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A very first draft! Just kind of
playing around, but gives you some idea of the kinds
of things I'm thinking about. I'll polish this next spring,
may be published 1-2 years down the road.

Kuntz disk 336, name = 2

BASALTIC VOLCANISM ALONG THE GREAT RIFT, IDAHO; II MODELS FOR STORAGE,
ASCENT, AND ERUPTION OF MAGMA

by

Whoever contributes

INTRODUCTION

This paper presents a model that accounts for the formation, storage, ascent, and eruption of magma along the Great Rift. Similar models have been formulated for volcanoes that have had several episodes of well-monitored activity, such as for Kilauea (e.g. Shaw, 1980; Ryan and others, 1981; Klein, 1982;), Mt. St. Helens (papers in Lipman and Mullineaux, 1981), and Mt. Etna (e.g. Wadge, 1977). A prodigious amount of geophysical and geological data for precursor events and observations of the processes and products of several eruptions formed the framework for the models for these volcanoes. These models have also been based on physical properties of magmas and host rocks and mathematical constraints that provide both insight and guidelines to volcanic processes (e.g. Shaw, 1980; Wilson and Head, 1981; Delaney and Pollard, 1982).

The volcanism along the Great Rift is not particularly amenable to a similar kind of analysis because eruptions have not been witnessed and seismic activity that accompanies migration and eruption of magma are not available. Thus, the model for Great Rift volcanism presented here is based both on what is currently known or reasonably inferred about the volcanism there and on models and constraints developed from well-monitored basaltic eruptions.

GENERAL MODEL

The model presented here involves generation of magma in the upper mantle, storage of magma in a lower crustal reservoir, and periodic release of magma from the reservoir by propagation of fractures to the surface. Dike propagation is governed by generation of magmatic driving pressure in the reservoir that produces failure by exceeding the tensile strength of the surrounding rocks. Magma then rises in the fractures due to buoyant forces.

The principal source of magmatic driving pressure (Johnson and Pollard, 1973) is the density contrast between magma stored in a reservoir and the rocks that surround and lie above the reservoir. Non hydrostatic processes, such as convective creep of mantle or lower crust and volume changes on melting and partial crystallization are not considered important and are ignored (Wilson and Head, 1981). Wilson and Head show that gas exsolution typically occurs in magmas at depths of less than 2 km, thus the ascent of bubble-free magma can be treated almost completely by the magmatic pressure-buoyant rise model. These general aspects reflect ideas presented previously by Weertman (1971), Shaw (1980), Wilson and Head (1981), and Wadge (1982)

FIELD CHARACTERISTICS OF GREAT RIFT VOLCANISM AND INFERRED ERUPTIVE PROCESSES

The field characteristics of vent areas and flows along the Great Rift are remarkably similar to comparable features along the southeast and southwest rift zones of Kilauea, Hawaii. By analogy, these similar volcanic features are assumed to have formed by similar volcanic processes.

Well-documented Hawaiian eruptions (Macdonald and Abbott, 1970; Richter, 1970; Swanson and others, 1979) show a typical sequence of events that begins with harmonic tremor, ground-cracking, and steam explosions. Propagation of eruptive fissures is followed by "curtain of fire" lava fountains that extend

for as much as 10 km along an eruptive fissure or series of en echelon fissures. This first stage of fissure eruptions leads to the development of spatter ramparts, relatively small volumes of shelly pahoehoe flows, and downwind blankets of fine-grained tephra.

A second stage of eruptive activity begins if the eruption is sustained for several hours. The length of the eruptive fissure(s) generally diminishes, fountains become localized along a short segment(s) of a fissure, and spatter ramparts are succeeded by cinder-spatter cones which form around lava fountains. Diminished fountaining is followed by quiet, more voluminous outpouring of lava over and through cinder-spatter cones.

A third stage, involving prolonged overflow for periods of weeks or months, produces a lava cone composed largely of pahoehoe flows that mantle the older cinder-spatter cone. Large lava cones or shield volcanoes are produced when sustained eruptions extend for months or a few years at a single vent.

In general, during historic Hawaiian activity and, by inference, during prehistoric activity along the Great Rift, eruptions proceeded in a sequence from fissure-curtain of fire activity through flow from one or several pipe-like vents along the eruptive fissure that produced spatter-cinder cone activity, followed by sustained eruptions from pipe-like vents that produce vent complexes atop lava cones and shield volcanoes. Consideration of magma flow rates, feeder dike thickness, and solidification of magma on the walls of feeder fissures, led Delaney and Pollard (1982) to conclude that growth of one or several central vents above pipe-like conduits, after solidification of magma in other parts of the dike, is a necessary condition for Hawaiian eruptions to last more than a few days. They show that localization of flow in dikes results from local thickening of conduits due to brecciation and

removal of wallrocks associated with increased rates of magma flow and general thinning of the dike due to growth of a layer of solidified magma along the walls of the dike associated with decreased rates of magma flow. They note that eruptions terminate after a few hours or days because the flow rate was too slow or the dike was too narrow to overcome the inhibiting effects of magma solidification on the walls of the feeding fissure.

Certain eruptions along the Great Rift are believed to represent only the first stage of the sequence, other eruptions probably passed through all of the stages. For example, flows of the Kings Bowl lava field, and the Trench Mortar Flat, Vermillion Chasm, and the Deadhorse flows were formed during the first, eruptive fissure stage. The Broken Top, Sawtooth, Silent Cone, and Grassy flows are believed to have formed during stage 2 eruptions from short fissures and (or) pipe-like vents in cinder cones. The Wapi lava field and the Blue Dragon, Devils Cauldron, and Minidoka flows represent eruptions that probably lasted several months or possibly a few years from a central vent complex, perhaps similar to the Mauna Ulu eruptions at Kilauea during the period from 1969 to 1972 (Swanson and others, 1979).

^{well-preserved and well-exposed}
Characteristics of vents and flows for several of the most recent eruptions along the Great Rift are summarized in table 1. Eruptions are

Table 1--near here.

classified as typical of stages 1, 2, or 3 of the eruptive sequence described above, based on their vent characteristics, flow types, and flow volumes.

The eruption of relatively high silica, contaminated lava flows along the Great Rift is not easily related to the sequence of basaltic eruptions. These eruptions were explosive because the flows contain monoliths of bedded cinders

that represent the broken fragments of crater walls that originally surrounded their vents. Vents for the Serrate, Devils Orchard, Highway, Indian Wells north, Indian Wells south, Sawtooth, and Silent Cone flows are all located near the center of large, breached cinder cones, which suggests that these flows were fed from pipe-like conduits and emplaced at the end of a second stage of activity that may have been preceded by fissure eruptions. All vents of the siliceous flows are covered by colluvial cinders derived from the inner walls of the cones, but location of the vents near the centers of the central depressions of the cinder cones suggest pipe-like rather than fissure-like conduits.

Most COM fissure sets of stage 1 eruptions typically consist of 2 to as many as a dozen individual eruptive fissure segments that form a fissure system 1 to 10 km long. Individual eruptive fissure segments range from 50 m to about 1000 m long and the average length of the eruptive fissure systems is about 5 km. Delaney and Pollard (1981, 1982) note that fissure segments probably coalesce at shallow depths to form a single dike.

Most cinder cones along the Great Rift are in fact cinder cone complexes that consist of several nested cones that formed from pulsating lava fountains. The cinder cone complexes were constructed during stage 2 activity and probably formed above pipe-like conduits. Lava cones and small shield volcanoes along the Great Rift that formed during stage 3 activity typically have vent areas with central depressions that contain pit craters, lava lakes, and small cinder-spatter cones.

The width of dikes that fed eruptions along the Great Rift is not known with certainty. We know of no exposures of dikes in the COM lava field, but a few dikes are exposed in the Kings Bowl lava field. At the latter locality, dikes are 0.5 to 1 m wide (Greeley and others, 1977, figures 11-14 and 11-16;

Greeley, 1977, figure 3-19). The width of dikes in the COM lava field were estimated from exposures in eruptive fissures and spatter cones that were judged to have suffered little erosion and enlargement during eruption. The estimated widths are in the range 0.5 to 2 m, most widths are about 1 to 1.5 m (table 1). This width is comparable to the average width of dikes in other regions of basaltic volcanism: Hawaii = <1 m, the Tertiary dikes of Scotland = 1.8 m, dikes in Icelandic eruptive fissures 100 m below the surface are = 4 ± 2 m, and the average width of 1000 dikes in eastern Iceland is 3 m (in McConnell, 1967).

STRUCTURE AND DENSITY OF THE CRUST NEAR THE GREAT RIFT AND MAGMATIC DRIVING PRESSURES

The forced intrusion model is dependent on the density contrast between magma and surrounding rocks, thus the structure and density of the crust in the vicinity of the Great Rift are important factors in constraining the model. Recent seismic studies (Smith and others, 1982; Braile and others, 1982; Sparlin and others, 1982) have provided data on velocity and density distributions of the lithosphere in the eastern SRP (fig 1). The

Figure 1--near here.

uppermost layer consists of basaltic lava flows and interlayered sediments (1-2 km, $K_p = 3.2$ km/sec, density = $2,000 \text{ kg/m}^3$) that represents the Pliocene-Pleistocene basaltic volcanic rocks of the eastern and central parts of the SRP, of which the volcanism along the Great Rift represents the latest phase. Below the uppermost layer is a thicker layer of volcanic rocks and interbedded sediments (5-8 km, $K_p = 4.0-5.3$ km/sec, density = $2,450-2,550 \text{ kg/m}^3$),

which is interpreted as rhyolitic tuffs and flows that formed during the late Miocene-Pliocene period of rhyolitic volcanism in the eastern SRP, described in Part I and summarized by Christiansen and McKee (1978). The two volcanic layers lie above a greatly thinned, anomalous, upper crustal layer (1-10 km, $K_p=6.15$ km/sec, density= $2,670 \text{ kg/m}^3$). The thinning of the upper crustal layer is believed to be associated with the time progressive rhyolitic volcanism. A middle crustal layer (5-10 km, $K_p=6.53$ km/sec, density= $2,880 \text{ kg/m}^3$), localized in the crust beneath the eastern SRP, is interpreted to represent intrusions of mafic material into upper crustal rocks. The lower crust (5-10 km, $K_p=6.80$ km/sec, density= $3,000 \text{ kg/m}^3$) has a relatively uniform thickness in the eastern SRP, appears to be relatively undisturbed, and extends into adjacent areas. The Moho occurs at a depth of about 40 km.

Greensfelder and Kovach (1982) attributed a decrease in shear wave velocity in the crustal depth range of 20-40 km (fig 1) to granulites that were depleted of quartz and, presumably, other low-temperature-melting minerals, during earlier episodes of partial melting. They discounted the presence of "significant amounts of partial melt" to explain the velocity decrease because they considered the lower crust to have a temperature below the solidus of dry granodiorite.

Priestly and Brune (1982) found a low velocity zone in the upper mantle that extends from about 50 to 180 km, thus the lithosphere is about 50 km thick in the vicinity of the Great Rift (fig 1). A similar low velocity zone beneath the Great Basin was interpreted as resulting from a zone of partial melt in the upper mantle (Priestly and Brune, 1978).

The rock composition for the upper layers of the crust is reasonably well constrained, based on field exposures and a 3-km-deep drill hole at the Idaho National Engineering Laboratory (Doherty and others, 1978), located about 50

km northeast of the Great Rift area. Greensfelder and Kovach (1982) suggested that a 3-km-thick, $K_p=6.00$ km/sec, density= $2,7000$ kg/m³ layer beneath the basaltic and rhyolitic layers may consist of pre-Cenozoic carbonate rocks like similar rocks that are widely exposed along both the northwest and southeast margins of the eastern Snake River Plain. Such an interpretation appears unreasonable in that inclusions of such rocks have not been found in any volcanic rocks in the region, but inclusions of rhyolitic volcanic rocks and granulites are common (Kuntz, unpublished field data). The composition of the deepest crustal layer is constrained by the character of granulite inclusions (Part I; Leeman, 1978), and by seismic properties (Greensfelder and Kovach, 1982).

Gravity, magnetic, and magnetotelluric investigations by D. R. Mabey and W. D. Stanley (in Kuntz and others, 1983) have revealed additional details of the upper crustal structure near the Great Rift. A positive magnetic anomaly of about 70 gammas is centered along the northern part of the Great Rift and is roughly coincident with an elongated gravity low (fig 2). The southeast

Figure 2--near here.

edge of the magnetic anomaly is sharp, oriented NE-SW parallel to the eastern SRP, and it coincides with the southeast limit of eruptive vents along the COM segment of the Great Rift. The magnetic anomaly extends northward beyond the topographic boundary of the SRP and is continuous with a northwest-trending zone of magnetic highs that is coincident with a zone of Tertiary intrusive bodies that extends for about 90 km northwest of the plain. The magnetic and gravity data suggest that a granitic intrusive body with a density of about $2,550$ kg/m³ (D. R. Mabey, pers. comm., 1983) underlies the Great Rift area.

Multibody inversion analysis of the regional magnetic field in the area of the Great Rift (Bhattacharya, 1980; Bhattacharya and Mabey, 1980) indicates that the large scale magnetic anomalies are produced by rocks in the upper crust and not by near-surface basalt flows. The multibody analysis shows that the plutonic body is highly magnetized with a top surface 1 to 3.5 km below sea level and a base 10.5 to 11.5 km below sea level (fig. 1). The study also showed that the base of the magnetized crust occurs at a depth of about 10 km below sea level at the northern part of the Great Rift and at a depth of about 18 km below sea level along the southern part of the Great Rift. Bhattacharya and Mabey (1980) and Stanley and others (1977) suggested that the base of the magnetized crust may coincide with the Curie-point isotherm, thus temperatures of 500-580°C. may exist at depths as shallow as 10 km beneath the northern part of the COM segment of the Great Rift and at depths of 18 km along the southern end of the COM segment (fig. 2). This data suggests that the northern part of the COM segment of the Great Rift has extremely high heat flow, comparable to that of the Battle Mountain, Nevada, area, and that the southern part of the Great Rift has heat flow that is comparable to normal heat flow for the Great Basin (Lachenbruch and Sass, 1978; fig 3). High heat

Figure 3--near here.

flow in the SRP is also suggested by the work of Blackwell and Brott (1981), and heat flow values of 2.6 HFU have been determined from the 3-km-deep borehole at the Idaho National Engineering Laboratory (D. A. Blackwell, pers. commun., 1983).

An obvious physical constraint on magma stored in a deep crustal reservoir for long periods of time is that it must retain enough heat so that

its temperature remains above the solidus at that depth. Figure 3 shows solidi and liquidus for McKinney and King Hill basalts (Thompson, 1971) from the Snake River Plain, representative of SRP olivine tholeiite and non-contaminated COM basalts, respectively. The experimental data suggest that temperatures are probably in the range of 1100°-1250°C at a depth of 35-40 km, corresponding to a geothermal gradient of about 2.2 to 2.5 HFU, roughly equivalent to the high Basin-Range gradient of Lachenbruch and Sass (1978).

Magma driving pressures (Johnson and Pollard, 1973) are calculated from the inferred crustal density-thickness distribution shown in fig. 2 and from magma densities according to the relation:

$$P_{mH} = (P_{L_s} - \rho_m g H) - P_{L_H}$$

where P_m = magma pressure at depth H above the source, P_{L_s} = lithostatic pressure at the source, ρ_m = magma density, $g = 9.81 \text{ m/sec}^2$, H = height above source, and P_{L_H} = lithostatic pressure of overlying column of rocks at depth H . This relation assumes Newtonian flow of magma, an assumption that is reasonably well justified, as described farther on. The crustal density distribution used for calculation of magma driving pressures includes the inferred pluton. The pluton lies below a 1-km-thick layer of basaltic rocks and a 1.5-km-thick layer of rhyolitic rocks, inferred from the magnetotelluric and seismic data. The thicknesses and densities of the middle and lower crustal layers are those proposed by Sparlin and others (1982). The average crustal density of the crustal model (fig. 1) is 2,855 kg/m³.

Densities for COM magmas at 1250°C., calculated by the method of Bottinga and Weill (1970) from average chemical compositions (Part I), range from 2,760 kg/m³ for the Sheep Trail Butte flow to 2,450 kg/m³ for the Highway flow. *The calculated density of an average Snake River tholeiite (Part I, table 1) is 2760 kg/m³.* The density for the Blue Dragon flow, representative of most of the large-volume flows of the COM lava field, is 2,690 kg/m³, and the density of the Sawtooth

flow, representative of most of the contaminated flows of the COM lava field, is $2,570 \text{ kg/m}^3$. A density of $2,700 \text{ kg/m}^3$ is taken as representative of most typical COM magmas and the density contrast between a typical COM magma and the average crustal density is about 150 kg/m^3 .

The calculated magma driving pressures (fig. 4) could have been

Figure 4--near here.

considerably less if the magmas had any shearing strength (Johnson and Pollard, 1973). Calculations farther on suggest that basaltic COM magmas probably had shearing strengths of about 30 N/m^2 ($3 \times 10^2 \text{ dynes/cm}^2$), which would tend to decrease the driving pressures by extremely small amounts. For example, the maximum driving pressure is 783 bars at a depth of 11 km for a magma of density 2700 kg/m^3 (fig. 4) and 774 bars at the same depth if the strength of the magma is also considered. The driving pressures shown in figure 4 were calculated for both non-contaminated (2700 kg/m^3) and contaminated (2570 kg/m^3) magmas that rise from a reservoir near the base of the crust at depths of about 30 to 40 km. Note that maximum driving pressures range from 0.5 to 0.8 kbar for the denser magma and from about 0.7 to 1.0 for the lighter magma.

Not all magmas are likely to reach the surface because rocks near the surface are not only less dense than rocks at depth, they are also less dense than most COM magmas. Thus, low density, near surface rocks represent buoyancy barriers that can be penetrated by magmas only under conditions of sufficient magmatic pressure (Elder, 1978) Note in figure 4 that the pluton creates a sharp decrease in the magma driving pressure, perhaps tending to impede the rise of the high density magmas beyond the depth of the base of the

pluton.

If contaminated magmas originated by mixing of basaltic magma and a minimum melt, as suggested in Part I, then these low density magmas must have originated at reasonably deep levels in the crust, where the source material has a density greater than that of the contaminated magmas, such as the $2,880 \text{ kg/m}^3$ deep crustal layer (fig. 1). Partial melting of the low density granitic pluton is unwarranted because these rocks and the overlying rocks present a density barrier to the rise of only the least dense of the contaminated COM magmas. The inferred geothermal gradients and the experimental data shown in figure 3 suggest that melting of the dry, lower crustal material probably occurs at depths in the range of 20-40 km beneath the COM segment of the Great Rift. The geothermal gradients inferred from the magnetic data of Bhattacharya and Mabey (1980) suggest a reason for the geographical restriction of the contaminated rocks to the northern segment of the Great Rift; temperatures are sufficient to melt the lower crust only beneath the northern segment and decrease to temperatures that are insufficient for melting beneath the southern segment.

STRESS ORIENTATION AND FAILURE CRITERIA

Shaw (1980) noted that the direction of magma injection paths is influenced mainly by regional tectonic stress orientations. Nakamura (1977) showed that volcanic fissures are large-scale magma fracturing experiments, where σ_3 is perpendicular to the plane of dikes and fissures and σ_1 and σ_2 lie in the plane. Hydraulic fracturing experiments show that σ_1 is commonly horizontal and increases monotonically at depth due to the weight of overburden (Howard and Fast, 1970; Pollard, 1976). The vertical orientation of fracture pathways and direction of magma transport indicate that the

direction of minimum stress axis in the crust is horizontal, as would be expected in the extensional tectonic regime of the Snake River Plain (Weaver and others, 1979).

Failure criteria for fracture propagation have been summarized by Roberts (1970), Pollard (1973), Wadge (1977), and Shaw (1980). Briefly, a triaxial stress condition is assumed in which σ_1 and σ_3 are horizontal, σ_2 is vertical, and σ_1 and σ_2 lie in the vertical plane of the fracture (Shaw, 1980). The magmatic driving pressure is assumed to produce a stress σ_p that is perpendicular to the walls of the magmatic reservoir and the tensile strength of the rocks enclosing the reservoir is τ_0 . A vertical fracture will be produced when the value of σ_p reaches τ_0 , such that:

$$\sigma_{pmin} = \tau_0 = 3\sigma_3 - \sigma_1 - P_m \quad (\text{Wadge, 1977})$$

where P_m = internal magmatic pressure. If $P_m > 3\sigma_3 - \sigma_1$, then σ_{pmin} is tensile and when $\sigma_{pmin} = \tau_0$, fracture occurs in a vertical plane along the direction of σ_2 and $P_m > 3\sigma_3 + \tau_0 - \sigma_1$. Once the fracture forms, subsequent fracturing comes from the tip of the magma body in the crack when the conditions of $P_m > 3\sigma_3 + \tau_0 - P_p$ are met (where P_p is the pore pressure), provided that

$$3\sigma_3 + \sigma_1 < 4P_m$$

It is important to note that σ_3 need not be tensile. If P_m is large compared to $\sigma_3 - P_m$, then σ_3 will be compressive (Pollard, 1973).

Shaw (1980) noted that magma injection by extensional fracture can occur at any depth in the crust or mantle when the ratio of fluid pressure to the vertical principal stress approaches unity. Pollard (1973) noted that intrusions propagate along extension fractures when:

$$\sigma_1 - \sigma_3 - P_m - P_p + (P_m - \sigma_3) 2L/T = \tau_0$$

where L and T are the length and thickness of the fracture. Propagation is predicted at even lower values of magma pressure than for the previous

criteria because the term ($P_m - S_3$) is multiplied by $2L/T$, a large number. As Pollard states, dikes have a tremendous mechanical advantage over the host rock because of large L/T values and numerous, critically oriented microscopic flaws, thus they will propagate at very small values of $P_m - S_3$. Excess magmatic pressures need not exceed 200 bars to cause fracture of reservoir walls at all reasonable depths (Wilson and Head, 1981).

ORIGIN OF MAGMA AND MAGMA SUPPLY RATE

The pressure-depth-temperature conditions of partial melting of the asthenosphere for the origin of SRP magmas has been discussed by several investigators. On the basis of experimental work, Thompson (1975) suggested that Mg-rich, SRP olivine tholeiites were formed by partial fusion of Fe-rich spinel lherzolite at 50-60 km. He also concluded that partial crystallization of the olivine tholeiites occurred at about 30 km, where ascent to the surface was delayed, and formed the more fractionated basalts. Stout and Nichols (1977) and Leeman and others (1976) reached similar conclusions and the latter authors concluded that the more fractionated COM magmas originated from more Mg-rich parental magmas by intermediate pressure (8-10 kbar, equivalent to depths of ~30 to 35 km, see fig. 1) fractionation of megacrysts that are observed in the Lava Creek flows, some of the most mafic COM flows.

Causes of melting and production of basalt magma in the SRP area have been ascribed to pressure-release melting in ascending mantle plumes (Morgan, 1971) and to stress relief at the base of the lithosphere due to tectonic extension (Christiansen and McKee, 1978). Christiansen and McKee suggested specifically that 1) greater amounts of extension, such as at the boundaries of tectonic provinces, produces higher percentages of partial melting at shallower depths to generate tholeiitic magmas, 2) that basalts rise buoyantly

at localities where favorable structures or large deviatoric stresses accomodate the rise, such as areas where extensional faulting occurs simultaneously with basaltic volcanism, and that 3) the absence of extensional surface faulting tends to favor long or terminal residence of basaltic magma in the crust. The Great Rift area is unique in the SRP in that multiple episodes of volcanism have occurred along the same fissure system (Kuntz and others, 1983). If Christiansen and McKee's ideas are applicable, then perhaps greater amounts of extensional stress and faulting have been concentrated along the Great Rift during the last 15,000 years that resulted in greater amounts of volcanism, compared to other areas of the SRP.

Upward migration of pockets and films of magma from a source region in the asthenosphere to a storage area in the crust occurs mainly due to buoyant forces and is guided by stress conditions in the asthenosphere (e.g. Waff and Bulau, 1979; Shaw, 1980; Turcotte, 1981). The failure cascade concept of Shaw (1980) relates stress conditions in the asthenosphere-lithosphere and fracture mechanisms for "kneading" melt upward from the source region. Shaw suggests that there is little or no storage of magma at or near the source area and that magma migrates and coalesces into large bodies soon after it forms. The average crustal residence time for basalt at Kilauea is 10 years, thus the asthenosphere is sampled directly and quickly with little modification of magma chemistry. A constant magma supply from an asthenospheric source region to a crustal storage reservoir has been postulated for Kilauea (Swanson, 1972; Aki and Koyanagi, 1981), Etna (Wadge, 1975), and Piton de la Fournaise (Ludden, 1977). Wadge (1982) states that production of magma at sources beneath polygenetic volcanoes is steady over periods of at least 100 years, thus a constant magma supply rate is assumed to exist beneath the Great Rift. If melt migrates quickly and is not stored in the asthenosphere beneath

the Great Rift, then the long repose intervals between eruptions suggests long residence times in the lower crust. This addition of magma to the lower crust tends to promote major changes in basalt magma chemistry through crystal fractionation and by contamination from partially melted wall rocks of the reservoir.

Fedotov (1981) analysed magma supply rates for volcanoes of various compositions and assumed cylindrical feeding channels and found that the rate depends mainly on geothermal gradient, magma composition, and the solidus temperature of the magma at the surface. He noted that rates of magma supply are variable for different types of volcanoes and that supply rates ultimately determine the type of volcanism. For basaltic volcanoes where magma is formed under oceanic geothermal gradients, magma supply rates are large ($7-8 \times 10^6 \text{ m}^3/\text{yr}$), they display nearly continuous activity, and they are fed nearly continuously from their source. Basaltic volcanoes with intermediate supply rates ($4-6 \times 10^6 \text{ m}^3/\text{yr}$) have intermittent surface activity and part of the magma supply accumulates in crustal chambers. Basaltic volcanoes with small magma supply rates ($2-4 \times 10^6 \text{ m}^3/\text{yr}$) are of the areal-cinder cone type and display sporadic activity. Volcanism along the Great Rift is of the latter type. If an average supply rate of Fedotov is assumed ($3 \times 10^6 \text{ m}^3/\text{yr}$) and the long-term output rate is taken as $2 \times 10^6 \text{ m}^3/\text{yr}$ (Part 1), then the extrusive-intrusive ratio of COM magmas is about 0.67. for the last 15,000 years.

RELEASE OF MAGMA AND VOLUME OF MAGMA RESERVOIR

The magma reservoir system beneath the Great Rift is "open" in that the influx and accumulation of magma is interrupted periodically by release of magma to the surface. After an eruptive period ends, the magma reservoir is assumed to have a specific volume, but continues to receive new magma from the

source region during the repose interval. The added volume of magma causes the magma pressure to increase above the lithostatic pressure to a value that exceeds the tensile strength of the reservoir walls and failure of the reservoir occurs then when a critical amount of magma accumulates in the reservoir.

The short repose times at Kilauea and Mauna Loa contrast markedly with the long repose times for the Great Rift (Part 1, table 5). The great difference in repose times between the Great Rift and Hawaiian volcanoes can probably be related to the great difference in magma supply rates. Klein (1982) describes the length of repose intervals at Kilauea and Mauna Loa as "random", ie, the probability of an eruption occurring within a span of time is a constant, independent of the times and nature of the previous eruptions. He states that Kilauea and Mauna Loa are in a "delicate balance", constantly ready to erupt, and that forces that are small in comparison to magmatic pressures or tectonic stresses are capable of triggering eruptions, producing a random model for variable repose times. Klein suggests that external forces that trigger Hawaiian eruptions are earth tides and earthquakes. By applying this principle to the Great Rift then, a delicate balance exists before eruptions between the stresses caused by increasing magmatic pressure that tend to promote eruption and the stresses on the reservoir walls produced by the interaction of the confining pressure of the enclosing rocks and the regional tectonic stresses that tend to inhibit eruptions. The balance is tipped in the favor of eruption when the magmatic pressure increases to some critical value or when unrelated factors, such as earthquakes, occur to upset the balance.

By assuming that the volume of the reservoir does not change and that the added volume of magma is accomodated by the compressibility of the stored

magma, Blake (1981) related the critical volume of added magma (ΔV), the volume of the magma reservoir (V_r), the tensile strength of the reservoir walls (σ), and the bulk modulus of the magma (b) as follows:

$$\frac{\Delta V}{V_r} = \frac{\sigma}{b}$$

The "active volume" of the magma reservoir may be estimated by using a relationship defined by Klein (1982) for Kilauea and Mauna Loa, where active volume is the product of the magma supply rate and the mean repose time. Using Fedotov's estimate of $\sim 3 \times 10^6 \text{ m}^3/\text{yr}$ for magma supply rate and 2,000 years for the average repose interval for COM eruptions, an active volume of $6 \times 10^9 \text{ m}^3$ or 6 km^3 is calculated, roughly equivalent to the volume of lava flows erupted in eruptive period B and about 150% of the average volume of lava flows in several other eruptive periods (Part 1, table 4)

For purposes of calculation, a feeder dike is considered to be 1 km long, 2 m wide, 40 km high, and has a volume of 0.08 km^3 , thus the volume of congealed magma remaining in feeder dikes is considered to be negligible compared to the volume of erupted magma and ΔV is taken to be $4 \times 10^9 \text{ m}^3$. Values for the tensile strength of most rocks are 10-50 MPa (Blake, 1981) and an average for basalt is about 15 MPa (Lama and Vutukuri, 1978). A value for the bulk modulus of basaltic magma can be calculated from seismic wave velocity and magma density (Howell, 1959, p. 158); a value of $V_p \sim 4 \text{ km/sec}$ is assumed for 100% liquid at 40 km (H. M. Iyer, pers. commun., 1983) and $2,700 \text{ kg/m}^3$ is used for the magma density. These values yield a bulk modulus of about 43 GPa and a magma reservoir volume of 10^4 km^3 , an extremely large volume, corresponding to a cube with an edge of 23 km or a sphere with a radius of about 14 km.

Blake (1981) states that the assumption that V_r remains constant during the continued accumulation of magma is probably not appropriate for volcanoes

where additional reservoir space is created by crack opening, filling, and propagation, as would be expected in the tensional tectonic regime at the base of the crust beneath the Great Rift. In this case, the relationship becomes:

$$\frac{\Delta V}{V_r} \approx \frac{P_0}{b} (1+s) + s$$

where s is the fractional volume expansion of the chamber. In cases where $s \gg \frac{P_0}{b}$, such as probably exists for the Great Rift magma reservoir system, then the assumption of $s \approx 0$ leads to overestimation of V_r by up to an order of magnitude. Because there is no way to estimate s for the Great Rift case, as might be possible from geodetic data, it is assumed then that V_r is probably on the order of 10^3 km^3 . By assuming that the surface dimensions of the Great Rift (~50 km long, ~4 km wide) reflect the geometry of the underlying magma reservoir, and using a value of $V_r = 10^3 \text{ km}^3$, a magma reservoir with dimensions of 50 by 4 by 5 km is suggested. The height and width of the reservoir are probably much greater than these values due to the variable width and density of individual dike-like magma reservoirs within the reservoir system.

The critical factors of magma supply rate and the pressure regime of the magma reservoir affect the type of volcanic eruptions, as discussed by Shaw (1980) and shown in figure 5. Two limiting cases of stress adjustment of the

Figure 5--near here.

reservoir walls affect the release of magma during eruption. Rapid adjustment produces a constant pressure potential so that nearly all magma is expelled from the reservoir and tends to promote long-lived, large-volume eruptions. Slow adjustment produces a progressive pressure drop in proportion to the volume of expelled magma and the eruption terminates when magmatic pressure

reaches a critical level, producing short lived, low-volume eruptions. Interrelationships between magma supply and stress adjustment factors are shown in figure 5. A low magma-supply rate and slow stress adjustment of reservoir walls on eruption are factors that appear to control volcanism along the Great Rift and account for long repose intervals and steady state volcanism.

Long repose intervals are attributed to nonsupply of magma or storage of magma in a pluton or subvolcanic dike system by Wadge (1982). The model of Shaw (1980) suggests that low rates of melting and magma supply to the storage reservoir and continuous expansion of the reservoir are the major factors that account for the long repose intervals between eruptive periods along the Great Rift. The above calculations from the model of Blake suggest that, under normal circumstances, about 6 km^3 of magma needs to be added to the magma reservoir to create the pressure conditions necessary to cause an eruption. A magma supply rate of $\sim 3 \times 10^6 \text{ m}^3/\text{yr}$ and an extrusive/intrusive ratio of 0.67 for the Great Rift suggests that about 6 km^3 of magma is delivered to the magma reservoir in about 2,000 years. If extension rates and magma supply rates are relatively constant and if the Blake model applies to the Great Rift volcanic system, then maximum repose intervals of about 2,000 years are suggested. Shorter repose intervals could be caused by factors that alter the tensile strength of reservoir rocks, such as greater pressure from greater buoyancy of low density magmas, earthquakes, etc.

DIKE GEOMETRY, MAGMA RISE VELOCITY AND MASS ERUPTION RATES

Koide and Bhattacharji (1975) showed that effective stresses are greatest at the top of magma reservoirs and, because fractures occur in the areas of largest effective stresses, magma reservoirs tend to develop upward. Thus,

magma bodies in the deep crust are typically dike-like in 3 dimensions, with large aspect ratios. Based on the notions that magmatic pressure acts across the walls of a reservoir and that magmas are lighter than enclosing rocks, crack propagation for delivery of magma to the surface probably occurs by the ductile faulting mechanism at the high confining pressure and temperature near the base of the crust and by brittle faulting and extension fracture mechanisms at higher levels in the crust (Pollard, 1973).

Minimum crack width at depth can be estimated by using the relationship of critical dike half width that just permits magma movement due to buoyant forces (Wilson and Head, 1981)

$$r_{ct} = \left[\frac{Z H A \eta}{8 g (\rho_m - \rho_h)} \right]^{1/4}$$

where Z is a constant, $1.25 \cdot 10^6 \text{ m}^2/\text{sec}$, H is dike height, A is a constant (24), η is magma viscosity, and $\rho_m - \rho_h$ is the density contrast between magma and a wall rocks. For conditions in the lower crust, and assuming the open conduit model, with H as 30 to 40 km, η as 300 Pa.s (Wilson and Head, 1981), and the density contrast as 300 kg/m^3 , r_{ct} (dike half-width) is about 0.4 m. For the multi-fracture model, with $H = 5 \text{ km}$, r_{ct} is about 0.2 m. At a depth of about 15 km, where the density contrast drops to about 100 kg/m^3 , r_{ct} is about 0.4 m for the open conduit model and r_{ct} is about 0.2 m to 0.3 m for the multifracture model with crack heights in the range of 1-5 km. These calculations suggest that minimum dike widths are about 0.4 to 0.8 m at depth, only slightly less than the observed widths of feeder fissures at the surface.

Width/length ratios of fissures at various depths can be estimated according to the relation of Fedotov (1981):

$$\frac{w}{L} = \frac{4(1-\sigma^2)\rho_m}{E}$$

where l is the half length, ν is Poisson's ratio, P_m is excess magmatic pressure, and E is Young's modulus. Poisson's ratio ranges from 0.23 to 0.31 in the lithosphere beneath the eastern SRP (Greensfelder and Kovach, 1982) and Young's modulus, calculated (Howell, 1959, p. 207) from Poisson's ratio, rock density, and the shear wave velocity (Greensfelder and Kovach, 1982), ranges from about 20 GPa near the surface to about 100 GPa at the base of the crust. Using a value of 15 MPa (150 bars) for P_m (fig. 4; Wilson and Head, 1981), width/length ratios range from about $3 \cdot 10^{-3}$ near the surface to about $6 \cdot 10^{-4}$ in the lower crust. If P_m is lower, perhaps 2 MPa (20 bars), as suggested by Aki and others (1977) for Kilauea, then width/length ratios for Great Rift fissures range from about $4 \cdot 10^{-4}$ near the surface to about $7.5 \cdot 10^{-5}$ in the lower crust. For surface widths of 1 and 2 m (table 1) and P_m of 15 MPa (150 bars), fissure lengths of about 0.6 to 1.2 km are predicted, equal to or slightly less than the observed lengths of eruptive fissures (table 1). For surface widths of 1 and 2 m and P_m of 2 MPa (20 bars), fissure lengths of 5 to 10 km are predicted, equal to or slightly greater than the observed lengths of eruptive fissures along the Great Rift.

The heights of magma-filled cracks at various depths can be estimated by a relationship given by Wadge (1982):

$$H \leq 2 \left[\frac{K_c}{\sqrt{\pi} \cdot g \cdot (\rho_m - \rho_h)} \right]^{2/3}$$

where K_c is fracture toughness (Pollard and Muller, 1976), ~~and fracture toughness is~~ a property of rocks that measures resistance to crack propagation (Delaney and Pollard, 1982). Experimental data indicate K_c values on the order of 1-5 MPa/m² for sedimentary rocks and K_c values of 100 MPa/m² were calculated for host rocks penetrated by dikes near Ship Rock, NM (Delaney and Pollard, 1982). Using the value $K_c = 100 \text{ MPa/m}^2$, and density contrasts

of 300 kg/m^3 for the lower crust and 150 kg/m^3 for the middle crust, crack heights of about 1.5 km to 2.3 km are predicted, respectively.

The data above suggest that magma-filled cracks in the lower crust have lengths of about $5-10 \text{ km}$, widths of about $0.5-2 \text{ m}$, and heights of about 2 km , yielding volumes on the order of $4 \cdot 10^6$ to $3 \cdot 10^7 \text{ m}^3$.

Magma in cracks at depth create a buoyant force equivalent to (Weertman, 1971):

$$F_b = (\rho_m - \rho_h) \cdot g \cdot V$$

where variables are as defined previously, and V is crack volume. For appropriate values for these variables, as discussed above, the buoyant forces are on the order of $4 \text{ to } 30 \text{ GN}$ for the most mafic COM magma and about $16-120 \text{ GN}$ for the most siliceous COM magma.

The minimum rise rate for magma in cracks of various widths can be estimated from relations provided by Fedotov (1978) and Wilson and Head (1981);

$$V_{ct} = \frac{ZH}{r^2}$$

where r is the conduit radius. For crack half widths of 1 and 0.5 m and H in the range of $30-40 \text{ km}$, V_{ct} values are about $4 \text{ to } 5 \cdot 10^{-2} \text{ m/sec}$ and $1.5 \text{ to } 2 \cdot 10^{-1} \text{ m/sec}$ respectively.

The velocity of magma rise in cracks at all Reynolds numbers is given by Wilson and Head (1981) as

$$V = \frac{A_n}{4K\rho_m r} \left\{ \left[1 + \frac{64 \cdot g \cdot r^3 \cdot (\rho_m - \rho_h) K \rho_m}{A^2 h^2} \right]^{1/2} - 1 \right\}$$

where K is a constant (0.01), r is the crack half width, and the other variables are as defined previously. The average crustal density contrast

for most COM basaltic magmas is 150 kg/m^3 and magma viscosity is taken as 300 Pa.s (Wilson and Head, 1981). For dike half widths of 1 m , 0.75 m , and 0.5 m , the respective velocities are about 1.6 , 0.9 , and 0.4 m/sec . These velocities allow calculation of the Reynolds number (Wilson and Head, 1981):

$$Re = \frac{2 \cdot r \cdot V \cdot \rho_m}{\eta}$$

The velocity values cited above all yield Reynolds numbers of less than 20, which indicate laminar flow (Bird and others, 1960).

As suggested by Wright and others (1979) for Kilauea, the absence of mantle-derived xenoliths in the COM basaltic lava flows suggests rapid transport of magma as discontinuous pulses in narrow fractures (analogous to capillary action) rather than in steady-state flow through large conduits. Additionally, lack of xenoliths can be related to long repose intervals when xenoliths may settle to the bottom of the magma reservoir.

The yield strength of COM magmas can be estimated by a relation given by Wilson and Head (1981) as:

$$S_y \ll \left[\frac{2 \cdot H \cdot A \cdot \eta \cdot g^3 (\rho_m - \rho_h)^3}{8 B^4} \right]^{1/4}$$

where all variables and constants are as previously defined and $B \sim 1$ for a long, narrow fissure. Wilson and Head suggest that the actual magma strength is less than the value calculated from this relation by a factor of at least 10. For values of an average density contrast of 150 kg/m^3 , a magma viscosity of 300 Pa.s and height of magma column of 30 to 40 km , the yield strength is probably less than 30 N/m^2 ($3 \times 10^2 \text{ dynes/cm}^2$). Wilson and Head (1981) state that the rise of magmas that possess yield strengths greater than 10^3 N/m^2 will be limited by the yield strength rather than by cooling. Thus, the assumption made earlier that the most mafic COM magmas probably behave as Newtonian

fluids seems reasonably well justified.

Mass eruption rates can be estimated from the relation given by Wilson and Head (1981) as:

$$\frac{M}{L} = \frac{A \eta}{2K} \left\{ \left[1 + \frac{64 \cdot g \cdot r^3 (\rho_m - \rho_h) K \rho_m}{A^2 \eta^2} \right]^{1/2} - 1 \right\}$$

where variables and constants are as previously defined. For values of $\eta = 300$ Pa.s, $r = 1$ m, a density contrast of 150 kg/m^3 , and a magma density of 2700 kg/m^3 , the mass eruption rate is $9 \times 10^3 \text{ kg/sec m}$. For comparison, Wilson and Head calculated that the eruption rates for the 1961 Askja eruption was $1 \times 10^4 \text{ kg/sec m}$, and that the rate for the 1783 Lakagigar eruption was at least $1.5 \times 10^3 \text{ kg/sec m}$. Swanson and others (1975) estimated a rate of about $3 \times 10^4 \text{ kg/sec m}$ for the Yakima member of the Columbia River Group.

By assuming that 1) the calculated COM eruption rate was constant (probably and unreasonable assumption, as discussed below), that 2) COM eruptions occurred along a fissure 5 km long, and that 3) COM flows have a density of 2500 kg/m^3 , then the large pahoehoe flows of the COM lava field, such as the Blue Dragon, Minidoka, and Carey-Sunset flows, each of which have average volumes of about 3 km^3 (Part 1, table), probably could not have formed in a period of less than 2 days.

The calculated mass eruption rate for the Kings Bowl lava field, assuming a fissure half width of 0.5 m, a magma density of 2800 kg/m^3 , and a density contrast of 100 kg/m^3 , is about $8 \times 10^2 \text{ kg/m sec}$. By assuming a constant eruption rate, an eruptive fissure 5 km long, a lava flow density of 2500 kg/m^3 , and a volume of 0.01 km^3 for the Kings Bowl lava field, the field could have formed in a period of about 2 hours.

To obtain a more realistic estimate of the times of emplacement of these flows, emplacement can be modelled by continuous flow from an eruptive

fissure, but with decreasing flow rate with time. Machado (1974) and Scandone (1979) noted that the flow rate for many eruptions decreases exponentially according to the relation:

$$Q = Q_0 (-At)$$

and

$$Q(t) = \frac{dV}{dt}$$

Substituting the first equation into the second and integrating both sides yields

$$\int_0^{\infty} Q_0 \cdot e^{-At} dt = \int_0^{\infty} dV$$

and

$$A = \frac{Q_0}{V_{tot}}$$

where V_{tot} is the total flow volume. For $Q_0 = 4.5 \times 10^7$ kg/sec for a 5-km-long fissure, lava flow volume = $3 \times 10^9 \text{ m}^3$, and lava density of $2,300 \text{ kg/m}^3$, $A = 3 \times 10^{-6}$ seconds. From the properties of the exponential curve, 63% of the total volume is extruded in $t = \frac{1}{A}$, or 3.3×10^5 seconds (~4 days) and 95% of the total volume is extruded in $t = \frac{3}{A}$, or 10^6 seconds (~12 days).

Applying the same model to the Kings Bowl lava field, and using $Q_0 = 4 \times 10^6$ kg/sec for a 5-km-long eruptive fissure and a flow volume of 2.5×10^{10} kg, A is 1.6×10^{-4} seconds. From the properties of the exponential curve, 63 % of the total volume of the Kings Bowl lava field could have formed in about 2 hours and 95 % of the total volume could have formed in about 5.5 hours.

CAUSES FOR THE TERMINATION OF VOLUME-PREDICTABLE ERUPTIONS

Equilibrium conditions exist for both the onset and termination of eruptions along the COM segment of the Great Rift. For the onset of eruptions, a period of time proportional to the volume of the preceding eruption elapses before sufficient magmatic pressure is produced to force eruption. If only a small amount of magma is withdrawn from the reservoir, then the magmatic driving pressure is reduced a small amount, but always to a value that approaches zero, as required by the buoyant rise model, because the eruption ends when hydrostatic equilibrium is established.

For the buoyant rise model to apply exactly to maintain a constant driving pressure, then there must be a steady deformation of reservoir wall rocks so that the reservoir volume decreases simultaneously as magma is withdrawn. Because COM eruptions occur in times that are short compared to inferred strain rates for the deep crust (Shaw and Swanson, 1975), then recovery of the reservoir walls probably does not keep pace with magma withdrawal, thus P_m approaches zero, the magmatic head decreases in proportion to the volume of erupted magma, and the eruption ceases when some fraction of available magma is withdrawn from the reservoir that is necessary to restore hydrostatic equilibrium. These relations are shown schematically in figure 6, where buildup of magmatic pressure is shown to increase at a linear

Figure 6--near here.

rate to a fairly broad region of critical P_m values. The figure shows that the critical region is reached in a period of about 2,000 years and may extend to as much as 3,000 years. The magnitude of P_m can not be defined exactly, but previous calculations suggest values in the range of 20 to 150 bars.

MODEL FOR MAGMA PLUMBING SYSTEM AND SUMMARY OF VOLCANIC PROCESSES

Seismic data related to magma generation and movement that may give clues to the magma plumbing system beneath the Great Rift are lacking, thus the inferred plumbing system presented here is based on conceptual models for magma plumbing systems developed for other volcanoes, the mathematical constraints discussed above, and from the geometry of the surface vent systems. .

Differences in the processes of magma storage and delivery for olivine tholeiite magmas of the SRP and for COM magmas may be inferred from their contrasting chemical and mineralogical characteristics (Part I). According to Stone (1967), Tilley and Thompson (1970), Thompson (1975), Leeman and Vitaliano (1976), and Stout and Nicholls (1977), SRP olivine tholeiites have chemical and mineralogical characteristics that suggest that they are the products of partial melting of the mantle at 50-60 km depths and that they experienced fractionation that is related to only limited residence time in the crust in reservoirs at pressures of <5 kbar, corresponding to depths of 20 km or less. In contrast, the compositions of COM basaltic rocks suggest longer storage times in crustal reservoirs that favor compositional heterogeneity by processes of magma mixing and (or) crystal fractionation.

The chemical characteristics of COM rocks and the geographic and chronologic aspects of COM eruptions (Part 1) and the constraints on the magmatic processes discussed in this paper shed some light on the size and shape of magmatic storage reservoirs beneath the Great Rift. From 2 to 6 km³ of magma have been erupted in each eruptive period, but the magma reservoir are much more voluminous. Because some COM magmas have become progressively more silicic, magma reservoirs are probably large enough to accomodate crystal fractionation and assimilation to produce a wide range of chemical

compositions, but not so small as to be completely emptied after each eruption. However, each eruptive period has produced relatively primitive COM-type magma with SiO_2 contents of less than 47% along with more silicic magmas. This suggests a plexus of dike-like reservoirs, each of which may be compositionally zoned in a vertical sense to a greater or lesser degree. Each eruptive period tends to approach evacuation of a magma reservoir by removing more silicic, fractionated magma at the top of the reservoir early in an eruptive period and followed by eruption of progressively more mafic magma at deeper levels in the reservoir as the eruption proceeds.

Segments of the Great Rift have experienced eruptions at different times and produced magmas of varying compositions, which suggests that the magma reservoirs may consist of discrete chambers, probably dikes, that lie at different levels and (or) lateral positions beneath the Great Rift.

Generalized magma plumbing systems for volcanoes of various types have been described by Machado (1974), Shaw (1980), Fedotov (1981), and Wadge (1982), among others. Common elements in these systems are: 1) a magmatic source region in the asthenosphere, 2) a magma "column" that leads upward from the source region, 3) a lower crustal or subcrustal magma reservoir located at or near the Moho, 4) magma-filled cracks that transfer magma upward from the lowest reservoir, 5) an upper level, subvolcanic reservoir, and 6) conduits that lead from the subvolcanic reservoir to volcanic vents. Only some of these features are present at any one volcano.

The best study of a magma plumbing system is that for Kilauea, summarized in Wright and others (1980), Shaw (1980), and Ryan and others (1981). For Kilauea, a magma source region lies at a depth of 65-40 km, an upward-converging, funnel-like, primary conduit leads from the source volume to a shallow reservoir at a depth of 6.5 to 1.5 km, where dike-like conduits that

transfer magma to the surface. Studies by Wadge (1977) for Mt. Etna and by Martin and Rose (1981) for Fuego suggest that these composite stratovolcanoes have similar plumbing systems; both include a magmatic source region in the asthenosphere, a primary magma conduit, magma-filled cracks that transfer magma to a higher level, a subvolcanic reservoir, and dike-like conduits that transfer magma from upper-level reservoir to volcanic vents at the surface. Martin and Rose included a lower crustal reservoir to the inferred plumbing system at Fuego.

Elements of these plumbing systems that are compatible with what is known or can be inferred about the plumbing system beneath Great Rift, are shown in figure 7. The source region of partial melting is at 50-70 km in a low

Figure 7--near here.

velocity zone in the upper mantle (Hill and Pakiser, 1967; Priestly and Brune, 1982; fig. 1 of this paper). The pressure at this depth is about 15-20 kbar, which corresponds to depths and pressures of partial melting suggested by Leeman and others (1976) and Stout and Nicholls (1977). The geothermal gradients beneath the Great Rift allow for melting of spinel lherzolite at depths of about 50-70 km (fig. 3). The repeated eruptions along the Great Rift during the last 15,000 years suggest that greater amounts of strain and partial melting are localized in the upper mantle beneath the Great Rift compared to other areas of the SRP. The concepts of Shaw (1980) suggest transfer of magma to the storage reservoir via multiple dikes that "knead" magma from the source region at a supply rate of $\sim 3 \times 10^6 \text{ m}^3/\text{yr}$.

The depth of the primary reservoir at about 30-40 km in the lower crust is based on experimental data by Thompson (1975), Tilley and Thompson (1970),

and Leeman and others (1976) that the most mafic COM, non-contaminated lavas may originate by fractional crystallization of more primitive SRP-type magma at pressures of 8-10 kbars. The simple possibility that the fissure vent systems directly reflect the geometry of the magmatic plumbing system at depth, where magma may be generated, stored, and transported to the surface along narrow, dike-like bodies in the upper mantle and crust, was suggested for the Columbia Plateau by Swanson and others (1975) and for Kilauea by Shaw (1980). Compositional differences in lava flows erupted from different parts of the northern segment of the Great Rift suggest that the magma reservoir consists of a plexus of dike-like bodies that are intersected by long, narrow feeder fissures. The volume of the magma reservoir is on the order of 10^3 km^3 . Dike-shaped bodies and magma pathways to the surface are predicted by failure criteria. Similar shapes were proposed by Lachenbruch and Sass (1978) for magma bodies in an extending crust that accomodates intrusion of melt from the asthenosphere. The deep crustal rocks that form the reservoir walls are probably granulite, based on seismic velocities (Greensfelder and Kovach, 1982) and inclusions in COM rocks (Leeman, 1978; Part 1). The greater geothermal gradients beneath the northern part of the Great Rift probably result from a larger magma supply rate at that locality, which promotes greater amounts of partial melting of reservoir rocks and production of siliceous magmas.

The orientation of feeder fissures indicate that σ_3 is horizontal and perpendicular to the Great Rift and that σ_1 and σ_2 are oriented in a plane parallel to the fissure system. Failure of the magma reservoir occurs due to increasing magma driving pressure by the constant addition of magma from the source region. The active volume of the reservoir is about 6 km^3 , which is added approximately every 2,000 years. Failure criteria suggest that fissures

that constitute the reservoir are tapped by propagation of cracks to the surface by extensional fracture. Crack geometry at depth is that of thin discs with widths of up to several meters and lengths and heights of several km. The magma rises in fissures due to buoyancy, driven by the density contrast between magma and wall rocks. Velocities are on the order of 1 m/sec and flow is laminar. Because of low yield strengths, magma movement is nearly Newtonian and is controlled mainly by cooling and accretion of chilled magma to fissure walls,

The low density rocks in the upper crust offer a buoyancy barrier to the upward movement of basalt magma and thus a secondary storage chamber is shown schematically in the form of dikes and sills in figure 7. Evidence for a shallow-level, secondary reservoir is weak, but Leeman and others (1976) suggest that low pressure (< 5 kbar) fractionation, corresponding to depths of < 20 km, accounts for compositional and mineralogical variation of SRP-type McKinney basalt, and Sparlin and others (1982) suggest dikes and sills of basalt at this level on the basis of seismic data.

Eruptive fissures have mass eruption rates that allow for emplacement of large volume ($\sim 3 \text{ km}^3$) flows in a period of several days. Eruptions probably end because of an exponential decrease in eruption rate due to the slow adjustment of the magma reservoir on magma withdrawal. The volume of the magma withdrawn in each eruption is proportional to the length of the previous repose interval. A volume of magma necessary to restore the lithostatic stress condition is erupted, resulting in the volume-predictable condition.

Figure captions

1. Structure, velocity of seismic waves, and density distributions of the crust and upper mantle of the eastern Snake River Plain, Idaho. Models from Braile and others (1982), Sparlin and others (1982), Priestly and Brune (1982), and D. R. Mabey and W. D. Stanley (in Kuntz and others, 1983). Density distribution model to right, taken largely from data by D. R. Mabey and W. D. Stanley (in Kuntz and others, 1983) is used in determination of magma driving pressures shown in figure 4.
2. Gravity and magnetic anomalies and inferred thermal characteristics of the crust in the Great Rift area, eastern Snake River Plain, Idaho. For all maps, cross-hatched and en echelon lines show locations of eruptive fissures and noneruptive fissures that constitute the Great Rift. Outlines of Craters of the Moon (COM), Kings Bowl (KB), and Wapi (WP) lava fields also shown. Map A shows regional gravity (from D. R. Mabey in Kuntz and others, 1983), Map B is residual magnetic map continued upward to 11.4 km above sea level, contour interval 20 gammas (from Bhattacharyya and Mabey, 1980). Map C is contour map to of the depth below sea level of the base of the magnetized crust, contour interval is 1 km (from Bhattacharyya and Mabey, 1980).

3. Geothermal gradients and pertinent experimental data relating to generation, storage, and migration of basaltic magmas along the Great Rift. Geothermal gradients from Lachenbruch and Sass (1978). Abbreviations used on geothermal gradients: BMH-Battle Mountain high gradient, HBR-high Basin and Range gradient, LBR-low Basin and Range gradient, SC-stable continent gradient. Numbers in parentheses on each curve are heat flow in HFU. Horizontally ruled area represents possible temperature-depth relations in crust along the COM segment of the Great Rift, as described in text. Dashed lines represent possible geothermal gradients for the northern (N) and southern (S) parts of the Great Rift. Abbreviations used for experimental data: GSS-granite water-saturated solidus (Merrill and others, 1970), GDIIS-granodiorite type II solidus (Robertson and Wyllie, 1971), GDDS-granodiorite dry solidus (Robertson and Wyllie, 1971), SLDS-spinel lherzolite dry solidus (Kuishi and others, 1968), ML-McKinney basalt dry liquidus (Thompson, 1972), MS-McKinney basalt dry solidus (Thompson, 1972), KHL-King Hill basalt dry liquidus (Thompson, 1972), KHS-King Hill basalt dry solidus (Thompson, 1972).
4. Lithostatic and magma driving pressures at various depths, calculated from relations discussed in text. To left, variation in lithostatic pressure (5) and magma driving pressure (6) calculated from the crustal density model shown. To right, magma driving pressures for magmas of densities 2700 kg/m^3 and 2570 kg/m^3 , rising from depths of 30 and 40 km.

5. Relations of magma supply rate, pressure regime of magma reservoir, and eruptive processes (after Shaw, 1980).
6. Relation between magma driving pressure and time in the development of the Craters of the Moon lava field. Ages for eruptive periods from Part 1. Dashed lines denote uncertainty in ages of eruptive periods and in the duration of repose intervals, as discussed in Part 1.
7. Schematic representation of magmatic plumbing system beneath the northern part of the Great Rift. See text for discussion.



Figure 1 (right side)

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Velocity (km/s)

Density

1-10/85

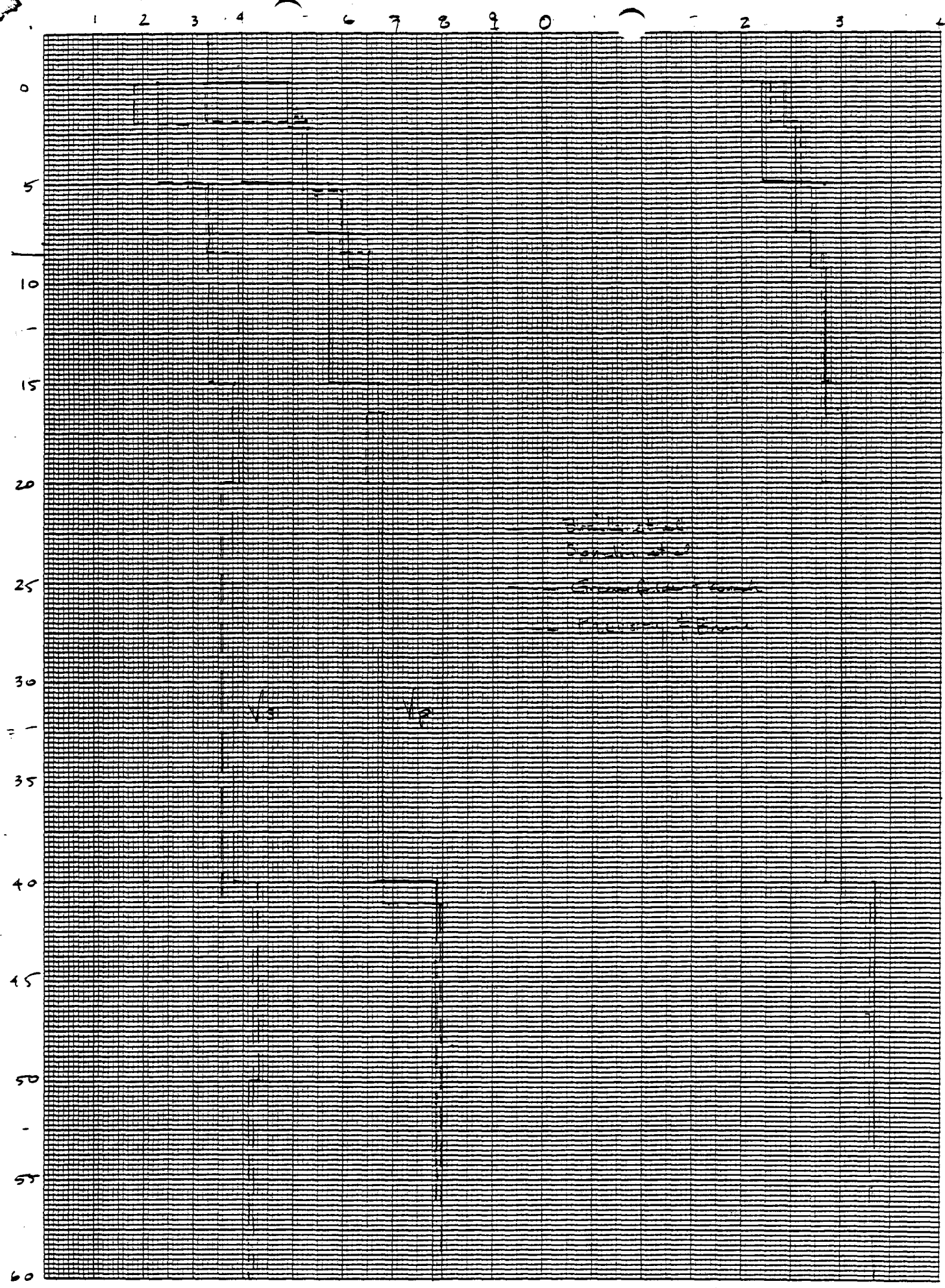
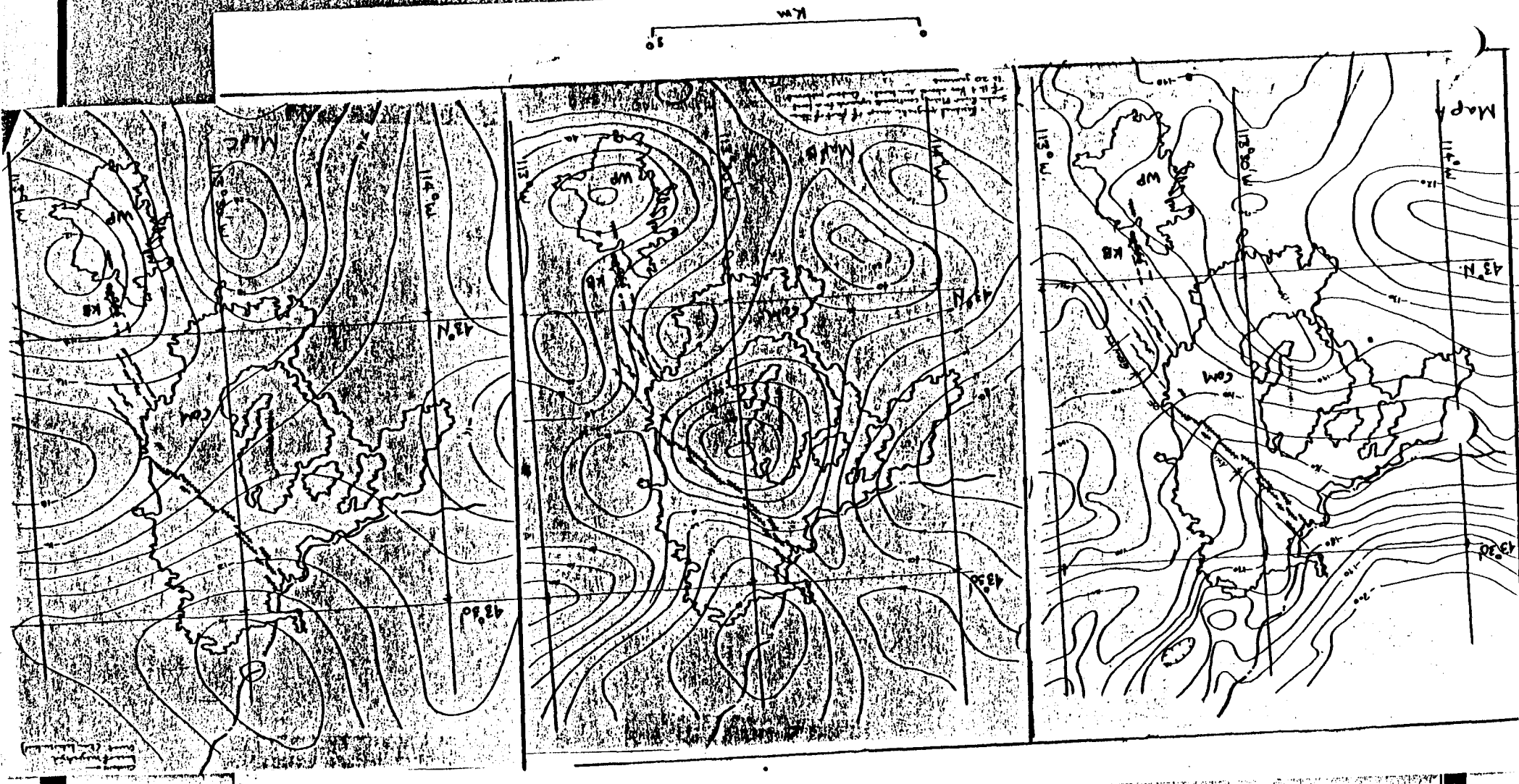


Figure 1 (left side)

Figure 2



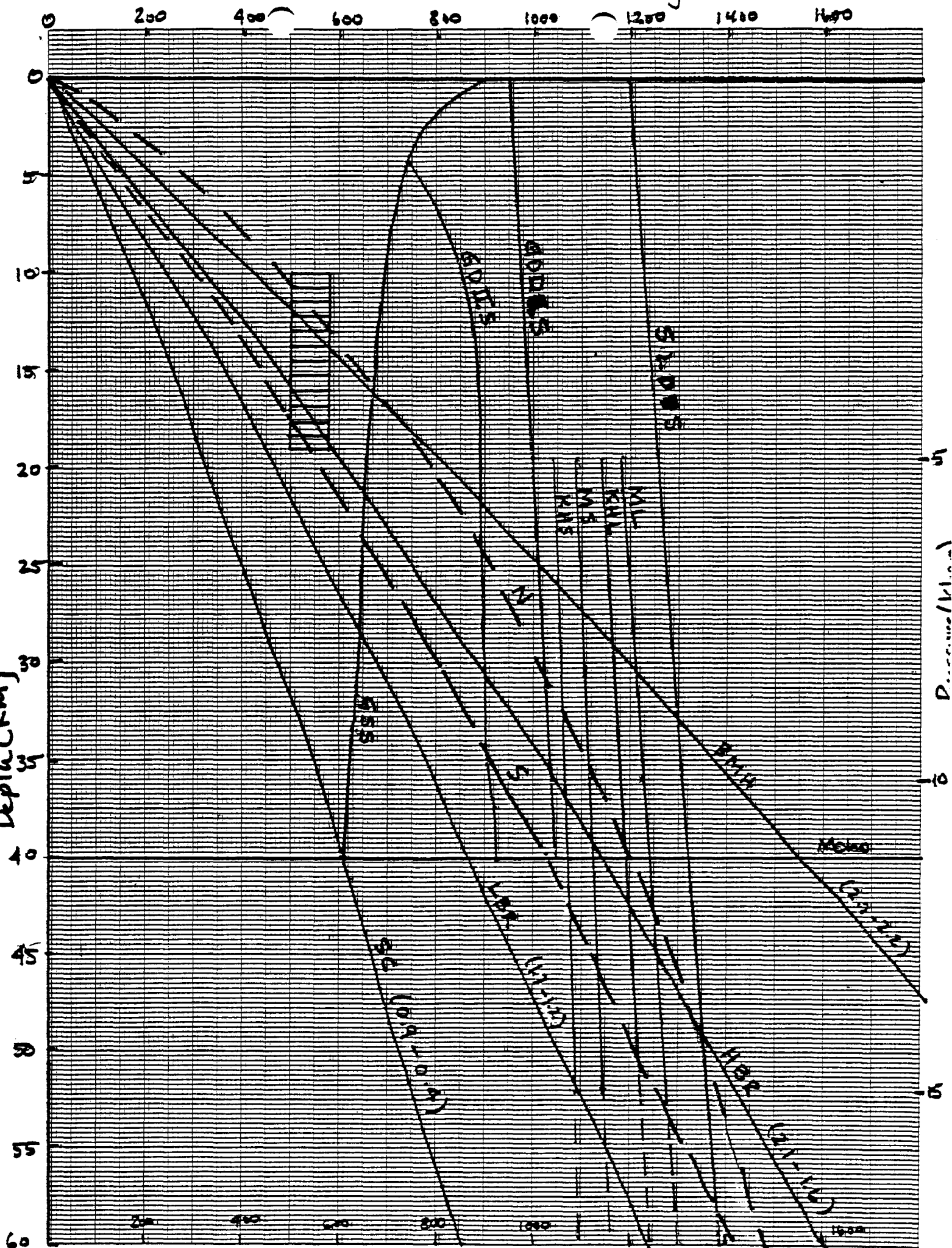


Figure 4

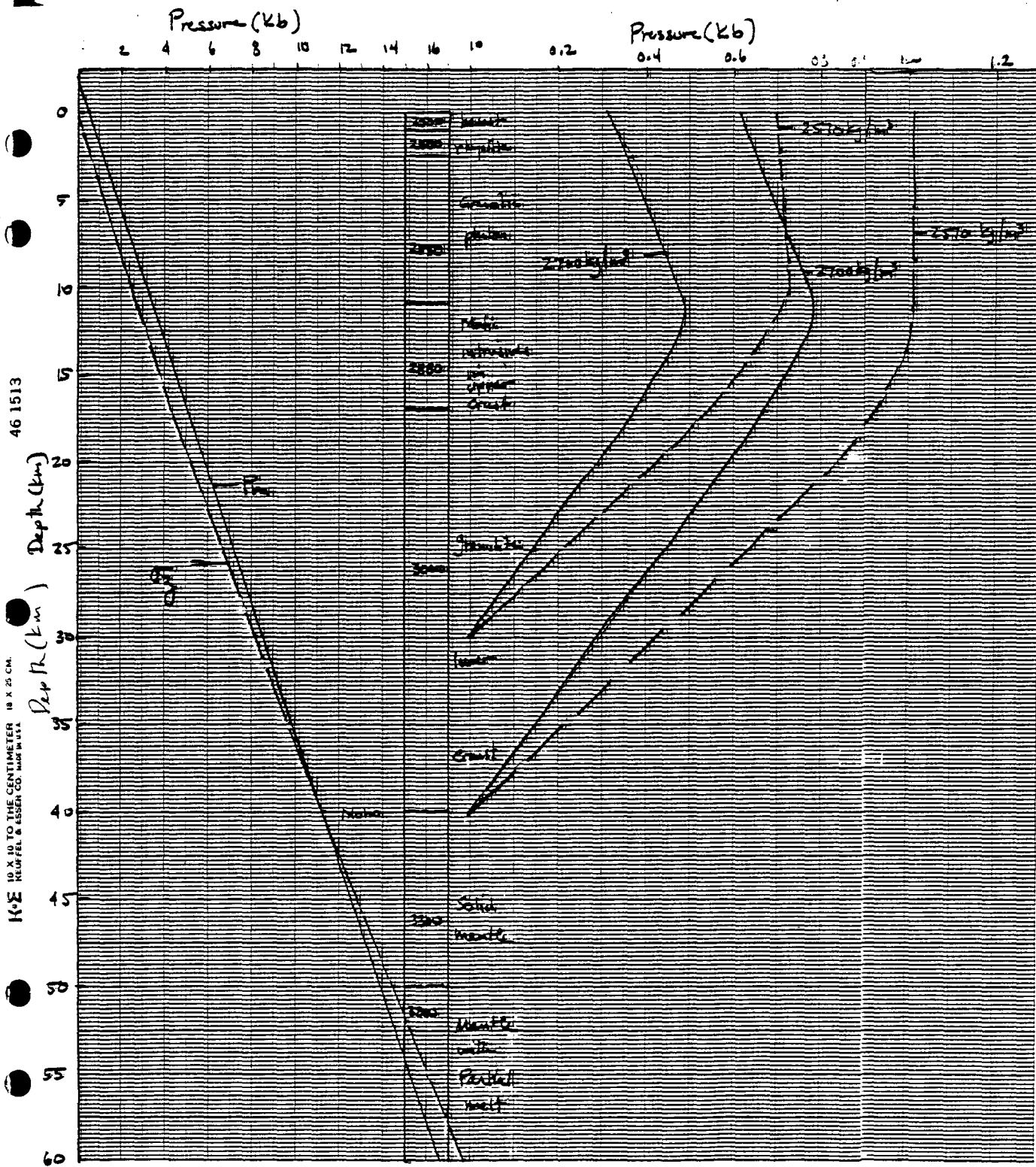


Figure 5

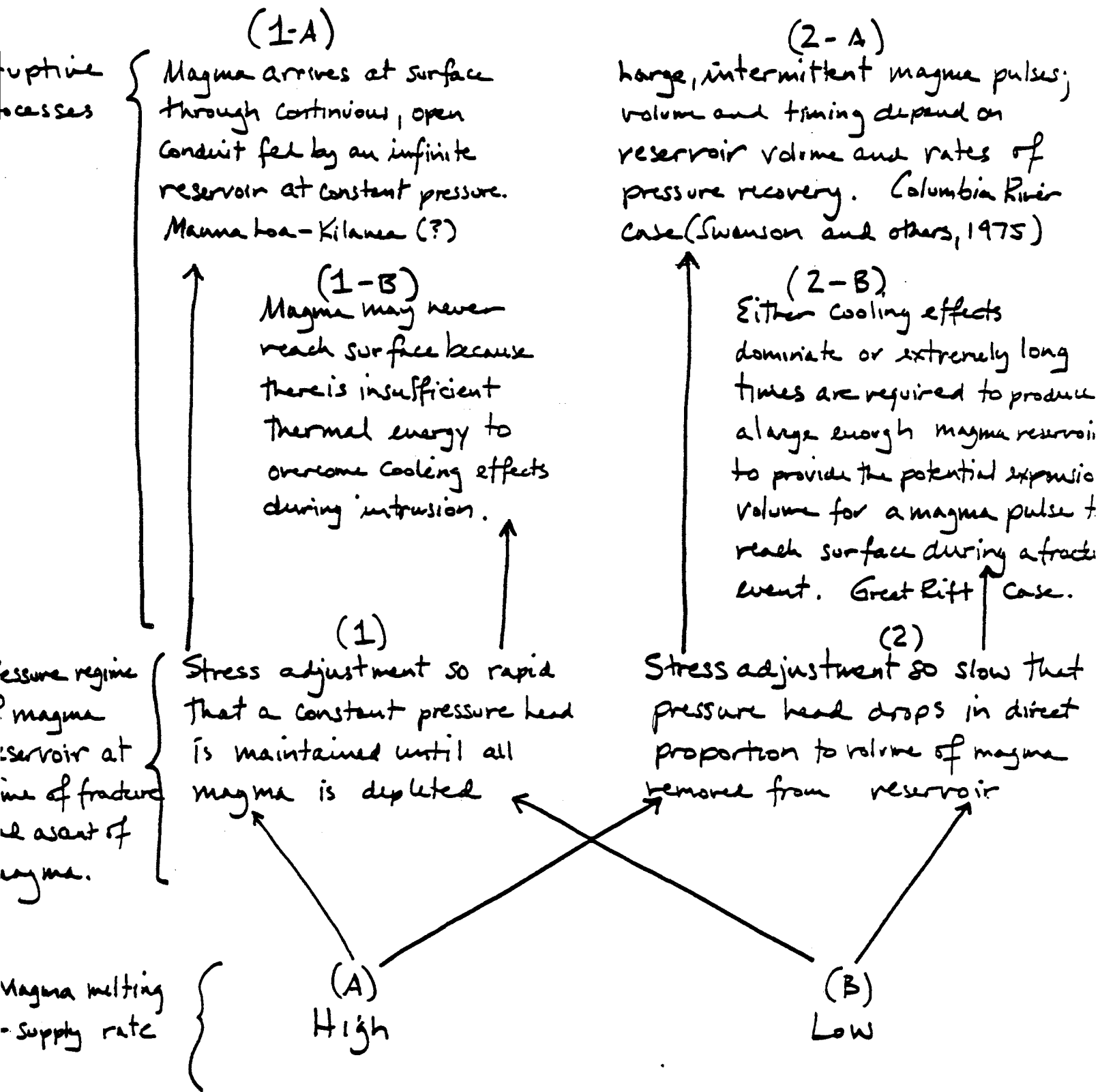
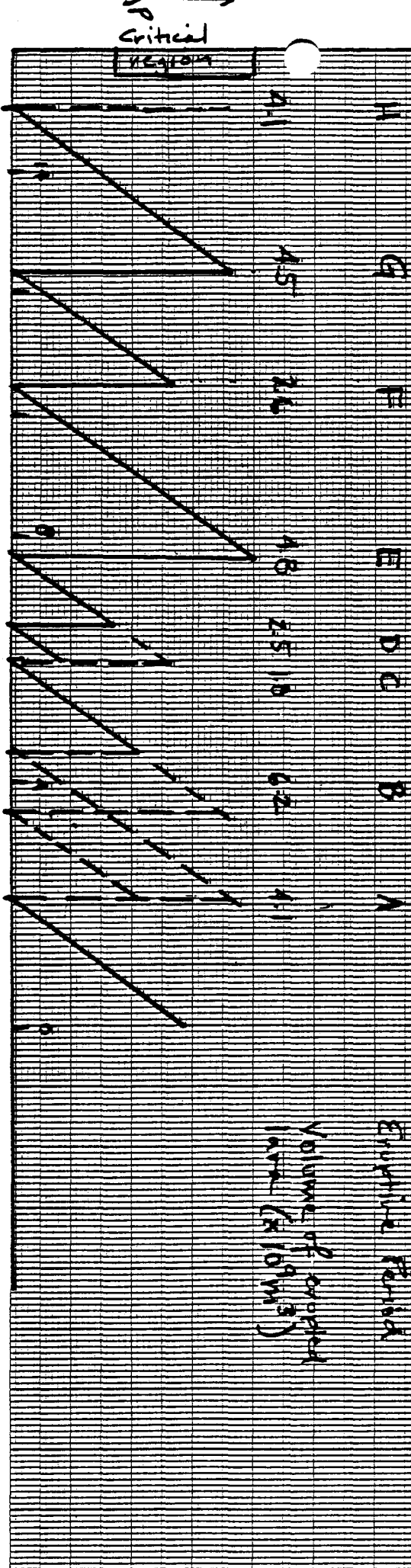


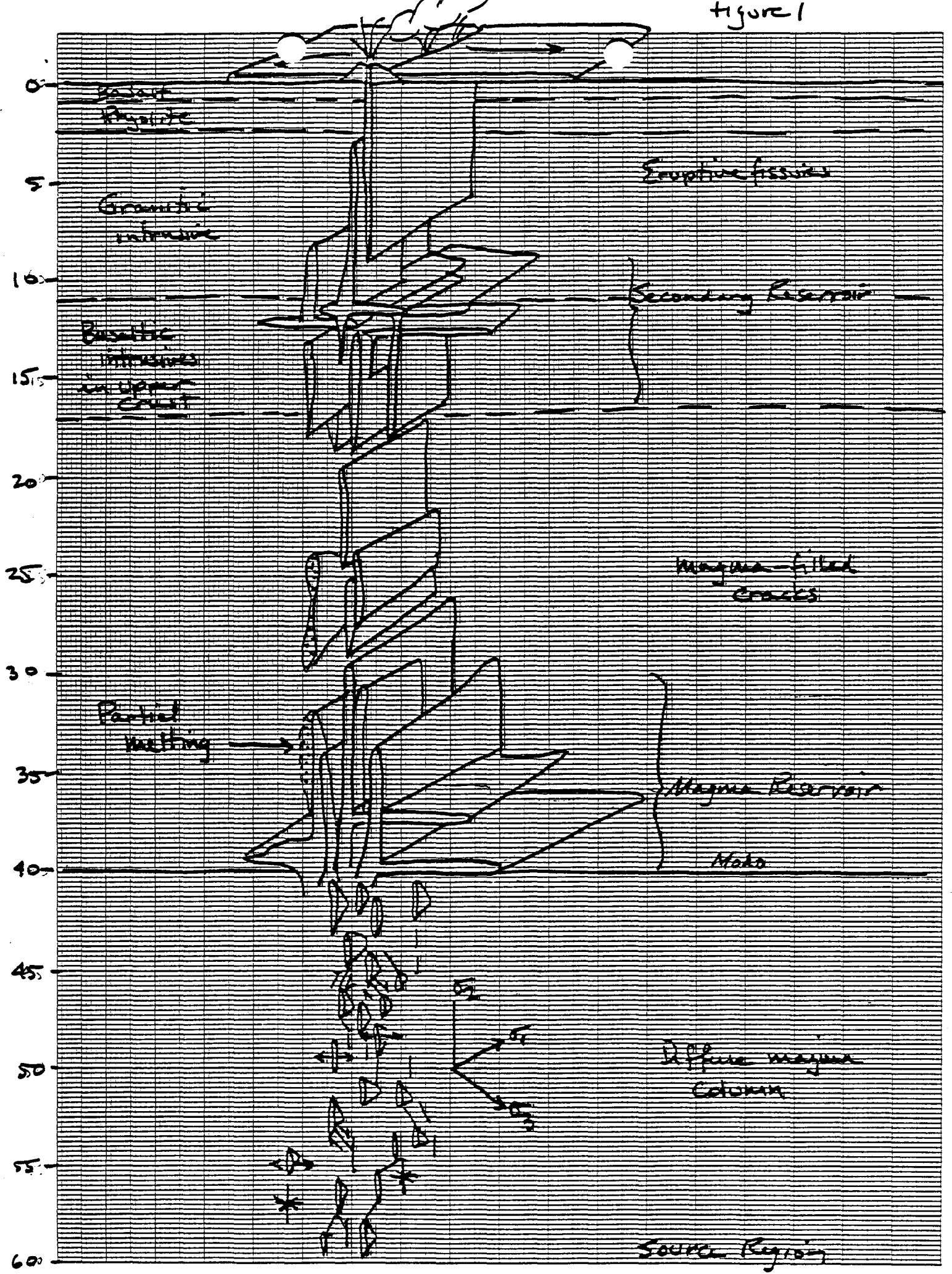
Figure 6



Years B.P. (10³)

figure 1

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Wapi, and Kings Bowl

Table 1.--Characteristics of vents and flows for selected eruptions in eruptive periods A and B, craters of the noon lava field

Flow name	Length of <u>single</u> eruptive fissure segments. Length of zone of fissures in parentheses (km)	Estimated fissure width (m)	Estimated flow volume (km ³)	Flow type	Comments
Broken Top	unk (<0.3)	unk	0.1	Surface and tube(?) fed	Intermediate volume stage 2 eruption. Flow issued from 2 obscure vents on east and south sides of Broken Top cinder cone. Vents now largely obscured by colluvium
Blue Dragon	0.5 (1)	1-2	3.4	Tube-fed pahoehoe flows	Large volume, sustained fissure eruption of stage 3. Eruptions from southernmost cinder cone in Big Crater cone complex, and from 0.5 km-long fissure south of Big Craters. Spatter cones aligned along fissure have internal diameters of 1-3 m. Fissure width estimated from smallest diameters
Flows of Trench Mortar Flat	0.3-1.3 (5)	<2	0.03	Shelly, thin pahoehoe flows	Small volume, stage 1 fissure eruption. Flows erupted from echelon segments (each 0.3 to 1.3 km long) in 5 km-long zone of eruptive fissures. Fissure widths estimated at localities where erosion-enlargement judged to be minimal.
North Crater	unk	unk	0.01	Surface-fed pahoehoe flows	Small volume eruption from pipe-like vent(?), vent now covered by colluvial cinders from miner wall of North Crater
Big Craters	<0.2 (0.9)	1-2	0.05	Surface-fed pahoehoe and slab pahoehoe flows	Small volume, stage 2 fissure eruption. Two sources fissures north of Big Craters cinder cone complex are short (<0.2 km), but Big Craters cinder cone complex is aligned along fissure system 0.9 km long. Fissure width estimated from deepest part of eastern eruptive fissure
Serrate, Devils Orchard, Highway	unk	unk	0.5 for all 3 flows	Bulbous aa and block flows	Vent area near North Crater largely destroyed by eruptions, also covered by younger North Crater flow. Eruption explosive because remnants of Crater walls are contained in flow. Eruptions may have been from central pipe-like vent in North Crater
Kings Bowl	0.1-0.5 (4.8)	<1.5 (measured)	0.01	Shelly and thin pahoehoe flows	Small volume, stage 1, fissure eruption. Eruptions from about a dozen fissures in a zone about 6.5 km long. Dikes <1.5 m thick exposed in fissure at Kings Bowl (see Greely and others, 1977, figures 11-14, 11-16; Greely, 1977, figure 3-19.)
Wapi	(>0.6)			Surface and pahoehoe flows	Large volume, stage 3 eruption. Vent complex consists of eleven eruptive centers, aligned over a buried eruptive fissure (Champion and Greely, 1977)
Vermillion Chasm	0.6-1 (2.9)	1-2	0.1	Shelly-thin pahoehoe flows	Small volume, stage 1 fissure eruption. Eruptive fissures enlarged at most localities by explosive venting. Fissure width estimated at deepest, narrowest part of fissure system. Fissure system is 2.9 km long; consists of 3 fissures that range from 0.6-1 km in length
Deadhorse	0.1-0.6 (10)	<1.5	0.04	Shelly-thin pahoehoe flows	Small volume, stage 1 fissure eruption. Many thin flows that were erupted from numerous (>13) echelon right-stepping eruptive fissures that range from 0.1-0.6 km in length. The Deadhorse fissure system is the longest fissure system that was active along the Great Rift during a single eruptive pulse
Devils Cauldron	(>0.3)	unk	0.9	Tube and surface fed pahoehoe flows	Intermediate volume sustained eruption from central vent with lava lakes perched above vents. Fissure and vent are obscured by flows and lava lakes
Minidoka	unk	unk	3.0	Tube-fed pahoehoe flows	Large volume, sustained stage 3 eruption from central vent complex that is now covered by flows

unk = unknown

Table 1.--Characteristics of vents and flows for selected eruptions in eruptive periods A and B, craters of the noon lava field.

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Flows of Trench Mortar Flat	0.3-1.3 (5)	<2	0.03	Shelly, thin pahoehoe flows	Small volume, stage 1 fissure eruption. Flows erupted from echelon segments (each 0.3 to 1.3 km long) in 5 km-long zone of eruptive fissures. Fissure widths estimated at localities where erosion-enlargement judged to be minimal.
North Crater	unk	unk	0.01	Surface-fed pahoehoe flows	Small volume eruption from pipe-like vent(?), vent now covered by colluvial cinders from miner wall of North Crater
Big Craters	<0.2 (0.9)	1-2	0.05	Surface-fed pahoehoe and slab pahoehoe flows	Small volume, stage 2 fissure eruption. Two sources fissures north of Big Craters cinder cone complex are short (<0.2 km), but Big Craters cinder cone complex is aligned along fissure system 0.9 km long. Fissure width estimated from deepest part of eastern eruptive fissure
Serrate, Devils Orchard, Highway	unk	unk	0.5 for all 3 flows	Bulbous aa and block flows	Vent area near North Crater largely destroyed by eruptions, also covered by younger North Crater flow. Eruption explosive because remnants of Crater walls are contained in flow. Eruptions may have been from central pipe-like vent in North Crater
Kings Bowl	0.1-0.5 (4.8)	<1.5 (measured)	0.01	Shelly and thin pahoehoe flows	Small volume, stage 1, fissure eruption. Eruptions from about a dozen fissures in a zone about 6.5 km long. Dikes <1.5 m thick exposed in fissure at Kings Bowl (see Greely and others, 1977, figures 11-14, 11-16; Greely, 1977, figure 3-19.)
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Vermillion Chasm	0.6-1 (2.9)	1-2	0.1	Shelly-thin pahoehoe flows	Small volume, stage 1 fissure eruption. Eruptive fissures enlarged at most localities by explosive venting. Fissure width estimated at deepest, narrowest part of fissure system. Fissure system is 2.9 km long; consists of 3 fissures that range from 0.6-1 km in length
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Devils Cauldron	(>0.3)	unk	0.9	Tube and surface fed pahoehoe flows	Intermediate volume sustained eruption from central vent with lava lakes perched above vents. Fissure and vent are obscured by flows and lava lakes
Minidoka	unk	unk	3.0	Tube-fed pahoehoe flows	Large volume, sustained stage 3 eruption from central vent complex that is now covered by flows

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Flows of Trench Mortar Flat	0.3-1.3 (5)	<2	0.03	Shelly, thin pahoehoe flows	Small volume, stage 1 fissure eruption. Flows erupted from echelon segments (each 0.3 to 1.3 km long) in 5 km-long zone of eruptive fissures. Fissure widths estimated at localities where erosion-enlargement judged to be minimal.
North Crater	unk	unk	0.01	Surface-fed pahoehoe flows	Small volume eruption from pipe-like vent(?), vent now covered by colluvial cinders from miner wall of North Crater
Big Craters	<0.2 (0.9)	1-2	0.05	Surface-fed pahoehoe and slab pahoehoe flows	Small volume, stage 2 fissure eruption. Two sources fissures north of Big Craters cinder cone complex are short (<0.2 km), but Big Craters cinder cone complex is aligned along fissure system 0.9 km long. Fissure width estimated from deepest part of eastern eruptive fissure
Serrate, Devils Orchard, Highway	unk	unk	0.5 for all 3 flows	Bulbous a-a and block flows	Vent area near North Crater largely destroyed by eruptions, also covered by younger North Crater flow. Eruption explosive because remnants of Crater walls are contained in flow. Eruptions may have been from central pipe-like vent in North Crater
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