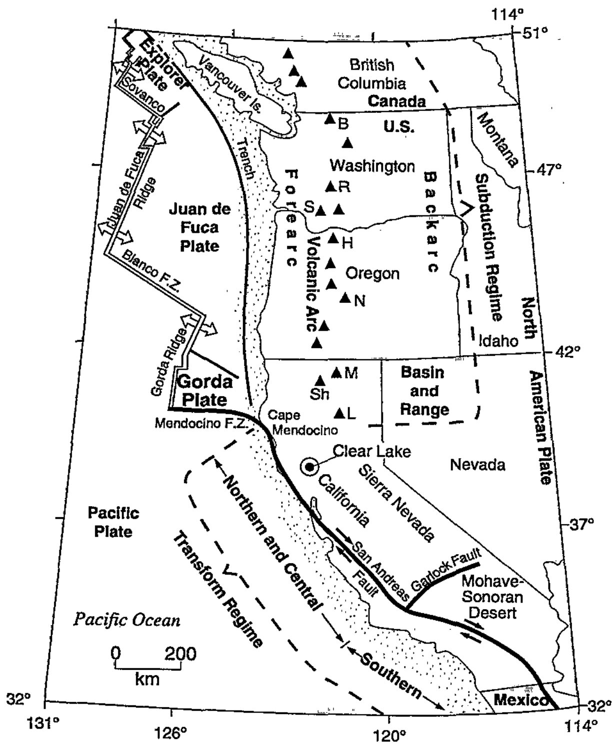

Fig. 1. Tectonic framework of the northwestern United States (modified from Mooney and Weaver, 1989). The region can be divided into a subduction regime north of Cape Mendocino, and a transform regime to the south. Clear Lake (solid circle) is located in the northern portion of the transform regime. Solid triangles mark active or dormant volcanoes of the Cascade range: L, Lassen; Sh, Shasta; M, Medicine Lake; N, Newberry Volcano; H, Mt. Hood; S, Mount St. Helens; R, Mount Ranier; B, Mt. Baker.

numerous subparallel faults and represents the boundary between the Pacific and North American plates south of the Mendocino Triple Junction (Jennings, 1992; Castillo and Ellsworth, 1993). Regional tectonic considerations and the approximate correlation in age between volcanism in the Coast Ranges and northward migration of the Mendocino Triple Junction have led many workers to conclude that magmatism in the region results from mantle upwelling at the southern edge of the Gorda Plate (Dickinson and Snyder, 1979; McLaughlin, 1981; Johnson and O'Neil, 1984; Fox et al., 1985; Benz et al., 1992). The Clear Lake volcanic field is the northernmost and youngest manifestation of this magmatism (Fig. 2). Although the state of stress in the region is largely transpressional, volcanism around Clear Lake is apparently localized in regions of transtension by complex structures of the San Andreas fault system (Hearn et al., 1988).

B. Magmatic History

Volcanism in the Clear Lake region occurred over an elongate NW-striking band about 20-40 km wide and 55 to 75 km long, with ages ranging from about 2.1 to 0.01 Ma (Hearn et al., 1976; Donnelly-Nolan et al., 1981) (Fig. 2). Spatial and temporal trends of volcanism within the Clear Lake region are similar to larger-scale trends in the northern Coast Ranges (Donnelly-Nolan et al., 1981; Stimac et al., 1992). That is, silicic volcanism shows a general migration to the north with time (Fig. 3). Mafic volcanism is more widespread initially, but is also concentrated progressively further north during the later episodes of volcanism.

Donnelly-Nolan et al. (1981), Hearn et al. (1981) and Stimac et al. (1992) outlined evidence for the existence of several discrete silicic centers through the life of the Clear Lake magmatic system. Aside from very small-volume eruptions at Pine Mountain (2.06 Ma), the earliest silicic rocks are rhyolite and dacite lavas of Cobb Mountain (1.1 Ma) and granitoid rocks that form the core of the Geysers geothermal field (>1.3 to 0.9 Ma; Donnelly-Nolan et al., 1981; Pulka, 1991; Dalrymple, 1992). Successively younger silicic centers are located at Mt. Hannah-Seiglar Mountain (<0.9 to 0.6 Ma), Mt. Konocti (0.65 to 0.30 Ma), and Borax Lake (0.09 Ma) (Donnelly-Nolan et al., 1981). Together, these centers display a pronounced northward progression of silicic magmatism through time (Fig. 3; Stimac et al., 1992; Donnelly-Nolan et al., 1993).

C. Petrologic Features

Volcanism at Clear Lake was dominated by eruption of numerous, small to moderate-volume

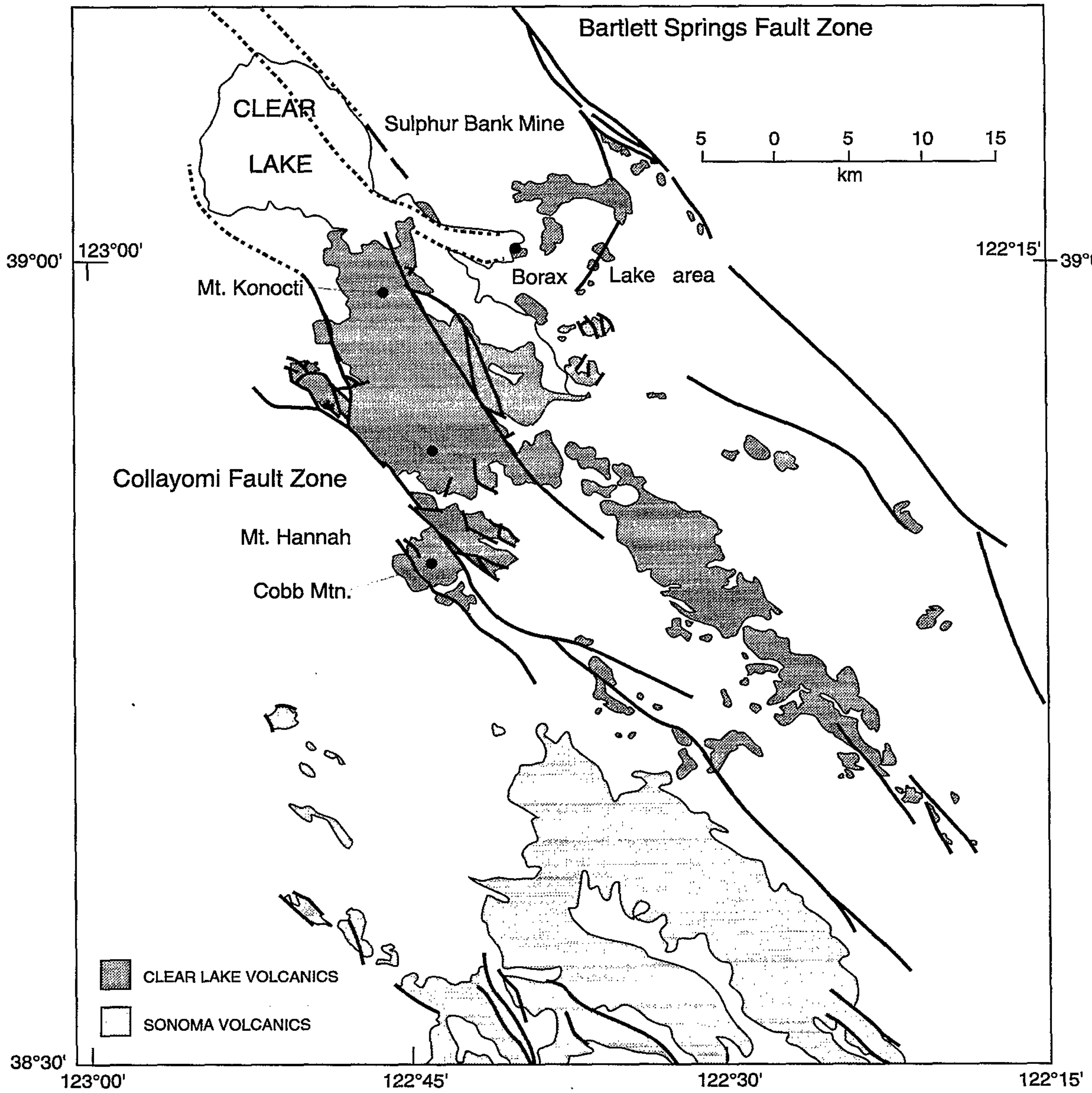

Fig. 2. Distribution of young volcanic rocks in the Clear Lake and adjacent Sonoma volcanic centers (modified from Wagner and Bortugno, 1982; Hearn et al., 1988; Donnelly-Nolan et al., 1993; and Jennings, 1993). Major faults of the San Andreas system are shown, but smaller structures are omitted. Silicic volcanic centers mentioned in the text are also shown.

lava flows and domes (generally <6 km^3 per eruption), with a total erupted volume of about 100 km^3 (Donnelly-Nolan et al., 1981; Hearn et al., 1981). Lesser pyroclastic deposits represent the products of magmatic, phreatic, and phreatomagmatic eruptions. Similar styles of volcanism (numerous small eruptions of lavas and subordinate pyroclastic material) are common to regions of active crustal extension such as the Coso volcanic field, California (Hildreth, 1981; Bacon, 1982).

The nature of upper crustal magma bodies at Clear Lake is inferred from studying sequences of closely related silicic lava flows in the Mt. Konocti, Cobb Mountain, and Borax Lake areas. Detailed petrographic and chemical studies indicate that these lavas formed by multi-stage mixing of silicic and mafic magma in clusters of

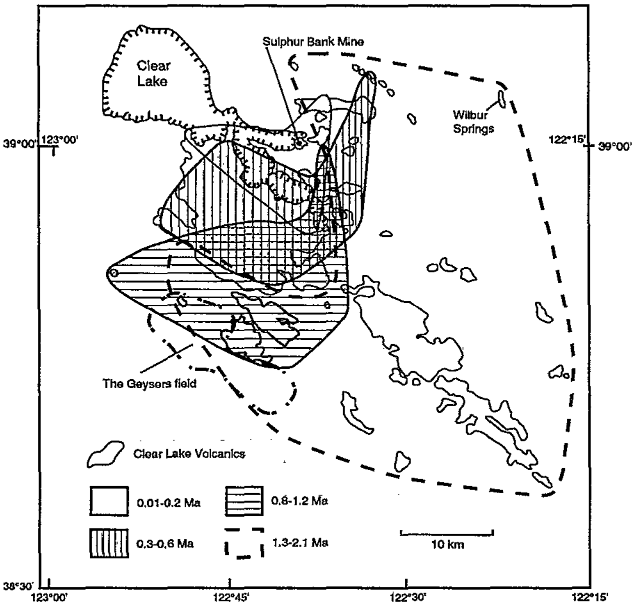

Fig. 3. Spatial and temporal distribution within the Clear Lake volcanic field (modified from Donnelly-Nolan et al., 1993). Silicic volcanism shows a general migration to the north with time.

shallow crustal chambers (Patterson-Latham, 1985; Stimac, 1991; Stimac and Pearce, 1992; Stimac et al., 1994). Eruption of silicic sequences commonly began with aphyric rhyolite (e.g. the rhyolites of Thurston Creek and Borax Lake), and progressed to more crystal-rich rhyolite and dacite lavas (Stimac, 1991).

Based on chemical, petrological, and isotopic data only a few mafic lavas in the Clear Lake region retain exclusively mantle signatures. Most mafic to intermediate lavas are variably contaminated by crustally-derived melt and crystalline debris (Futa et al., 1981; Stimac et al., 1993a). The most contaminated of these lavas contain crustal xenoliths which provide information about deeper levels of the magmatic system. The xenolith suite includes noritic to gabbroic plutonic rocks as well as high-grade metamorphic rocks. Together, these rocks probably represent fragments of gabbroic intrusions and their contact aureoles (Stimac, 1993; Stimac et al., 1993b). Thermobarometry on xenoliths yield temperature and pressure estimates of ~780-900°C and 4-6 kb respectively for metamorphic xenoliths, indicating that at least a portion of the mafic magmatic system at Clear Lake resided in the crust from 14 to 21 km depth. Metasedimentary xenoliths also exhibit evidence for partial melting and variable degrees of melt extraction, providing direct evidence for anatexis and assimilation in the lower crust.

Based in part on the above petrologic evidence, the Clear Lake magmatic system can be envisioned as a simplified two-level system, driven by triple junction migration and concomitant decompression melting within the underlying asthenosphere (Fig. 4). The focus of the resulting volcanism has shifted north and east with time (Donnelly-Nolan et al., 1981; Hearn et al., 1981; Stimac et al., 1992). The deep part of the system (lower to mid-crustal levels) is dominated by mafic intrusions, granulite-facies metamorphism, and local melting of metasedimentary protoliths. This deep system gave rise to variably contaminated and hybridized intermediate to silicic magmas, which eventually migrated to higher crustal levels (Stimac et al., 1992). The upper level system consisted primarily of numerous localized clusters of small silicic magma bodies, rather than a single large silicic chamber (Stimac, 1991; Stimac and Pearce, 1992). Although at least one moderately sized, upper crustal plutonic complex was created (Geysers felsite), most batches of silicic magma were relatively small, and probably crystallized rapidly unless recharged at regular intervals. This petrologic model for the Clear Lake magmatic system is broadly consistent with heat flow and geophysical evidence presented in the next section, but leaves open the possibility of relatively recent intrusion of silicic magma not represented by eruption.

D. Geophysical Features

1. Regional and Local Heat Flow

Heat flow and its implications for the structure of the crust in the northern Coast Ranges of California have been modeled and discussed by Lachenbruch and Sass (1980), Liu and Furlong

(1992), and Liu (1993). They stressed the importance of a northwardly-migrating thermal perturbation beneath the northern Coast Ranges caused by creation of a "slabless window" beneath the southern edge of the Gorda plate. Lachenbruch and Sass (1980) compiled a contour map of heat flow in the western U.S. (Fig. 5). The Coast Ranges of California south of the Mendocino Triple Junction are enclosed by the 1.5 HFU (64 mW/m^2) contour, and The Geysers-Clear Lake region is enclosed by the ≥2.5 HFU (104 mW/m^2) contour. Lachenbruch and Sass (1980) modeled the heat flow that would result from basaltic underplating, or basaltic intrusions at various crustal depths, and concluded that the *regional* heat flow anomaly could result from extensive underplating, or intrusion of basalt into the *lower* crust, but that intrusion of basalt into the upper crust would give rise to much higher regional heat flow than observed. We emphasize that their treatment is based on regional heat flow (averaging about 2 HFU), and does not explicitly address the origin of The Geysers-Clear Lake thermal anomaly.

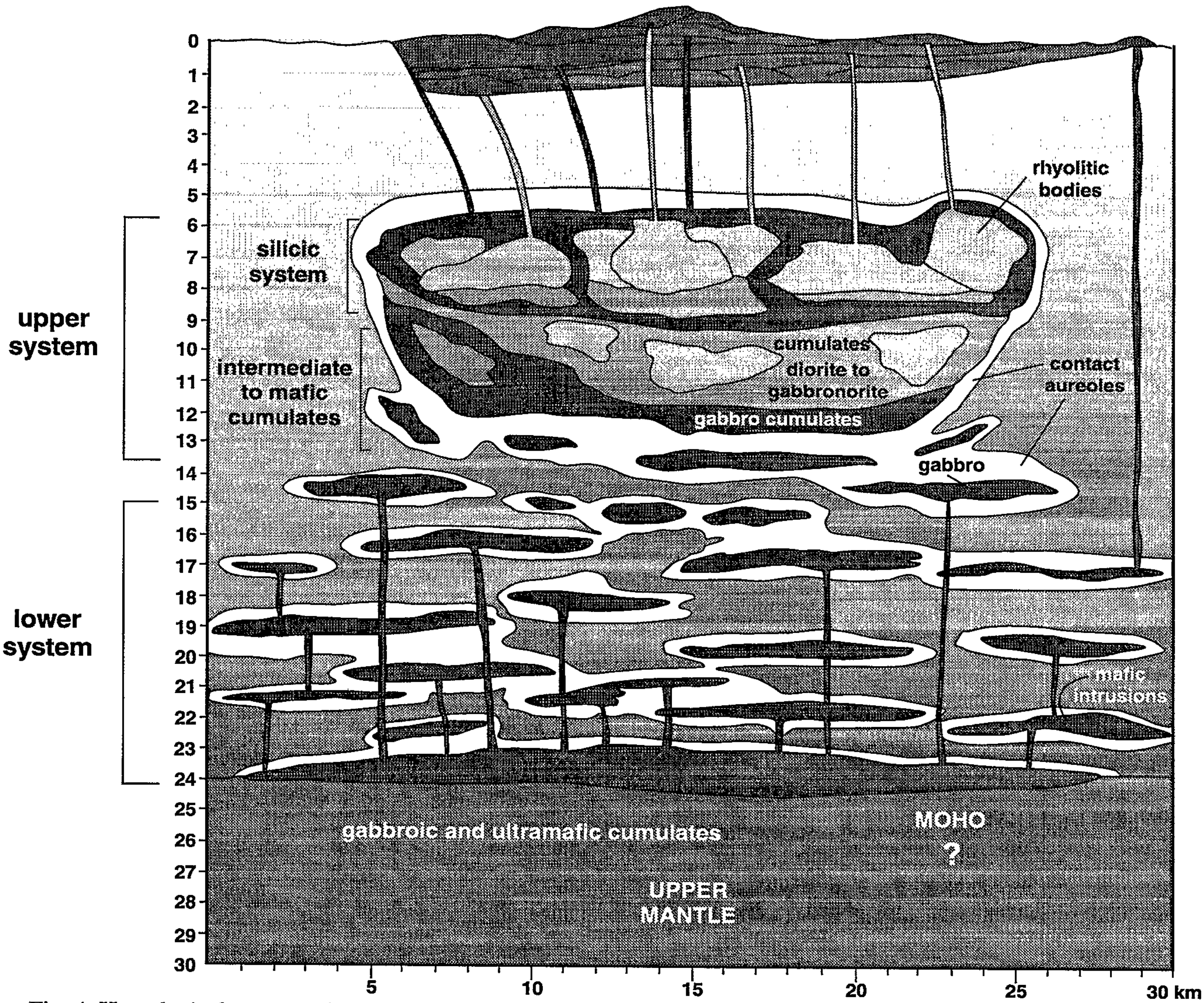

Fig. 4. Hypothetical cross-section of the Clear Lake magmatic system at a late stage in its evolution. This simplified view stresses the eventual development of a two level system of crustal magmatism consisting of a deep, dominantly-mafic system, and a shallow, dominantly-silicic system. The cartoon depicts the cumulative extent of the magmatic system from initiation at 2 to 3 Ma ago to ~200 ka ago.

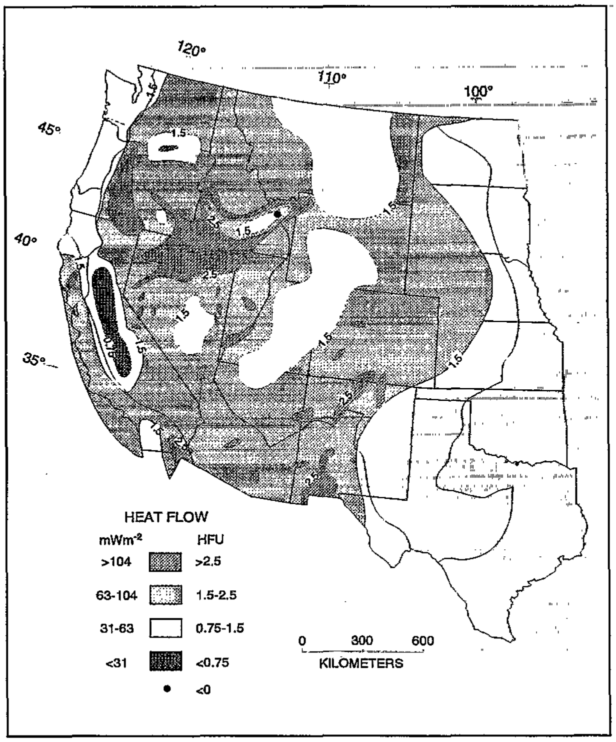

Fig. 5. Heat flow contour map of the western United States (modified from Lachenbruch and Sass, 1980). The California Coast Ranges south of the Mendocino triple junction is enclosed by the 1.5 to 2.5 HFU (heat flow units) contour, whereas the Clear Lake region is enclosed in the >2.5 HFU contour.

More detailed studies of The Geysers-Clear Lake region have documented heat flow greatly exceeding the regional average (Jamieson, 1976; Urban et al., 1976; Walters and Combs, 1989). Jamieson (1976) modeled the heat source for The Geysers geothermal field as a region on the southwest flank of the gravity anomaly centered over Mt. Hannah. He interpreted the heat flow to be consistent with hot (>700°C) intrusive rocks at a depth of about 8 km over a wide area, with heat flow from the source to the surface *primarily by conduction*. Superimposed on this conductive anomaly, Jamieson found extremely high temperature gradients in The Geysers area. These gradients reflect elevated near-surface temperatures in a conductive cap due to *convective* transfer of heat from depth by local steam and condensed hot-water systems along high-angle fractures. Within in this area Urban et al. (1976) found that the region between the surface and the shallow steam reservoir is mainly conductive. Heat flow in this interval measured in two holes was 7.5 and 9.3 HFU. None of the deep holes (up to 3.4 km) drilled in The Geysers have penetrated beneath the nearly-isothermal zone of steam production. This zone directly overlies and includes portions of a composite silicic plutonic complex known as the "Geysers felsite" (Hulen and Nielsen, 1993).

Walters and Combs (1989) published the most comprehensive study of heat flow in The Geysers-Clear Lake region. They documented a 4 HFU (168 mW/m^2) thermal anomaly over an area of at least 750 km^2 (Fig. 6). This heat flow anomaly encompasses The Geysers and that portion of the Clear Lake volcanic field containing silicic volcanic rocks or their intrusive equivalents. The Geysers steam field lies entirely within their 8 HFU contour (Fig. 6).

We summarize thermal gradient and calculated heat flow data for "deep" wells north of The Geysers in Table 1 and Fig. 7, and embellish the heat flow contours of Walters and Combs in Fig. 6. One point made clear from this data is that the majority of deep holes drilled in the Clear Lake region were dry or had limited fluid production. Another feature of well data illustrated in Fig. 8 is that thermal gradients in "deep" wells are conductive. These features indicate that heat transport north of The Geysers is primarily by conduction as previously suggested by Jamieson (1976) and Walters and Combs (1989).

2. Gravity and Magnetics Surveys

Geophysical surveys of the Clear Lake region have focused on identifying active magma bodies (see summaries by Isherwood, 1981 and Griscom et al., 1993). Early gravity and magnetics studies at Clear Lake identified a large negative (-24 mGal) gravity anomaly centered beneath Mt. Hannah (Fig. 6), and suggested it represents a magma chamber roughly 14 km in diameter at ≥7 km depth (Chapman, 1975;

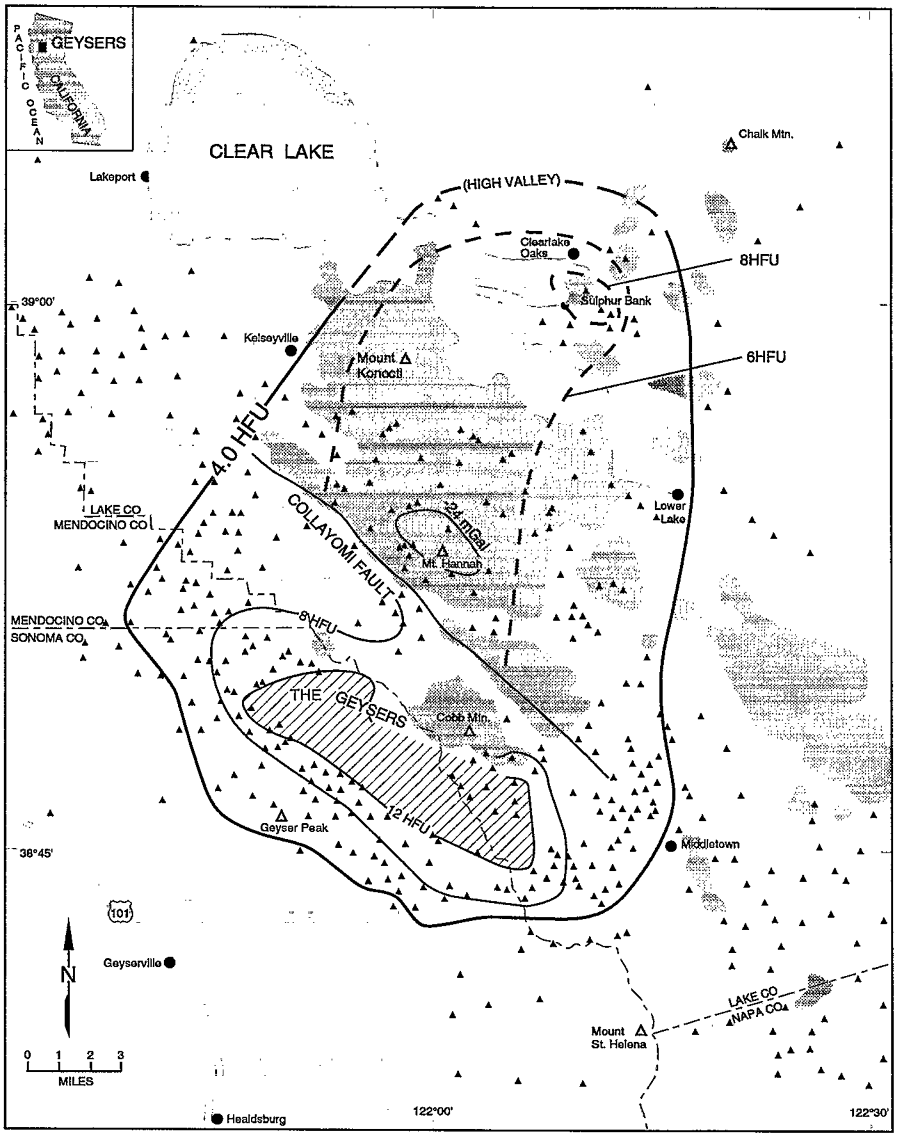

Fig. 6. Heat flow map of The Geysers-Clear Lake region (from Walters and Combs, 1989). The area enclosed by the 4 HFU contour coincides almost exactly with exposures of silicic volcanic and shallow plutonic rocks (Geysers felsite), but does not enclose early mafic lavas that comprise the southeastern portion of the volcanic field. Dashed 6 and 8 HFU contours are based on data presented in Table 1 and Fig. 7. The -24 mGal gravity low centered beneath Mt. Hannah is shown for reference. A clear heat flow discontinuity occurs across the Collayomi fault zone due to the presence of The Geysers steam field.

Isherwood, 1981). More recently Griscom et al. (1993) discussed various possible interpretations of the geophysical data (gravity, aeromagnetics, and magnetotelluric soundings) and summarized conflicting evidence that would either place a 5x23 km sill-like, silicic magma body within 4-5 km of the surface (favored by Griscom et al., 1993), or place it as deep as 15 km (favored by Blakely and Stanley, 1992 and Stanley and Blakely, 1995). Griscom et al. (1993) also cited the presence of tourmaline-hornfels similar to that making up the contact aureole of the Geysers felsite in deep drill holes (1.5-3.6 km) in the Clear Lake region, bolstering the case for a shallow magma body underlying the Mt. Hannah area.

Table 1. Information on "deep" geothermal exploration, gradient, and production wells in the Clear Lake region, California

Map No.	Name/Operator (Other Name/Operator)	Location	Depth (m)	BHT (°C)	Gradient (°C/km)	Calculated Heat Flow mW/m^2	HFU	Formation(s)	Comments
Geothermal Exploration Wells									
1	Audry #1/Phillips	T13N, R7W, S8	3063	297	141	375[1]	9.0[1]	Fran.	"Dry;" copious CO_2 and some water (Beall, 1985); temp. @ 2886 m.
2	Bianchi #1/Aminoil USA (B&R Unit 1)	T11N, R8W, S23	3431	230	—	—	—	Fran.	Dry.
3	Bianchi #2/Aminoil USA	T11N, R8W, S23	3050	135	54	146[2]	3.5[2]	Fran.	Limited fluid; no production.
4	B-J #1/ Aminoil USA	38°45' Lat., 122°35' Long.	3148	170	—	—	—	Fran.	Dry; CO_2 @ 300-400 psi.
5	Borax Lake #7/Phillips	T13N, R7W, S7	3077	≥250	96	256[1]	6.1[1]	Fran.	"Dry;" some water (Beall, 1985).
6	Bounsall #1/Shell	T10N, R7W, S16	2515	—	34	246[1]	5.9[1]	Fran.	"Dry;" 66°C @ 1180 m.
7	Bouscal #1/Republic (Guisti #1)	T12N, R7W, S30	2769	245	—	—	—	GV-Serp.-Fran.	"Dry;" some fluid.
8	Bouscal #1A/Republic (Occidental)	T12N, R7W, S30	2750	245	—	—	—	GV-Serp.-Fran.	Dry.
9	Bradley #1/Earth Energy	T13N, R7W, S5	643	208	(322)[3]	(869)[3]	(20.8)[3]	Fran.	Water; 218°C @ 503 m.
10	Bradley #2/Earth Energy	T13N, R7W, S5	1226	175	137	370[2]	8.8[2]	Fran.	Limited fluid.
11	Clear Lake #1/Thermal Power (Magma Sulph. Bank #1)	T13N, R7W, S5	460	135	—	—	—	Fran.	Water.
12	Davies #1/South. Union Prod.	T11N, R8W, S36	2195	177	—	—	—	Fran.	Dry.
13	Jorgenson #1/Union Oil (Thurston #1)	T13N, R8W, S36	3220	310	86	214[1]	5.1[1]	CLV-Fran.	"Dry;" some water (Beall, 1985); max. temp. @ 3050 m; base of volcanics @ 670 m.
14	Kettenhoffen #1/Getty Oil (Eureka-Magma #1/Eur. Mag. Explor.)	T13N, R8W, S36	2610	≥200	—	—	—	CLV-GV-Serp.-Fran.	Dry; temp. est. from Goff et al. (1977); base of volcanics @ 744 m.
15	Magma #1/Magma Power	T14N, R5W, S29	274	122	—	—	—	GV-Serp.	Small water entry.
16	Magma-Watson #1/Magma-Dow Chem.	T13N, R8W, S20	1754	>200	—	—	—	CLV-GV-Serp.-Fran.	Dry; base of volcanics @ 1265 m.
17	Neasham #1/Republic-Occidental	T12N, R8W, S8	3673	360?	—	232[1]	5.5[1]	CLV-Serp.-Fran.	"Dry;" small water entry @ 2615 m, 275°C; base of volcanics @ 728 m.
18	Newfield #1-33/McCulloch	T12N, R8W, S33	2739	≥220	—	—	—	CLV-Serp.	Dry; base of volcanics @ 10 m.
19	S&B #1/Aminoil USA (Smith and Breazeale #1)	T11N, R8W, S23	3431	232	—	—	—	Fran.	Dry.
20	Sullivan #1/E.B. Towne	T12N, R8W, S18	1871	190	273	737[2]	17.6[2]	Serp.	Dry; gradient "kick" at bottom of well.
21	Sulphur Bank #1/ SB Geoth. Power (Hawaii Sulph. Bank #1)	T13N, R7W, S5	168	186	—	—	—	Fran.	Water.
22	Tellyer #1/MCR Geothermal	T11N, R8W, S13	3717	238	—	—	—	Serp.	Dry.

Table 1. (Cont.)

Map No.	Name/Operator (Other Name/Operator)	Location	Depth (m)	BHT (°C)	Gradient (°C/km)	Calculated Heat Flow mW/m²	HFU	Formations(s)	Comments
23	Wilbur Hot Springs #1/Cordero-Sun	T14N, R5W, S29	1146	134	32	86[2]	2.1[2]	GV-Serp.	Small water entry @1020 m, 138°C.
24	Wilbur Hot Springs #2/Sunoco-Bailey Min.	T14N, R5W, S29	2776	≤140	—	—	—	GV-Serp.	Small water entry.
25	Wilson #1/Republic-Occidental	T12N, R8W, S8	3048	325	95	232[1]	5.5[1]	CLV-Serp.-Fran.	"Dry;" small water entry; max temp. @ 3354 m; base of volcanics @ 1219 m.
Geothermal Gradient Wells									
26	Boggs 77-1/Republic	T12N, R8W, S8	741	—	>90	>243[2]	5.8[2]	CLV-Serp.-Fran.	Dry; base of volcanics @ 664 m.
27	Boggs Mtn #1/Geothermal Kinetics	T11N, R7W, S18	1309	50	30	81[2]	1.9[2]	GV	Dry.
28	Hannah Strat-1/Phillips	T12N, R8W, S22	342	38	103	278[2]	6.6[2]	CLV	Dry.
29	Hannah Strat-2/Phillips	T12N, R9W, S1	611	77	110	297[2]	7.1[2]	CLV-GV?	Dry.
30	Mt. Hannah #1/USGS	T12, R8W, S15	313	20	34?	—	—	CLV	Many cold water entries; not conductive.
31	Unnamed #1/Phillips	T13N, R7W, S9	502	33	16	43[2]	1.0[2]	Fran.	Cold water entries; not conductive.
Geothermal Production Wells									
32	Ag. Park #1/Lake County (Geotheral Agriculture Heat Center)	T13N, R8W, S33	492	31	—	—	—	CLV	Space heating well used for reinjection; max. temp 64°C @ 70 m.
33	Ag. Park #2/Lake County	T13N, R8W, S32	180	52	—	—	—	CLV	Space heating well presently unused; max. temp. 76°C @ 137 m.
34	Ag. Park #3/Lake County	T13N, R8W, S32	152	—	—	—	—	CLV	Space heating well, main producer; max temp 68°C @ 120 m.

[1] From Walters & Combs, 1989

[2] Assuming $K_{avg} = 2.7$

[3] Convective heat flow; values should be disregarded for conductive calculations.

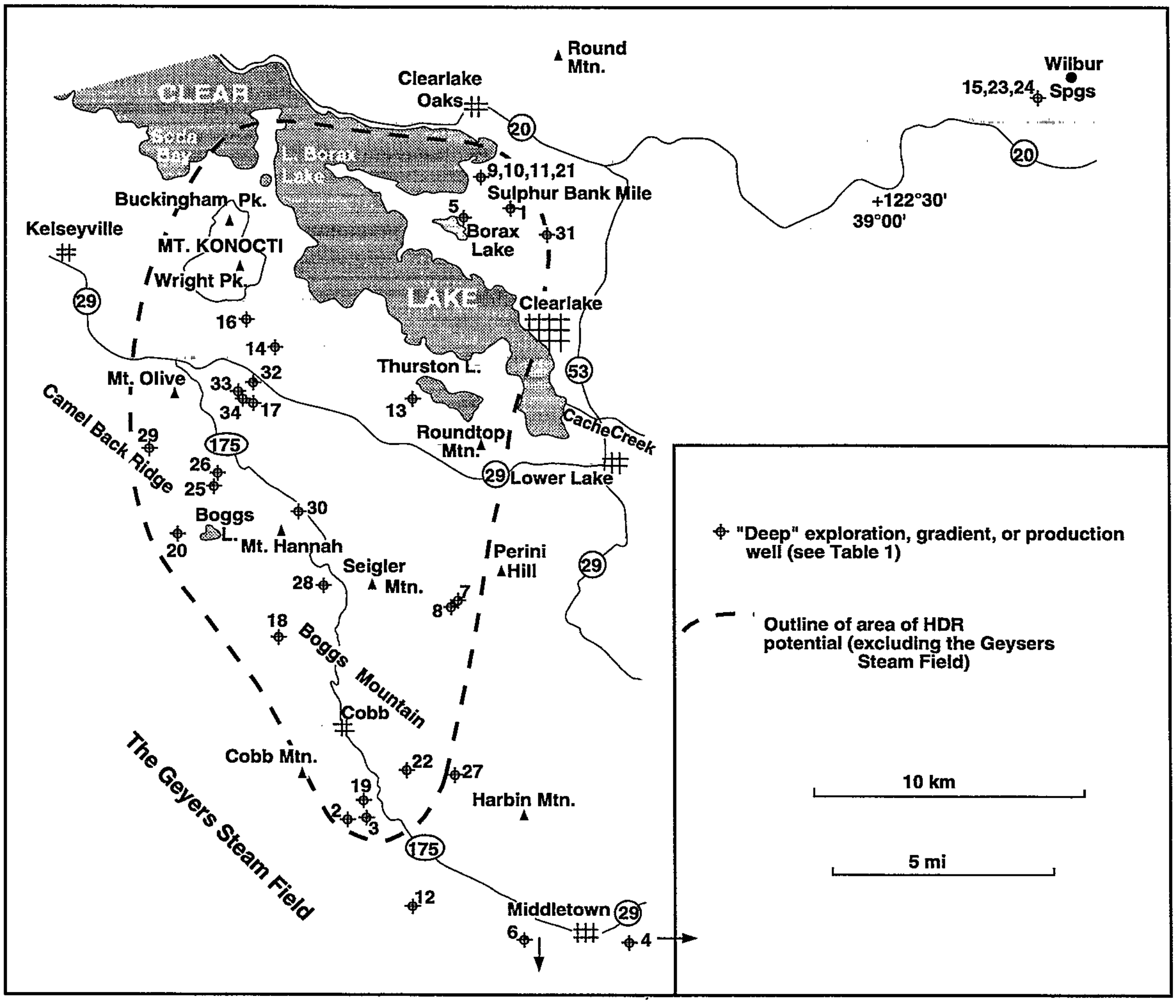

Fig. 7. Location map of "deep" wells drilled in The Geysers-Clear Lake region (data in Table 1).

3. Seismic Studies and Crustal Structure

Regional seismic studies suggest that the crust in the Clear Lake region is from 24 to 30 km thick, and is directly underlain by hot asthenosphere (Eberhart-Phillips, 1986; Mooney and Weaver, 1989; Benz et al., 1992; Castillo and Ellsworth, 1993). Inferred crustal cross sections of the region typically consist of 12 to 18 km of Franciscan-like material underlain by 12 to 14 km of gabbroic and mafic crystalline rocks (Fig. 9). However, recent wedge-tectonic models for the upper and middle crust beneath the Coast Ranges, based on seismic reflection and refraction studies (Fuis and Mooney, 1990; Wentworth and Zoback, 1990; Wentworth et al., 1984; Unruh and Moores, 1992) suggest that a metamorphic basement from the western Sierra Nevada and Klamath Mountains could underlie the Franciscan Complex in the Clear Lake region. Seismic velocities (~6.8 km/s) of the present lower crust (24-30 km) are consistent with it containing a significant amount of underplated or intraplated mafic crystalline rocks, but are probably too low for it to consist entirely of mantle material accreted to the base of the "slabless window" (cf. Furlong and Fountain, 1986).

In a study of P-wave velocity of northern California and southern Oregon, Benz et al. (1992) observed the largest low velocity anomaly in the region beneath Clear Lake, and attributed it to the presence of magma in both

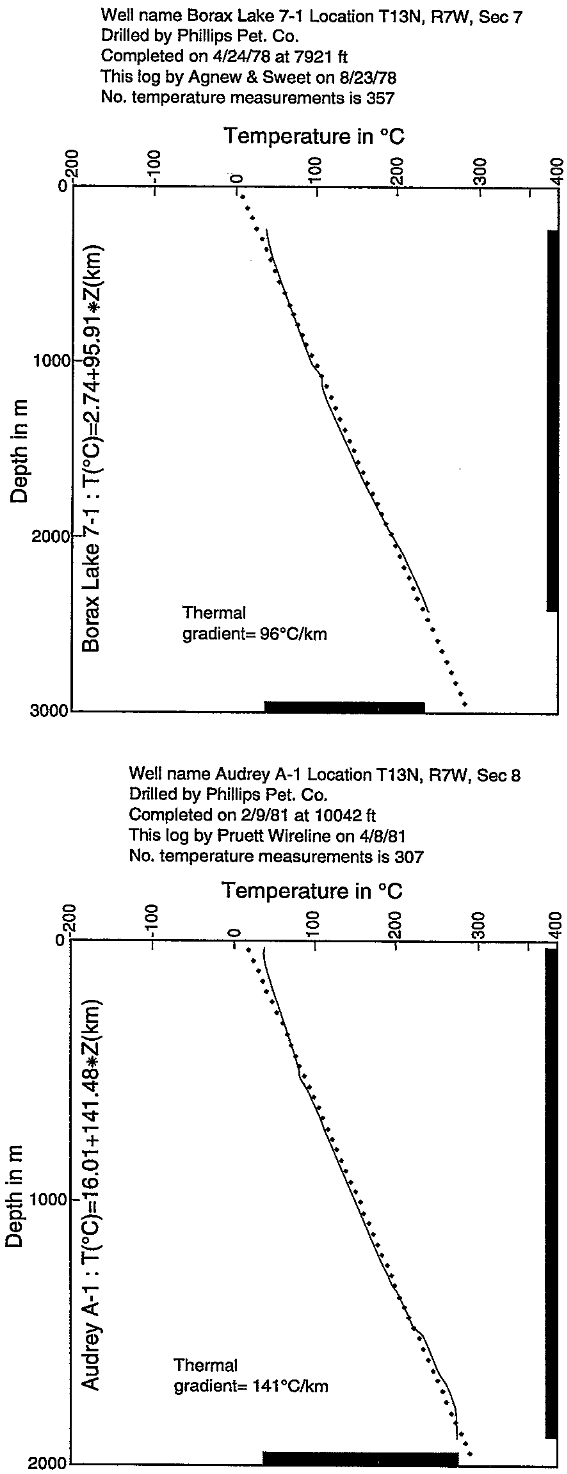

Fig. 8. Temperature profiles in the Audrey #1 and Borax Lake #7-1 wells (modified from Burns, unpub. Los Alamos Report). See Fig. 7 for well locations. Both wells display linear increases of temperature with depth and conductive thermal gradients approaching or exceeding 100°C/km. Dots show slope of the best fit gradient.

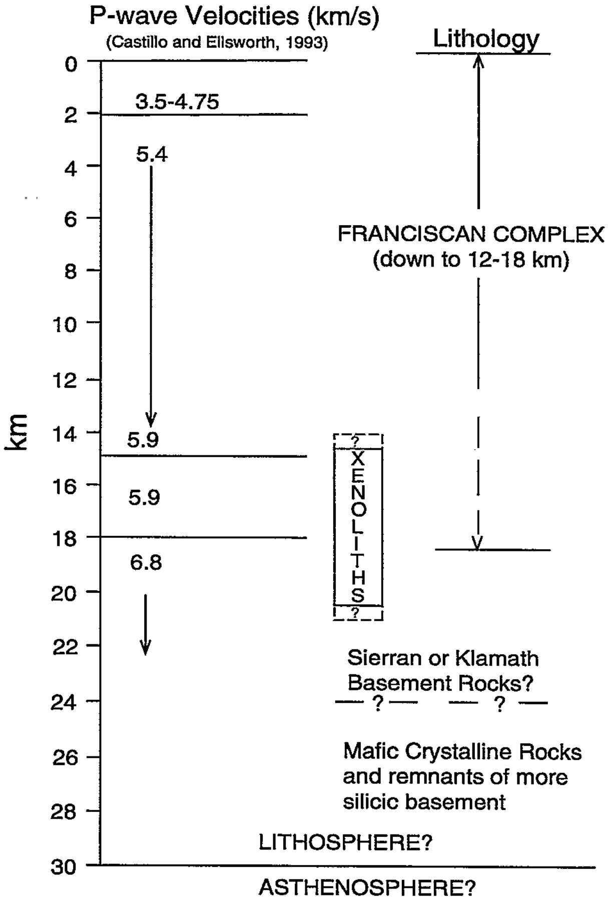

Fig. 9. Inferred crustal structure of the Clear Lake region. The P-wave velocities shown are from the model of Castillo and Ellsworth (1993), which was derived from study of earthquakes in the vicinity of the Maacama and Bartlett Springs Faults. The lithologies are generalized from those inferred or suggested by various authors (Wentworth et al., 1984; Wentworth and Zoback, 1990; Fuis and Mooney, 1990; Eberhart-Phillips, 1986; Mooney and Weaver, 1989; Castillo and Ellsworth, 1993). The crustal depths represented by xenoliths are inferred from thermobarometry (Stimac, 1993; Stimac, unpub. data).

the upper and lower crust. Similarly, an earlier seismic study of the Clear Lake region by Iyer et al. (1981) suggested the presence of magma extending from 4 to 30 km depth, but Eberhart-Phillips (1986) found no evidence for magma shallower than 7 km depth. Figure 10 shows velocity perturbations in The Geysers-Clear Lake area from inversion of teleseismic residuals (Iyer, 1988). The largest upper crustal P-wave anomalies (upper 30 km) correspond well to the negative gravity anomaly centered beneath Mt.

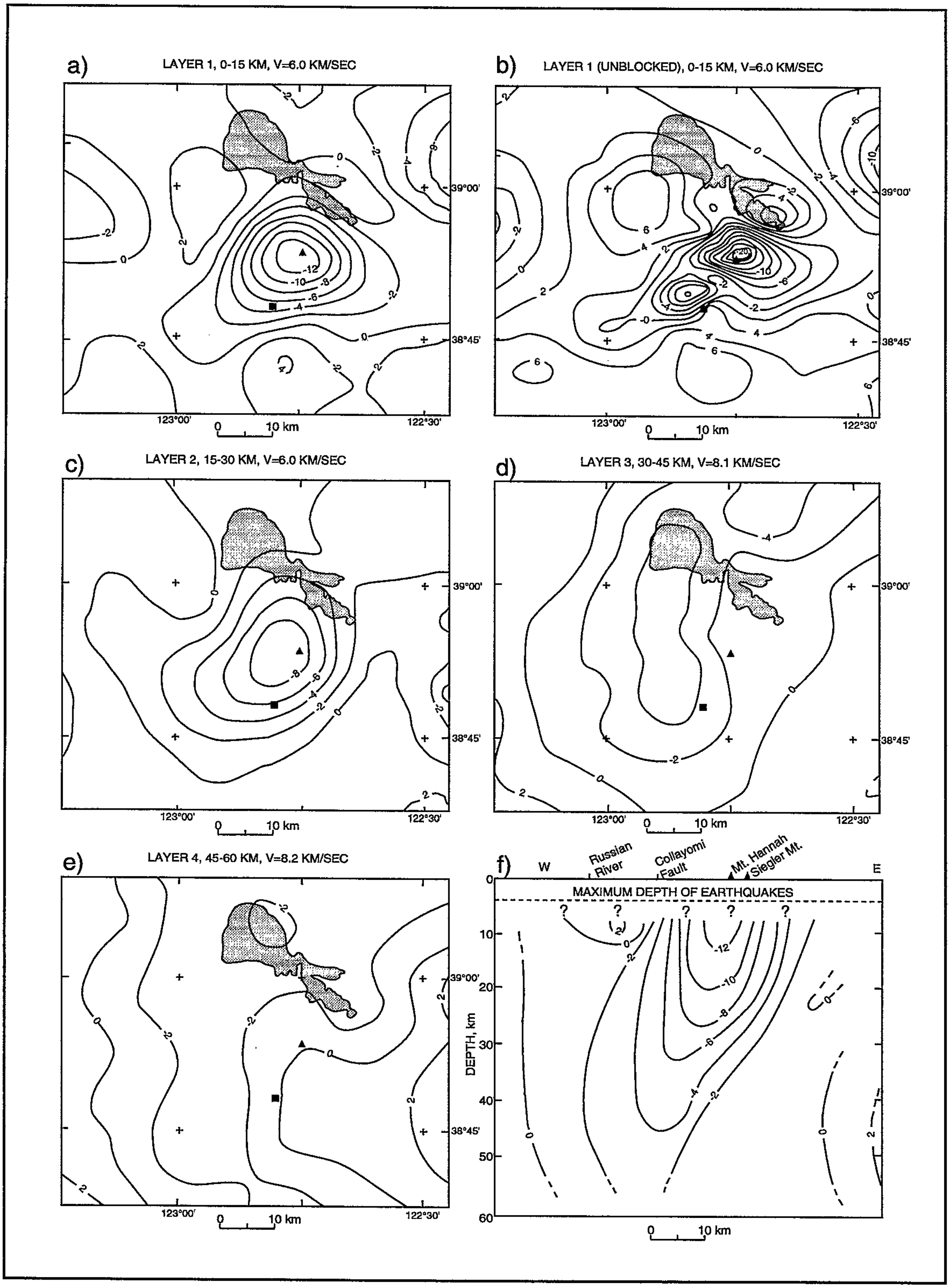

Fig. 10. Contour maps of teleseismic velocity residuals in The Geysers-Clear Lake region (from Iyer, 1988). Data were modeled using a 3-D inversion of four 15 km-thick layers. Individual blocks in the model were 10x10 km. The solid square and triangle mark the locations of The Geysers steam field and Mt. Hannah, respectively. (a) first layer in block format; (b) alternate first layer assigning each station a unique block (see Iyer, 1988 for details); (c-e) second to fourth layer, block format; (f) east-west cross section of the velocity anomaly.

Hannah, and with smaller perturbations in The Geysers area attributed to the steam reservoir. Velocity anomalies in the Mt. Konocti and Borax Lake areas are relatively small compared to the Mt. Hannah anomaly, and do not indicate the presence of large, active magmatic systems.

Recent small-scale tomography studies of The Geysers indicate that shallow silicic plutons such as the felsite have little velocity contrast with basement rocks dominated by graywacke, and thus other young granitoid plutonic rocks of the region would be difficult to image by this method (Ross et al., 1993; Zucca, pers. comm., 1994).

E. Thermal Waters and Gas Discharges

Water and gas samples from thermal springs, fumaroles, gas seeps, and nonthermal sources in The Geysers-Clear Lake region have among the widest range of chemical and isotopic compositions documented in any one region of geothermal activity (White et al., 1973; Barnes et al., 1973; Goff et al., 1977; Peters, 1991; Thompson et al., 1992; Donnelly-Nolan et al., 1993; Goff et al., 1993a, b; Goff and Janik, 1993; Janik et al., 1993). The distribution of springs along known faults, together with the wide range in water and gas chemistry indicate that no large hydrothermal system exists north of the Collayomi fault zone (Goff et al., 1993a, b). Small fault-localized reservoirs with inter-related hydrothermal fluids occur at Sulphur Bank Mine (~218°C), Wilbur Springs District (~140°C), and Sulphur Mound Mine (about 70°C). Moreover, the paucity of hydrothermal fluids in most exploration wells (Table 1) indicate relatively dry conditions in the deep subsurface. Interestingly, $^3He/^4He$ ratios from throughout the Clear Lake region suggest that a magmatic ^3He component is present locally, being highest beneath the Sulphur Bank Mine-Borax Lake area (R/R_A of 7.5 to 7.9; Janik et al., 1993; Goff et al., 1995).

III. THERMAL MODELING

A. Approaches and Assumptions

In order to test the effect of magma bodies on thermal gradients measured near the surface, we have numerically simulated temporal and spatial variation of crustal heat flow with time. Because numerous authors have shown results for analytical solution of 1-D linearized expressions of thermal diffusion, our goals were to explore thermal diffusion in a heterogenous media, where diffusion not only reflects local thermal gradients, but where diffusivities also vary spatially as a function of rock type, temperature, and emplacement history. For two dimensions where heat content is converted to temperature with division by heat capacity and density, we desire the solution for:

$$\frac{\partial T}{\partial t} = \frac{\partial k_x}{\partial x}\frac{\partial T}{\partial x} + \frac{\partial k_y}{\partial y}\frac{\partial T}{\partial y} + k_x\frac{\partial^2 T}{\partial x^2} + k_y\frac{\partial^2 T}{\partial y^2}, \quad (1)$$

where T is temperature, t is time, x and y denote lateral and vertical distance, respectively, and k_x and k_y are the lateral and vertical thermal conductivities, respectively. This equation is nonlinear, so we developed a user-interactive graphic interfaced numerical code to get an explicit solution (Wohletz and Heiken, 1992) by a finite differing technique. With continual checking of the solution stability, the simulated geological structure can be changed at any point during the simulation to achieve any one of a multitude of geologic scenarios Because we calculate heat flow in two dimensions, we have to assume an axisymmetrical system, which is discussed later.

The temperature dependence of thermal conductivities is modeled after Chapman and Furlong (1990) where:

$$k(t,z) = k_0\left(\frac{1+cz}{1+bT}\right). \quad (2)$$

For this equation thermal conductivity [$k(T, z)$] is a function of crustal depth (z) and temperature (T), k_0 = conductivity at 0°C, c is the crustal depth constant equal to $1.5x10^{-3}$/km, and *b* is the thermal constant equal to $1.5x10^{-3}$/km for the upper crust and $1.0x10^{-4}$/km for the lower crust. Other important aspects of this simulation are the initial thermal gradient, effects of the latent heats of fusion and crystallization of magma, and thermal convection by magmas and hydrothermal fluids.

The thermal gradient is obviously an important part of the solution as shown in the above heat flow equation, and it plays an important role for initial heat flow from a newly emplaced magma body as a function of emplacement depth. Without a realistic initial gradient, solutions would be unrealistic, especially for times shortly after magma emplacement. A user specified initial gradient is modeled by optimizing the heat flow into the bottom of the calculational mesh such that the desired gradient is stabilized within the mesh. This procedure includes constant horizontal gradients at the sides of the mesh and maintenance of a specified surface temperature (20°C).

The latent heats of fusion and crystallization are calculated between the temperatures of 650 and 1000°C, which represents the average solidification temperatures for a wide range of magma compositions most common in the Clear Lake region. Over this range, a linear variation in melt fraction with temperature is assumed, such that 350 kJ/kg of heat is either released or consumed for crystallization or melting, respectively. This assumption is a simplification in that the melt fraction and latent heat are strongly a function of phase composition, but this aspect of the problem is beyond the scope of our present study.

Magma is allowed to convect heat as a function of its composition and temperature. Where fully molten, magma convection is modeled with an arbitrary effective Nusselt number of 3 for silicic magmas and 10 for mafic ones, representing conservative average values for thermal Rayleigh numbers between 10^3 and 10^5. Where hydrothermal convection is specified in the host rock, again an effective Nusselt number is specified as 100. In this manner nearly isothermal gradients develop rapidly in hydrothermal systems, and over a prolonged time in magma bodies.

Numerous combinations of initial thermal gradient, magma body composition, structure, emplacement sequence, and emplacement depth were studied to verify consistency of the results with those of previous workers and observed thermal gradients in other volcanic terranes (e.g. Spera, 1980). The general conclusions reached from this survey are summarized in the next section. Assuming that the modeled geologic framework represents planes along axes of structural symmetry, the results of these 2-D simulations should be fairly representative of their 3-D analogues. This is also a simplification because of the complicated structure found in the Clear Lake region (McLaughlin, 1981; McLaughlin and Ohlin, 1984).

B. Key Variables and their Effects on the System

The thermal evolution of a magmatic system depends on many characteristics of the plutons and their country rocks (Spera, 1980; Furlong et al., 1991). The main factors related to the pluton include: (1) size and shape; (2) composition (including volatile content); (3) temperature of emplacement; (4) depth of emplacement; (5) emplacement history (single or multiple intrusions); (6) nature of internal heat flow (convection or conduction); (7) amount of radiogenic heat production; and (8) amount and timing of release of latent heat of crystallization. The main factors related to country rocks include: (1) ambient geothermal gradient; (2) thermal conductivity as a function of composition and depth; (3) crustal composition and structure (permeability and porosity); and (4) heat of fusion and endothermic reactions for rocks within the contact aureole. If advective or convective systems operate, then the nature of crustal structure (permeability and porosity) and geothermal fluid (1 or 2 phase) become more critical. The most important of these factors can be modeled numerically, but any model, no matter how sophisticated, is only as good as the physical constraints on the system. Furthermore, it is important to realize that numerical modeling of geologic heat flow cannot prove a specific petrologic history, it can only show which specific histories are physically plausible, and worthy of further consideration.

The following set of figures illustrate the effects of the major variables mentioned above on thermal gradients near the surface as a function of time for a system similar to Clear Lake (Figs. 11 to 17). These figures track the change in thermal gradient as measured in the upper 2 km of the crust above a cooling pluton. In most cases these simulations are for an axisymmetric magma body that is 7 km thick and 15 km long, emplaced with its roof at 7 km depth. The thermal gradients plotted are for a point directly over the center of the pluton. This point has the *maximum* thermal gradient for the given model conditions, and the gradient drops off toward the edge of the pluton at a rate dependent mainly on the geometry of the pluton, its depth of emplacement, and time after emplacement.

Pluton composition controls other variables such as emplacement temperature, internal flow (convection), heat of crystallization, and radiogenic heat production. Silicic plutons are generally emplaced at lower temperatures than mafic plutons. They are also less likely to convect due to much higher viscosity, and have much higher radiogenic heat productions. Thus silicic plutons generally cool more slowly than mafic plutons of equal size and have lower heat contents, despite having higher thermal conductivity. In our models we assign a single temperature-dependent thermal conductivity to plutons of silicic (3.1 W/m°C) and mafic (2.1 W/m°C) compositions that are likely to reflect a reasonable combination of these factors. Convection in magmas hotter than 650°C was simulated as described in the previous section. For plutons

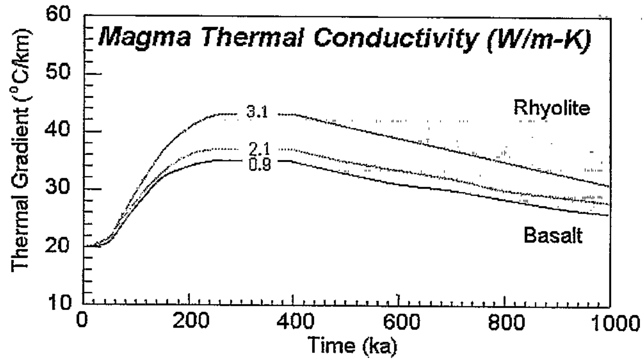

Fig. 11. Near-surface thermal gradient versus time for magmas of varying thermal conductivity. Abbreviations in this and the following figures are: Tgi= initial thermal gradient; Tim= initial magma temperature; Kh=host rock thermal conductivity; Km=magma thermal conductivity; Vm= cross-sectional magma volume in km^2; TOP=depth from surface to the roof of the magma body. In this model Tgi=20°C, Tim=1000°C, Kh=2.9W/m°C, Vm=7x15 km, and TOP=7 km.

emplaced at the same temperature and depth, varying the convection history and thermal conductivity of the pluton from 0.9 to 3.1 W/m°C has a small effect on the near-surface thermal gradient (Fig. 11).

The size and initial temperature of the pluton determine how much heat is added to the crust (Fig. 12). For example, a 1x15 km sill with a roof at 6 km produces about an 8°C increase in the near-surface thermal gradient in about 200 ka, whereas a 5x15 km sill increases surface thermal gradient over 20°C in 3200 ka. The shape of a pluton influences cooling rate. Shapes with high volume to surface area ratios (i.e. spheres, cubes) cool more slowly than shapes with low volume to surface area ratios (i.e. discs), but shape has little affect on the near-surface thermal gradient unless the pluton is shallow (<6 km). The emplacement temperature has a moderately large effect on near-surface thermal gradient (about a factor of 2 in our models for initial temperatures spanning those possible for magmas) (Fig. 13).

The depth of emplacement dramatically affects the thermal gradient at the surface (Fig. 14). Shallow plutons yield the highest peak thermal

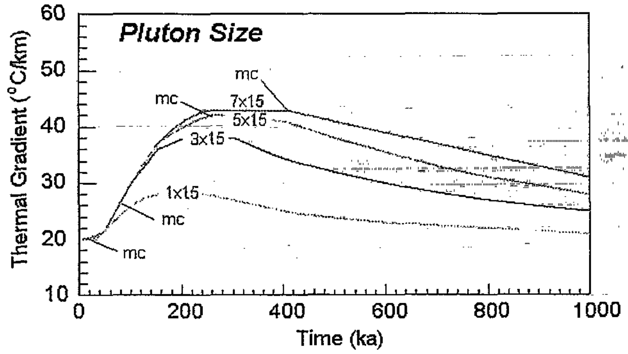

Fig. 12. Near-surface thermal gradient versus time for sills of varying volume and aspect ratios ranging from 1x15 km to 7x15 km. All plutons cool through the wet granite solidus (mc=magma crystallized) at times ranging from 10 to 500 ka. Km=3.1 W/m°C; other model parameters are the same as in Fig. 11.

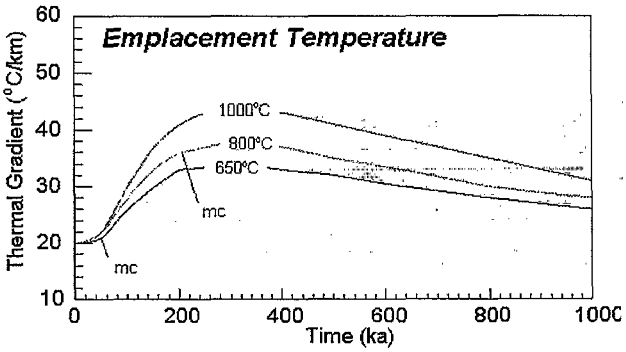

Fig. 13. Near-surface thermal gradient versus time for 7x15 km volume magma bodies with initial temperatures ranging from 650 to 1000°C. This temperature range is that typical for Clear Lake silicic magmas based on thermometry of Clear Lake silicic lavas (Stimac and Goff, 1994). Other model parameters are the same as in Fig. 11.

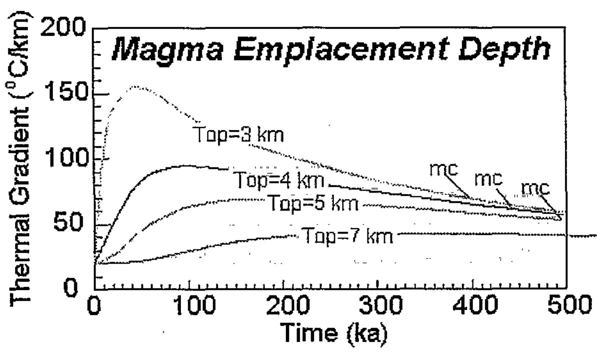

Fig. 14. Near-surface thermal gradient versus time for a 7x15 km magma body emplaced with its roof ranging from 3 to 7 km depth. "mc" indicates the time at which all the initial magma is below its solidus. Other model parameters are the same as in Fig. 11.

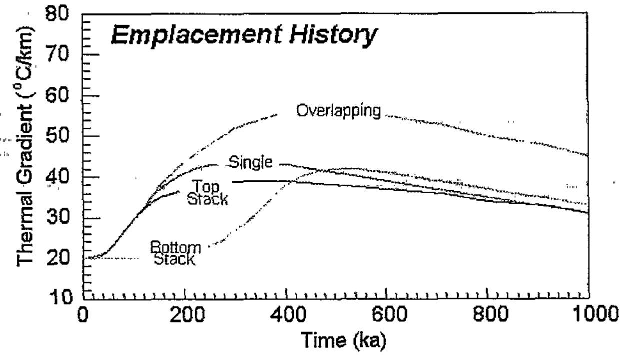

Fig. 15. Near-surface thermal gradient versus time for a 7x15 km magma body emplaced as a single event, and as four separate sills over a period of 400 ka. The separate sills were emplaced as adjacent, successively deeper intrusions. Other model parameters are the same as in Fig. 11.

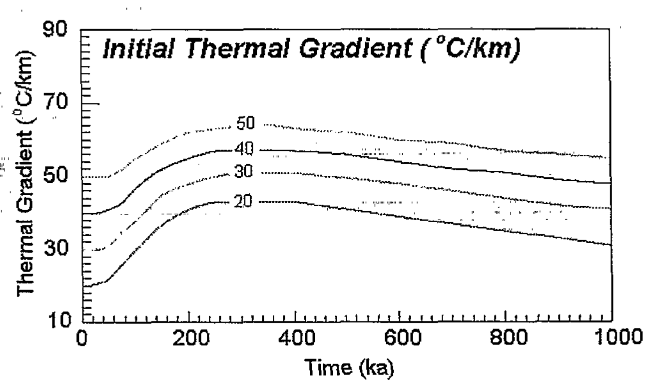

Fig. 16. Near-surface thermal gradient versus time for a 7x15 km magma body emplaced in rocks with an initial thermal gradient varying from 20 to 50°C/km. Other model parameters are the same as in Fig. 11.

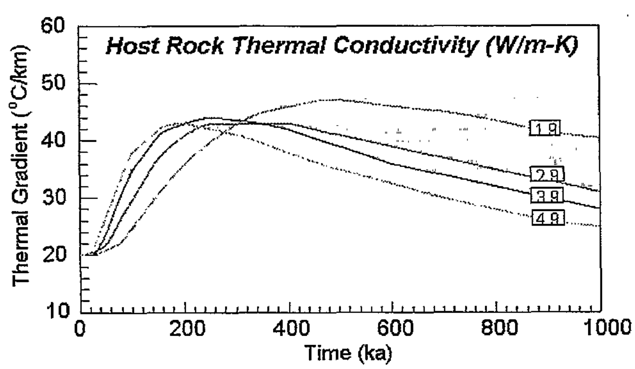

Fig. 17. Near-surface thermal gradient versus time for 7x15 km magma bodies emplaced in host rocks with thermal conductivities varying from 1.9 to 4.9 W/m°C. Other model parameters are the same as in Fig. 11.

gradient, but cool relatively quickly. Plutons emplaced at deeper levels produce lower peak thermal gradients at later times, but deliver a more sustained pulse of heat to the surface. Magmatic emplacement history has a variable affect on near-surface thermal gradients, depending on the geometry and time interval of emplacement (Fig. 15). For example, a single 7x15 km intrusion emplaced at 7 km depth yields a similar near-surface gradient as multiple 1-km-thick sills emplace over the same depth range from the top down (top stack), but a lower gradient than overlapping intrusions emplaced at 7 to 8 km (Fig. 15). This is because overlapping emplacements allow more heat to be added nearer the surface. Multiple 1-km intrusions stacked from the bottom up (bottom stack) yields near-surface gradients of a similar magnitude, but at a later time. Emplacement histories spanning more than 500 ka were not investigated in this study because silicic volcanic centers at Clear Lake generally have shorter age spans (Donnelly-Nolan et al., 1981).

Increasing the initial thermal gradient has the effect of shifting the thermal gradient versus time curve to higher gradients and shortening the time to maximum heat production at the surface (Fig. 16). Increasing the thermal conductivity of the host rocks has the effect of significantly shortening the time for heat to diffuse to the surface, and slightly lowering the peak thermal gradient observed because heat is lost to the surface more rapidly (Fig. 17). This effect leads to crossovers in the thermal gradient versus time plots. In other words, rocks with low conductivities are insulators, and therefore heat moves more slowly and is more effectively retained. The main result of changing host rock thermal conductivity is to dramatically change the time of arrival of the thermal peak at the surface, without significantly changing its magnitude. For example, varying thermal conductivity (K) between 1.9 and 2.9 W/m°C results in a 140 ka-shift in the thermal peak measured near the surface (Fig. 17).

C. Values of Key Parameters for the Clear Lake System

Now we wish to specifically model the Clear Lake magmatic system incorporating more complex histories consistent with the petrologic and geophysical constraints. These constraints allow a variety of scenarios ranging from single emplacement conductive models, to more complex models that include multiple emplacement of magma and zones of hydrothermal convection.

1. Basement Rock Types and Thermal Conductivities

Basement rocks in the Clear Lake region include the Late Jurassic to Late Cretaceous Franciscan Complex, the Middle Jurassic Coast Range ophiolite, the Late Jurassic to Early Cretaceous Great Valley sequence, and the 1.5 to <2.9 Ma Cache Formation (McLaughlin, 1981; Hearn et al., 1976, 1995). These rocks are overlain by a variable thickness (0 to 1200 m) of Pliocene to Pleistocene volcanic rocks. The typical thickness of volcanic rocks encountered in drill holes is about 600 m (Table 1). Accounting for the lower average thermal conductivity of young volcanic rocks (Walters and Combs, 1989) tends to delay the transmission of heat to the surface, and is incorporated in some models that follow.

In the Clear Lake region, the Franciscan Complex is composed of a heterogeneous assemblage of intensely deformed and weakly-to-strongly metamorphosed graywacke, shale, chert, and mafic igneous rocks (McLaughlin, 1981; McLaughlin and Ohlin, 1984). Discrete fault-bounded portions of the Franciscan Complex range from weakly metamorphosed (prehnite-pumpellyite facies) to strongly recrystallized (blueschist facies), but low temperature and pressure assemblages predominate. In the Clear Lake region the Great Valley Sequence consists of a basal zone of reworked detritus from the Coast

Range Ophiolite and Franciscan Complex, overlain by graywacke, mudstone, and comglomerate. The Coast Range Ophiolite consists primarily of variably serpentinized gabbro, metabasalt, and ultramafic rocks with the latter most abundant.

Thermal conductivity measurements for the Clear Lake region summarized by Walters and Combs (1989) are reproduced in Table 2. They showed that the thermal conductivity of graywacke, the most abundant rock type in deep wells, displayed no systematic variation from the surface to a depth of at least 600 m (and in some cases 3000 m), however, measurements were made at room temperature, and therefore do not reflect changes in thermal conductivity as a function of increased temperature and pressure with depth. We used the equation of Furlong and Chapman (1991) to approximate the affect of temperature on thermal conductivity in our modeling (Eq. 2). Their equation does not differ considerably from other approximations (e.g. Balling, 1976).

Table 2. Summary of rock thermal conductivities (W/m°C) from The Geysers-Clear Lake region.

Lithology	n	Range	Median	Mean	Error
Basalt	12	1.42-2.18	1.72	1.72	0.06
Andesite	27	1.21-2.14	1.59	1.68	0.05
Dacite	6	1.59-1.80	1.68	1.68	0.03
Rhyolite	23	0.75-1.80	1.00	1.09	0.01
Sandstone	59	2.21-3.30	2.60	2.68	0.03
Shale	35	1.92-2.68	2.22	2.22	0.03
Serpentine	85	1.84-3.43	2.60	2.64	0.04
Greenstone	73	1.88-3.14	2.47	2.47	0.04
Graywacke	563	2.43-3.77	3.02	2.97	0.01
Chert	21	3.02-4.15	3.56	3.56	0.06

Data from Walters and Combs (1989).

Because the structural relations and distribution of these rock types in the subsurface are not well known, we used a weighted average of conductivities in all of our models. We assumed that, on average, basement rocks consist of 80% graywacke, 8% serpentinite, 7% greenstone, 3% shale, and 2% chert. This yields an average thermal conductivity of ~2.9 W/m°C. We included low thermal conductivity volcanic cover rocks in some models of the Mt. Konocti area, using a weighted average Kh of ~1.6 W/m°C based on values for volcanic rocks in Table 2 and their relative abundances in outcrop. Horizontal and vertical components of this conductivity were assumed to be equal in all conductive models, since there is no evidence for large convective regimes north of the Collayomi fault zone and beneath the main Clear Lake volcanic field (Goff and Decker, 1983; Walters and Combs, 1989; Thompson et al., 1992; Goff et al., 1993a, b).

2. Magma Compositions and Thermal Conductivites

The thermal conductivities of magmas are less well known than those of rocks. The only data for molten rocks are from Murase and McBirney (1973), supplemented by recent data on molten diopside (Davaille et al., 1993). Conductivities for magmas are a function of two mechanisms of heat transfer (lattice conduction and radiative conduction), which are in turn dependent on magma composition and temperature. Thermal conductivities generally decline with increasing temperature (declining crystal content) up to temperatures approaching the liquidus for any given magma. Thermal conductivities increase with increasing temperature once a given magma reaches temperatures near its liquidus. Silicic to intermediate glasses show a smooth increase in thermal conductivity with increasing temperature (Murase and McBirney, 1973). Suspended crystals would act to lower the thermal conductivity in silicic magmas. In our modeling we assumed thermal conductivities of 3.1 W/m°C for rhyolitic magmas (similar to typical granitic rocks) and 2.1 W/m°C for mafic magmas. A recent study by Davaille et al. (1993) on molten diopside suggests that these thermal conductivities could be up to an order of magnitude too high, however we consider it safer to use the larger data set of Murase and McBirney (1973) until corroboration of lower conductivities come from additional sources. In general, lowering thermal conductivities of basaltic magmas by an order of magnitude would result in cooling times 4 to 5 times longer for mafic intrusions (Davaille et al., 1993). However as shown in Fig. 11 thermal conductivity of magma has relatively little affect on near-surface gradients in our models.

3. Magma Emplacement Temperatures

Magma emplacement temperatures were inferred from geothermometry based on mineral assemblages and compositions (Stimac, 1991; Stimac and Goff, unpub. data). These result indicate that Clear Lake mafic magmas were generally erupted at temperatures from 1015 to 1170°C (most from 1150 to 1170°C), whereas silicic magmas were erupted at a wide range of temperatures, correlating with their crystal content and the extent of their interaction with more mafic magmas (Stimac, 1991; Stimac and Goff, unpub. data). Crystal-poor rhyolite and rhyodacite lavas that show evidence of interaction with mafic magma just prior to eruption yield temperatures >1000°C, whereas crystal-rich rhyolitic and rhyodacitic lavas yield temperatures from 650 to 750°C. Since silicic volcanism in a given center commonly began with eruption of crystal-poor rhyolite, initial magma temperatures in the models ranged from 900 to 1000°C, whereas mafic magmas were emplaced at 1150 to 1170°C.

4. Regional Thermal Gradient

The best estimate of the regional thermal gradient can be derived from heat flow data presented in Lachenbruch and Sass (1980). As discussed above, they showed that the Coast Ranges have a relatively high heat flow compared to adjacent areas (Fig. 5). Except for local anomalies, heat flow throughout the province is relatively uniform, averaging about 2 HFU (83 mW/m^2). If we assume an average thermal conductivity of 2.9 W/C°m, then this implies thermal gradients of about 28°C/km. Lachenbruch and Sass (1980) ascribe this relatively high regional heat flow to thin crust and unusually hot mantle, with additional dike injection, and possible minor contributions from frictional heating along faults of the San Andreas system.

In The Geysers-Clear Lake region heat flow is greater than or equal to 4 HFU (167 mW/m^2), at least twice the regional average (Walters and Combs, 1989). This heat flow anomaly is almost certainly due to a combination of deep mafic intrusion, and eventual development of more shallow silicic magma bodies, and implies a thermal gradient of at least 58°C/km assuming a thermal conductivity of 2.9 W/C°m. We have constructed 6 and 8 HFU contours in the region based on a number of deep wells extending from Mt. Hannah to Borax Lake (Table 1 and Fig. 7).

An alternative approach to establishing the initial thermal gradient for models of the silicic magmatic system is to simulate the affect of mafic intrusion into the deep crust based on tectonic models of mafic magma production (Liu and Furlong, 1992). Numerical simulations by Liu and Furlong (1992) indicate that mantle upwelling would result in 30 to 40% partial melting of peridotite at the top of the slabless window, depending on the thickness of the overlying crustal lid. This could result in generation of a layer of basaltic magma 4 to 5 km thick in the wake of the Mendocino Triple Junction. Simple conductive models for emplacement of a mafic sill 4 km thick and 25 km long at 12, 15, and 20 km depth imply that mafic intrusion must occur at relatively shallow levels to significantly affect surface thermal gradient on a reasonable time scale (Fig. 18). For example, a 4x25 km mafic sill emplaced with its roof at 12 km raises surface thermal gradient from 20 to 45°C in about 2 Ma. Limited evidence from xenoliths in Clear Lake lavas indicates that mafic intrusion occurred at depths as shallow as 14 to 21 km (Stimac, 1993; Fig. 4). Assuming basaltic magmas were produced by decompression melting in the slabless window, and intruded in significant volume at this depth range, mafic magmas could have raised the thermal gradient to between 25 to 40°C/km over about 2 to 3 Ma, overlapping above the 28°C/km estimated from regional heat flow measurements described above.

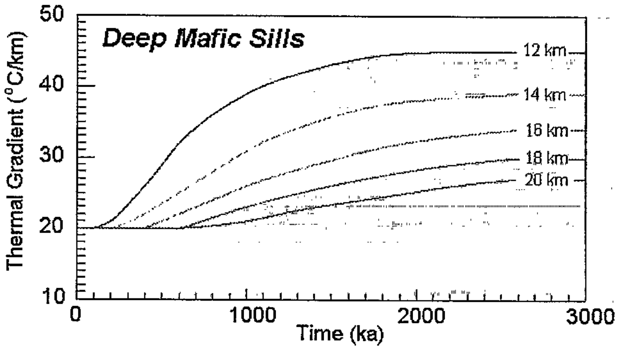

Fig. 18. Near-surface thermal gradient versus time for a 4x25 km volume mafic magma bodies emplaced with their tops at depths ranging from 12 to 20 km.

Based on the nearest point of overlap of the two approaches outlined above, we have used initial thermal gradients of 28-30°C/km in our models. As shown earlier, varying the initial thermal gradient between 20 and 50°C/km results in a relatively small change in the timing and magnitude of peak near-surface thermal gradient compared to variables such as magma body size and depth of emplacement (compare Figs. 14 and 16).

D. Constraints on Magma Production

1. Empirical Estimates

Estimates of magma production rates were summarized by Shaw (1985), who concluded that 0.01 km^3/yr (10^4 km^3/Ma) is an average value for all compositions. Shaw (1985) further estimated average rate of production of silicic magma at 0.001 km^3/yr (10^3 km^3/Ma), however rates for large silicic systems such as the Timber Mountain/Oasis Valley and Yellowstone centers may be as high as 0.03 km^3/yr (e.g. Christiansen, 1984).

2. Estimates for the Clear Lake Region Based on Tectonic Models

Tectonic models for mafic magma production in the Clear Lake area can be extended to silicic compositions assuming that the depth of mafic intrusion is known and silicic magmas are generated by anatexis. Liu and Furlong (1992) estimated the amount of silicic crustal melt that could be produced by underplating or intrusion of basalt into the crust based on the slabless window hypothesis. Assuming a migration rate of 5 cm/yr, basaltic *underplating* was predicted to yield peak silicic magma production rates of about 2000 km^3/Ma (0.002 km^3/yr), whereas intrusion of basaltic magma *into* the crust would generate a much higher volume of crustal melt. Their model for injection of a 2 km-thick sill at 20 km depth yielded peak silicic magma production rates up to 8000 km^3/Ma (0.008 km^3/yr) (Fig. 19). The total volume implied by these rates are about 1010 km^3 and 3200 km^3 for underplating and intracrustal intrusion, respectively, with over 90% of this being produced within 1 Ma of basalt intrusion. This yields an average rate of 0.001 to 0.003 km^3/yr over this short period. The calculations of Liu and Furlong (1992) are broadly consistent with other heat flow work (Jamieson, 1976), and with regional seismic studies that indicate the crust in the Clear Lake region is <30 km thick, and contains anomalous low-velocity zones due to the presence of magma in both its upper and lower portions (Benz et al., 1992).

E. Percentage of Magma Erupted

Smith and Shaw (1975, 1978) and Smith (1979) suggested that there is a correlation between caldera size, ash-flow tuff volume, and the volume of the source magma chamber (Fig. 20). They concluded that silicic magma bodies are *typically* ten times the volume of the dense-rock-equivalent of their eruptive products. Crisp and

Spera (1984) concluded that ratio of magma erupted to magma emplaced for intracontinental silicic systems ranges from about 1:4 to 1:16,

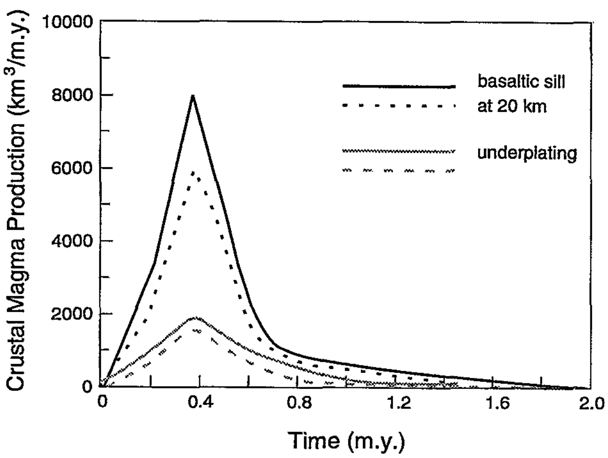

Fig. 19. Predicted rates of silicic magma production based on thermal modeling of Liu and Furlong (1992). The two production rates are based on: (1) basaltic underplating at 30 km depth, and (2) a 2-km-thick sill injected at 20 km depth. The solid and dashed curves assume nonlinear and linear melting relationships respectively (see Liu and Furlong, 1992). These rates should be considered maximums for the given scenarios because the lower crusts was assumed to consist of metapelitic rocks. A lower crust consisting in part of mafic crystalline rocks would undergo significantly less partial melting.

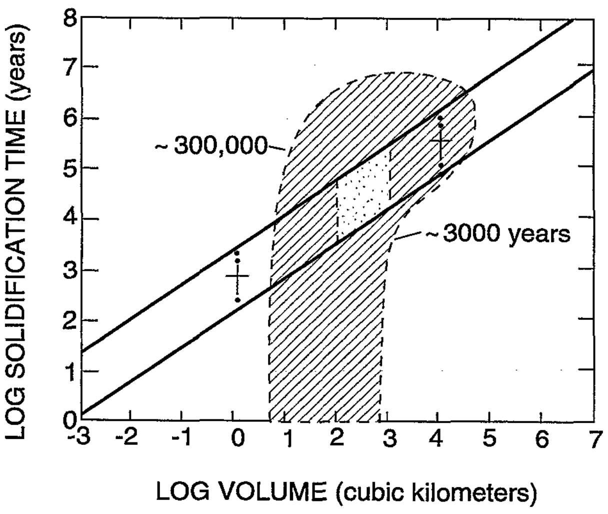

Fig. 20. Log solidification time versus log volume (modified from Fig. 1b from Shaw, 1985). Solidification refers to crystallization from 850 to 650°C for a magma body with its roof at 4 km depth.

whereas this volume ratio for subduction-related systems range from 1:6 to 1:13.

It is important to note that these estimates do not include the volume of mafic root zones associated with these systems. A number of workers have argued that all continental magmatism is ultimately related to intrusion of mantle-derived basalt (e.g. Hildreth, 1981). If one includes the volume of mafic magma that may be involved in genesis of more silicic magmas, then the volume of magma emplaced in the crust may be from 10 to 1000 times that erupted, depending on the proportion of underplating to intracrustal intrusion (Shaw, 1980, 1985; Hildreth, 1981).

We suggest that large systems such as the Timber Mountain and Valles calderas emphasized in these studies tend to erupt a higher proportion of the total magma emplaced than smaller systems such as Clear Lake. For example, the significant isotopic differences between individual members of rapidly emplaced large-volume ash-flow tuff sequences from the Timber Mountain volcanic center are consistent with a high proportion of the magma present in the original chambers being evacuated before emplacement of subsequent magmas (Sawyer et al., 1990; Farmer et al., 1991). Similarly, the observation that post-caldera rhyolite lavas of the Valles caldera appear to represent small, discrete batches of magmas chemically and isotopically unrelated to the preceding caldera-forming eruption (Spell et al., 1993) suggests that little "eruptible" magma remained in the Bandelier chamber after formation of the Tshirege member ignimbrites.

On the other hand, silicic centers at Clear Lake produced sequences consisting of numerous chemically and isotopically related lavas, with a total eruptive volume <100 km^3. This volume is less than that estimated for the Geysers felsite alone. As shown below, it is impossible to model the high heat flow associated with the Clear Lake system assuming that 10% of the total magma emplaced was erupted.

F. The Mt. Konocti Volcanic Center

1. Introduction

Eruption of rhyolite and dacite magma in the Mt. Konocti area from 0.65 to 0.25 Ma produced about 45% of the total volume of magma erupted in the Clear Lake volcanic field (Donnelly-Nolan et al., 1981). This episode of volcanism was chosen for thermal modeling because the ages and petrologic character of individual eruptions are well studied (Donnelly-Nolan et al., 1981; Hearn et al., 1981; Stimac, 1991). A large number of magma emplacement models were tried for the Mt. Konocti episode of volcanism, some of which are summarized in Tables 3 and 4. These models can be divided into three major groups: (1) single emplacement, conductive models (MK1); (2) multiple emplacement, conductive models (MK2); and (3) single emplacement, convective models (MK3). We focus on the case of an episodically recharged magma body with dimensions of 7x15x15 km. This volume was chosen for detailed models because it broadly fits the geophysical and petrologic models for magma bodies at Clear Lake, and generates the observed near-surface thermal gradient in some models that include volcanic cover and late mafic intrusion. A 7x15x15 km magma body yields a extrusion:intrusion ratio of 1:35, which is considered a good estimate of magma volume associated with the Mt. Konocti system, including mafic contributions (Stimac, 1991).

The insulating affect of volcanic rocks with low thermal conductivity is generally not an important factor in the Clear Lake volcanic field, but could be in the Mt. Konocti area, where the thickness of the volcanic pile is locally over 1 km (Hearn et al., 1976, 1995), or where substantial thickness of shale is present (Stanley and Blakely, 1995). Many of the models described below were run with and without volcanic cover

Table 3. Summary of Mt. Konocti models.

Model	Summary of Model Conditions	Figure
	Single-emplacement, conductive models	
MK1a	2x15 km with roof at 6 km depth	21
MK1b	7x15 km with roof at 6 km depth	21
MK1c	10x15 km with roof at 6 km depth	21
MK1e	2x15 km with roof at 4 km depth	22
MK1i	7x15 km with roof at 3 km depth	22
MK1f	7x15 km with roof at 4 km depth	22
MK1g	10x15 km with roof at 4 km depth	22
	Multiple-emplacement, conductive models	
MK1d	same as MK1c, with 1 km low-k cover added at 200 ka elapsed time	21
MK1h	same as MK1f with 1 km low-k cover added at 200 ka elapsed time	22
MK1j	same as MK1i with 1 km low-k cover added at 250 ka elapsed time	22
MK2a	7x15 km with roof at 4 km by multiple silicic emplacements according to schedule in Table 4	21
MK2-MD1	7x15 km with roof at 4 km depth with 4 equally-spaced 1-km-thick mafic dike added at 500 ka elapsed time	23
MK2-MD2	same as MK2-MD1 with 1 km of low-k cover added at 250 ka elapsed time	23
MK2-MD3	same as MK2-MD1 with 2 km of low-k cover added at 0 and 250 ka elapsed time	23
MK2-SS1	7x15 km with roof at 4 km depth with a 2x15 km silicic sill added at 4 km depth at 500 ka elapsed time, and with 1 km of low-k cover added at 250 ka	23
	Single-emplacement, convective models	
MK3-1a,b	same as MK1a with a 2-km-thick convective zone over top magma from 0 to 400 ka elapsed time	24
MK3-2a,b	same as MK1a with three 1-km wide convective "faults" from magma body roof to surface	24
MK3-2c,d	same as MK3-2a,b with 1 km low-k cover added at 250 ka elapsed time	24

All times are "elapsed times" from initial emplacement 600,000 years ago.

Table 4. Emplacement schedule of multiple intrusion model MK2a.

Elapsed Time (ka)	Dimensions (km)	Magma Type and Temp. (°C)	Volume (km^3)
0	1x15(x15)	Silicic, 950	225
40	1x4(x15)	Mafic, 1170	60
50	1x4(x15)	Mafic, 1170	60
60	1x4(x15)	Mafic, 1170	60
100	1x8(x15)	Mafic, 1170	120
140	1x8(x15)	Mafic, 1170	120
160	1x8(x15)	Mafic, 1170	120
200	1x8(x15)	Mafic, 1170	120
240	2x8(x15)	Mafic, 1170	240
260	2x8(x15)	Mafic, 1170	240
290	2x8(x15)	Mafic, 1170	240

Volumes assuming axisymmetric magma body and replenishment geometry.

rocks, and as will be seen, adding low thermal conductivity rocks to the upper portion of the mesh significantly delays the timing of the peak near-surface thermal gradient, as well as increasing its magnitude. First we will consider single emplacement, conductive models that assume emplacement schedules based on the observed timing of eruptions in the Mt. Konocti area (Donnelly-Nolan et al., 1981).

2. Conductive Models

Conductive models assuming a wide variety of magma volumes and emplacement depths, indicates that observed near-surface gradients in the Mt. Konocti area cannot be reproduced by any simple emplacement scheme cued to eruption history and volume, unless low thermal conductivity cover rocks, or late intrusion are also assumed (see Table 3 for a summary of models shown in Figs. 21 to 24). For the Mt. Konocti area, the eruption volume (about 45 km^3) implies a magma body of 450 km^3 using the "1:10 rule" of Smith and Shaw (1975, 1978), or 2x15x15 if scaled to the spacial distribution of lavas erupted from 650,000 years ago to the present. A single magma body of this size emplaced at 6 km depth would crystallize in only 60,000 years, and yield a maximum near-surface thermal gradient of 51°C/km about 160,000 after emplacement (see Model MK1a in Fig. 21). The maximum near-surface thermal gradient drops to 37°C/km after 600,000 years, *equivalent to the present time in all Mt. Konocti models.* This is well below the observed near-surface thermal gradients in the Mt. Konocti area today (Walters and Combs, 1989).

Even much larger magma bodies emplaced at 6 km depth fail to produce observed gradients. Models with dimensions of 7x15x15 (1575 km^3) and 10x15x15 (2250 km^3) yield peak near-surface gradients of only about 60°C/km at about 300,000 years, dropping to 52 to 54°C/km at 600,000 years elapsed time (Model MK1b and MK1c in Fig. 21). Adding 1 km of low thermal conductivity cover at 250 ka increases gradients by about 10°C/km and delays the peak gradient after about 100,000 years, but still fails to produce observed gradients (Model MK1d in Fig. 21).

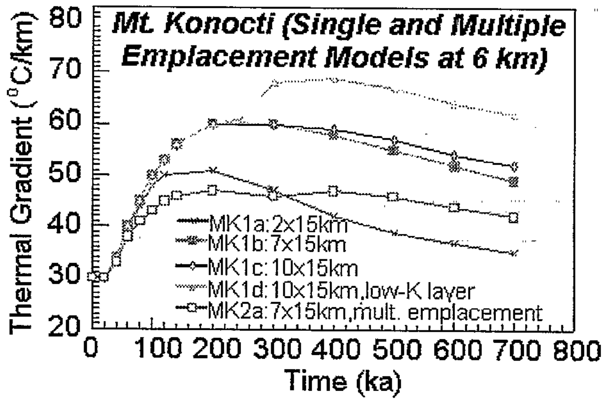

Fig. 21. Near-surface thermal gradient versus time for conductive models of the Mt. Konocti volcanic center with magma bodies emplaced at 6 km depth and 2-d volumes ranging from 2x15 to 10x15 km (see Table 3 for model summaries). Models MK1a to MK1d are single-emplacement models, whereas model MK2a is a multiple emplacement model with emplacements qued to the spatial and temporal eruption history of the Mt. Konocti center (see Table 4 for summary). Model MK1d is the same as model MK1c, but with a 1 km-thick layer of low thermal conductivity rocks added at 250 ka elapsed time. Other models parameters are Tgi=30°C/km, Tim=1000°C, Km=3.1 W/m°C, and Kh=2.9 W/m°C.

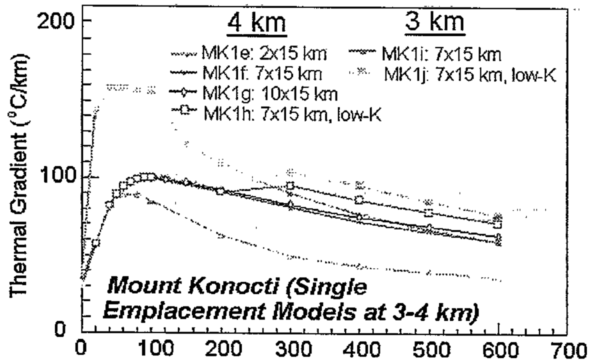

Fig. 22. Near-surface thermal gradients versus time for single-emplacement conductive models of the Mt. Konocti volcanic center with magma bodies emplaced at 3 to 4 km depth and 2-D volumes ranging from 2x15 to 10x15 km (see Table 3 for model summaries). One-km-thick layers of low thermal conductivity volcanic cover were added to models MK1h and MK1j at 250 ka elapsed time. Other model parameters are the same as in Fig. 21.

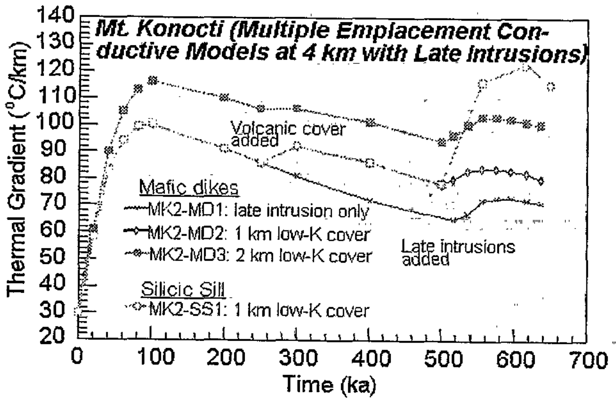

Fig. 23. Near-surface thermal gradients versus time for multiple-emplacement conductive models of the Mt. Konocti volcanic center with late intrusion. Magma volume and emplacement depth are 7x15 km and 4 km in all models. Model MK2-MD1 to simulate intrusion of three mafic dikes at 500 ka elapsed time, with 0 to 2 km of low thermal conductivity cover (see Table 3 for model summaries). Model MK2-SS1 simulates intrusion of a 2x15 km silicic sill at 4 km depth at 500 ka, with 1 km of low thermal conductivity cover added at 250 ka. Other model parameters are the same as in Fig. 21.

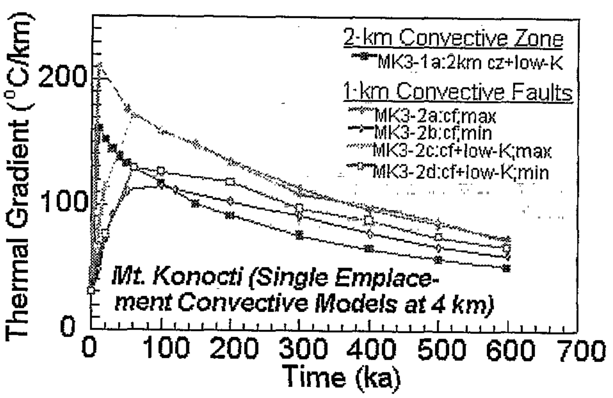

Fig. 24. Near-surface thermal gradients versus time for single-emplacement convective models of the Mt. Konocti volcanic center. Magma volumes and emplacements depth are the same as in MK1a. Model MK3-1 simulates a 2-km-thick zone of convection above the magma body. Convection begins at 600 ka and shuts off at 200 ka. This model also has a 1-km-thick layer of low thermal conductivity volcanic rocks at the surface. Models MK3-2a to MK3-2d share three vertically-directed, convective zones 1-km-wide extending from the roof of the magma body to the surface. These zones are simulate faults. Curve MK3-2a is the maximum gradient (nearest grid block to fault) and curve MK3-2b is the minimum gradient observed over the magma body (but farthest grid block from fault). MK3-2c and d are the same as MK3-2a and b, but have a 1-km-thick layer of low conductivity volcanic rocks at the surface.

Assuming the same magma volumes, but shallower emplacement depths (3 to 4 km), produces peak gradients in the range of those observed in the Mt. Konocti area, but at elapsed times of ~300,000 years (Fig. 22). By the present (or 600,000 years elapsed time), thermal gradients have dropped to ~63°C/km without volcanic cover (Models MK1e, f, g, and i), and ~74°C/km with addition of 1 km of low thermal conductivity cover at 250 ka (Model MK1h and j). These values are close to, but slightly below the observed thermal gradients observed in "deep" wells in the Mt. Konocti area, which range from 86-110°C/km (Table 1 and Fig. 7).

We also investigated the thermal affects of later mafic intrusion into the system, as indicated by eruption of mafic lavas whose ages are estimated at from 100,000 to 10,000 years B.P. in the Mt. Konocti area (Donnelly-Nolan et al., 1981; Hearn et al., 1988). Some of these lavas contain sanidine and quartz xenocrysts similar to those found in earlier erupted dacites, suggesting that the mafic magmas mixed with, or assimilated silicic magma before eruption. In models designated MK2-MD, we added four 1-km-wide mafic dikes to a 7x15 km chamber with its roof at 4 km (Fig. 23). The dikes, which were added at 500 ka, project to within 4 km of the surface and are spaced 4 km apart. It can be seen that these additions prolong the period of high near-surface gradients, but only achieve the observed gradients when low thermal conductivity volcanic cover is also assumed (MK2-MD2 and 3 in Fig. 23). In model MK2-MD2, a 1 km layer of low thermal conductivity rocks was added at 250 ka, whereas in MK2-MD3, two 1 km-layers were added at 0 and 250 ka, respectively.

The affects of late intrusion of a large silicic sill at shallow depth is shown in model MK2-SS1 in Fig. 23. The model incorporates the addition of a 2 km-thick sill at 4 km depth at 500 ka. The model illustrates that late, shallow intrusion of silicic magma not manifested by eruption could dramatically increase near-surface gradients within 100 ka elapsed time. Although this model achieves observed thermal gradients, it should be noted that there is no clear geologic or geophysical evidence for an active magma body underlying the Mt. Konocti area.

3. Convective Models

Model MK3-1 is also similar to MK1a, but incorporates a 2-km-thick zone of hydrothermal convection directly above the magma body for the first 400 ka after emplacement (Fig. 24). This zone is inferred to result from fracturing and hydrothermal convection in a fixed region above the pluton, analogous to that observed above The Geysers felsite (Hulen et al., 1993). The time and depth constraints are imposed because while there is some evidence of past hydrothermal activity (e.g. Bell Mine, Sulphur Mound Mine), no hydrothermal systems of any consequence are currently active in the Mt. Konocti area, even in holes as deep as 3 km (Table 1; Goff and Decker, 1980; Goff et al. 1993a, b; Goff and Janik, 1993). Because convective transport is very rapid compared to conductive transport, the results of this model are very similar to that of MK1j. That is, the overall effect of the hydrothermal zone is similar to moving the magma body up nearly an equivalent distance in the crust (Fig. 24).

Model MK3-2 uses the same magmatic system as MK1a but incorporates three convective zones 1 km wide above a magma body. These convective zones simulate broad, fault-controlled hydrothermal systems and sustain high near-surface thermal gradients and very high temperatures for the duration of the model (Fig. 24). Regions of high gradient begin to develop adjacent to these zones through a combination of convective and conductive heat transport, significantly increasing thermal gradients between convective zones for the duration of the model. The possible role of hydrothermal convection at Clear Lake is explored in more detail below in models of the Borax Lake area.

4. Conclusions from Mt. Konocti Models

After considering a large number of possible models, we conclude that some combination of shallow silicic magma bodies, low eruption:intrusion ratios, late intrusion of mafic or silicic magma without silicic eruption, and low thermal conductivity volcanic cover are required to explain the observed thermal gradients of the Mt. Konocti area. Convective heat transport may have also played a role in the past, but evidence cited earlier indicates that no large hydrothermal system is currently active in the area. Convection also produces sharper lateral gradients, thus widespread convection would be necessary to explain the large aerial extent of the Clear Lake thermal anomaly.

G. Models of the Borax Lake Volcanic Center

The Borax Lake-Sulphur Bank Mine area contains the youngest silicic lavas in the Clear Lake region (Donnelly-Nolan et al., 1981), and the most vigorous hydrothermal activity recognized in the Clear Lake region exclusive of The Geysers (White and Roberson, 1962). The Borax Lake lavas consists of a zoned sequence grading upward from basaltic andesite to rhyolite, which appears to have resulted from incomplete mixing of basaltic and rhyolitic endmembers (Bowman et al., 1973; Patterson-Latham, 1985). The rhyolite which caps this sequence has been dated at about 90,000 years B.P. (Donnelly-Nolan et al., 1981), and represents <1 km^3 of magma. A number of other young basaltic andesite and andesite lavas and scoria cone deposits were emplaced in the Sulphur Bank Mine area, and to the east along two nearly N-S linear trends (Hearn et al., 1976, 1981), but attempts to date these units by K-Ar were largely unsuccessful (Donnelly-Nolan et al., 1981). The age of the andesite of Sulphur Bank Mine has been estimated at 44,500 years B.P. based on a carbon-14 date of stump material found in sediments beneath the flow. Similarly, the age of ash layers interbedded with peat in Clear Lake is bracketed from about 90,000 to 10,000 years B.P. by carbon-14 dates (Sims et al., 1981).

The Sulphur Bank Mine is the site of the most vigorous geothermal system identified north of the Colliyami fault zone (White and Roberson, 1962; White et al., 1973; Beall, 1985; Goff and Janik, 1993; Goff et al., 1993b; Goff et al., 1995). Wells drilled near this site yielded conductive thermal gradients of 96 to 141°C/km at depths of 1.0 to 1.3 km. A maximum downhole temperature of 218°C at 500 m was measured in the hydrothermal system (Table 1). The hydrothermal system at Sulphur Bank mine appears to be fault-controlled, extending over a maximum width of 200 m, and has chemical and isotopic characteristics consistent with some magmatic components being present in the fluids (Goff and Janik, 1993; Goff et al., 1993b; Goff et al., 1995), although traditionally Sulphur Bank hydrothermal fluids have been interpreted as metamorphic and/or connate waters (cf. White and Roberson, 1962; Goff and Janik, 1993).

A series of conductive and convective models were constructed based on the Borax Lake-Sulphur Bank Mine area. The models were scaled to represent N-NW cross-sections extending from exposures of the Borax Lake sequence lavas to Sulphur Bank Mine, a distance of approximately 5 km. A 2x5x5 km magma body was emplaced with its roof at 4 km. Because of the very small volume of magma erupted, this magma chamber size yields a volume ratio of <1:125. A magma body consistent with the 1:10 "rule of thumb" of Smith and Shaw (1975, 1978) would certainly fail to produce the observed high heat flow in this area even if heat transport were dominantly by convection.

Models assuming all heat transport by conduction can approach measured thermal gradients in the area if the depth of emplacement is very shallow (3 km), and magma body size greatly exceeds volume ratios of 1:10 (Fig. 25). Models combining the effects of local convection and regional conduction were run based on the available geologic constraints (Fig. 26). A 0.2-km-wide zone of hydrothermal convection was modeled as extending vertically from the magma body to the site of the Sulphur Bank Mine. These models show that thermal gradients reach high values in the convective zone within 5,000 years elapsed time (Fig. 26). At 40,000 years elapsed time, the thermal gradient near the convective zone reached a steady-state value of about 200°C/km (Fig. 26). Conductive heat transport around the convective zone has led to a inverted funnel-shaped region of higher temperatures. In other words, isotherms around the convective zone are bowed upward due to a combination of convective and conductive heat transport. By 90,000 years elapsed time, this zone has widened, creating a region about 1 km wide where near-surface gradients approach or exceed 100°C/km (Fig. 26). Virtually all of the original magma has also crystallized by 90,000 years elapsed time. This model is generally consistent with observed gradients cited above for the Borax Lake-Sulphur Bank area.

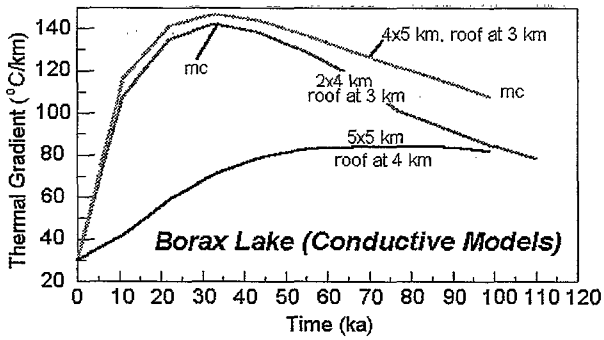

Fig. 25. Near-surface thermal gradient versus time for conductive models of the Borax Lake volcanic center with magma bodies emplaced from 3 to 4 km depth, and 2-D volumes ranging from 2x4 to 5x5 km. All models are single-emplacement, and none have low conductivity volcanic cover. Other models parameters are Tgi=30°C/km, Tim=1000°C, Km=3.1 W/m°C, and Kh=2.9 W/m°C.

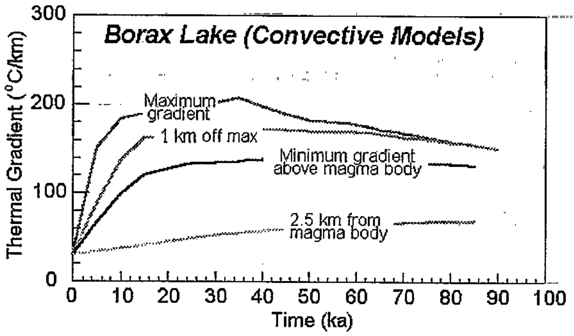

Fig. 26. Near-surface thermal gradient versus time for convective models of the Borax Lake volcanic center with a 2x5 km magma body emplaced at 4 km depth. A 200-m-wide convective zone extending from the magma body to the surface is located near one side of the magma body to simulate a convective fault. Other models parameters are Tgi=30°C/km, Tim=1000°C, Km=3.1 W/m°C, and Kh=2.9 W/m°C.

IV. DISCUSSION AND CONCLUSIONS

The conductive and convective heat transport simulations of cooling magma bodies described above illustrate that the three most important factors in determining near-surface thermal gradients are depth of magma emplacement, magma chamber volume, and the relative importance of conductive and convective heat transport in the upper crust. We have only indirect constraints on these key variables, but they provide the necessary boundary conditions to model the magma-hydrothermal system at Clear Lake. Each is discussed below in light of our modeling and other pertinent data below.

A. Depth of Emplacement

Lines of evidence for shallow intrusion in The Geysers-Clear Lake region include: (1) The Geysers felsite is very shallow and was probably emplaced at depths of <3 km; (2) hornfelsic rocks similar to those that make the contact aureole of The Geysers felsite were encountered in deep wells beneath the main Clear Lake region; (3) purely conductive thermal models necessitate shallow depth of emplacement (3 to 4 km) to explain the very high regional heat flow data; and (4) interpretations of the gravity and magnetics anomalies centered beneath Mt. Hannah include magma emplacement scenarios as shallow as 2-3 km (see Griscom et al., 1993). Of these arguments, the heat flow modeling would be most compelling if one could completely eliminate the possibility of hydrothermal convection. As discussed above, the case for dominantly conductive heat transport in the Clear Lake region is better than virtually any other region of young magmatism that we know of, but evidence for small, fault-controlled hydrothermal systems in the past and present can be found.

B. Magma Chamber Volume and Ratio of Intrusion to Extrusion

Comparison of volcanic and plutonic rocks provides some insight into the issue of the fraction of magma erupted in a given setting (Crisp and Spera, 1984). Recent study of the subsurface geology of The Geysers geothermal field has revealed the occurrence of a shallow, silicic, composite pluton known as the "Geysers felsite" (Schreiner and Suemnicht, 1981; Thompson, 1989; Hulen and Nielson, 1993). This body ranges in composition from granite to granodiorite. The granite has been dated at >1.3 Ma (Dalrymple, 1992), and the granodiorite at 1.19 and 0.95 Ma (Pulka, 1991). Based on drilling, the felsite must have a volume of over 100 km^3, equal to the *entire* eruptive volume of Clear Lake volcanic rocks (Donnelly-Nolan et al., 1993). Hulen and Nielson (1993) have suggested that granodiorite phases of the pluton may be equivalent to dacitic lavas of the Cobb Mountain sequence (Goff and McLaughlin, 1976; Hearn et al., 1976) dated at 1.06-1.08 Ma (Donnelly-Nolan et al., 1981). Preliminary petrographic and microprobe study of core samples from the granodiorite phase (Stimac and Hulen, unpub. data) is consistent with this interpretation. If true, then the volume ratio of extrusive to intrusive rock in this case must be at most 1:20, and could be considerably lower.

As mentioned earlier modeling based on the slabless window scenario by Liu and Furlong (1992) suggests that between 1000 and 3200 km^3 of silicic magma could be produced by basaltic underplating or intrusion into the deep crust. Comparing these figures to the total volume of silicic volcanic rocks (about 50 km^3) yields ratios ranging from 1:20 to 1:64. This range exceeds both the 1:10 "rule of thumb" of Smith and Shaw (1975, 1978) and estimates of Crisp and Spera (1984), which range up to 1:16. Thermal models which are broadly consistent with features of volcanic rocks, thermal waters, and geophysical surveys imply volume ratios of 1:35 for Mt. Konocti and <1:50 for Borax Lake. Our calculations are similar to estimates for the Coso geothermal system (Bacon, 1982) and may be typical of bimodal volcanic fields dominated by lava and dome eruptions.

C. Relative Importance of Conductive and Convective Heat Transport

The largest thermal anomalies in the upper crust are typically the result of hydrothermal convection above cooling magma bodies. This is because convecting hydrothermal fluids can potentially transmit heat orders of magnitude faster than impermeable rock. Such systems commonly consist of deeply circulating meteoric water (plus lesser magmatic fluids) that form neutral-chloride geothermal reservoirs with or without hydrothermal outflow plumes. A classic example is the Valles caldera hydrothermal system (Goff et al., 1988, 1992). Such systems can dramatically affect subsurface temperatures adjacent to their flow paths, by a combination of local convection and regional conduction, such as at the Fenton Hill HDR test site just outside the west rim of the Valles caldera (Harrison et al., 1986; Sass and Morgan, 1988). Thus the role of hydrothermal convection must be considered in any area of anomalous heat flow.

Several lines of evidence presented earlier indicate that no large hydrothermal system is present in the Clear Lake region north of the Collayomi fault zone (Goff et al., 1993a, b, Goff and Janik, 1993), yet this is a zone of highly anomalous heat flow (Jamieson, 1976; Walters and Combs, 1989). We believe the role of convection in the Clear Lake region must be limited to: (1) earlier convective systems that are no longer active (e.g. Bell Mine near Mt. Konocti), (2) deep convective systems that have no surface manifestations, or (3) presently active, fault-controlled systems of small size (e.g. Sulphur Bank and Sulphur Mound Mine areas). Small hot spring areas such as Seiglar Springs and Howard Hot Spring have been shown by drilling to be vanishingly small in volume. Geochemistry of spring waters also indicates that these systems are isolated and relatively shallow, thus they are ignored in our modeling.

A number of models were run that simulated these possibilities, and the primary conclusion that can be drawn is that even limited convection measurably enhances transmission of heat into the upper several kilometers of the crust. Thus the magmatic heat source may be somewhat deeper (up to 6 km) and smaller than in purely conductive models, but is generally shorter lived. Another important result of convective models is that lateral variations in thermal gradient are much sharper than in conductive models, thus regions of elevated thermal gradient largely reflect the geometry of the convective system, unless it is old and deep. Considering the large aerial extent of the heat flow anomaly in the Clear Lake region and the apparent lack of high-temperature geothermal activity, we believe heat transport by conduction to be the dominant process in the region, being modified only locally by convective transport systems.

D. Implications for HDR

The results of thermal modeling support previous assessments defining the excellent HDR potential and poor conventional geothermal potential of the Clear Lake region (e.g. Goff and Decker, 1983). Taken in the context of other observations, these models suggest that Clear Lake is unusual in at least three respects. First, it appears that magma bodies in the Clear Lake region were emplaced at unusually shallow levels (<3-6 km). Second, it appears that only a small proportion of the magma emplaced was erupted compared to larger magmatic systems emphasized by Smith and Shaw (1975, 1978). Third, it seems likely that a combination of host rock type, early tectonic and diagenetic history, and a dominantly compressional stress regime have limited hydrothermal processes north of the Collayomi fault, despite high thermal gradients.

Conductive thermal models suggest the possibility that granitic bodies similar to the Geysers felsite underlie much of the Clear Lake region (up to 750 km^2) at depths as shallow as 3-4 km. This is significant because future HDR reservoirs could potentially be sited in young granitoid plutons rather than in structurally complex Franciscan basement rocks.

ACKNOWLEDGMENTS

This report benefitted from numerous discussions with J. Donnelly-Nolan, B. Carter Hearn, Jr., R. J. McLaughlin (all at the U.S. Geological Survey), D. Jacobs (Unocal Corp.), J. Hulen (UURI), and K. Burns (Los Alamos National Lab). Invaluable assistance with manuscript preparation was given by Lanny Piotrowski. James Archuleta, Anthony Garcia, and Barry Moore prepared some of the figures. This project was funded by a Director's post-doctoral appointment to JAS at Los Alamos National Laboratory and by a U.S. Department of Energy grant to FG (Exploration Geochemistry). Additional assistance was received through a grant from the City of Clear Lake and the California Energy Commission.

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Appendix I

Acronyms
HDR- hot dry rock geothermal
HFU- heat flow unit
mc- magma crystallized

Dimensionality
2-D- two-dimensional
3-D- three-dimensional

Age
Ma- millions of years (1 Ma=1,000,000 years)
ka- thousands of years (1 ka=1000 years)
B.P.- before present

Distance and Volume
km- kilometer
mi- mile
Vm- volume of magma in cubic kilometers
TOP- depth to top of magma body in kilometers

Temperature and Geothermal Gradient
T- temperature in degrees Centigrade (°C)
Tim- initial magma temperature (°C)
Tgi- initial geothermal gradient in °C/km

Thermal Conductivity
K, k- thermal conductivity in W/m°C (Watts/meter/degree Centigrade)
Kh- thermal conductivity of host rock
Km- thermal conductivity of magma

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