THE CHEMICAL EVOLUTION OF LAUPAHOEHOE GROUP ALKALIC LAVAS, SOUTH RIFT, MAUNA KEA, HAWAII

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ABSTRACT

Mauna Kea volcano consists of early shield-building lavas (the Hamakua Group), composed primarily of basalt, overlain by a thin alkalic cap of hawaiite (the Laupahoehoe Group). 37 samples representing these alkalic cap lavas were collected from the south rift of the volcano and analyzed for major elements and selected trace elements (Sc, Ni, Cu, Zn, Sr, Rb, Nb, Zr, Y). These lavas were collected from cones whose temporal relations are known and which span the entire stratigraphic sequence of the alkalic cap. Using stratigraphic relations, the chemical evolution of the Laupahoehoe Group was examined. The overall compositional variation was utilized to evaluate crystal tractionation versus partial melting control. In addition, compositional variation away from the summit was examined to evaluate possible concentric zonation.

The generally aphyric nature of Laupahoehoe Group lavas suggests that they are representative of liquid compositions. The lavas probably do not represent primary mantle melts, having mg numbers ranging from 37.6 to 48.5. These low mg numbers indicate that the lavas have undergone extensive fractionation. Major and trace element abundances suggest that the overall compositional variation in the alkalic cap lavas is primarily the result of crystal fractionation. Petrographic evidence and chemical variation suggest fractionation of plagioclase and Ti-rich magnetite at low pressures.

Chemical variation also suggests that clinopyroxene separation has occurred. To account for the absence of clinopyroxene phenocrysts in these lavas and the presence of xenoliths inferred to be mantle-derived, a dynamic flow crystallization model is proposed. Although olivine fractionation was not significant in controlling the compositional variation of Laupahoehoe Group lavas, olivine fractionation at some point in their evolution is indicated by the extremely low Ni abundances and low mg numbers.

Although the compositional variation observed in Laupahoehoe Group lavas is due primarily to crystal fractionation, the lack of any systematic chemical variation with time suggests that they were not the product of a single, continuously fractionating magma chamber being periodically tapped. These lavas may represent batches of compositionally similar magmas having undergone similar fractionation histories. The observed chemical variation is not consistent with the lavas being generated by successively decreasing amounts of partial melting, but some variation in partial melting may be indicated. The lack of systematic chemical variation with distance from the summit of the volcano suggests that any horizontal or vertical zonation present in the underlying mantle did not control the evolution of the Laupahoehoe Group.

The distinct compositional gap which separates the alkalic cap lavas of the Laupahoehoe Group from the underlying shield-building lavas of the Hamakua Group cannot be explained by crystal fractionation or by successively decreasing amounts of partial melting of a common source. It is tentatively suggested that the two groups represent independent magmas generated from distinct mantle sources.

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INTRODUCTION

Mauna Kea ("White Mountain") is the highest of the five principle volcanoes on the island of Hawaii, rising to a height of 4205 m (Fig. 1). Alkalic lavas consisting of hawaiite and mugearite form a thin cap over the underlying tholeiitic and alkalic basalts which form the bulk of the volcanic pile. The purpose of this study is to examine the chemical evolution of the alkalic lavas and propose a model to explain the observed compositional variation. Major elements will be used to determine whether the alkalic sequence on Mauna Kea can be explained as the result of crystal fractionation or partial melting variation. In addition, chemical variation with distance from the summit will be examined to test Jackson and Wright's (1970) concentric zonation hypothesis.

A Summary of Previous Investigations and Excursions to Mauna Kea

This section is a brief survey of early excursions to Mauna Kea and of subsequent investigations and observations by laymen and geologists on the nature of the volcano. The earliest investigations were carried out by missionaries and crew members of visiting ships. Later, geologists like C. E. Dutton ascended the volcano and made primarily descriptive observations. From the turn of the century to present date, geological investigations have included petrographic and chemical studies and also studies of products of previous episodes of glaciation.

Mauna Kea has been extensively visited since its first discovery by American and British exploration groups. Native Hawaiians did not visit the mountain often, probably out of deference to (and fear of) the goddesses who live there. Joseph Goodrich was the first recorded non-native to ascend Mauna Kea. He made two ascents, one in 1823 and another in 1832 (Goodrich, 1824, 1829, 1833). He made hand-specimen descriptions and commented on the nature of the lavas. He also described granite encased in shells, probably referring to the abundant gabbroic nodules with lava coatings (Goodrich, 1929). In 1825, James Macrae, a botanist on Lord Byron's H.M.S. Blonde, climbed the volcano and recorded his observations (Wilson, 1922). David Douglas, the English naturalist and botanist for whom the Douglas Fir is named, climbed Mauna

Kea in 1834 (Wilson, 1919). He died shortly afterwards in a wild bull pit on the flank of the volcano under clouded circumstances. Charles Pickering and W. D. Brackenridge, members of the U.S. Exploring Expedition under Naval Lieutenant Charles Wilkes, made the ascent in 1841 (Wilkes, 1845). The first accounts of ascents and descriptions of general volcanic features of Mauna Kea were made by non-geologists and are somewhat vague. They did, however, inspire future expeditions and paved the way for continued investigation of the volcano.

Cohen (1880) made the first petrographic observations of Mauna Kea lavas on two samples collected by Hillebrand. He recognized two lava types: basalts and pyroxene andesites (along with lavas transitional between the two). Dutton (1884) was the first trained geologist to examine the volcano. He climbed the mountain in 1882 from the Humu'ula saddle and spent a few hours at the summit making hand-specimen descriptions and general observations. He recognized the effects of frost action and also noted the lighter color of summit lavas compared to those exposed along the lower flanks of the volcano. Hitchcock (1909) climbed Mauna Kea in 1886 and made several observations. He suggested that Mauna Kea was older than Mauna Loa on the basis of erosional valleys along the base of the volcano being deeper and more developed. He also described dunite nodules included in volcanic bombs from the summit cones and made observations on the general nature of the lavas. Baldwin (1889) climbed to the summit in 1889 and, like Dutton, recognized the lighter color of summit lavas. Merrill (1893) made petrographic descriptions of samples collected by Preston. Cross (1915) reexamined these samples and classified them all as olivine-plagioclase basalts. He commented on the dominance of plagioclase and clinopyroxene phenocrysts over olivine.

In 1901, Daly climbed the south slope of Mauna Kea and collected two samples, one from the southeast slope (3353 meters elevation) and another from the summit area at Pu'u Poliahu. The first chemical analyses of Mauna Kea lavas were done on these two samples by Steiger (Daly, 1911a). Daly classified the samples as andesitic basalt and trachydolerite, respectively. Alternatively, Washington (1923) classified them as andesine andesites and Macdonald (1949a) referred to them as andesites. Daly (1911b) suggested that vertical magmatic differentiation resulted in olivine basalt comprising the bulk of the mountain below 1500 meters elevation and between 1500 and 3500 meters elevation. Above the 3500 meter level, trachydolerite and andesitic basalt or "basic augite basalt" form the bulk of the largely pyroclastic material. Daly (1911a) made note of lherzolite nodules contained in the lavas and suggested that they represent accumulations of minerals separated from their host magmas by crystal fractionation. Daly (1910) was also the first to recognize evidence of previous glaciation on Mauna Kea.

Bryan (1915) described the general geologic features of Mauna Kea and suggested that volcanism on the upper slopes continued for a longer period of time than on the lower section of the mountain. He based this conclusion on observations that large gulches are restricted to the lower flanks. Powers (1920) examined Mauna Kea lavas exposed along the Hamakua Coast and reported an olivine nodule from lavas collected at Kapulena in the Waipi'o area. He believed that the lava flow which formed the Laupahoehoe Point peninsula issued from fissures near the mouth of the valley and was the most recent lava erupted near the base of the mountain.

Washington (1923) made the first comprehensive survey of Hawaiian lavas. He analyzed 10 samples collected along the highways encircling Mauna Kea and made petrographic descriptions of the lavas. He classified four of these samples as andesine and andesites and andesine basalts of various texture (e.g. aphyric, ophitic). Other samples include aphyric basalt, feldspar-phyric basalt, crysophyric oligoclase basalt and picrite basalt. Barth (1931) made more detailed petrographic descriptions of nine of Washington's Mauna Kea samples. P_2O_5 concentrations of Washington's samples are not consistent with later analyses of lavas collected from similar localities. Washington recognized the interfingering of Mauna Kea lavas with those of Hualalai and suggested that the contact lay beneath recent Mauna Loa lavas. He thought the contact was indicated by an abrupt change in the red soil near the Keamuku sheep station along the Kona-Waimea road.

Thomas A. Jaggar, founder of the Hawaiian Volcano Observatory in 1912, visited the south slope and summit area in 1925. He supported Daly's conclusion of previous glaciation, citing the large alluvial fans as evidence (Jaggar, 1925a, 1925b). Gregory and Wentworth (1937) reported additional, more detailed evidence for glaciation. Wentworth and Powers (1941) spent five weeks on the upper slopes and suggested that four separate stages of glaciation had occurred. Stearns (1945), however, believed two of these deposits to be explosion breccias and another to be a fanglomerate deposited during flooding caused by volcanic melting of glacial material. Glacial features on Mauna Kea have also been described by Wentworth (1940, 1966), Wentworth and Powers (1943), Porter (1973, 1979a) and Porter et al.,(1977).

Wentworth (1938) examined ash-deposits and made petrographic and hand-specimen descriptions of tuffs of rocks from cinder cones on Mauna Kea. Chemical analyses were done for several palagonitized ash samples. Further studies of the tephra units on Mauna Kea have been made by Porter (1971, 1972a, 1973, 1979a).

Studies by Macdonald (1945, 1949a, 1949b) and Stearns and Macdonald (1946) represented the first reports on the geology of the entire volcano and included descriptions of the geomorphology, pyroclastic geology, petrography, stratigraphy and composition of the lavas. Previously published analyses (Daly, 1911a; Washington, 1923; Wentworth, 1938) were also included but no new analyses were performed.

Trace element studies by Wager and Mitchell (1953) included a sample previously analyzed for major elements by Washington (1923) and a new sample collected from the west slope of Mauna Kea. Four other previously analyzed Mauna Kea samples (Washington, 1923) were analyzed for trace elements by Nuckolds and Allen (1954). Kuno et al. (1957) performed trace element analyses of an andesine andesite collected by Washington and another Mauna Kea sample.

The first in-depth examination of the chemistry of lavas from Mauna Kea was done by Macdonald and Katsura (1964). They collected and analyzed major elements for 10 new samples collected along the lower flanks of the volcano. They also made use of the silica-alkali diagram first used by Tilley (1950). Macdonald (1968) later collected and presented major element analyses of 9 additional samples and made conclusions on the possible origin of the diversity of Hawaiian lavas and their interrelationships. Bence (1982) reanalyzed two of Macdonald's samples for both major and trace elements.

Evolutionary Sequence of Hawaiian Volcanoes

The island of Hawaii is the youngest island in the Hawaiian Chain. The age of volcanism along the chain generally becomes progressively younger from the northwest to the southeast (Dalrymple _et al_., 1981). Numerous hypotheses have been proposed to explain this age progression, but a stationary mantle "hot-spot" (Morgan, 1971, 1972a, b; Basu, 1975) is the most widely accepted.

Hawaiian volcanism generally follows a well-defined evolutionary sequence (Macdonald and Abbott, 1970). The vast majority of Hawaiian lavas fall into two suites defined chemically and petrographically by Tilley (1950): (1) the tholeiitic suite and (2) the alkalic suite. Macdonald (1968; Macdonald and Katsura, 1964) divided Hawaiian volcanism into a succession of stages. The earliest stage of volcanism has been designated the shield-building stage and is thought to represent the bulk of the volcano. It is during this 0.2 to 1.3 million year period (data from Macdonald and Abbott, 1970; Lanphere and Dalrymple, 1980) that 97-100% of the shield is constructed by frequent eruptions of fluid tholeiitic lavas from both summit and rift zone vents. Lavas of the tholeiitic suite consist predominantly of tholeiitic basalt, tholeiitic olivine basalt and oceanite with only one occurrence of rhyodacite (Stearns and Vaksvik, 1935; Macdonald, 1940; Macdonald and Katsura, 1964) and icelandite (Sinton, 1981) in the Wai'anae Volcano, O'ahu. Small segregation veinlets and patches of Fe-rich basalt, pegmatoid and granophyre are also present (Kuno et al., 1957; Macdonald, 1968). Rocks from the tholeiitic suite are generally dark gray to black in color and commonly contain phenocrysts of forsteritic olivine, plagioclase (An 50-70) and less common augite (Macdonald, 1949a). Accessory phases include magnetite, apatite, ilmenite and rare rutile. The rare coarse grained plutonic xenoliths collected from tholeiitic lavas are cumulates consisting primarily of olivine with plagioclase and clinopyroxene (Jackson, 1968). The cumulate nature of the xenoliths suggests they were formed in floored magma chambers (Jackson, 1968). Although seismic evidence (Eaton, 1962, 1967; Estil and Dunnebier, 1979) supports a mantle origin for Hawaiian tholeiites, mantle-derived xenoliths are not present, probably due primarily to settling during magma storage in crustal level reservoirs. Disaggregation and resorption may also be important in preventing the incorporation of mantle-derived xenoliths in the lavas.

Alkalic and tholeiitic basalts have been dredged from Lo'ihi, a submarine volcano located southeast of the island of Hawaii and thought to be at a very primitive stage of evolution. The presence of alkalic lavas may indicate that a period of alkalic volcanism precedes tholeiitic volcanism (Moore et al., 1979, 1982; Clague et al., 1981a). This is supported by age dating measurements using palagonite growth rates which indicated that the alkalic basalts are older (about 500 years b.p.). It has been suggested (Green, 1971;

Moore et al., 1982) that during the initial stages of Hawaiian volcanism, small amounts of partial melting may create alkalic lavas. As the amount of partial melting increases, tholeiitic lavas predominate and, finally, during the waning stages of activity, the amount of partial melting decreases and alkalic lavas again are erupted. Frey and Clague (1981), however, suggest that on the basis of minor and trace element data, Lo'ihi tholeiites and alkalic basalts are not derived from a compositionally homogeneous source by different amounts of partial melting. A change in the depth of melting or a heterogeneous source, may be responsible for the variation in the magma type.

The alkalic cap stage follows the principle shield-building stage and is characterized by less frequent and generally more explosive eruptions. The lavas consist of alkalic basalts, hawaiites, mugearites, ankaramites, benmoreites and trachytes (Macdonald and Katsura, 1964). Eruptive activity is not confined to rift zones as during the tholeiitic stage and lavas tend to be principally erupted from cinder cones scattered over the summit and flanks of the volcano. Xenoliths of both cumulate and metamorphic texture have been collected from lavas belonging to the alkalic suite (Jackson, 1968). Cumulates include gabbro, dunite, wehrlite, anorthosite, clinopyroxenite and troctolite. Nodules exhibiting obvious metamorphic textures are predominantly dunites and wehrlites. The presence of mantle-derived xenoliths in alkalic lavas suggests a rapid rise of magmas to the surface with little or no crustal level storage. It has been generally assumed that Hawaiian alkalic magmas are generated at greater depths than tholeiitic magmas on the basis of mantle derived inclusions and experimental evidence (Macdonald, 1968; Ringwood, 1975; Yoder, 1976).

A period of volcanic inactivity takes place following the alkalic cap stage, lasting from 0.5 to 2 million years, during which extensive erosion of the volcanic edifice occurs. On some Hawaiian volcanoes (Ni'ihau, Kaua'i, East Moloka'i, West Maui, Ko'olau), this period of inactivity is followed by small, explosive eruptions of more strongly silica undersaturated lavas (nephelinites, melilitites, basanites and basanitoids). This period of reviewed volcanism has been termed the "post-erosional" stage (Macdonald, 1968). These eruptions tend to be somewhat concentrated along rift zones which transect the general alignment of the Hawaiian Ridge (e.g. the Honolulu Volcanic Series on O'ahu, Winchell, 1947). It has been suggested on the basis of isotopic and trace element date (Clague and Frey, 1982) that the lavas were generated by small amounts of partial melting (5-11 %) of a garnet lherzolite source. The presence of garnet-bearing ultramafic nodules of mantle origin (Green, 1966; Jackson, 1966, 1968; White, 1966; Kuno, 1969; Beeson and Jackson, 1970; Jackson and Wright, 1970; Leeman, 1979) has been proposed to indicate a deeper source for the post-erosional volcanics than for the alkalic or tholeiitic lavas.

The general evolutionary sequence is incomplete on several Hawaiian volcanoes. Mauna Loa and Kilauea have not yet left the tholeiitic shield-building stage, although Mauna Loa may be close to it judging by a single occurrence of alkalic basalt (Mattey, personal communication). The volcanoes of Hualalai, Mauna Kea, Kohala, Lana'i, West Moloka'i and Wai'anae have not experienced post-erosional volcanism. The Ko'olau volcano has no alkalic cap.

The Tholeiite-Alkalic Basalt Transition

There are presently two general schools of thought concerning the nature of the transition between the tholeiitic and alkalic suites in Hawaii: (1) fractional crystallization alone, at one or more than one depth, and (2) successively smaller amounts of partial melting. Other processes have also been previously called upon by various authors. However, mechanisms such as limestone assimilation (Daly, 1911a, 1944; Macdonald, 1949b), igneous carbonatite assimilation (Macdonald, 1968), thermodiffusion and volatile migration (Macdonald, 1968) have largely been ruled out as being significant in producing the transition.

Fractional Crystallization

Much of the arguments in favor of derivation of the alkalic suite from a primitive tholeiite parent magma have centered on volume relationships, chemical gradation and interbedding on the two rock types (Macdonald and Katsura, 1962, 1964).

Workers first investigating Hawaiian volcanism (Cross, 1915; Washington and Keyes, 1928; Barth, 1931; Kennedy, 1933) considered all Hawaiian magma types to be derived from olivine basalt solely by crystal fractionation. Later studies by Macdonald (1949a, 1949b, 1968; Macdonald and Katsura, 1964), Tilley (1950), and Wager and Mitchell (1953) proposed that alkalic olivine basalts are probably derived from olivine tholeiite magmas. Utilizing modal abundances, major and trace element variation, and high pressure experimental data (Tilley _et al_., 1965; Green and Ringwood, 1967), Macdonald (1968) and Wager and Mitchell (1953) proposed that olivine, orthopyroxene (at high pressures, Macdonald, 1949b) and also clinopyroxene and plagioclase were the primary fractionating phases.

A major drawback to these studies has been the selection of an appropriate parent magma composition. Primitive tholeiitic olivine basalt has been the most commonly called upon candidate (e.g. Macdonald and Katsura, 1961). However, several workers (e.g. Powers, 1935, 1955; Kuno _et al_., 1957) have pointed out that undersaturated alkalic basalts cannot be derived from a saturated tholeiite parent magma by crystal fractionation at low pressures. Experimental evidence reported by Yoder and Tilley (1957, 1962) on a low pressure thermal divide in the region of the Fo-Ab-Di join in the simplified basalt tetrahedron (Ne-Fo-Ab-Di-Si) supports this. Alternatively, Green and Ringwood (1967) proposed that alkalic olivine basalt may be derived from an olivine tholeiite parent magma by subtraction of aluminous enstatite at high pressures (13.5 to 18.0 kbar, 35-70 km depth) and, thus, drive liquid compositions across the low pressure thermal divide. The parent composition was taken from Macdonald and Katsura (1961) who averaged two samples from the 1959 Kilauea Iki eruption. The Kilauea Iki lavas were proposed to be closer to a more "primitive" composition than other Kilauea lavas because the high discharge rate and high temperatures of the erupted lavas (1190-1200°C) suggested rapid rise of the magma. However, Green and Ringwood's parental tholeiite composition is probably inappropriate because the 1959 Kilauea Iki lavas are highly olivine-phyric and, thus, probably do not represent liquids. Bence (1981) has pointed out that this parent composition also does not fall along the olivine control line on the Ol-Pl-Si join as would be expected for a magma parental to Hawaiian tholeiites. A more reasonable parent composition (one which falls along the olivine control line - has higher silica, lower aluminum, TiO_2 and CaO; see Bence, 1981, table 1.2.6.14 for comparison) will not fractionate towards alkalic basalt (Bence, 1981).

Partial Melting

Presnall _et al_., (1979) proposed that the tholeiite-alakalic basalt transition may be the result of successively decreasing the amount of partial melting of a source that migrates in depth with time. Using the data of Beeson (1976) from East Moloka'i, norms of phenocryst-free compositions were plotted on the basalt tetrahedron. Compositions become less silica saturated with decreasing age and, thus, trend away from the silica apex (Fig. 2). This trend cannot be explained by crystal fractionation. Superimposed on this trend away from the silica apex is an opposite trend due to crystal fractionation. This can be seen on the tetrahedron by the shift of compositions back towards the silica apex. Samples falling along this superimposed trend show decreasing mg numbers (100 Mg/(Mg+Fe^{2+})) and increasing Na_20, K_20, $Ti0_2$ and P_20_5 concentrations with decreasing age, thus supporting the proposition that crystal fractionation did occur. The trend towards and across the critical plane into the nepheline normative field may be explained by successively decreasing the amount of partial melting.

The origin of alkalic olivine basalt magmas by small amounts of partial melting of a garnet lherzolite source composition at high pressures has been proposed by Gast (1968) on the theoretical grounds and by Kushiro (1972) on experimental evidence. Frey (1979), in reviewing trace element studies of alkalic suites, pointed out that small amounts of partial melting (less than 15 %) of garnet peridotite is consistantly proposed. Leeman and Budahn (1979) proposed, on trace element considerations, that Hawaiian tholeiites may be generated by greater amounts of partial melting (15-20 %).

Kohala and Haleakala Types of Alkalic Cap Volcanism

Macdonald (1959b) recognized two basic patterns of alkalic cap stage volcanism which follow different evolutionary trends: the "Kohala" type and the "Haleakala" type. In Kohala type volcanoes, the transition between slightly alkalic basalts and more differentiated alkalic lavas (principally mugearite, benmoreite and trachyte with rare hawaiite) is characterized by a sharp erosional unconformity (e.g. Kohala, West Maui, East Molokai). K-Ar dating of basalts and mugearites from Kohala (Mcdougall and Swanson, 1972) suggests that the hiatus in eruptive activity is probably brief, lasting between 0.1 and 0.2 million years. Haleakala type volcanoes are characterized by a gradual change from basalt to differentiated alkalic lavas, predominantly hawaiite with minor mugearite (e.g. Haleakala, Mauna Kea, Wai'anae, Kaua'i, West Moloka'i). The reason why some volcanoes develop Kohala type and other Haleakala type volcanism is presently unknown.

Compositional Variation Within the Akalic Cap Stage

Major and trace element chemical variation within Hawaiian alkalic suites is generally considered to be the result of fractionl crystallization, primarily of plagioclase, clinopyroxene and, also, olivine, titanomagnetite and chrome spinel (Macdonald, 1949b, 1968; Powers, 1955; Kuno *et al*., 1957; Muir and Tilley, 1961; Macdonald and Katsura, 1962, 1964; Tilley *et al*., 1965) Well-developed alkalic sequences that range in composition from alkalic olivine basalt to trachyte can be found on Kohala (Stearns and Macdonald, 1946; Macdonald and Katsura, 1964; Macdonald, 1968; Feigenson et al., in prep.), West Maui (Stearns and Macdonald, 1942; Macdonald and Katsura, 1964; Macdonald, 1968) and Haleakala (Stearns and Macdonald, 1942; Macdonald and Powers, 1946, 1968; Macdonald and Katsura, 1964).

Compositional variation within the alkalic suite generally involves an increase in alkalies, silica, phosphorus and incompatible trace elements (Th, Zr, Rb, Ba) coupled with a decrease in MgO, CaO and compatible trace elements (Sr, Ni, Sc). This chemical variation probably reflects the subtraction of clinopyroxene, olivine and plagioclase. This is supported by the observed modal abundances of these phases. Part of the observed chemical variation, however, is probably due to partial melting variation. At some point in the alkalic sequence, magnetite becomes more abundant in the lavas. A corresponding decrease in TiO_2 and total iron is also observed. Towards the latter stages of differentiation, apatite may also be present in the lavas to a greater extent and is accompanied by a decrease in P_2O_5. Highly differentiated members of the alkalic suite (e.g. benmoreites and trachytes) may also contain amphibole phenocrysts (West Maui, Kohala). Macdonald (1968) considered chemical variation within the alkalic suite to be controlled mainly by low pressure crystal fractionation. However, the purported absence of clinopyroxene phenocrysts in Kohala lavas in spite of contrary geochemical evidence has led Furst and Weigand (1979) and Feigenson *et al*., (in prep.) to postulate the importance of high pressure crystal fractionation. However, clinopyroxene phenocrysts are not uncommon in alkalic basalts from the upper section of the Pololu Formation of Kohala. Thus, high pressure fractionation may not be necessary to explain the observed chemical variation.

REGIONAL GEOLOGY OF MAUNA KEA

Mauna Kea is the second oldest volcano on the island of Hawaii (from youngest to oldest: Kilauea, Mauna Loa, Haulalai, Mauna Kea and Kohala). K-Ar dating (Porter *et al.*, 1977; Porter, 1979a) indicates that volcanism probably began at least 400,000 years b.p., creating the bulk of the volcano (Fig. 3). Later volcanism appears to have been concentrated along three principle rift zones that radiate outwards from the summit towards the west, northeast and south. The most recent eruptions probably occurred about 4400 years b.p. (Porter, 1971, 1979a).

Previous petrologic and field studies (Macdonald, 1945, 1949a, 1949b, 1968; Stearns and Macdonald, 1946; Macdonald and Katsura, 1964) have indicated that volcanism on Mauna Kea can be separated into two major stages of activity. The two stages consist of any early, shield-building stage (the Hamakua Group) which comprises the majority of the volcanic pile and a later, alkalic cap stage (the Laupahoehoe Group) which forms a thin veneer of lavas (100-300 meters above sea-level). Porter (1979a) has dated the uppermost Hamakua Group lava exposed on the upper slopes of the volcano at 278,500 ± 68,500 years b.p..

The Hamakua Group

Stearns and Macdonald (1946) subdivided the Hamakua Volcanic Series into a lower member consisting of olivine basalt, less abundant picrite and minor basalt and an upper member consisting of olivine basalt with interbedded hawaiites and ankaramites. Macdonald (1968) has suggested that the contact between the lower and upper members, generally defined by a reddish soil layer, occurs at different stratigraphic levels depending on the location. However, the occurrence of many similar soil zones interspersed throughout a single section (e.g. Laupahoehoe Gulch) makes the division of Hamakua lavas into an upper and lower member on this basis problematic. Porter (1979a) renamed the Hamakua Volcanic Series the Hamakua Group and designated the upper member as the Hopukani Formation. Porter's terminology will be followed throughout this paper.

The lower member of the Hamakua Group is exposed only along the Hamakua Coast in sea-cliffs and in some roadcuts. Field studies by Garcia (personal communication) have indicated that lower member lavas consist of olivine basalt with much less abundant picritic basalt and minor basalt. The lavas occur as thinly bedded pahoehoe and aa flows. Both of the lower member lavas analyzed by Macdonald and Katsura (1964) are alkalic olivine basalts (C-77, C-78). Both samples have moderate K_2O concentrations (0.81 and 0.82) and plot in the alkalic field of the alkali-silica diagram as defined by Macdonald and Katsura (1964).

Lavas of the upper member of the Hamakua Group, the Hopukani Formation (Porter, 1979a), are exposed in kipukas, gulches on the upper south slope, and in flows and cinder cones on the lower flanks of the volcano. The lavas are well exposed in highway cuts and in sea-cliffs along the Hamakua Coast. Upper member lavas consist of olivine basalt with much less abundant augite-rich picrite basalt, ankaramite and hawaiite. They occur as thin-bedded (1 to 8 m thick) pahoehoe flows and thicker (8 to 12 m thick) aa flows, locally up to 35 m thick. The thickest flow measured by Garcia (personal communication) on the Hamakua Coast, however, was 19 m. Abundant olivine occurs in many of these lavas and a few of the flows are plagioclase-phyric. Along the Hamakua Coast, the thickness of Pahala ash was utilized to distinguish Hamakua Group flows from Laupahoehoe Group flows. Hamakua Group lavas generally have a greater than 2 m of ash and soil cover whereas Laupahoehoe Group lavas have less than 1.5 m of cover. However, a clear distinction between the two groups on this basis is often difficult. The type-section of Macdonald (upper member of the Hamakua Volcanic Series) is located on the Hamakua Coast along the south wall of Laupahoehoe Gulch. On the upper slopes of Mauna Kea, the two groups are separated by a layer of glacial till (the Pohakuloa Formation). Porter (1979a) designated the sections exposed in Waikahalulu and Pohakuloa gulches below the till as the type-sections for the Hopukani Formation.

The Laupahoehoe Group

The alkalic cap stage of volcanism on Mauna Kea has been designated the Laupahoehoe Volcanic Series by Macdonald (1945; Stearns and Macdonald, 1946). It was named for the intracanyon flow which formed a delta as it entered the ocean and created the Laupahoehoe Point peninsula. Laupahoehoe lavas form a cap on the volcano and are well exposed on the upper slopes and within some gulches on the Hamakua Coast (Fig. 4) (Porter, 1979b). The total thickness of the cap is unknown. Stearns and Macdonald (1946) suggested that the thickness of Laupahoehoe lavas on the flanks of the volcano is less than 100 m, possibly as little as 10 or so meters thick, locally. The thickness of the Laupahoehoe lavas in Pohakuloa Gulch is 110 m at 3600 m elevation and 10 m at 2200 m elevation (Porter, 1972b). If no caldera is present beneath the summit lavas, the Laupahoehoe section is probably about 200 m thick. The roughly concentric distribution of some vents around the summit suggests the presence of arcuate fissures occupied by ring dikes or cone sheets at some depth (Macdonald, 1945; Stearns and Macdonald, 1946; Porter, 1972b). These deep-seated fractures may represent the boundary of a buried caldera. If a caldera is assumed and Laupahoehoe lavas are the only lavas to fill it, the total thickness of the cap at the summit may be between 400 and 550 m (Porter, 1972b, see Fig. 2) Stearns and Macdonald originally estimated the thickness to be between 610 and 914 m.

Stearns and Macdonald (1946) separated the Laupahoehoe Volcanic Series into two members. The lower member, Pleistocene in age, comprises the bulk of the series and is exposed over the flanks of the volcano. The upper member, Holocene in age, consists of post-glacial cinder cones and associated lava flows that are exposed only along the south slope of Mauna Kea. The type-section of Stearns and Macdonald is located in Laupahoehoe Gulch where Laupahoehoe lavas are separated from Hamakua lavas by a marked erosional unconformity. There is, however, no widespread unconformity separating the two groups. At other locations, local erosional unconformities are located at different stratigraphic levels.

Porter (1979a) renamed the Laupahoehoe Volcanic Series the Laupahoehoe Group and dispensed with the upper and lower member distinctions of Stearns and Macdonald. Porter's type-sections are located in Pohakuloa and Waikahalulu gulches. In these gulches, Laupahoehoe Group lavas are separated from the underlying Hamakua Group lavas by glacial drift of the Pohakuloa Formation. This relationship is best illustrated in Waikahalulu Gulch where, along the west wall, a layer of interbedded boulders and gravels of inferred glacial origin (Porter _et al_., 1977; Porter, 1979a) separates the underlying Hopukani Formation lavas of the Hamakua Group from the overlying Liloean Stage lavas of the Laupahoehoe Group.

Porter (1979a; Porter et al., 1977) divided the Laupahoehoe Group into six stages of volcanic activity based on the occurrence of three glacial units interbedded with the lavas and pyrodastic material. From youngest to oldest, the six volcanic stages of the Laupahoehoe Group are the Loaloan, Kemolean, Kuupahaan, Hanaipoean, Poliahuan and Liloean stages (Fig. 3). The Loaloan Stage is equivalent to the upper member of the Laupahoehoe Volcanic Series of Stearns and Macdonald.

The Laupahoehoe Group lavas are young (Fig. 3). Radiometric age dates for lavas from the group range from 174,000± 37,400 years b.p. (Poliahuan Stage) to 4400 ± 110 years b.p. (Loaloan Stage). Ages of Loaloan and Kemolean Stage lavas were determined by radiocarbon methods (Porter, 1971, 1979a). Radiometric ages for Laupahoehoe Group lavas are difficult to determine by K-Ar methods because they are young and some appear to have undergone radiogenic argon loss during emplacement of adjacent or overlying flows. For example, the Liloean Stage is stratigraphically lower than the overlying Poliahuan Stage, but K-Ar dating indicates the Poliahuan Stage to be older (Porter, 1979a). The uppermost flow of the Hopukani Formation has been dated by K-Ar methods (Porter, 1979a) at 278,000 ± 68,500 years b.p., which sets a maximum age for the Laupahoehoe Group.

PREVIOUS GEOCHEMICAL STUDIES OF LAUPAHOEHOE GROUP LAVAS

Nine samples of Laupahoehoe lavas were analyzed by Macdonald and Katsura (1964) and Macdonald (1968). Following the criteria of Coombs and Wilkinson (1969) using differentiation index (D.I.) and normative An-content, the samples include six hawaiites (C-72, C-73, C-79, C-201, C-206, C-207), one mugearite (C-208) and a basalt (tholeiite transitional to alkalic olivine basalt) (C-205). The mugearite represents the most differentiated lava sample analyzed from Mauna Kea (D.I. 64.6, An-content 20.4). Macdonald (1968) also included an ankaramite (C-209) in the Laupahoehoe Volcanic Series. However, Porter (1979a) remapped the flow and reassigned it to the Hopukani Formation. C-205 appears to be an anomaly. Macdonald (1968) calls it a tholeiite transitional to alkalic olivine basalt and it plots in the tholeiite field of the alkali-silica diagram (Fig. 5). It was collected from a dome on the Hamakua Coast near O'okala. It has a thin (less than one meter) layer of ash and soil covering it, clearly indicative of a youthful age compared to the Hopukani lavas which have thicker ash and soil layers (greater than 2 meters) (Garcia, personal communication). A sample recollected from the same locality by Garcia is very different in outward appearance, having about 5% olivine as compared to C-205 which has only microphenocrysts of plagioclase. Collection and analysis of another sample should clarify whether this sample is anomalous or if a sample mixup occurred. One sample (C-207) is listed by Macdonald (1968) as hawaiite transitional to alkalic olivine basalt. It plots compositionally within the general cluster formed by nearly all of the Hamakua analyses and does not resemble typical Laupahoehoe compositions. Bence (1981) reanalyzed C-71, an ankaramite from the group, and C-72 for major elements and some trace elements, including rare earths. The only major discrepancy between Macdonald's and Bence's analyses is P_2O_5. Macdonald's analyses show significantly lower P_2O_5 abundances which may be due to analytical error.

PRESENT WORK

The purpose of this study was to determine the compositional variation among south rift Laupahoehoe Group lavas with age and distance from the summit and to test possible models (crystal fractionation vs. partial melting) that might be responsible for the variation. The processes controlling the chemical evolution of the alkalic cap on Mauna Kea can, thus, be better constrained. The south rift zone (Fig. 4) of Mauna Kea is the best area for examining temporal variations in the Laupahoehoe Group are present, as are lavas from the upper portion of the Hamakua Group (the Hopukani Formation). Lavas from all six stages were collected from cones along the south rift (Fig. 46). Elevations on the south rift range from 4206 m at the summit to 1950 m in the saddle. Unfortunately, vents from each stage are not evenly distributed on the slopes of the volcano. Middle stages are most common at the summit; early and late stages are most common on the flanks of the mountain.

In order to examine the chemical evolution of Hawaiian volcanoes, suites of lavas whose temporal relations are well-established must be used. The majority of studies, however, do not take the stratigraphy of the lavas into account and examine only the overall chemical diversity. Recent studies on Hawaiian volcanoes (Beeson, 1976; Clague and Beeson, 1980; Feigenson _et_ _al_., in prep.) have helped constrain the fractionation histories of the Hawaiian alkalic suite by utilizing stratigraphically controlled lava sequences.

Although the chemistry of lavas from Mauna Kea has been examined by many workers, stratigraphic control has, generally, not been utilized. Porter's (1979b) map of the stratigraphy of the upper slopes of Mauna Kea has made it possible to collect lavas whose temporal relations are known. These relations have been used to examine the chemical evolution of the alkalic suite on Mauna Kea. By examining the chemical evolution and chemical diversity of the lavas, it should be possible to determine whether the observed chemical variation is due primarily to crystal fractionation (at low or higher pressures) or if variations in the amount of partial melting must also be considered. Open-system fractional crystallization (O'Hara, 1977; O'Hara and Mathews, 1981) can be examined by observing the behavior of incompatible trace elements with time.

PETROGRAPHY

Laupahoehoe Group south rift lavas are generally aphyric with rare plagioclase (0-10%) and very rare olivine (less than 5%) and magnetite (less than 2%) microphenocrysts. The groundmass is composed primarily of plagioclase (30-50%), opaques (predominantly magnetite) (15-25%), olivine (2-8%) and a fine-grained matrix consisting of indeterminant phases too small for microscopic identification. Apatite is a rare groundmass phase in some samples. Megacrysts of olivine plagioclase and clinopyroxene are very rare. The uppermost lava from the Hopukani Formation (H-6) is aphyric and indistinguishable in hand specimen from the majority of Laupahoehoe lavas. The freshness of the lavas is not related to age but is inversely related to distance from the summit. Lavas exposed near to the saddle are subject to locally heavy rainfall and are generally more weathered. Most of the Recent lavas (Loaloan Stage) occur in the saddle area and are somewhat weathered on the surface. Nevertheless, only fresh samples were analyzed. Hopukani Formation lavas exposed in Waikahalulu Gulch are generally fresh.

RESULTS

36 new samples from the Laupahoehoe Group and one from the Hopukani Formation were collected from the south rift and analyzed by XRF for major and trace elements (Ni, Sc, Zn, Cu, Sr, Rb, Y, Zr and Nb) (see Table 1 for analyses). 13 samples were analyzed from the Loaloan Stage, 6 from the Kemolean, 6 from the Kuupahaan, 4 from the Hanaipoean, 2 from the Poliahuan and 5 from the Liloean Stage. The Uppermost flow of the Hopukani Formation exposed in Waikahalulu Gulch was also analyzed. 7 samples of Laupahoehoe Group lavas collected by Macdonald (Macdonald and Katsura, 1964; Macdonald, 1968) were also analyzed for Rb, Sr, Y, Zr and Nb (C-72, C-73, C-201, C-205, C-206, C-207 and C-208). Major elements were analyzed by XRF, atomic absorption (for Mg and Nz) and wet chemical techniques (for H_20, $C0_2$, Fe0 and P_20_5) at the University of Manitoba and trace elements were analyzed by XRF at the University of Hawaii (see Appendix 1 for details).

All chemical analyses (Table 1) have been recalculated anhydrous to 100%, with $Fe^{3+}/ (Fe^{2+} + Fe^{3+})$ set to 0.15 (Table 2). Norms have been calculated on this basis. The overall chemical variability of Laupahoehoe Group south rift lavas is small, but coherent trends are evident for many compatible and incompatible elements. Laupahoehoe south rift lavas plot in the alkalic field of the alkali-silica diagram (Fig. 5) and form a typical Hawaiian alkalic trend (Macdonald, 1968). Laupahoehoe Group lavas are separated from lavas of the Hamakua Group by distinct compositional "gap".

The majority of south rift lavas are nepheline normative but many are hypersthene normative. Only the Hanaipoean Stage contains no nepheline normative rocks. The Hopukani Formation sample collected from the south rift (H-6) is nepheline normative as well but is higher in MgO, CaO, TiO_2, Ni and Sc and lower in silica, alkalies, aluminum, Sr, Zr, and Rb than south rift Laupahoehoe lavas. Two south rift samples (one from the Loaloan Stage (ML-20)) and one from the Poliahuan Stage (Mp-1) are quartz normative. These samples were collected from near the summit and near the saddle, respectively. This illustrates the lack of correlation between degree of silica saturation and either stratigraphic position or distance from the summit.

South rift lavas form a positive linear trend towards increasing D.I. (differentiation index) and decreasing An-content (Fig. 6) typical of all Hawaiian volcanic suites. Overall, the Laupahoehoe Group trend increases slightly more positively than the Hamakua Group trend and a "kink" separates them. The lavas range in composition from hawaiite (An 30-50) to mugearite (An 15-30) based on normative An-content and D.I. (Coombs and Wilkinson, 1969). 5 analyses plot in the mugearite field but are separated by a large gap from the more evolved C-208 (mugearite, Macdonald, 1968).

On the Ne-Pl-Ol-Si ternary diagram from the generalized basalt tetrahedron (Fig. 7), Laupahoehoe south rift compositions generally cluster in the critically undersaturated field near the Ol-Pl join, and form a trend which crosses the boundary into the undersaturated field, generally following the one atmosphere cotectic (O'Donnell and Presnall, 1980). This trend continues slightly into the silica oversaturated field.

Major Elements

Zr, a moderately incompatible trace element in basaltic compositions, was chosen as an indicator of the degree of fractionation. Plots using MgO, K_2O, P_2O_5 and mg number all show more scatter than similar plots using Zr.

MgO, CaO, TiO_2 and total iron decrease progressively with increasing Zr abundance (Figs. 8, 9, 10, 11). MgO and CaO depletion may be the result of clinopyroxene or olivine + plagioclase fractionation. The decrease in TiO_2 may be the result of Ti-rich magnetite fractionation. Successively decreasing the amount of partial melting would cause TiO_2 to progressively increase.

SiO_2, K_2O and P_2O_5 all increase steadily with increasing Zr abundance (Figs. 12, 13, 14). Phosphorus exhibits the most linear enrichment trend. K_2O and P_2O_5 enrichment might be the result of either crystal fractionation or successively decreasing amounts of partial melting. An increase in silica is not consistant with successively decreasing amounts of partial melting, however. Al_2O_3 and Na_2O data are scattered but generally increase (Figs. 15, 16).

Trace Elements

Nb, Rb and Y behave incompatibly with respect to Zr abundance (Figs. 17, 18, 19). This increase in abundance may be due to either crystal fractionation or changes in the degree of partial melting. The observed trends intersect the ordinate, indicating that the bulk distribution coefficient for Zr is less than the bulk distribution coefficients of other trace elements measured.

Sr and Sc both decrease with increasing Zr abundance (Figs. 20, 21). Sr decrease may be due to crystal fractionation of plagioclase, apatite or alkali feldspar, but is inconsistant with decreasing the amount of partial melting. Sc depletion could be due to crystal fractionation of clinopyroxene or amphibole or to a decrease in the amount of partial melting. Ni abundances are very low (0-9 ppm) and appear to remain relatively constant with increasing Zr abundance (Fig. 22). This constant trend suggests that Ni is not being significantly removed by any fractionating phase, but this may be due to the very low abundances.

DISCUSSION

The Non-Primary Nature of Laupahoehoe Group Lavas

Normally, Hawaiian lavas contain phenocrysts, commonly olivine, plagioclase and/or augite. Thus, whole-rock analyses of these lavas might not represent liquid compositions. In contrast, Laupahoehoe south rift lavas are virtually aphyric and, therefore, should be good indicators of liquid compositions. This is supported by the occurrence of xenoliths, inferred to be mantle-derived, in some of the lavas (Jackson et al., 1982). The occurrence of these mantle-derived xenoliths suggests that the Laupahoehoe lavas traveled from mantle depths to the surface without significant shallow-level crustal storage and, hence, probably have not undergone appreciable shallow-level fractionation. However, despite the lack of phenocrysts and the presence of mantle xenoliths in these lavas, the mg numbers of south rift lavas are considerably lower than would be expected for magmas in equilibrium with mantle peridotite.

Irving and Green (1976) suggest that primary liquids in equilibrium with mantle olivine of Fo 86-90 should have mg values between 66 and 75. This is based on experimentally determined equilibrium values of the Fe/Mg distribution coefficient between olivine and liquid ($K_D = (X_{Mg}/X_{Fe}) / (X^{ol}_{Mg}/X^{ol}_{Fe})$). Values for this distribution coefficient have been determined by Roedder and Emslie (1970) and O'Hara et al.

(1975) for terrestrial basaltic and ultramafic melt compositions and range from 0.28 to 0.36. The coefficient appears to be independent of temperature, pressure and composition (Irving and Green, 1976; Irvine, 1977; Langmuir and Hanson, 1980). These distribution coefficient values have been used to calculate mg numbers which would be in equilibrium with likely mantle olivine composition. The calculated mg numbers are supported by melting studies of anhydrous peridotite which produced melts with mg numbers of 69-86 with residual olivine compositions of Fo 88-95 (Jaques and Green, 1980).

Mg numbers for Laupahoehoe south rift lavas range from 37.6 to 48.5. These relatively low values strongly suggest they have experienced significant fractionation and are, thus, not representative of primary mantle melts.

Chemical Variation Within the Laupahoehoe Group

The major and trace element and normative compositional variations exhibited by Laupahoehoe Group south rift lavas are examined below to help constrain the process(es) that might be responsible for producing the observed trends. Major and trace element variations are used to indicate what mineral phases were most likely involved in creating the observed variations.

Fractional Crystallization

Petrographic evidence suggests that plagioclase and magnetite fractionation may have occurred. Chemical variation supports this. However, chemical variation also indicates that fractionation of phases (clinopyroxne, olivine) which are not present in the lavas has also occurred. Evidence for crystal fractionation of plagioclase, magnetite, clinopyroxene and olivine will be discussed below. The evidence against significant fractionation of apatite, alkali feldspar, mica and amphibole will be presented as well.

Plagioclase

Plagioclase is present as a minor microphenocryst phase in some Laupahoehoe Group south rift lavas. Plagioclase fractionation is supported by the decrease in CaO and Sr with increasing Zr abundance (Figs. 9, 20). Previously determined Sr partition coefficients (Philpotts and Schnetzler, 1970; Allegre *et al.*, 1977; Villemant *et al.*, 1981) indicate that the phases which most strongly incorporate Sr ($D_{Sr}>1$) are plagioclase, apatite and alkali feldspar. However, apatite and alkali feldspar fractionation are unlikely to be strongly involved in removing Sr and supporting evidence is presented in a later section.

Ti-rich Magnetite

Magnetite is a significant groundmass phase in Laupahoehoe lavas and microphenocrysts are also present in some lavas. Significant magnetite removal is supported by the progressive depletion of TiO_2 and total iron with increasing Zr (Figs. 10, 11). Distribution coefficients of Ti in magnetite have been calculated for two Hualalai alkalic olivine basalts using the data of Bence (1981) and give values of 6.04 and 9.46. TiO_2 distribution coefficients for iron-oxides in hawaiites fall within this range: hawaiite from Mauna Kea (C-72), 8.08; hawaiite from Kohala (C-68), 7.57. It is unlikely that clinopyroxene fractionation can explain the observed TiO_2 depletion trend because the partition coefficient of Ti in clinopyroxene is probably not much higher than 1.0 (Allegre _et_ _al_., 1977). Distribution coefficients for TiO_2 in Haleakala alkalic lavas using Fodor _et_ _al_. (1975) clinopyroxene data and Macdonald and Katsura (1964) and Macdonald and Powers (1968) whole-rock data are: alkali olivine basalt, 0.34; hawaiite, 0.65; and mugearite, 0.90. Calculations for clinopyroxene phenocrysts in alkali olivine basalts from Hualalai (0.34 and 1.00) and Kohala (0.38) give similar results (Bence, 1981). The decrease in TiO_2/Al_2O_3 ratios with increasing Zr abundance (Fig. 23) also supports magnetite fractionation. Clinopyroxene fractionation would cause an increase in this ratio. This is based on the low TiO_2 contents (0.70-2.39 wt. %) of clinopyroxenes analyzed from alkalic olivine basalts from Hualalai and Kohala and from an ankaramite belonging to the Hamakua Group (C-71) (Bence, 1981). However, these compositions may not be representative of clinopyroxenes fractionating within Mauna Kea hawaiites, although TiO_2/Al_2O_3 ratios are not likely to be greater than 1.

Clinopyroxene

The decrease in MgO, CaO and Sc and the increase in Al_2O_3/CaO ratios with increasing differentiation of south rift lavas (Figs. 8, 9, 21, 24) strongly suggests that significant clinopyroxene fractionation has occurred. Clinopyroxene is the phase most capable of incorporating Sc in most basaltic melt compositions. D_{Sc}^{Cpx} values range from 3.0 to 3.5 (for review of partition and distribution coefficient data, see Allegre *et al*., 1981). The Sc depletion trend observed for south rift lavas (Fig. 21) supports the role of clinopyroxene fractionation. The change in the ratio of Al_2O_3/CaO is a good indicator of the relative importance of clinopyroxene versus plagioclase. Within Laupahoehoe Group south rift lavas, Al_2O_3/CaO ratios increase with increasing Zr abundance (Fig. 24) suggesting that clinopyroxene fractionation was dominant.

Zr/Y ratios of south rift lavas show a slight enrichment trend with increasing P_2O_5 (Fig. 25). This increase suggests either removal of a phase capable of fractionating Zr/Y ratios (probably clinopyroxene or garnet) or successively decreasing amounts of partial melting. Clinopyroxene fractionation is supported by the positive correlation between Zr/Y and Al_2O_3/CaO ratios.

Major and trace element variation trends support clinopyroxene fractionation, but clinopyroxene phenocrysts are not present in south rift lavas. The lack of clinopyroxene phenocrysts, despite supportative geochemical evidence, and the presence of inferred mantle xenoliths, suggests that clinopyroxene removal may be the result of dynamic flow crystallization and not gravitational settling in crustal magma chambers (Irving, 1978, 1980). In this model, clinopyroxene would precipitate from the magma and be plated along cooler conduit walls and pre-existing rock surfaces during transit of the magma to the surface. This process would not necessarily prevent entrained xenoliths from being transported. Pulses of ascending liquid might cause accumulated crystals to be plucked from conduit walls and included in the magma as cumulate nodules and xenolithic megacrysts (Irving, 1980). Clinopyroxene fractionation is supported by the common occurrence of clinopyroxene-rich cumulate xenoliths in pyroclastic ejecta and lavas from the south rift (although the relationship between the xenoliths and lavas has not been determined).

Thompson (1974) has experimentally demonstrated that clinopyroxene is on the liquidus of a hawaiite composition from Skye at pressures greater than 18 kb. This would be consistent with dynamic flow crystallization at mantle depths.

However, a hawaiite composition probably does not represent a likely parent composition for Laupahoehoe Group south rift lavas, especially since olivine is not on the liquidus at any pressure. Alkalic olivine basalt may also not be a suitable parent composition because olivine is on the liquidus only below 14 kb, followed by clinopyroxene to 30 kb (Takahashi, 1980). This is inconsistent with compositional trends indicating clinopyroxene fractionation and not olivine fractionation. Although these compositions may not represent Laupahoehoe Group south rift parent magma compositions, they do provide constraints on fractionation. A model such as dynamic flow crystallization is probably a more reasonable explanation for crystal separation in the upper mantle than an open magma chamber at these elevated pressures. At lower pressures, plagioclase replaces clinopyroxene as the liquidus phase of the Skye hawaiite. This is consistent with the observed plagioclase fractionation in Laupahoehoe south rift lavas. At lower temperatures, olivine joins plagioclase as a crystallizing phase. The plagioclase-olivine groundmass of south rift lavas supports this relationship.

It is possible that the "mantle-derived" xenoliths described by Jackson (1968) and Jackson et al. (1982) are not from the mantle. If these xenoliths were formed in crustal magma chambers, then, based on experimental evidence on hawaiites (Thompson, 1974), plagioclase would be the primary liquidus phase followed by plagioclase + olivine and finally by plagioclase + olivine + clinopyroxene. This would suggest that clinopyrexene is a late crystallizing (groundmass) phase and the almost total crystallization would be necessary to get significant clinopyroxene fractionation. It is, therefore, unlikely that extensive fractionation of clinopyroxene could occur at low pressures in a hawaiite composition. Limited clinopyroxene fractionation may not be able to explain the low Sc abundance (9-14 ppm) and the observed Al_2O_3/CaO enrichment trend in south rift lavas. Thus, it appears more likely that clinopyroxene fractionation in the hawaiites from Mauna Kea occurred at higher pressures. It is, therefore, suggested that separation of clinopyroxene via dynamic flow crystallization in the upper mantle may be responsible for some of the observed compositional variations.

Olivine

Olivine is present as a rare microphenocryst phase in Laupahoehoe Group lavas collected from the south rift. These lavas exhibit low (near the detection limit) and constant Ni with increasing Zr abundance (Fig. 22). Ni is preferentially incorporated into olivine to a greater extent than any other common basaltic phase (Hart and Davis, 1978). The nearly uniform Ni abundances and the scarcity of olivine phenocrysts in Laupahoehoe lavas indicate that olivine is no longer a primary crystallizing phase and was probably not significant in controlling the compositional variation of these lavas. The extremely low Ni concentrations of the lavas is strong evidence for substantial olivine fractionation at some point in their evolution. The common occurrence of dunite xenoliths of cumulate texture in south rift cinder cones and lavas supports the role of olivine. This is tentative, however, since the relationship of the xenoliths to the lavas has not been determined. The observed groundmass olivine is consistant with experimental evidence (Thompson, 1974) of a hawaiite from Skye which has olivine appearing with plagioclase at lower temperatures and pressures as a late crystallizing phase.

Apatite, Alkali Feldspar, Mica and Amphibole

Alkali feldspar and amphibole have not been identified in Mauna Kea lavas. Apatite is present as a minor groundmass phase in some lavas. Micas are present in some south rift lavas as small crystals distributed along fractures (tension gashes) and are possibly the result of volatile streaming during the final stages of crystallization.

Major and trace element variation of elements strongly incorporated into apatite (P_2O_5), alkili feldspar (K_2O), mica (K_2O, Rb) and amphibole (Rb, Zr) further suggests that these phases were not significantly involved in the fractionation of Laupahoehoe Group south rift lavas. The incompatible behavior of P_2O_5 (Fig. 14) and K_2O (Fig. 13) supports the proposition that apatite and alkali feldspar were not crystallizing phases. Rb is preferentially incorporated into micas (D = 3, Philpotts and Schnetzler, 1970; D = 1.9, Villemant et al., 1981). The incompatible behavior of Rb (Fig. 28) supports the observation that micas are not present as phenocrysts in Laupahoehoe south rift lavas and strongly suggests mica fractionation did not occur, Amphibole is also not likely to be a significant fractionating phase since ther is no corresponding decrease in Rb of Zr ($D_{Zr}^{Amph} = 1.2$, Villemant et al ., 1981). Sc depletion might be explained by amphibole as well clinopyroxene fractionation, but previously analyzed amphiboles from Kohala (Bence, 1981) have Al_2O_3 contents too high relative to CaO to completely account for the observed Al_2O_3/CaO increase in south rift lavas (assuming that Kohala amphiboles would be representative of possible fractionating amphiboles from Mauna Kea).

Selective depletion of volatile elements (Rb, K) by leaching via a migrating fluid or gas phase may have caused the anomalously low Rb and K_2O values in sample Mk-5. The presence of mica in tension gashes in Mauna Kea lavas supports this interpretation.

Summary of Evidence for Fractional Crystallization

The low mg numbers of Laupahoehoe south rift lavas suggest that they probably do not represent primary mantle melts and have undergone significant fractionation. Petrographic evidence and the systematic decrease in CaO, Sr, TiO_2, total iron and TiO_2/Al_2O_3 suggest that plagioclase and Ti-rich magnetite fractionation have occurred. This fractionation is suggested to have occured at low pressures and is supported by experimental studies of a hawaiite composition (Thompson, 1974) and the trend of some south rift lavas along the 1 atm., cotectic on the Ne-Pl-Ol-Si diagram (Fig. 7). Clinopyroxene fractionation is supported by the decrease in Sc, CaO, Al_2O_3/CaO and Zr/Y. Assuming that south rift magmas traveled directly to the surface from the upper mantle (based on the occurrence of presumed mantle xenoliths), the lack of clinopyroxene phenocrysts in the lavas may be due to dynamic flow crystallization at mantle pressures. High pressure clinopyroxene fractionation is supported by experimental studies of a hawaiite from Skye (Thompson, 1974) and an alkalic olivine basalt from Japan (Takahashi, 1980). The very rare microphenocrysts of olivine indicate thate olivine fractionation was no longer significant in controlling compositional variations in south rift lavas. The low Ni abundances and mg numbers of these lavas, however, suggest they have undergone substantial olivine fractionation at some point prior to eruption, probably at high pressures. The lack of apatite, alkali feldspar, mica and amphibole phenocrysts suggests that they probably were not significant fractionating phases. The incompatible behavior of P_2O_5, K_2O, Rb and Zr supports this.

Crystal Fractionation and Partial Melting

Presnall et al. (1979) proposed that the general trend of East Moloka'i lavas (Beeson, 1976) away from the silica apex with decreasing age on the Pl-Ol-Si ternary diagram (Presnall, unpub.) indicates the lavas were generated by successively smaller amounts of partial melting. Laupahoehoe south rift lavas generally plot along the one atmosphere cotectic, following a trend similar to that exhibited by East Moloka'i lavas (Fig. 7). However, south rift lavas show no temporal correlation with position along this trend. Successively smaller amounts of partial melting would result in magmas with successively lower mg numbers and silica and higher incompatible element abundances. This is not the case for south rift lavas. There is no systematic increase in incompatible elements (Zr, Rb, P_2O_5, K_2O) with time (Figs. 26-30), nor is there any observed temporal decrease in silica or MgO (Figs. 26, 27). Thus, it does not appear that successively smaller amounts of partial melting is responsible for generating the chemical variation observed in the Laupahoehoe Group.

South rift lavas plotted on the Ne-Pl-Ol-Si diagram (Fig. 6) form a trend which crosses from the critically undersaturated field into the silica oversaturated field. The thermal divide cannot be crossed through equilibrium crystallization at low pressures except under unusual conditions. If significant oxidation of iron occurs, the barrier may be crossed (Yoder and Tilley, 1962), but Fe_2O_3 concentrations of south rift lavas (Table 2) do not support such a high oxidation state. It is also unlikely that oxidation resulting in the observed magnetite fractionation could drive compositions from the critically undersaturated field all the way past the plane of silica saturation. If this trend is the result of the periodic tapping of a single magma body at high pressures, then compositions along the trend should exhibit incompatible element enrichment and compatible element depletion, typical of crystal fractionation processes. However, as pointed out, there is no correlation of position along this trend with mg number or incompatible element abundance. Thus, it does not appear that the trend is the result of the continuous fractionation of a single parent magma at either low or high pressures.

It is possible that the trend represents batches of compositionally similar magmas which reached the one atmosphere cotectic primarily through crystal fractionation of olivine and clinopyroxene. If this is the case, then some Laupahoehoe Group parental magmas may have been silica undersaturated and others critically undersaturated with respect to silica. This would imply that there was some variation in the amount of partial melting but that the variation was not systematic. This variation in the amount of partial melting is supported by the positive linear trend on the Zr/Nb vs. Zr (Fig. 31). This trend may be the result of batch partial melting processes (Minster and Allegre, 1978). The increase in Zr/Nb may also be the result of magnetite fractionation since Nb is preferentially incorporated relative to Zr (Pearce and Norry, 1979). In any case, Zr/Nb ratios do not change progressively with age, thus implying that partial melting variation was not systematic.

Compositional Variation vs Time

There is no systematic compositional variation with time exhibited by Laupahoehoe Group south rift lavas (Figs. 26-30). If the lavas were erupted from a continuously fractionating magma chamber, then compositions should follow a systematic chemical variation trend (i.e. an increase in incompatible elements and a decrease in compatible elements). It hte magma chamber approximated an open-system (i.e. was being periodically replenished), then there would most likely be a substantial increase in incompatible element abundance with time per refilling event. The lack of any systematic increase in incompatible elements or decrease in compatible elements with time suggests that these lavas are not directly related to each other by crystal fractionation of a single parent magma (open- or closed system). This is supported by the lack of systematic variation of silica saturation with time.

Despite the lack of coherent temporal variation, the overall chemical diversity of south rift lavas does show systematic enrichment and depletion trends which appear to correspond to crystal fractionation effects. As previously proposed, this may be the result of batches of compositionally similar magmas from a similar or common source. These batches, having undergone similar fractionation histories, would produce a compositional trend characteristic of crystal fractionation, but without temporal correlation. This appears to be the case for Laupahoehoe Group south rift lavas.

The Loaloan Rift

Eruptive activity within individual Laupahoehoe Group stages along the south rift is generally not confined to well-defined rifts or linear fractures. This may, in part, be due to vents along such rifts being covered by subsequent eruptions or glacial debris. The most recent volcanic activity on Mauna Kea is an exception. The majority of vents of Loaloan Stage age are concentrated along a rift roughly trending N 27° E (Fig. 32). The order of eruptions along this rift has been previously noted by Porter (1973). The vents become progressively younger from northeast to southwest. This suggests that the rift propagated in this direction (numbers on Fig. 33 refer to the order of eruption).

There is no systematic compositional variation along the Loaloan Stage rift (Figs. 33-37), except for Sr (Fig. 37). Sr content progressively increases with decreasing age along the rift. If this trend is real, it may be the result of plagioclase flotation or resorption, or it may possibly reflect the progressively decreasing role of plagioclase fractionation with time as the rift propagated. The lack of corresponding incompatible element enrichment or CaO variation, however, makes interpretation of the observed Sr enrichment trend problematic.

Compositional Variation vs Distance From Summit

A concentric zonation of Honolulu Volcanic Series lava type (and xenolith type) outward from the Ko'olau caldera on Oahu has been proposed by Jackson and Wright (1970). Although such a zonation has now been discounted for the Honolulu Volcanic Series (Clague & Frey, 1982), the possibility of it occurring on other Hawaiian volcanoes has not been examined. Jackson (1968) and Jackson and Wright (1970) suggested that suck a zonation may reflect horizontal differences in mantle composition away from the location of the original tholeiitic magma column. However, lava type does not vary systematically on Mauna Kea from the summit to the saddle along the south rift (a distance of nearly 14 km.). Vitrually all of the lavas are hawaiites with rare mugearites. The lack of any systematic chemical variation with respect to summitt proximity (Figs. 38-42) suggests that south rift lava compositions are not controlled by horizontal mantle zonation. However, the south rift is only a part of the overall extent of the volcano and may represent a single, relatively homogeneous zone.

Vertical stratification of pooled or ascending magmas due to density differences has been proposed by Daly (1911b), Macdonald (1949b) and Sparks _et al_. (1980). The lack of systematic chemical variation in south rift lavas outward from the summit and, thus, with decreasing eleveation (from 4205 to 1951 meters elevation), suggests that such stratification was not a controlling factor with respect to the evolution of the Laupahoehoe Group on the upper slopes of Mauna Kea. Although density differential may not control the variation within compositionally similar Laupahoehoe south rift lavas, it may be important in controlling the distribution of Laupahoehoe Group eruptive vents. Laupahoehoe Group (hawaiite) vents are more densely concentrated on the upper slopes of Mauna Kea. These hawaiites may have been able to ascend to the upper section of the volcano more readily than denser, olivine-phyric Hamakua lavas and, thus, these vents may be preferentially concentrated on the upper slopes. However, very little is known of the distribution of Hamakua vents (and their rock types) on the upper slopes of the mountain due to the covering of alkalic cap lavas. Thus, sny conclusion as to the cause of Laupahoehoe versus Hamakua Group vent distribution on Mauna Kea is purely speculative.

Relationship between the Laupahoehoe and Hamakua Groups

Macdonald (1949b, 1968) suggested that the transition between the tholeiitic and alkalic sequences on Mauna Kea occurred gradually with no significant period of volcanic inactivity separating them. On the flanks of the mountain, Hamakua Group lavas are generally covered by 2 m or more of soil. Laupahoehoe Group lavas generally have soil covers 1.5 m or less. This suggests that the hiatus in volcanic activity was probably not great. Nevertheless, the tholeiite-alkalic basalt transition was compositionally abrupt (i.e. there is no compositional gradation between the two groups). A compositional gap separating the Hamakua and Laupahoehoe groups is evident in major and trace element plots and in plots of normative compositions. This gap is inconsistant with compositional variations resulting solely from crystal fractionation. Also, the gap on diagrams of normative compositions cannot be explained by successively decreasing the amount of partial melting of a common or similar source.

The compositional gap between the Laupahoehoe and Hamakua groups cannot be satisfactorily explained simply by a hiatus in eruptive activity during which crystal fractionation continued. The two groups form separate trends on ternary diagrams representing faces of the basalt tetrahedron (Figs. 7, 43). Hopukani Formation lavas collected from the Hamakua Coast by Garcia and Wise (unpub.) and samples collected by Macdonald (1968; Macdonald and Katsura, 1964) do not closely follow the one atmosphere cotectic as do Laupahoehoe south rift lavas. To minimize the effects of olivine accumulation, only lavas with MgO contents under 7 wt. % were plotted on the Ne-Fo-Ab-Si diagram (Fig. 43). Separate trends are still evident suggesting the dichotomy may be real.

The distinct trends shown by south rift normative compositions are well illustrated on the D.I. versus normative nepheline (Fig. 44) and D.I. versus normative hypersthene (Fig. 45) diagrams. The Laupahoehoe and Hamakua groups form two virtually parallel trends that are separated by about a value of 10 on the D.I. scale. Crystal fractionation alone cannot account for the two separate, parallel trends. Nepheline normative Laupahoehoe south rift lavas are distinct from nepheline normative Hamakua lavas on these two normative composition diagrams. Derivation of south rift lavas by crystal fractionation from a Hamakua parent composition would demand fractionation of phases resulting in a decrease in normative olivine without an accompanying change in normative nepheline. Subtraction of olivine from a critically undersaturated parent magma would result in an increase in normative nepheline. However, nepheline normative Laupahoehoe and Hamakua lavas have similar values. This suggests that the two groups are not related by crystal fractionation. Smaller amounts of partial melting of a common or similar source is also inconsistant with the observed parallel trends since such a decrease would also cause initial melt compositions to be more nepheline normative. Thus, these trends suggest that the two groups represent two separate lineages derived from different sources.

The separation of the Laupahoehoe and Hamakua groups is also illustrated by the "kink" which divides the two groups on the D.I. versus An-content diagram (Fig. 46). Groups of co-magmatic rocks, related by crystal fractionation, from Gough Island, Tristan da Cunha, the Hebridean and East Otago, plot on nearly linear trends which are slightly displaced, but nearly parallel to each other (Coombs and Wilkinson, 1969). There is no observable kink in any of these trends. If the Laupahoehoe and Hamakua groups are related, this kink indicates that D.I. abruptly increases without an accompanying increase in An-content. This is inconsistant since, by definition, if D.I. increases, An-content must decrease since both quantities reflect the trend of liquids towards petrogeny's residua system upon crystal fractionation. Neither D.I. nor An-content would be affected by an increase in silica saturation for quartz and hypersthene normative compositions. The primary effect of extensive magnetite fractionation would be to increase silica saturation. Thus, magnetite fractionation is also not likely to explain the observed kink.

The uppermost lava of the Hopukani Formation (H-6) has an mg number of about 51, not much different than the most primitive south rift Laupahoehoe lava (48.5) despite significant differences in Ni abundances. This difference in mg number of about 2.5 is inconsistant with the observed difference in Ni of 70 ppm if the two are related by crystal fractionation. For contrast, in Hualalai alkalic olivine basalts in which olivine is a primary fractionating phase (Clague *et al.*, 1981b), a corresponding mg number decrease of 2.5 is accompanied by a Ni decrease of only 22 ppm. If H-6 is indeed representative of the most differentiated Hamakua Group composition (with respect to mg number and Ni content) and the most primitive Laupahoehoe Group composition is approximately that suggested, then the observed similarity in mg number while Ni abundances contrast markedly may suggest that Hamakua Group lavas and Laupahoehoe south rift lavas are not related by crystal fractionation.

From the above arguments, it is tentatively suggested that the Laupahoehoe and Hamakua groups represent independent magmas generated from distinct mantle sources. Unfortunately incompatible element ratios available for the south rift lavas (Zr/Nb, K/Rb, Zr/Y, Zr/Rb, K_2O/P_2O_5) may not be suitable for testing this hypothesis. Incompatible element ratios (e.g. Zr/Nb, Zr/Ti, Zr/Y) have been useful in constraining mantle source compositions of lavas approximating primary magma compositions (Erlank and Kable, 1976; Flower *et al.*, 1977; Pearce and Norry, 1979). However, south rift lavas probably do not represent primary magmas and certain elements may have been affected by fractionation of phases such as magnetite (Nb) and clinopyroxene (Y). Other elements (i.e. K, Rb) may have been subject to mobilization and contamination (Frey, personal communication).

SUMMARY

(1) The presence of inferred mantle-derived xenoliths suggests that Laupahoehoe Group south rift lavas originated in the mantle. However, mg numbers indicate that these lavas are probably not primary mantle melts but have undergone crystal fractionation. The aphyric nature of south rift lavas implies that the fractionating minerals were removed from the melt but the xenoliths were retained. A dynamic flow crystallization hypothesis is a method which might account for the simultaneous separation of fractionating mineral phases and retention of mantle xenoliths.

(2) Petrographic evidence and chemical variation within south rift lavas is consistant with low pressure crystal fractionation involving plagioclase and Ti-rich magnetite. The absence of clinopyroxene phenocrysts despite geochemical evidence supports a dynamic flow crystallization model occurring at mantle pressures. High pressure clinopyroxene fractionation is supported by experimental studies. Previous olivine fractionation is suggested by the low Ni abundances and mg numbers and probably occurred at high pressures.

(3) Successively decreasing amounts of partial melting of a common source cannot explain the observed chemical variation within south rift lavas. Some variation in

the amount of partial melting may have occurred, but this variation was not systematic with respect to time.

(4) The lack of systematic chemical variation with time suggests that south rift lavas were erupted as compositionally similar magma batches and are not the result of the periodic tapping of a single magma body (open- or closed-system).

(5) Compositions do not vary systematically with respect to summit proximity or elevation. Thus, any density and horizontal zonation present in the underlying mantle probably did not control the chemical evolution of Laupahoehoe Group south rift lavas.

(6) The compositional "gap" separating the Laupahoehoe and Hamakua groups cannot be explained by crystal fractionation or successively decreasing amounts of partial melting of a common source. However, insufficient trace element and isotopic data make any further evaluation of their relationship problematic.

TABLE 1

LOALOAN STAGE

	1	2	3	4	5	6	7
	M1-2	M1-3	M1-4	M1-5	M1-6	M1-8	M1-11
SiO_2	50.90	49.30	50.30	50.10	50.40	51.60	51.45
Al_2O_3	17.34	17.06	17.24	17.44	17.49	17.40	17.14
Fe_2O_3	3.34	5.31	2.92	3.24	5.62	9.81	2.92
FeO	7.40	5.48	7.72	7.56	5.16	0.88	7.10
MgO	2.99	4.42	4.20	4.16	3.99	3.42	3.59
CaO	6.81	7.30	6.67	6.69	6.61	6.01	6.31
Na_2O	4.78	5.35	4.70	4.68	4.73	4.80	5.08
K_2O	1.95	1.92	1.91	1.90	1.93	2.08	2.22
H_2O^+	0.30	0.21	0.28	0.30	0.23	0.53	0.40
H_2O^-	0.15	0.10	0.11	0.16	0.08	0.26	0.20
CO_2	0.03	0.09	0.06	0.03	0.08	0.06	0.10
TiO_2	2.62	2.52	2.70	2.64	2.67	2.12	2.24
P_2O_5	0.889	0.918	0.865	0.879	0.879	0.973	1.048
MnO	0.219	0.223	0.220	0.224	0.219	0.223	0.234
TOTAL	99.72	100.20	99.90	100.00	100.09	99.80	100.03

TABLE 1 (CONT.)

	8	9	10	11	12	13
	M1-13	M1-14	M1-15	M1-18	M1-19	M1-20
SiO_2	51.80	50.15	50.45	50.55	50.45	53.05
Al_2O_3	16.96	16.74	17.01	17.13	17.09	16.24
Fe_2O_3	5.68	5.09	3.62	4.51	4.24	4.63
FeO	4.36	5.92	6.97	5.85	6.44	5.88
MgO	3.45	3.97	4.11	4.12	4.15	3.53
CaO	6.45	6.66	6.81	6.83	6.68	6.14
Na_2O	5.03	4.93	4.84	4.85	4.63	4.04
K_2O	2.19	1.88	2.03	1.93	1.92	2.04
H_2O^+	0.38	0.38	0.24	0.37	0.32	0.27
H_2O^-	0.13	0.19	0.09	0.14	0.10	0.14
CO_2	0.04	0.11	0.03	0.03	0.09	0.11
TiO_2	2.28	2.66	2.71	2.65	2.62	2.38
P_2O_5	1.074	0.903	0.909	0.897	0.893	0.97
MnO	0.233	0.220	0.224	0.225	0.221	0.215
TOTAL	100.05	99.80	100.04	100.09	99.84	99.64

TABLE 1 (CONT.)

	KEMOLEAN STAGE					
	14	15	16	17	18	19
	Mk-1	Mk-2	Mk-3	Mk-4	Mk-5	Mk-6
SiO_2	50.50	51.10	51.05	51.20	50.15	52.05
Al_2O_3	16.53	17.04	17.14	16.99	16.96	17.54
Fe_2O_3	3.90	4.26	4.91	3.53	4.78	3.14
FeO	7.48	5.84	5.16	6.64	6.56	6.36
MgO	4.17	3.90	3.79	3.86	3.68	3.27
CaO	7.08	6.72	6.50	6.56	6.66	5.92
Na_2O	3.99	4.95	4.93	4.90	3.94	4.95
K_2O	1.76	1.98	2.04	1.98	1.77	2.05
H_2O^+	0.20	0.36	0.37	0.32	0.53	0.70
H_2O^-	0.19	0.08	0.09	0.18	0.69	0.27
CO_2	0.10	0.02	0.06	0.06	0.09	0.05
TiO_2	2.84	2.56	2.44	2.49	2.69	2.30
P_2O_5	0.88	0.913	0.932	0.944	1.10	1.035
MnO	0.211	0.214	0.211	0.214	0.210	0.219
TOTAL	99.83	99.94	99.62	99.86	99.81	99.86

TABLE 1 (CONT.)

KUUPAHAAN STAGE

	20 Me-1	21 Me-2	22 Me-3	23 Me-4	24 Me-5	25 Me-6
SiO_2	50.50	50.30	51.25	51.95	51.70	50.55
Al_2O_3	17.22	16.84	17.36	17.34	17.37	16.77
Fe_2O_3	3.81	3.37	3.23	3.32	3.71	3.99
FeO	6.68	7.32	6.80	6.62	6.02	7.24
MgO	4.07	4.13	3.74	3.47	3.33	4.17
CaO	6.69	7.06	6.49	6.38	6.29	6.79
Na_2O	4.37	4.48	4.88	4.95	5.15	3.93
K_2O	1.94	1.87	1.99	2.01	2.05	1.81
H_2O^+	0.40	0.12	0.15	0.30	0.23	0.40
H_2O^-	0.18	0.11	0.06	0.17	0.08	0.17
CO_2	0.04	0.09	0.03	0.07	0.04	0.07
TiO_2	3.02	2.88	2.66	2.33	2.40	2.80
P_2O_5	0.830	0.87	0.896	0.944	1.090	0.85
MnO	0.211	0.206	0.218	0.227	0.225	0.215
TOTAL	99.96	99.65	99.76	100.07	99.69	99.76

TABLE 1 (CONT.)

	HANAIPOEAN STAGE				POLIAHUAN STAGE	
	26	27	28	29	30	31
	Mh-1	Mh-2	Mh-3	Mh-4	Mp-1	Mp-2
SiO_2	52.65	51.60	52.10	51.65	51.35	49.35
Al_2O_3	16.79	16.85	17.31	17.16	16.36	16.90
Fe_2O_3	3.96	4.29	2.92	4.21	4.66	3.08
FeO	6.00	6.16	6.84	6.28	6.04	7.76
MgO	3.40	3.53	3.47	3.27	3.70	4.69
CaO	6.38	6.57	6.06	6.02	6.38	7.27
Na_2O	4.44	4.18	4.97	5.01	3.82	4.83
K_2O	2.12	2.07	2.10	2.12	2.12	1.89
H_2O^+	0.08	0.30	0.43	0.28	0.78	0.10
H_2O^-	0.09	0.22	0.10	0.10	0.72	0.05
CO_2	0.08	0.07	0.06	0.03	0.13	0.02
TiO_2	2.39	2.42	2.26	2.19	2.42	2.82
P_2O_5	1.13	1.33	0.885	1.275	1.09	0.859
MnO	0.218	0.217	0.211	0.225	0.219	0.218
TOTAL	99.73	99.81	99.72	99.83	99.79	99.84

TABLE 1 (CONT.)

	LILOEAN STAGE					HOPUKANI FORM.
	32	33	34	35	36	37
	Mi-1	Mi-2	Mi-3	Mi-4	Mi-5	H-6
SiO_2	52.55	50.65	51.85	52.00	49.90	47.35
Al_2O_3	17.58	17.22	17.36	17.78	16.86	14.10
Fe_2O_3	3.73	4.56	6.93	3.05	3.22	3.35
FeO	5.64	6.16	2.96	6.40	7.34	9.40
MgO	2.98	3.91	2.92	3.06	4.59	6.15
CaO	6.00	6.54	6.03	6.02	7.09	11.04
Na_2O	5.23	4.48	5.48	4.93	4.58	3.14
K_2O	2.18	1.90	2.08	2.05	1.86	0.96
H_2O^+	0.31	0.44	0.54	0.75	0.32	0.14
H_2O^-	0.09	0.18	0.19	0.29	0.08	0.03
CO_2	0.01	0.09	0.20	0.06	0.06	0.03
TiO_2	2.02	2.76	2.02	2.00	3.01	3.58
P_2O_5	1.196	0.909	1.210	1.186	0.798	0.505
MnO	0.228	0.211	0.227	0.227	0.216	0.185
TOTAL	99.75	100.01	100.00	99.81	99.93	99.97

TABLE 2

LOALOAN STAGE

		1	2	3	4	5	6	7
		M1-2	M1-3	M1-4	M1-5	M1-6	M1-8	M1-11
SiO_2		51.37	49.58	50.64	50.42	50.75	52.58	51.87
Al_2O_3		17.50	17.16	17.36	17.55	17.61	17.36	17.28
Fe_2O_3		1.75	1.72	1.74	1.76	1.72	1.65	1.63
FeO		8.93	8.77	8.85	8.96	8.74	8.41	8.33
MgO		3.02	4.44	4.23	4.19	4.02	3.49	3.62
CaO		6.87	7.34	6.71	6.73	6.66	6.12	6.36
Na_2O		4.82	5.38	4.73	4.71	4.76	4.89	5.12
K_2O		1.97	1.93	1.92	1.91	1.94	2.12	2.24
TiO_2		2.64	2.53	2.72	2.66	2.69	2.16	2.26
P_2O_5		0.90	0.93	0.88	0.89	0.89	0.99	1.06
MnO		0.22	0.22	0.22	0.22	0.22	0.22	0.23
TOTAL		99.99	100.00	100.00	100.00	100.00	99.99	100.00
Sc		13	13	13	12	13	10	11
Ni		2	0	4	3	2	3	0
Cu		5	5	7	8	3	5	0
Zn		115	123	125	121	118	129	131
Rb		35	35	34	32	37	39	40
Sr		1347	1327	1347	1305	1299	1355	1316
Y		44	47	48	46	48	50	49
Zr		436	431	439	463	437	499	498
Nb		-	-	59	58	60	68	67
CIPW NORMS								
Or		11.6	11.4	11.4	11.3	11.5	12.6	13.2
Ab		40.2	31.2	37.9	37.3	38.4	41.4	40.1
An		20.3	17.0	20.0	21.1	20.9	19.2	17.5
Ne		0.4	7.8	1.2	1.4	1.0		1.8
Di	Wo	3.3	5.6	3.0	2.7	2.6	2.0	3.0
	En	1.4	2.8	1.5	1.3	1.3	2.8	1.3
	Fs	2.0	2.7	1.5	1.4	1.3	3.5	1.6
Hy	En						1.9	
	Fs						2.4	
Ol	Fo	4.3	5.8	6.4	6.4	6.1	4.1	5.3
	Fa	7.0	6.3	7.2	7.5	7.2	5.7	7.0
Mt		2.5	2.5	2.5	2.5	2.5	2.4	2.4
Il		5.0	4.8	5.2	5.0	5.1	4.1	4.3
Ap		2.1	2.2	2.1	2.1	2.1	2.3	2.5

TABLE 2 (CONT.)

		8	9	10	11	12	13
		M1-13	M1-14	M1-15	M1-18	M1-19	M1-20
SiO_2		52.28	50.77	50.71	50.93	50.92	53.68
Al_2O_3		17.12	16.95	17.10	17.26	17.25	16.43
Fe_2O_3		1.59	1.77	1.71	1.66	1.73	1.69
FeO		8.12	9.03	8.74	8.48	8.80	8.64
MgO		3.48	4.02	4.13	4.15	4.19	3.57
CaO		6.51	6.74	6.85	6.88	6.74	6.21
Na_2O		5.08	4.99	4.86	4.89	4.67	4.09
K_2O		2.21	1.90	2.04	1.94	1.94	2.06
TiO_2		2.30	2.69	2.72	2.67	2.64	2.41
P_2O_5		1.08	0.91	0.91	0.91	0.90	0.98
MnO		0.23	0.22	0.22	0.23	0.22	0.22
TOTAL		100.00	99.99	99.99	100.00	100.00	99.98
Sc		9	13	12	13	12	12
Ni		5	2	2	2	2	4
Cu		4	3	3	7	8	5
Zn		120	127	120	125	120	121
Rb		40	31	35	33	35	31
Sr		1357	1251	1334	1273	1306	1257
Y		46	42	48	43	46	44
Zr		492	438	430	456	435	442
Nb		70	56	53	54	-	62
CIPW NORMS							
Q							2.0
Or		13.1	11.2	12.1	11.5	11.5	12.2
Ab		41.5	38.0	36.9	37.9	38.7	34.6
An		17.4	18.2	18.8	19.4	20.4	20.4
Ne		0.8	2.3	2.3	1.9	0.5	
Di	Wo	3.3	3.9	3.8	3.7	3.0	1.7
	En	1.5	1.8	1.9	1.8	1.5	8.9
	Fs	1.8	2.0	1.9	1.8	1.5	10.9
Hy	En						8.1
	Fs						10.0
Ol	Fo	5.0	5.7	5.9	6.0	6.3	
	Fa	6.5	7.0	6.7	6.5	7.2	
Mt		2.3	2.6	2.5	2.4	2.5	2.5
Il		4.4	5.1	5.2	5.1	5.0	4.6
Ap		2.6	2.2	2.2	2.1	2.1	2.3

TABLE 2 (CONT.)

	KEMOLEAN STAGE					
	14	15	16	17	18	19
	Mk-1	Mk-2	Mk-3	Mk-4	Mk-5	Mk-6
SiO_2	50.94	51.51	51.69	51.66	51.07	52.75
Al_2O_3	16.67	17.18	17.35	17.14	17.27	17.77
Fe_2O_3	1.85	1.63	1.62	1.65	1.84	1.55
FeO	9.42	8.29	8.24	8.42	9.40	7.91
MgO	4.21	3.93	3.84	3.89	3.75	3.31
CaO	7.14	6.77	6.58	6.62	6.78	6.00
Na_2O	4.02	4.99	4.99	4.94	4.01	5.02
K_2O	1.78	2.00	2.07	2.00	1.80	2.08
TiO_2	2.86	2.58	2.47	2.51	2.74	2.33
P_2O_5	0.89	0.92	0.94	0.95	1.12	1.05
MnO	0.21	0.21	0.21	0.21	0.21	0.22
TOTAL	99.99	100.01	100.00	99.99	99.99	99.99

Sc	13	12	11	12	12	10
Ni	1	5	6	6	4	7
Cu	0	1	3	1	4	3
Zn	124	116	115	119	132	150
Rb	30	36	39	36	22	37
Sr	1347	1323	1309	1321	1376	1258
Y	44	48	47	49	53	53
Zr	435	459	464	472	508	566
Nb	64	59	-	62	67	73

CIPW NORMS

Or		10.5	11.8	12.2	11.8	10.7	12.4
Ab		34.1	39.4	40.0	40.4	33.9	42.4
An		22.2	18.6	18.8	18.7	23.8	19.8
Ne			1.6	1.2	0.8		
Di	Wo	3.1	3.8	3.2	3.3	1.1	1.3
	En	6.9	1.8	1.5	1.6	1.8	3.0
	Fs	7.6	1.9	1.6	1.7	9.8	3.5
Hy	En	5.5				7.4	2.4
	Fs	6.0				9.2	2.8
Ol	Fo	2.5	5.6	5.6	5.7	1.0	3.7
	Fa	3.0	6.3	6.6	6.7	1.4	4.9
Mt		2.7	2.4	2.3	2.4	2.7	2.3
Il		5.4	4.9	4.7	4.8	5.2	4.4
Ap		2.1	2.2	2.2	2.2	2.7	2.5

TABLE 2 (CONT.)

KUUPAHAAN STAGE

	20	21	22	23	24	25
	Me-1	Me-2	Me-3	Me-4	Me-5	Me-6
SiO_2	50.94	50.72	51.58	52.28	52.16	51.11
Al_2O_3	17.37	16.98	17.47	17.45	17.52	16.96
Fe_2O_3	1.70	1.74	1.63	1.61	1.57	1.83
FeO	8.67	8.87	8.30	8.22	8.02	9.31
MgO	4.11	4.16	3.76	3.49	3.36	4.22
CaO	6.75	7.12	6.53	6.42	6.35	6.87
Na_2O	4.41	4.52	4.91	4.98	5.20	3.97
K_2O	1.96	1.89	2.00	2.02	2.07	1.83
TiO_2	3.05	2.90	2.68	2.34	2.42	2.83
P_2O_5	0.84	0.88	0.91	0.95	1.10	0.86
MnO	0.21	0.21	0.22	0.23	0.23	0.22
TOTAL	100.01	99.99	99.99	99.99	100.00	100.01
Sc	12	14	12	10	11	12
Ni	2	7	5	3	6	2
Cu	5	9	4	2	5	1
Zn	120	105	117	125	121	119
Rb	33	26	34	35	39	30
Sr	1293	1326	1292	1299	1285	1340
Y	44	46	46	48	52	47
Zr	448	410	452	503	522	409
Nb	60	55	61	66	66	61

CIPW NORMS

Or		11.6	11.1	11.8	12.0	12.2	10.8
Ab		37.3	38.1	40.6	42.2	42.7	33.6
An		21.8	20.5	19.7	19.3	18.4	23.0
Ne			0.1	0.5		0.7	
Di	Wo	2.6	3.8	2.8	2.7	2.5	2.3
	En	3.4	1.9	1.4	1.6	1.1	7.6
	Fs	3.2	1.9	1.4	1.9	1.3	8.2
Hy	En	2.1			0.4		6.6
	Fs	2.0			0.4		7.1
Ol	Fo	4.8	6.0	5.6	5.0	5.1	2.0
	Fa	5.1	6.6	6.5	6.5	6.6	2.4
Mt		2.5	2.5	2.4	2.3	2.3	2.6
Il		5.8	5.5	5.1	4.5	4.6	5.4
Ap		2.0	2.1	2.1	2.2	2.6	2.0

TABLE 2 (CONT.)

		HANAIPOEAN STAGE				POLIAHUAN STAGE	
		26	27	28	29	30	31
		Mh-1	Mh-2	Mh-3	Mh-4	Mp-1	Mp-2
SiO_2		53.05	52.14	52.63	52.08	52.47	49.58
Al_2O_3		16.92	17.03	17.49	17.30	16.72	16.98
Fe_2O_3		1.61	1.69	1.59	1.69	1.74	1.76
FeO		8.19	8.61	8.13	8.63	8.89	8.99
MgO		3.43	3.57	3.51	3.30	3.78	4.71
CaO		6.43	6.64	6.12	6.07	6.52	7.30
Na_2O		4.47	4.22	5.02	5.05	3.90	4.85
K_2O		2.14	2.09	2.12	2.14	2.17	1.90
TiO_2		2.41	2.45	2.28	2.21	2.47	2.83
P_2O_5		1.14	1.34	0.90	1.29	1.11	0.86
MnO		0.22	0.22	0.21	0.23	0.22	0.22
TOTAL		100.01	100.00	100.00	99.99	99.99	99.98
Sc		11	11	11	11	11	13
Ni		4	3	7	4	5	3
Cu		8	0	0	2	6	12
Zn		119	124	117	131	125	112
Rb		43	38	40	40	40	34
Sr		1293	1280	1274	1161	1263	1328
Y		51	55	47	53	51	48
Zr		524	513	508	605	508	428
Nb		66	66	66	-	69	53
CIPW NORMS							
Q						0.6	
Or		12.6	12.4	12.5	12.6	12.8	11.2
Ab		37.9	35.7	42.5	42.8	33.0	32.7
An		19.8	21.3	18.9	18.2	21.7	18.9
Ne							4.5
Di	Wo	2.0	1.2	2.3	1.4	1.4	4.9
	En	7.8	8.0	1.8	1.4	9.4	2.5
	Fs	9.3	9.8	2.1	2.0	11.2	2.3
Hy	En	6.9	7.5	0.7	0.8	8.8	
	Fs	8.2	9.1	0.8	1.1	10.5	
Ol	Fo	0.5	0.6	4.9	4.7		6.5
	Fa	0.7	0.8	6.3	7.1		6.6
Mt		2.3	2.4	2.3	2.5	2.5	2.6
Il		4.6	4.6	4.3	4.2	4.7	5.4
Ap		2.7	3.2	2.1	3.1	2.6	2.0

TABLE 2 (CONT.)

		LILOEAN STAGE					HOPUKANI FORM.
		32	33	34	35	36	37
		Mi-1	Mi-2	Mi-3	Mi-4	Mi-5	H-6
SiO_2		53.02	51.15	52.62	52.76	50.24	47.52
Al_2O_3		17.74	17.39	17.62	18.04	16.98	14.15
Fe_2O_3		1.51	1.73	1.56	1.55	1.72	2.08
FeO		7.71	8.81	7.93	7.89	8.76	10.59
MgO		3.01	3.95	2.96	3.10	4.62	6.17
CaO		6.05	6.61	6.12	6.11	7.14	11.08
Na_2O		5.28	4.52	5.56	5.00	4.61	3.15
K_2O		2.20	1.92	2.11	2.08	1.87	0.96
TiO_2		2.04	2.79	2.05	2.03	3.03	3.59
P_2O_5		1.21	0.92	1.23	1.21	0.81	0.51
MnO		0.23	0.21	0.23	0.23	0.22	0.19
TOTAL		100.00	100.00	99.99	100.00	100.00	99.99
Sc		10	13	11	11	13	27
Ni		8	8	6	4	8	78
Cu		5	2	8	2	3	71
Zn		130	118	132	133	119	106
Rb		41	31	36	40	29	14
Sr		1219	1388	1220	1226	1290	693
Y		54	46	54	53	43	30
Zr		576	426	603	584	381	224
Nb		-	60	71	70	58	32
CIPW NORMS							
Or		13.0	11.3	12.5	12.3	11.1	5.7
Ab		44.6	38.3	44.0	42.3	35.9	24.7
An		18.2	21.5	16.9	20.6	20.1	21.6
Ne				1.7		1.7	1.1
Di	Wo	1.6	2.2	2.3	0.7	4.2	12.5
	En	1.1	3.2	1.0	2.9	2.2	6.8
	Fs	1.5	3.4	1.3	3.9	1.9	5.3
Hy	En	0.4	2.1		2.6		
	Fs	0.6	2.3		3.4		
Ol	Fo	4.5	4.7	4.5	3.4	6.5	6.0
	Fa	6.5	5.5	6.9	5.0	6.3	5.3
Mt		2.2	2.5	2.3	2.2	2.5	3.0
Il		3.9	5.3	3.9	3.9	5.8	6.8
Ap		2.9	2.2	2.9	2.9	1.9	1.2

Figure 1: Map of Hawaiian Island chain with inset of the island of Hawaii. Small square on island of Hawaii represents studied area on Mauna Kea.

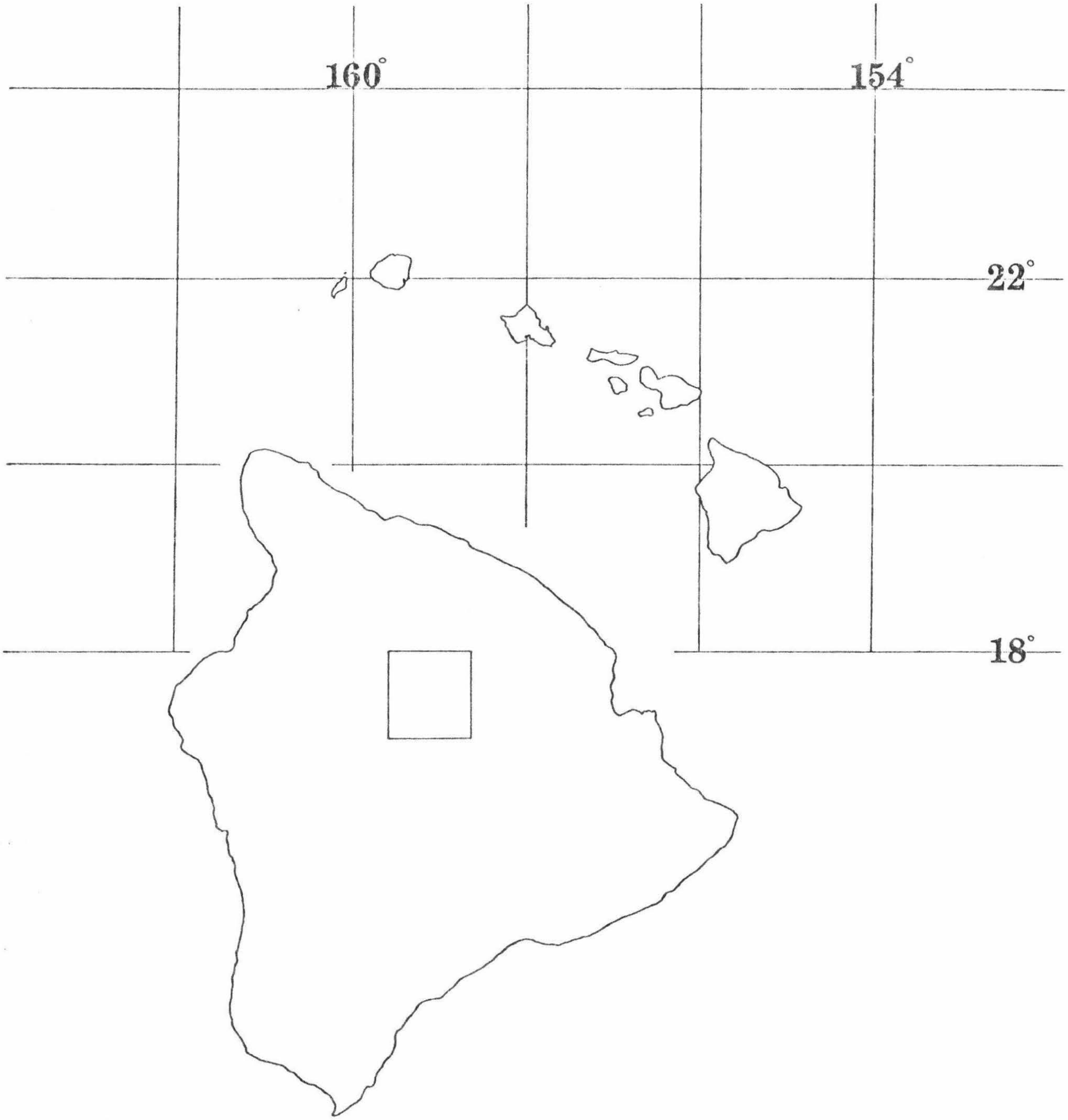

FIGURE 1

Figure 2: Normative Ne-Pl-Ol-Q diagram in mole %, with $Fe^{2+}/Fe^{2+} + Fe^{3+} = 0.89$. Heavy solid lines are isobaric univariant liquidus boundary lines along which forsterite, diopside and anorthite are in equilibrium with the liquid (Presnall et al., 1979). Boxes represent compositions of lavas from the East Molokai Volcano, Hawaii. Lavas are numbered in stratigraphic order, with #1 as the oldest lava (Beeson, 1976; Presnall, unpub.). (Figure courtesy of M. O. Garcia).

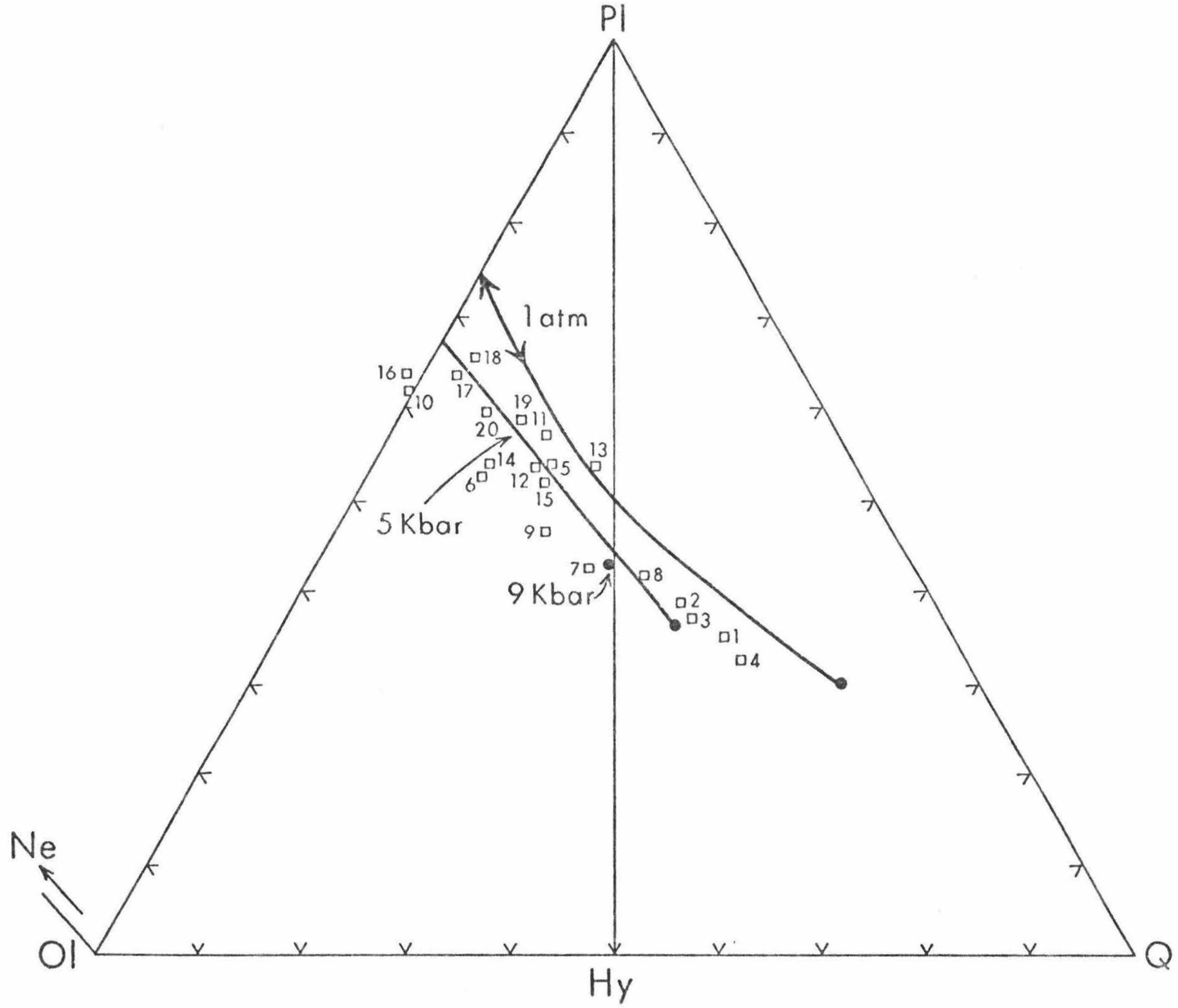
Pl
Ne
Ol
Hy
Q
1 atm
5 Kbar
9 Kbar
16
10
18
17
19
11
20
14
6
12
15
5
13
9
7
8
2
3
1
4

FIGURE 2

Figure 3: Time-stratigraphic and rock-stratigraphic units for Mauna Kea (Porter, 1979b).

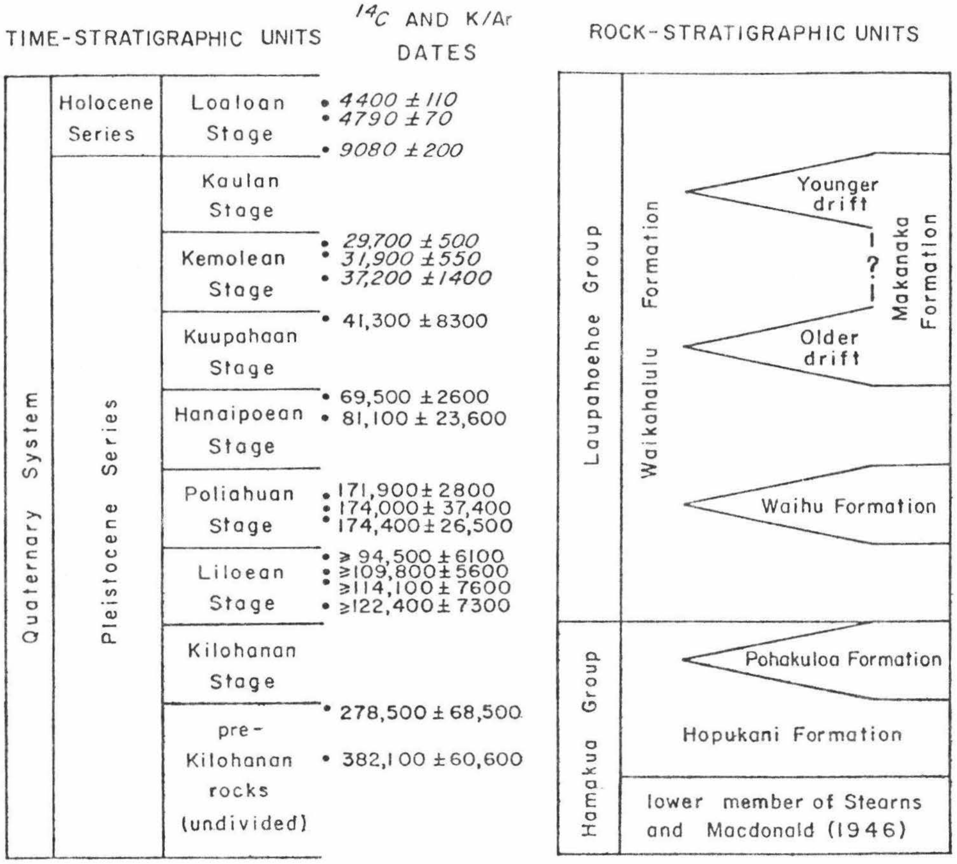
TIME-STRATIGRAPHIC UNITS
^{14}C AND K/Ar DATES
ROCK-STRATIGRAPHIC UNITS
Quaternary System
Holocene Series
Pleistocene Series
Loaloan Stage
Kaulan Stage
Kemolean Stage
Kuupahaan Stage
Hanaipoean Stage
Poliahuan Stage
Liloean Stage
Kilohanan Stage
pre-Kilohanan rocks (undivided)
4400 ± 110
4790 ± 70
9080 ± 200
29,700 ± 500
31,900 ± 550
37,200 ± 1400
41,300 ± 8300
69,500 ± 2600
81,100 ± 23,600
171,900 ± 2800
174,000 ± 37,400
174,400 ± 26,500
≥ 94,500 ± 6100
≥109,800 ± 5600
≥114,100 ± 7600
≥122,400 ± 7300
278,500 ± 68,500
382,100 ± 60,600
Laupahoehoe Group
Waikahalulu Formation
Younger drift
Makanaka Formation
?
Older drift
Waihu Formation
Hamakua Group
Pohakuloa Formation
Hopukani Formation
lower member of Stearns and Macdonald (1946)

FIGURE 3

Figure 4: Geologic map of the south rift area, upper slopes of Mauna Kea (Porter, 1979b).


Whl
Wml
Wul
Mkom
Wmc
Wll
Wpl
Wel
Wuc
Wlc
Wc
Woc
Wam
Wkl
Hol
Qub
Kpl
Kps
Le
Khlg
Humuula Saddle

FIGURE 4

Figure 4b: Sample location map for Laupahoehoe Group south rift lavas, Mauna Kea. Numbers refer to the order chemical analyses are presented on Tables 1 and 2.

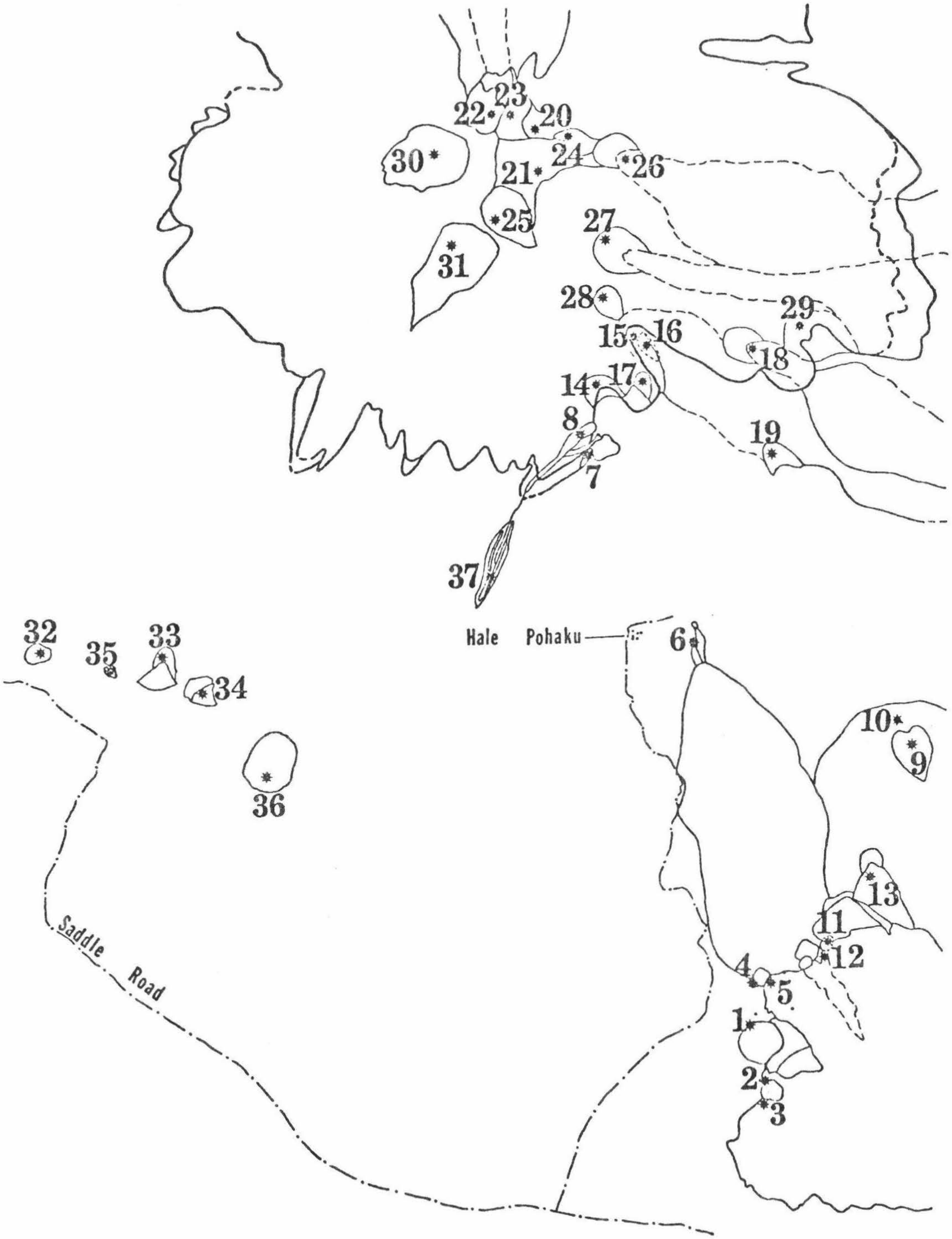
23
22
20
30
24
26
21
25
27
31
28
29
15
16
18
14
17
8
19
7
37
Hale Pohaku
6
32
35
33
34
10
9
36
13
11
12
4
5
1
2
3
Saddle Road

FIGURE 4b

Figure 5: Silica-Alkali diagram for Laupahoehoe and Hamakua group lavas. Numbers refer to stages within the Laupahoehoe Group: 1 = Loaloan Stage, 2 = Kemolean Stage, 3 = Kuupahaan Stage, 4 = Hanaipoean Stage, 5 = Poliahuan Stage, 6 = Liloean Stage. H = the uppermost lava flow of the Hopukani Formation, collected in Waikahalulu Gulch. Solid circles refer to Hamakua lavas collected from Laupahoehoe Gulch, 4-spoke solid circles to Hamakua lavas collected from Maulua Gulch, 8-spoke solid circles from Hamakua Group lavas collected by Macdonald (1968; Macdonald and Katsura, 1964), solid circle + arrow from Hamakua lavas from Ka'awali'i Gulch, 8-spoke open circles from lavas collected from the west slope of Mauna Kea (includes lavas from both groups), and 4-spoke open circles from Laupahoehoe Group lavas collected by Macdonald. The line represents the region of low-population thought to divide tholeiitic from alkalic lavas (Macdonald and Katsura, 1964).

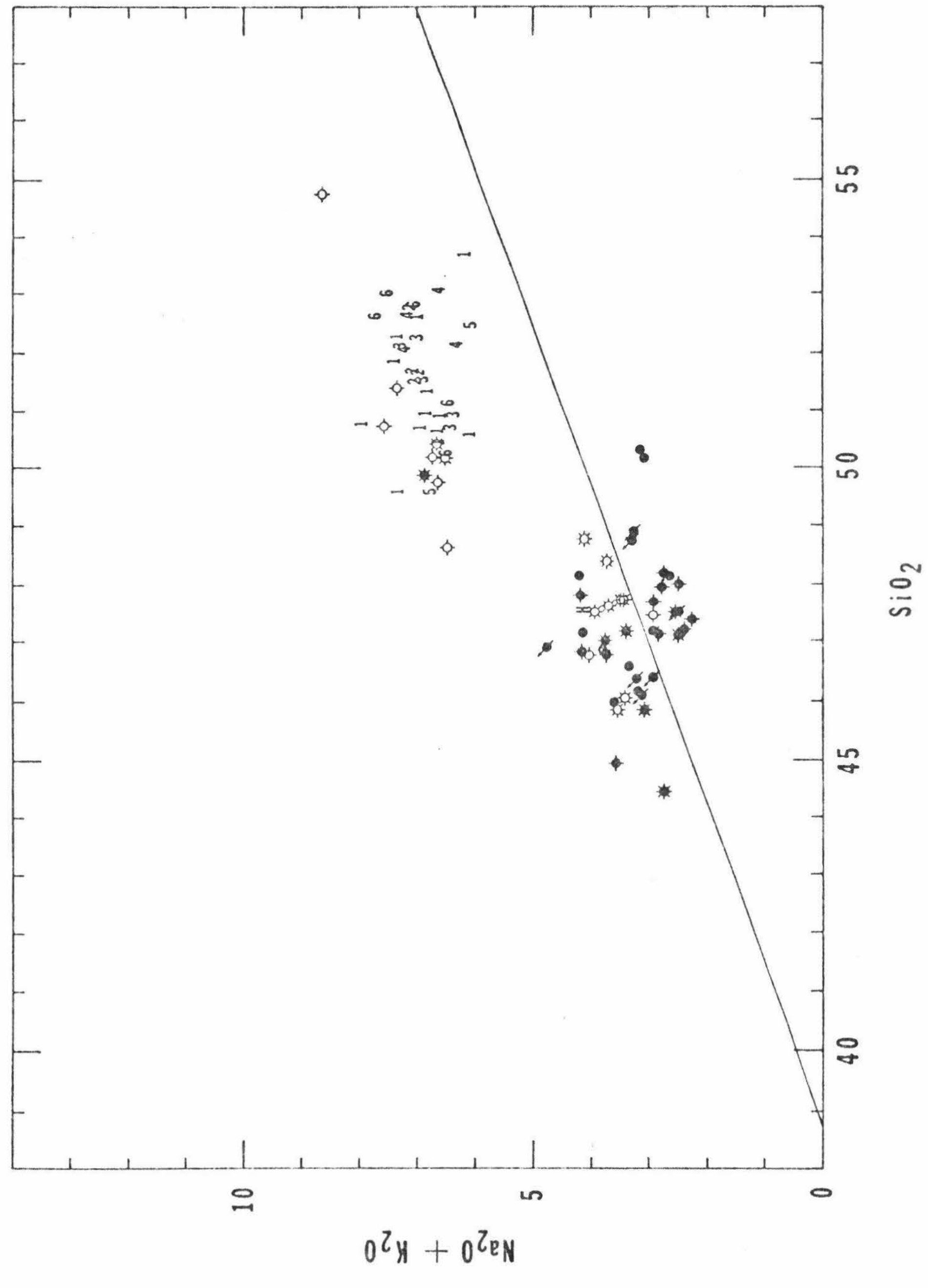

FIGURE 5

Figure 6: D.I. (normative quartz + orthoclase + nepheline + kalsilite + leucite + albite) versus % An (100 An/(An + Ab)). Squares = Loaloan Stage, circles = Kemolean Stage, triangles = Kuupahaan Stage, "+" = Hanaipoean Stage, "X" = Poliahuan Stage, diamonds = Liloean Stage, tree = uppermost flow of the Hopukani Formation. Hawaiite and mugearite fields follow the criteria of Coombs and Wilkinson (1969).

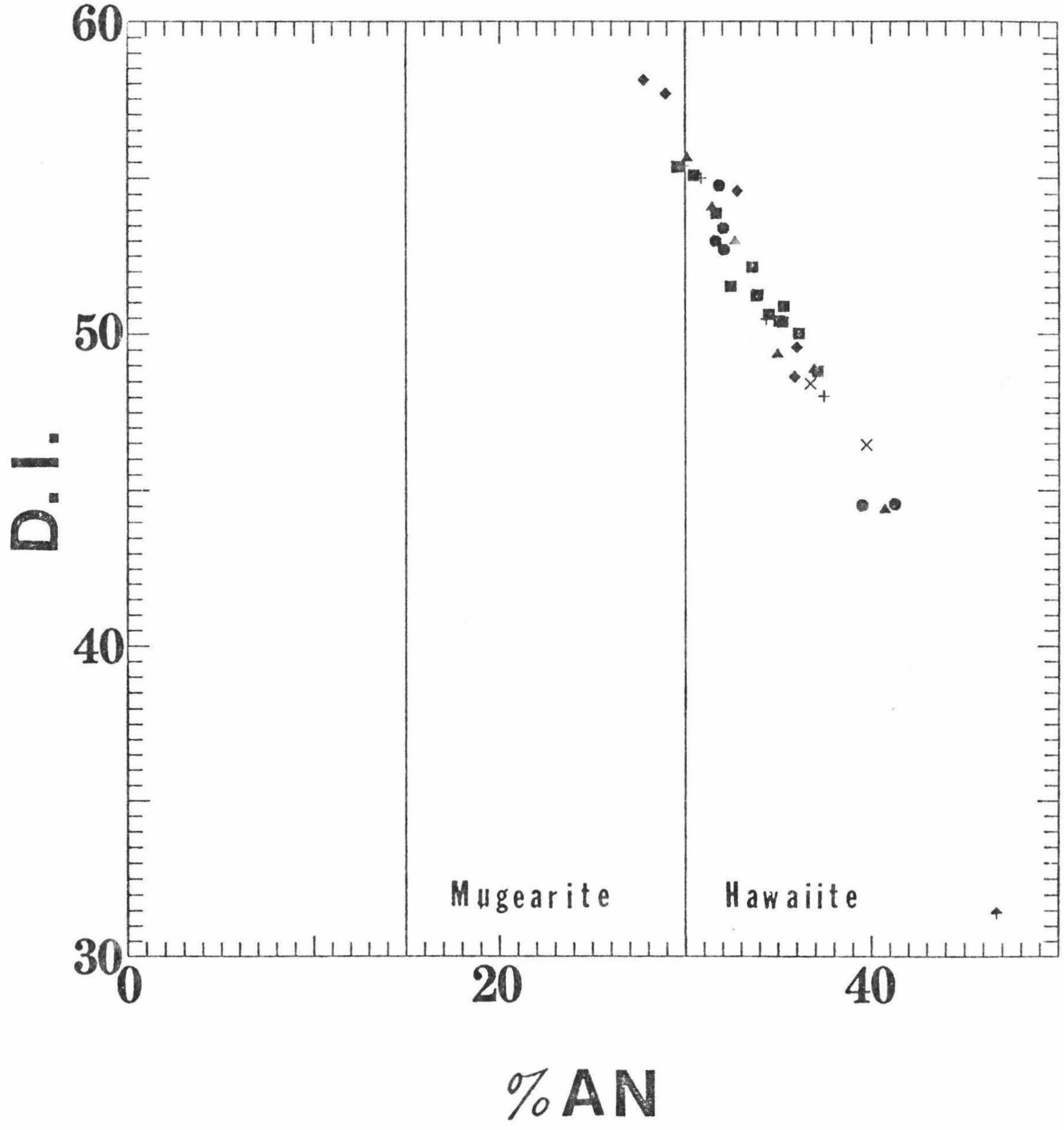

FIGURE 6

Figure 7: Ne-Pl-Ol-Q ternary diagram similar to Figure 2. Symbols same as in Figure 5. One atmosphere cotectic is from O'Donnell and Presnall (1980).

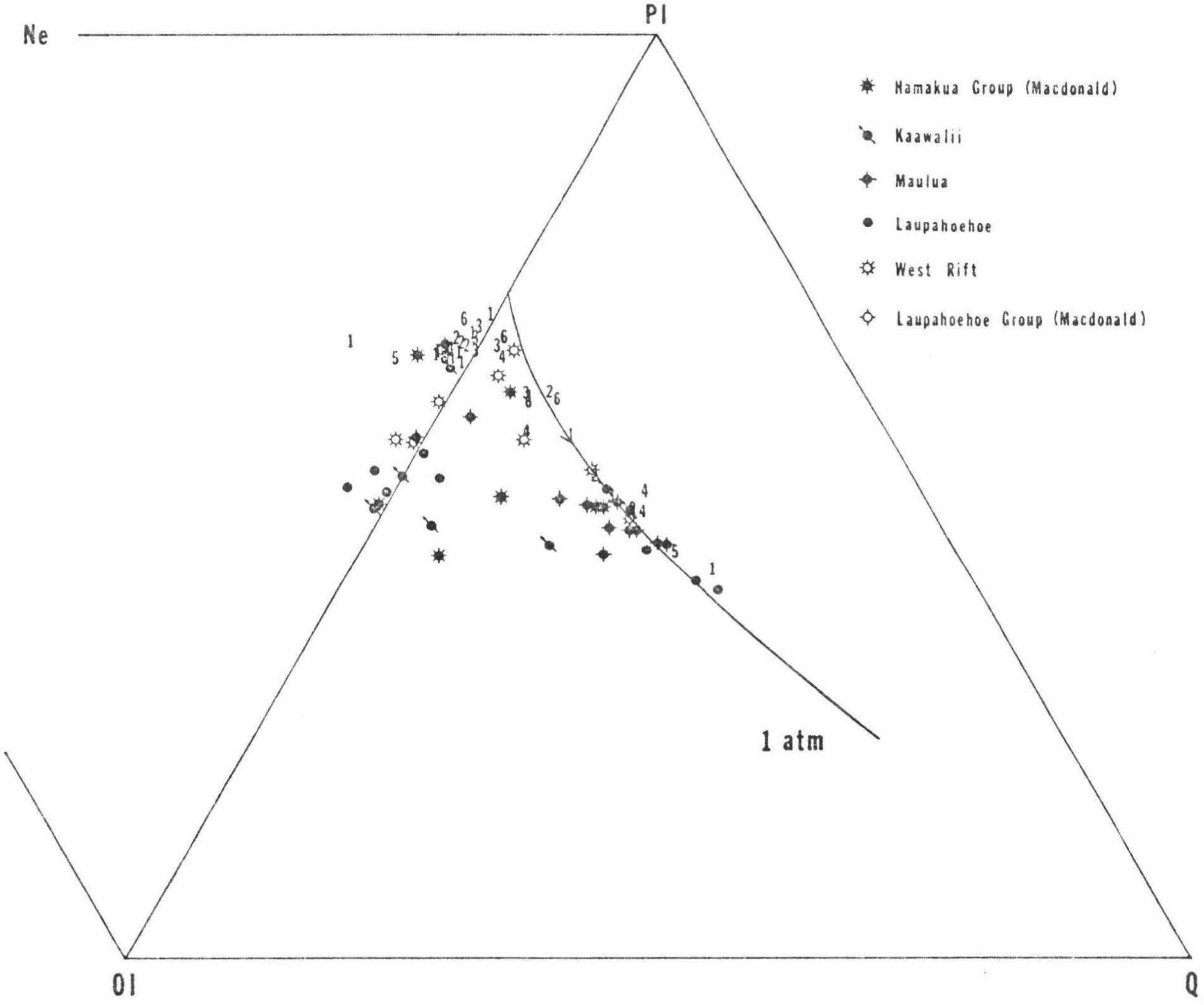
Ne
Pl
Ol
Q
1 atm
Hamakua Group (Macdonald)
Kaawalii
Maulua
Laupahoehoe
West Rift
Laupahoehoe Group (Macdonald)

FIGURE 7

Figures 8-16: Oxide-Zr variation diagrams. Symbols same as in Figure 6.

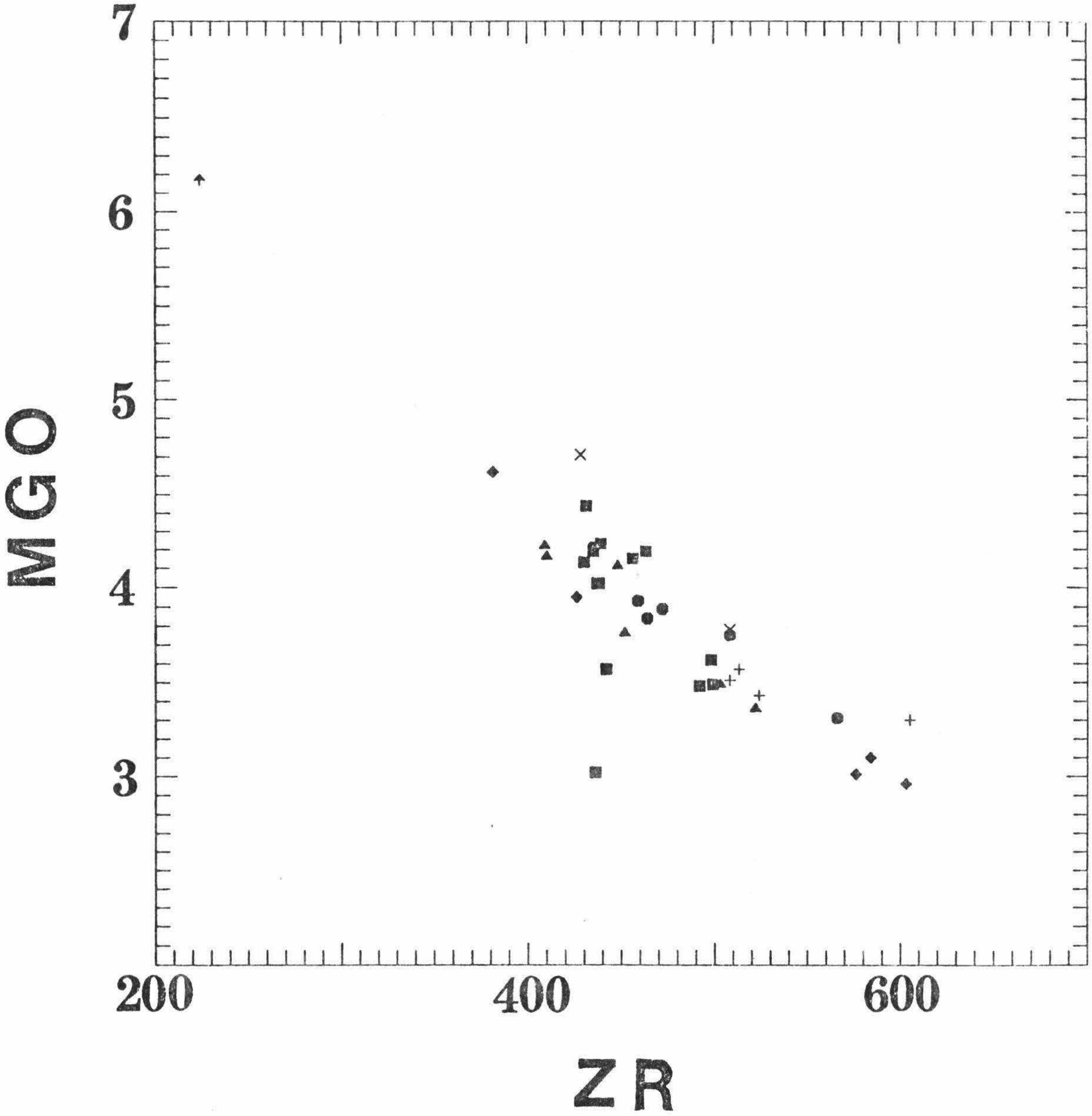

FIGURE 8

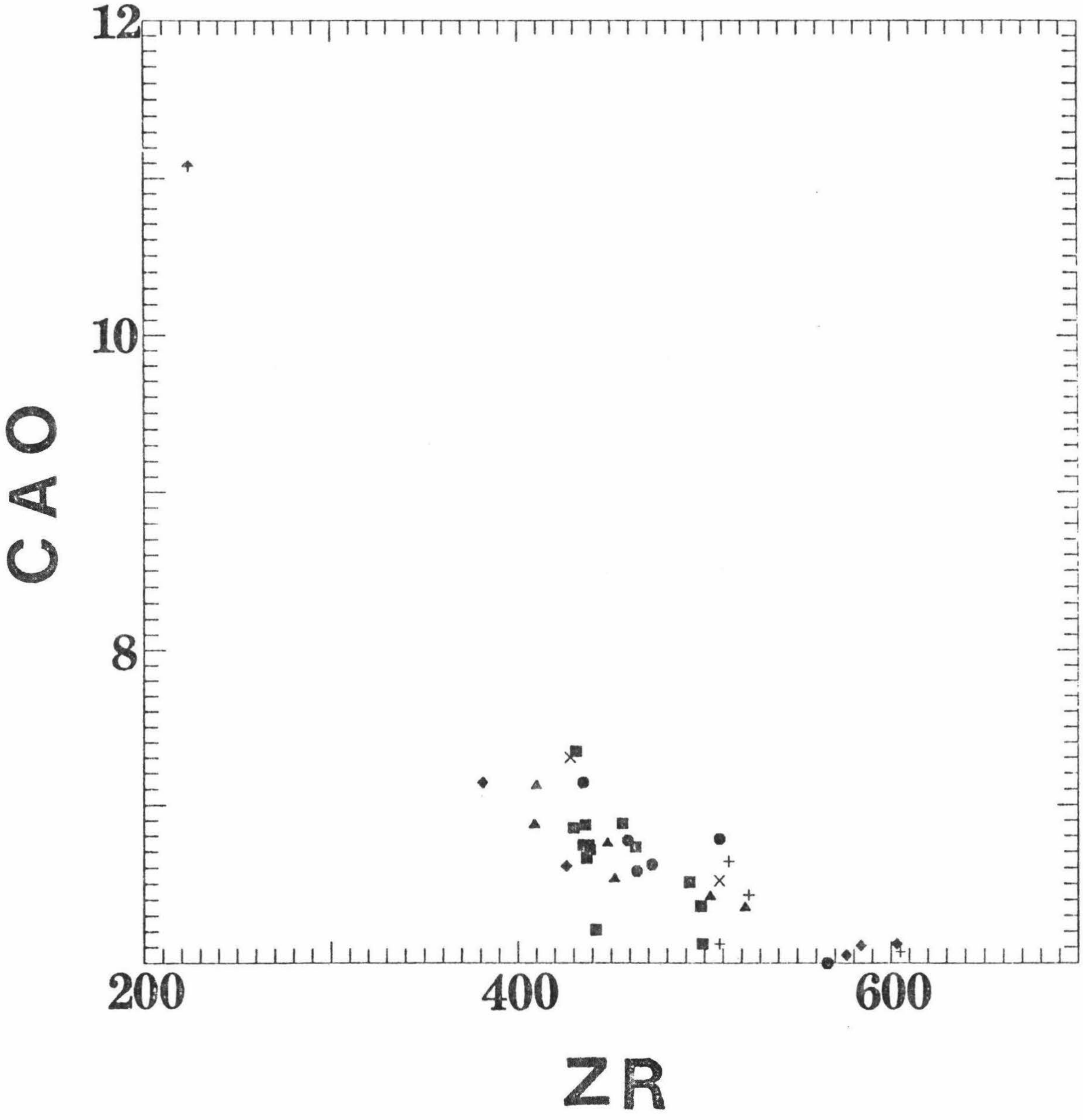

FIGURE 9

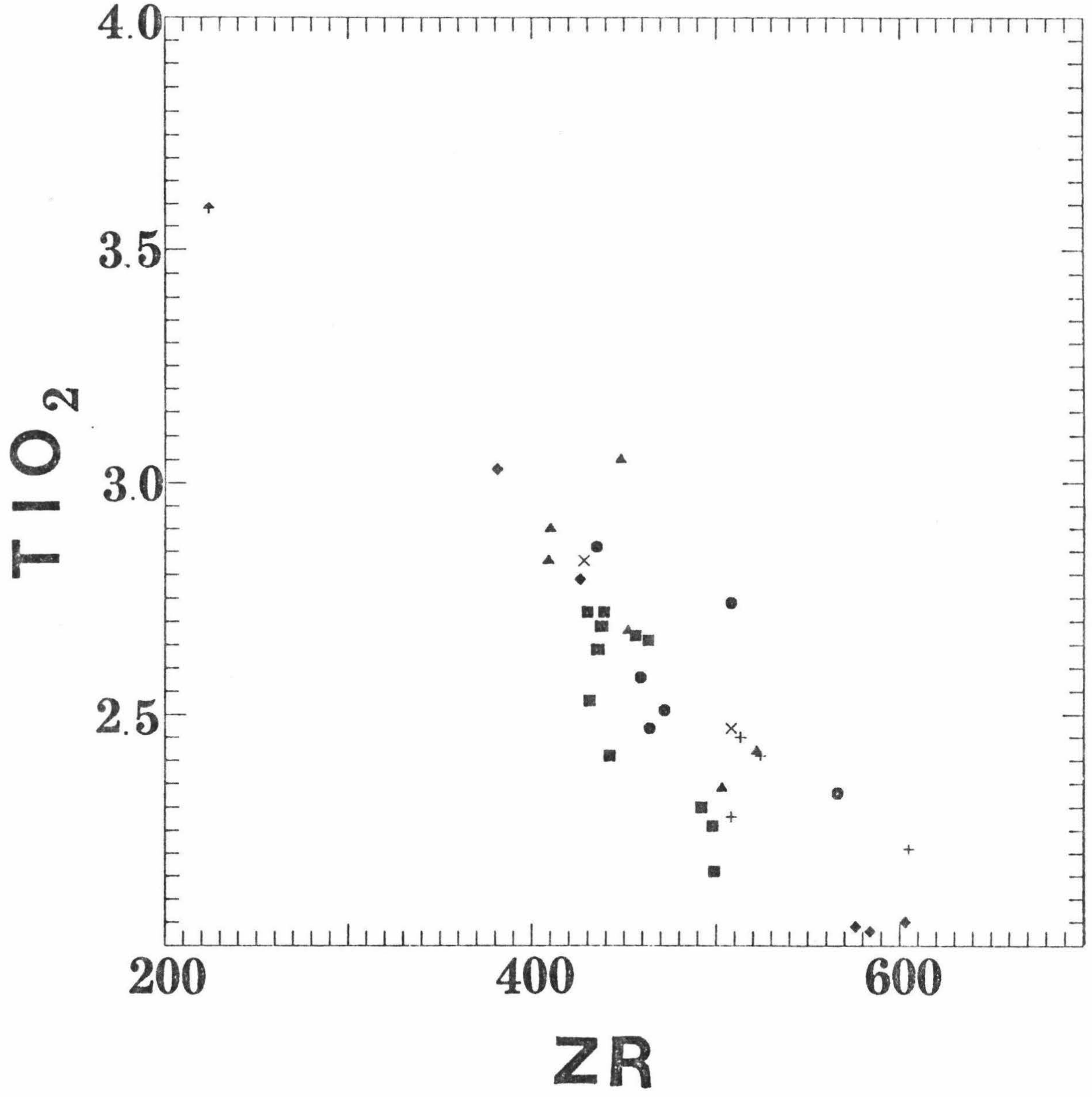

FIGURE 10

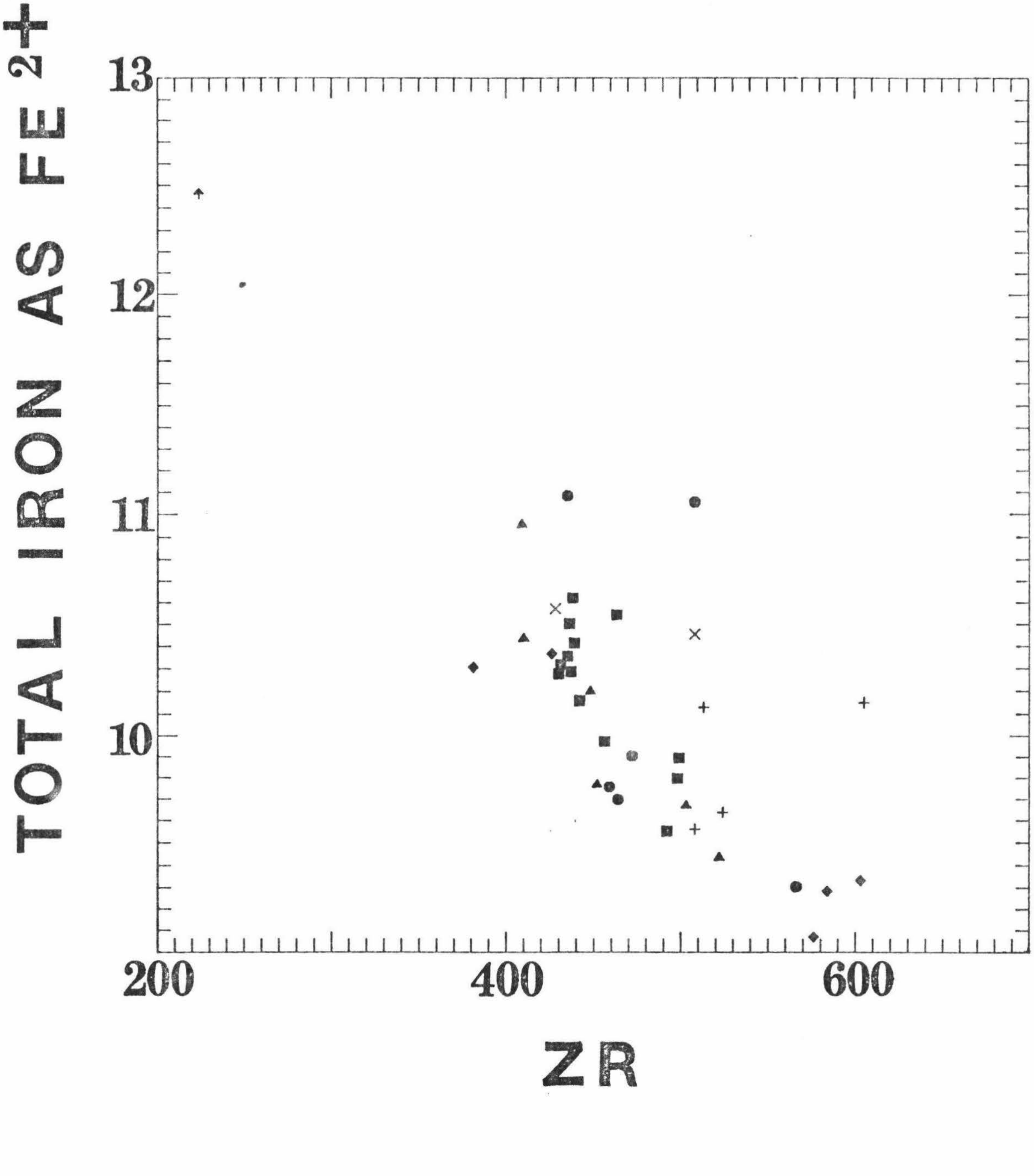

FIGURE 11

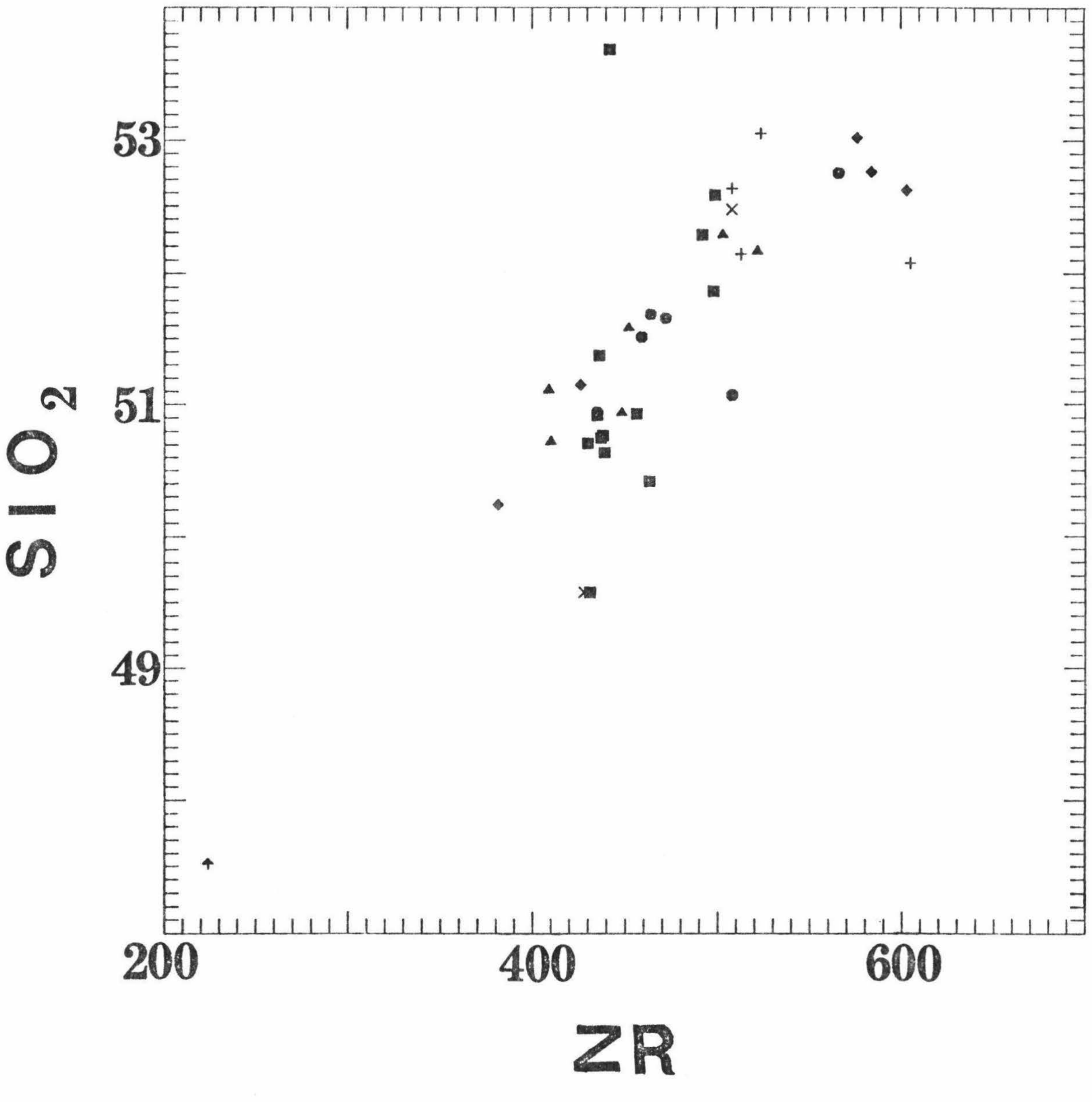

FIGURE 12

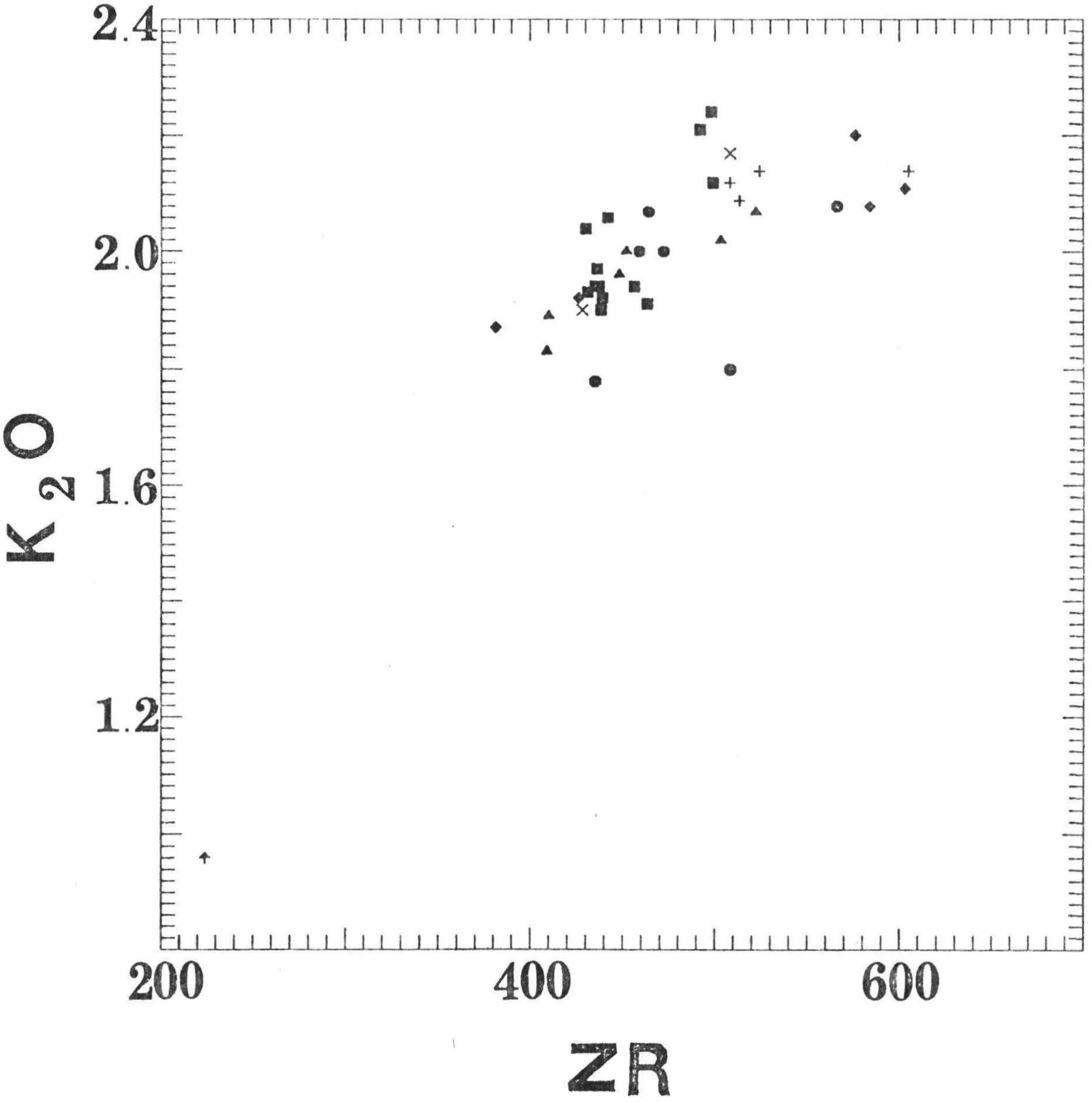

FIGURE 13

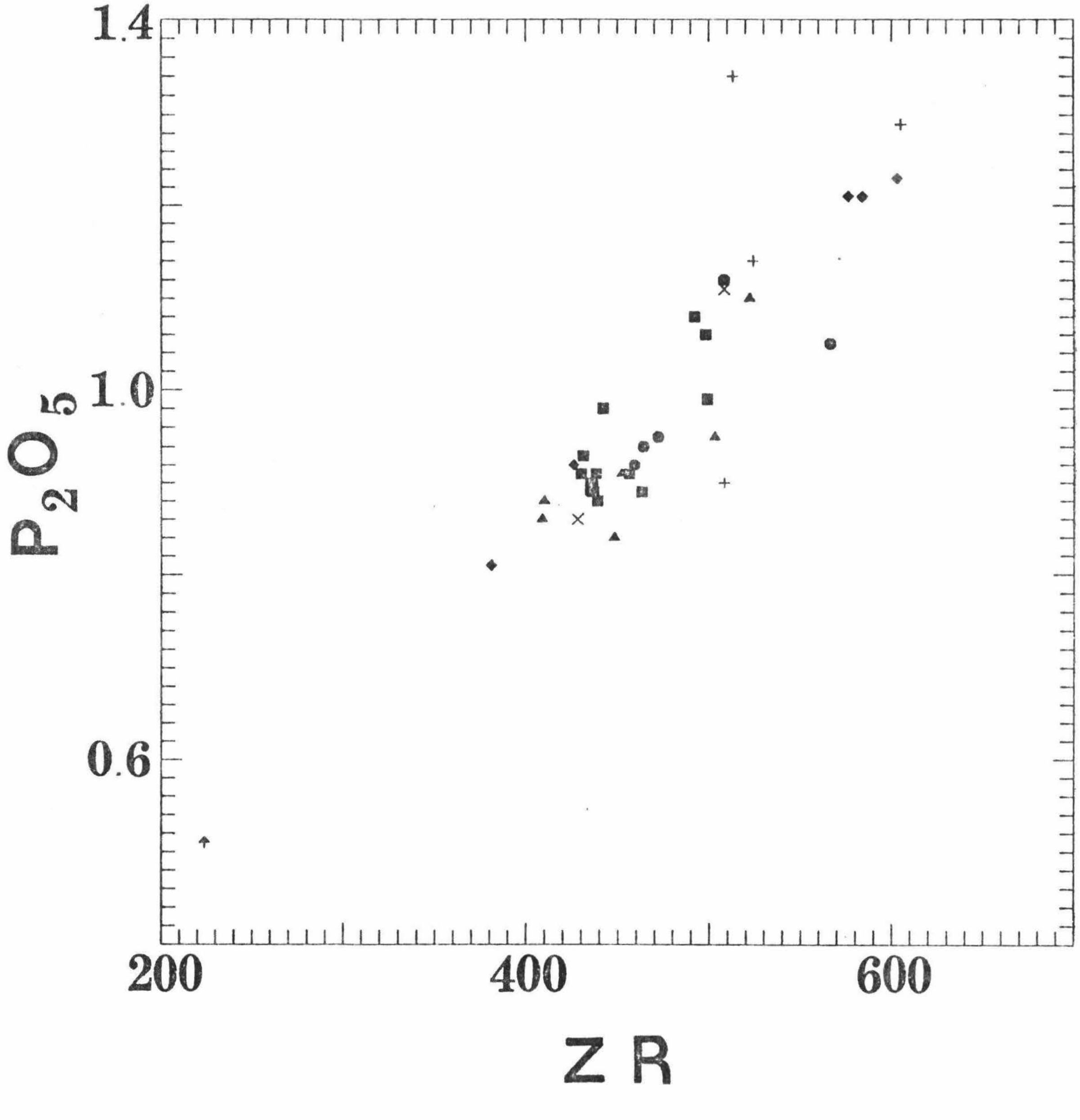

FIGURE 14

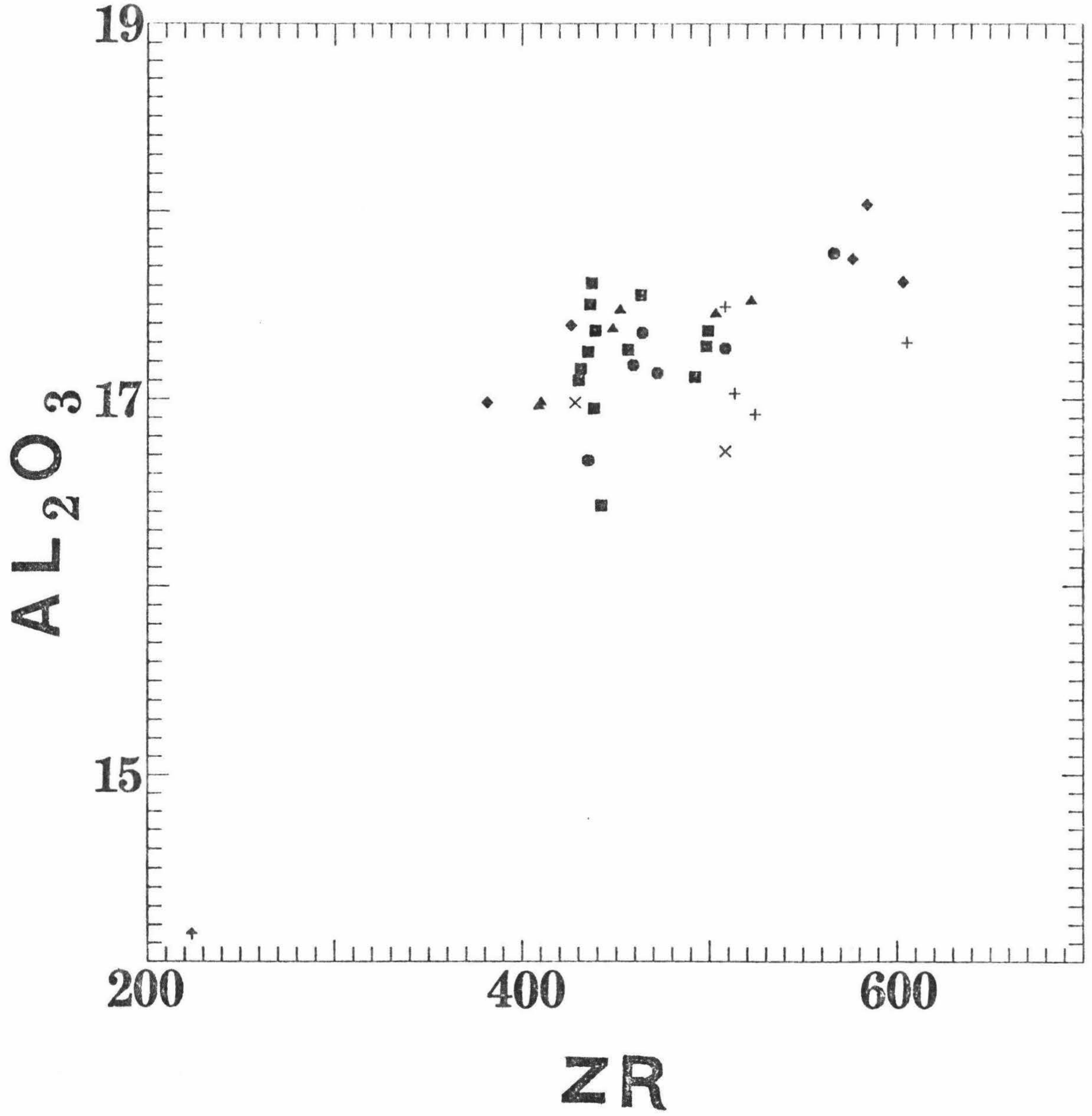

FIGURE 15

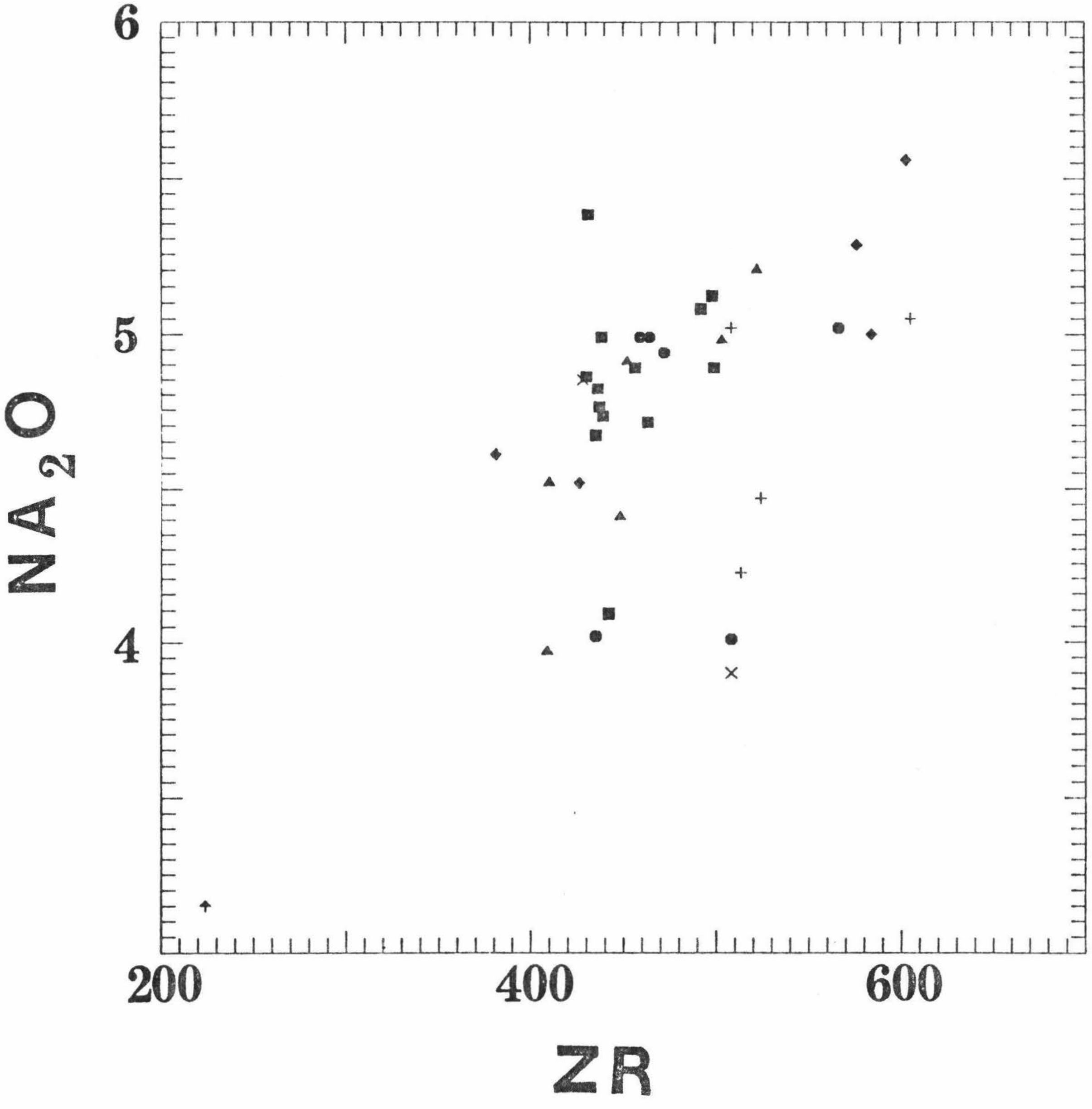

FIGURE 16

Figures 17-22: Trace element-Zr variation diagrams. Symbols same as in Figure 6. Analytical uncertainty is ± 2 ppm. Ni abundances are very close to the detection limit of the instrument.

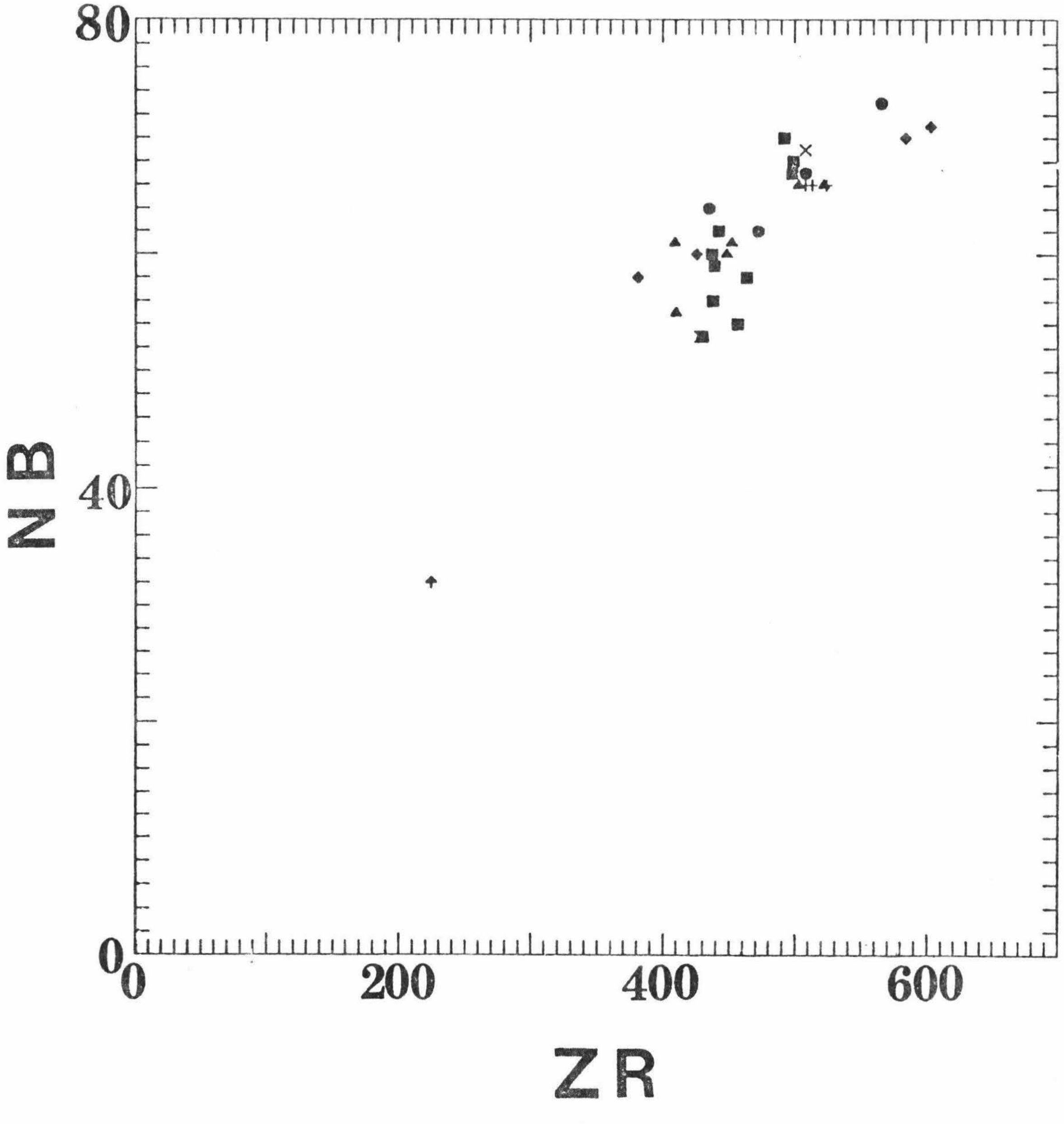

FIGURE 17

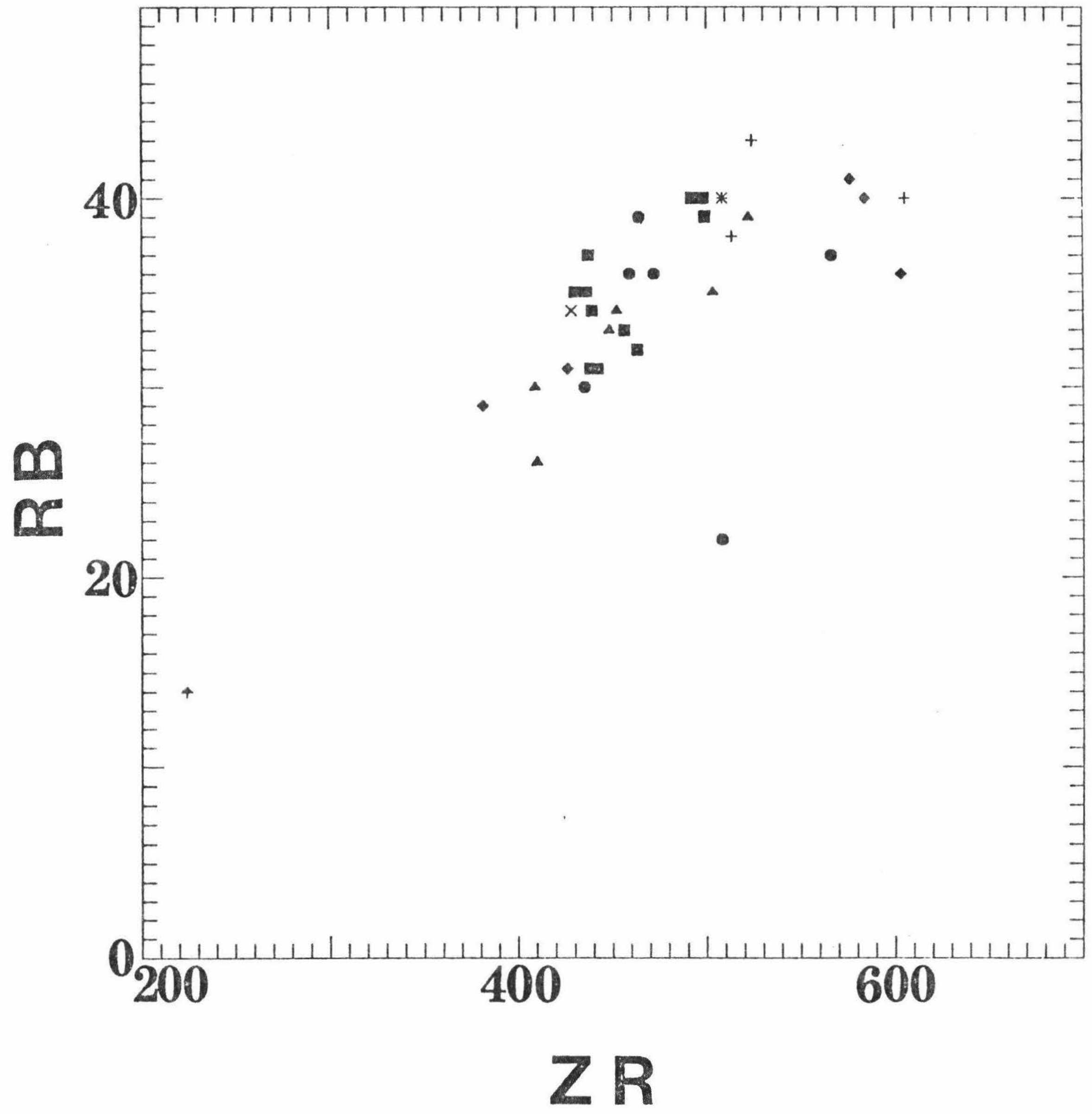

FIGURE 18

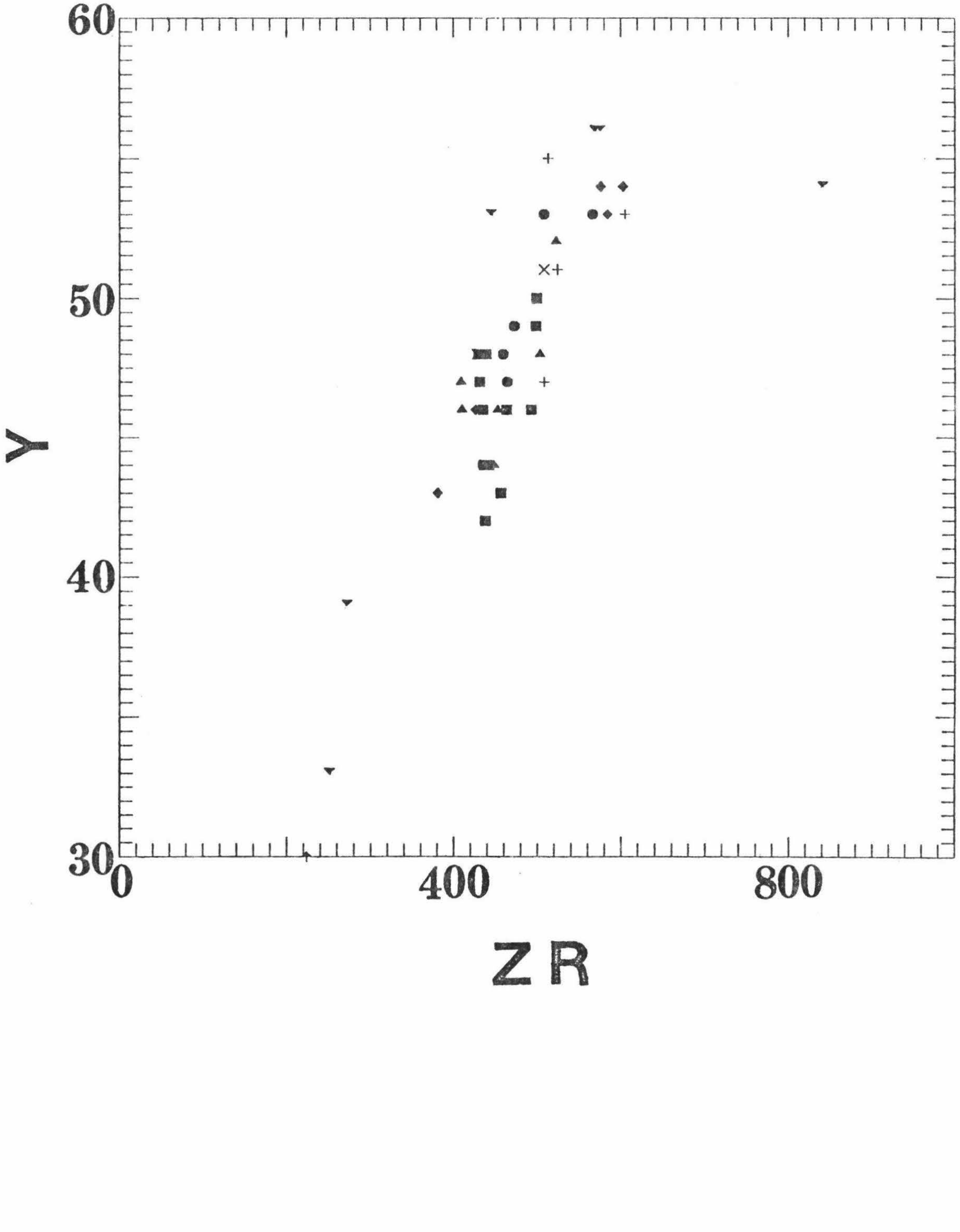

FIGURE 19

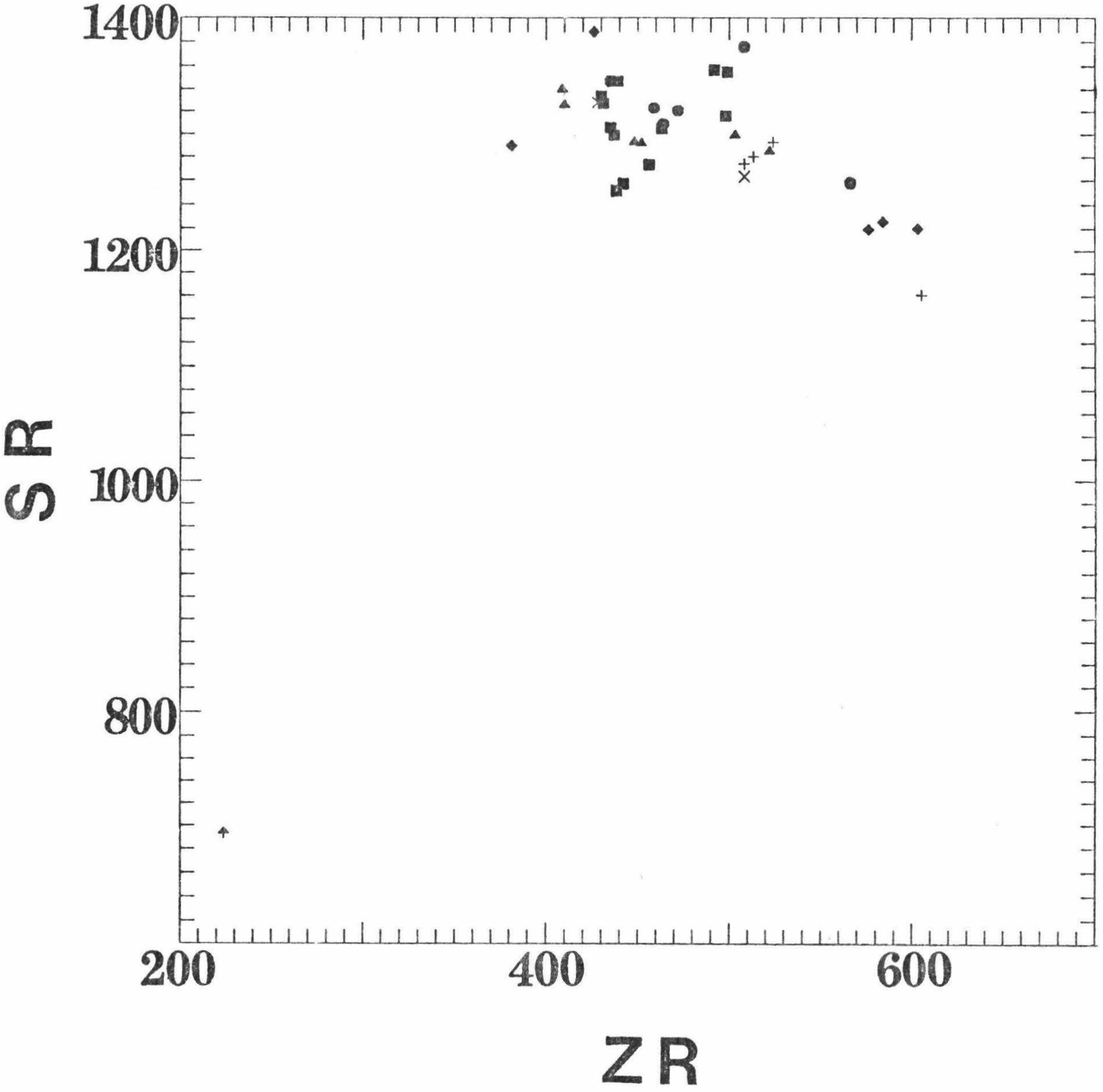

FIGURE 20

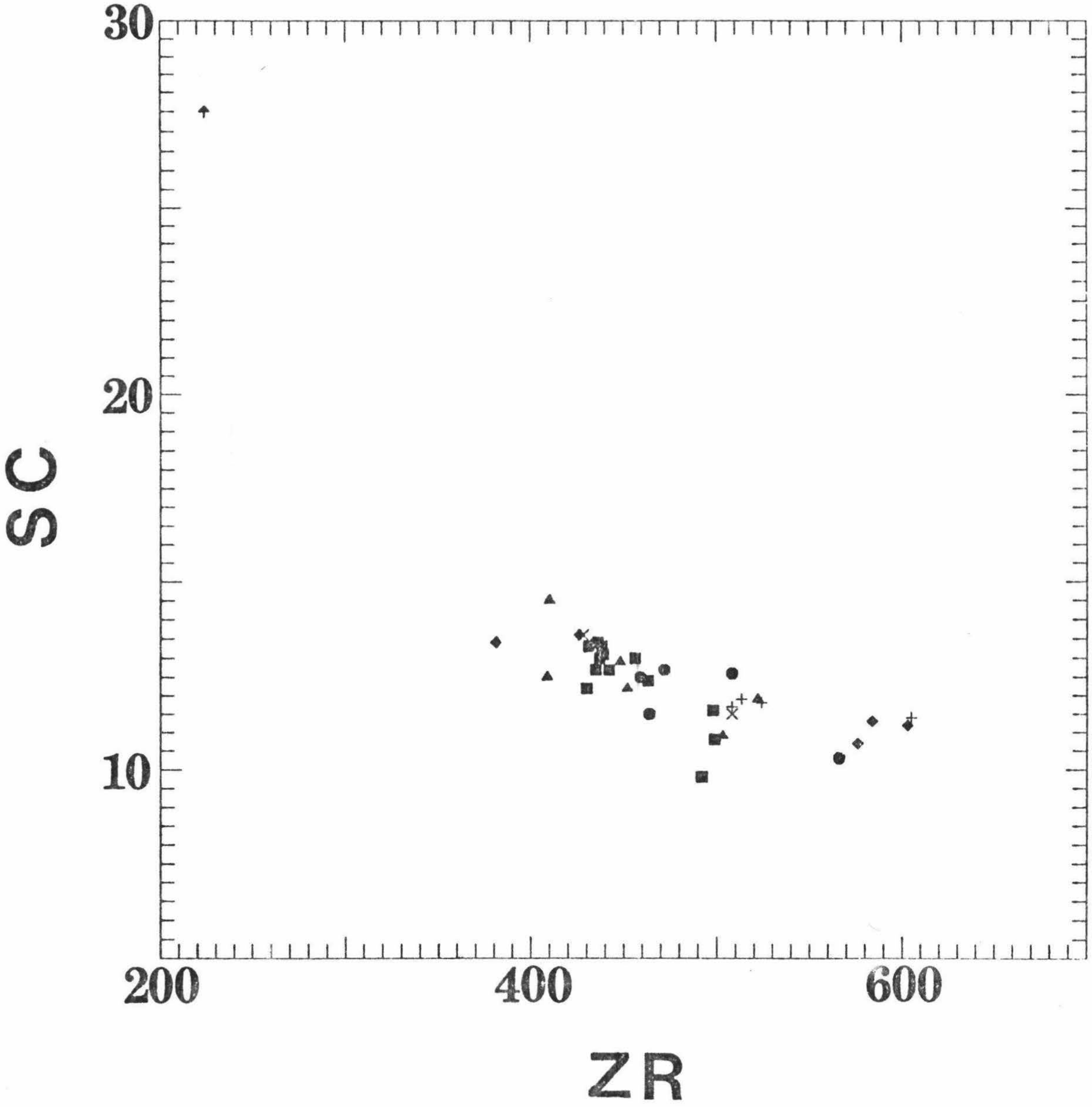

FIGURE 21

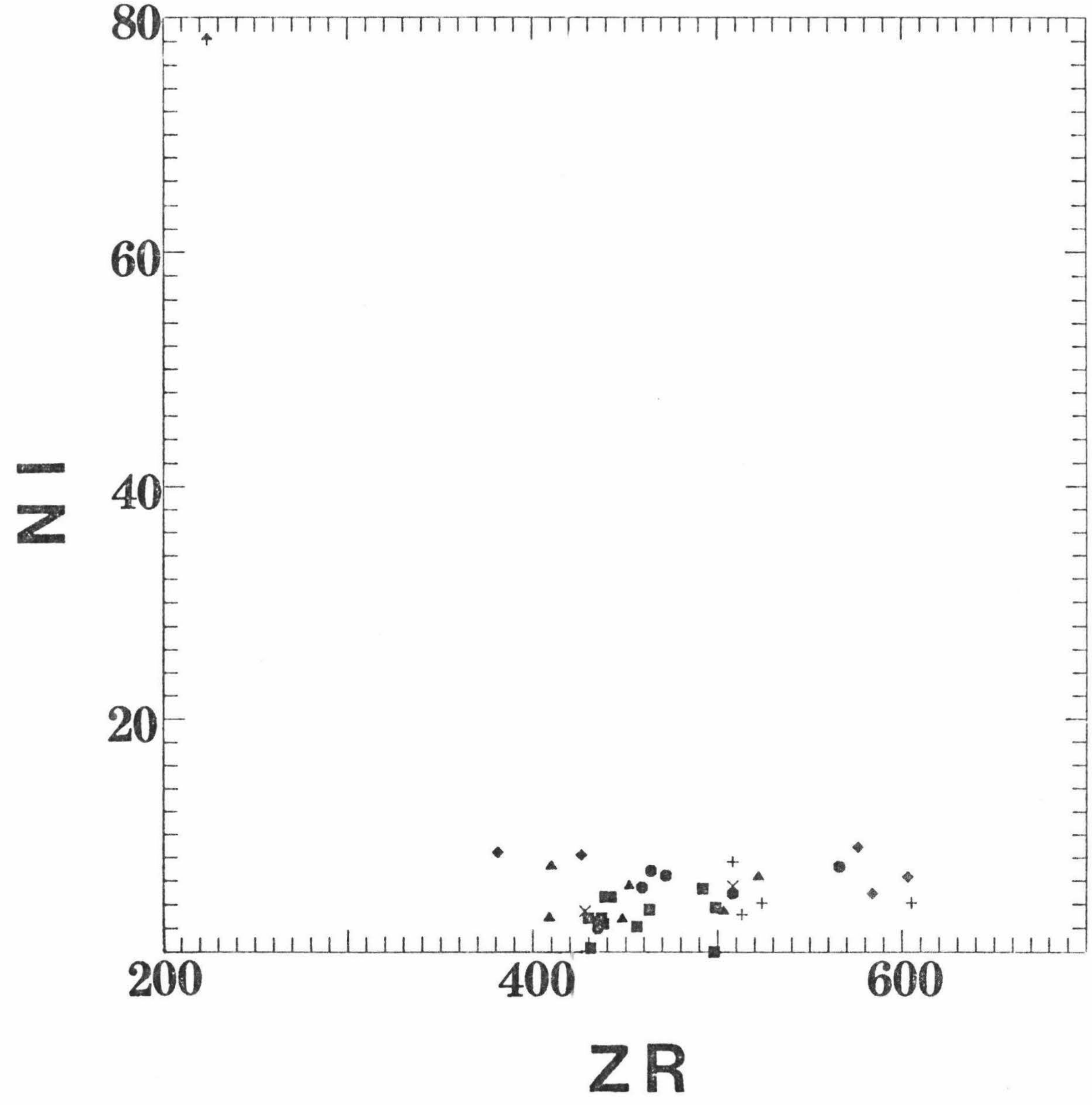

FIGURE 22

Figure 23: TiO_2/Al_2O_3 versus Zr for south rift lavas. If TiO_2 were being controlled by clinopyroxene, the trend would increase. The observed decrease probably indicates that Ti-rich magnetite is controlling TiO_2 abundances. Symbols same as in Figure 6.

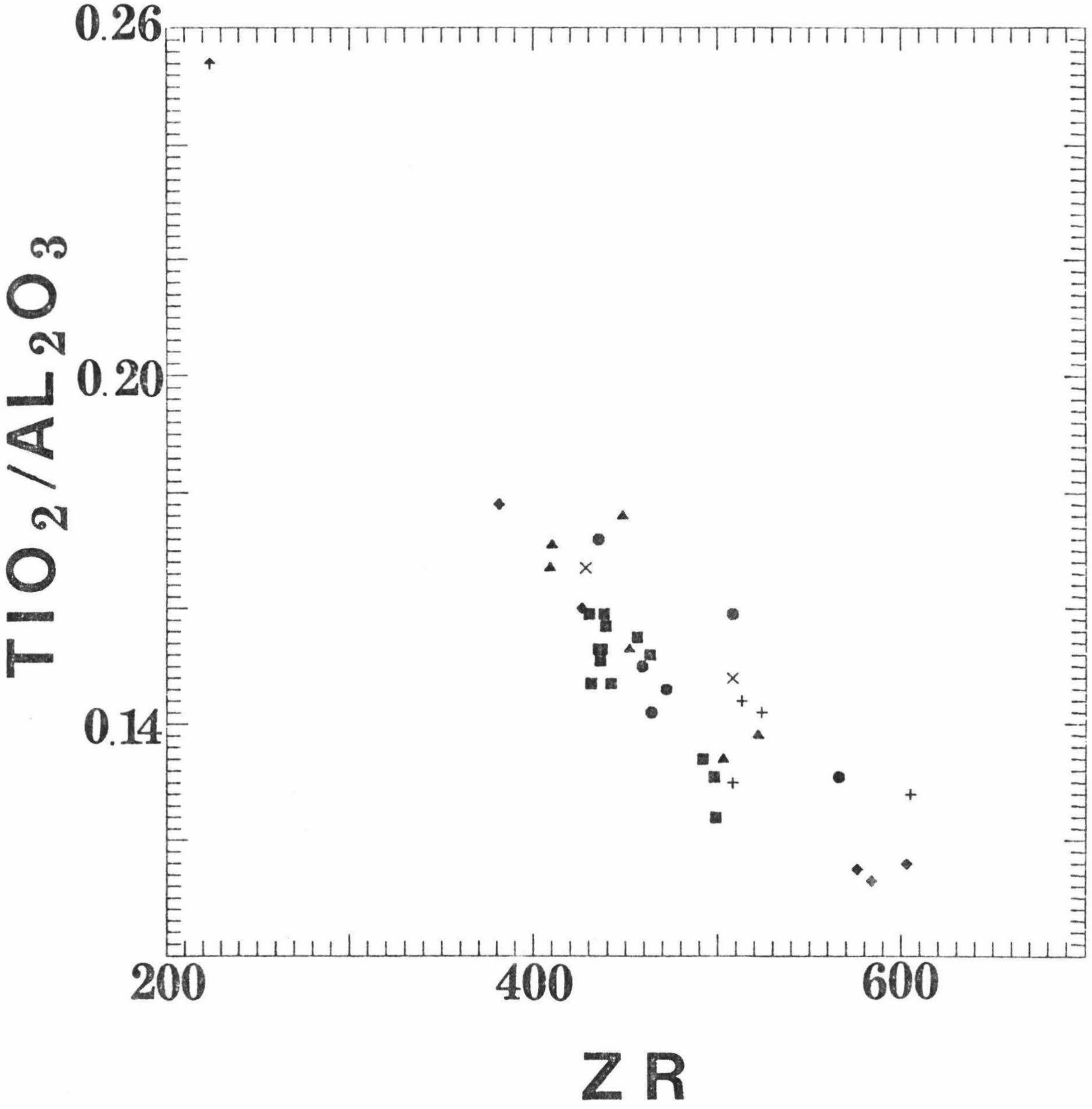

FIGURE 23

Figure 24: Al_2O_3/CaO versus Zr for south rift lavas. The observed increase suggests control by clinopyroxene. Plagioclase fractionation would cause this ratio to decrease. Symbols same in Figure 6.

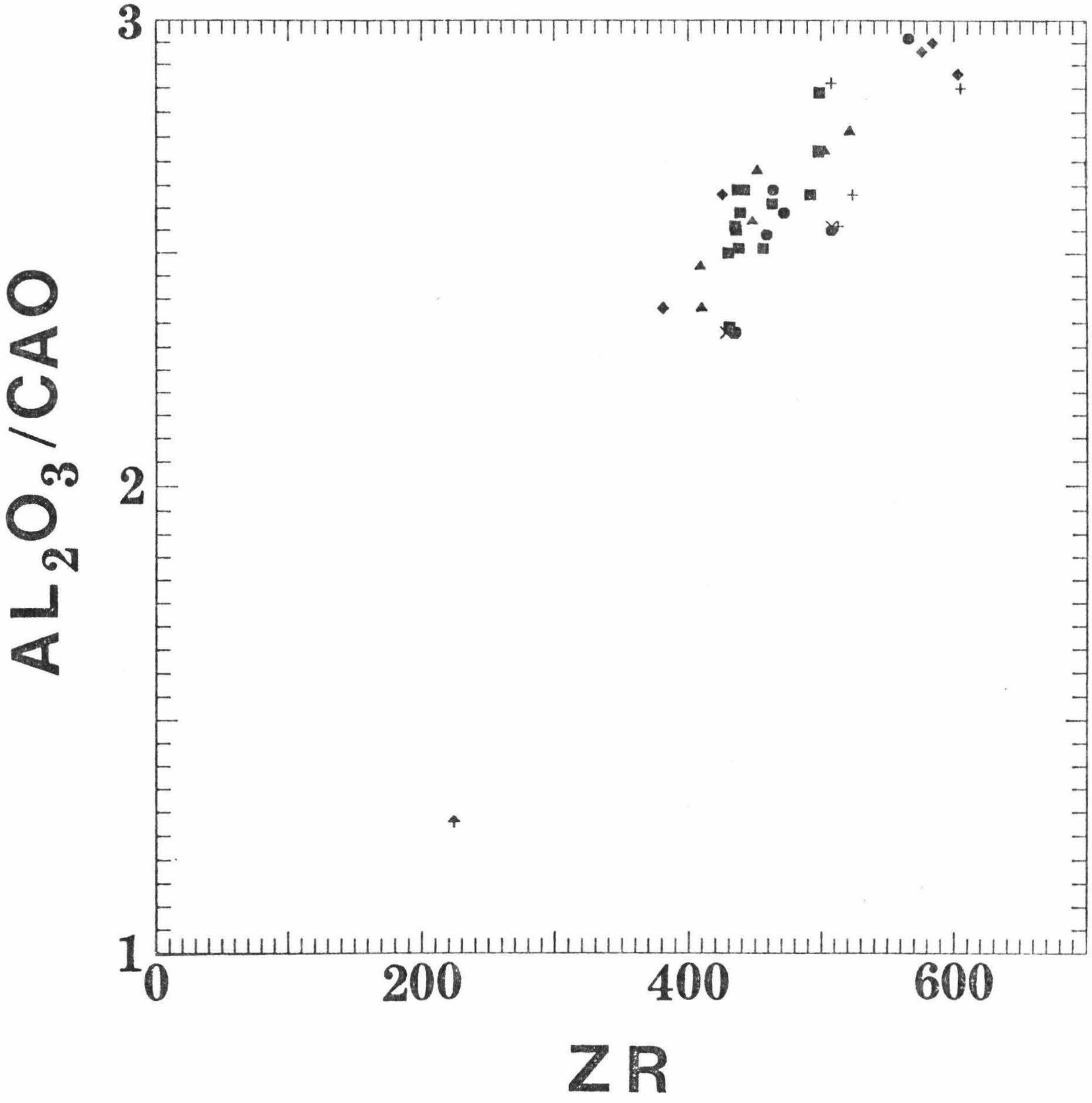

FIGURE 24

Figure 25: Zr/Y versus P_2O_5 for south rift lavas. The increase in ratios suggests that a phase capable of removing Y is being fractionated (probably clinopyroxene). However, successively decreasing amounts of partial melting could also account for the observed increase. Symbols are same as in Figure 6. Picnic table symbols represent Laupahoehoe Group samples collected by Macdonald (1968; Macdonald and Katsura, 1964).

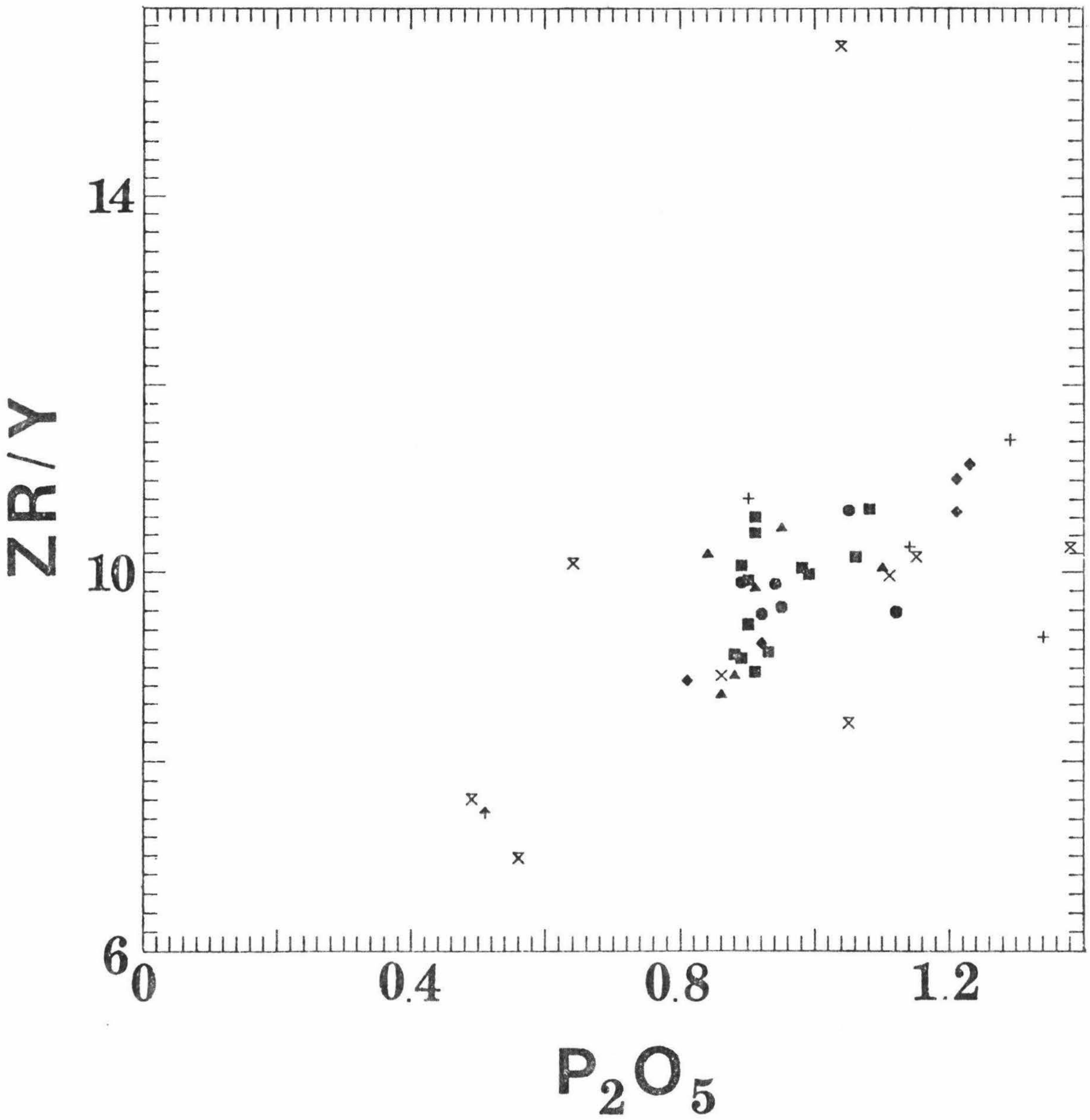

FIGURE 25

Figures 26-30; Oxide and trace element variation plotted against time. Age decreases to the right. Stages of the Laupahoehoe Group are separated from each other by spaces. Lavas which could not be related stratigraphically within each stage are plotted as the same age. Ml = Loaloan Stage, Mk = Kemolean Stage, Me = Kuupahaan Stage, Mh = Hanaipoean Stage, Mp = Poliahuan Stage, Mi = Liloean Stage.

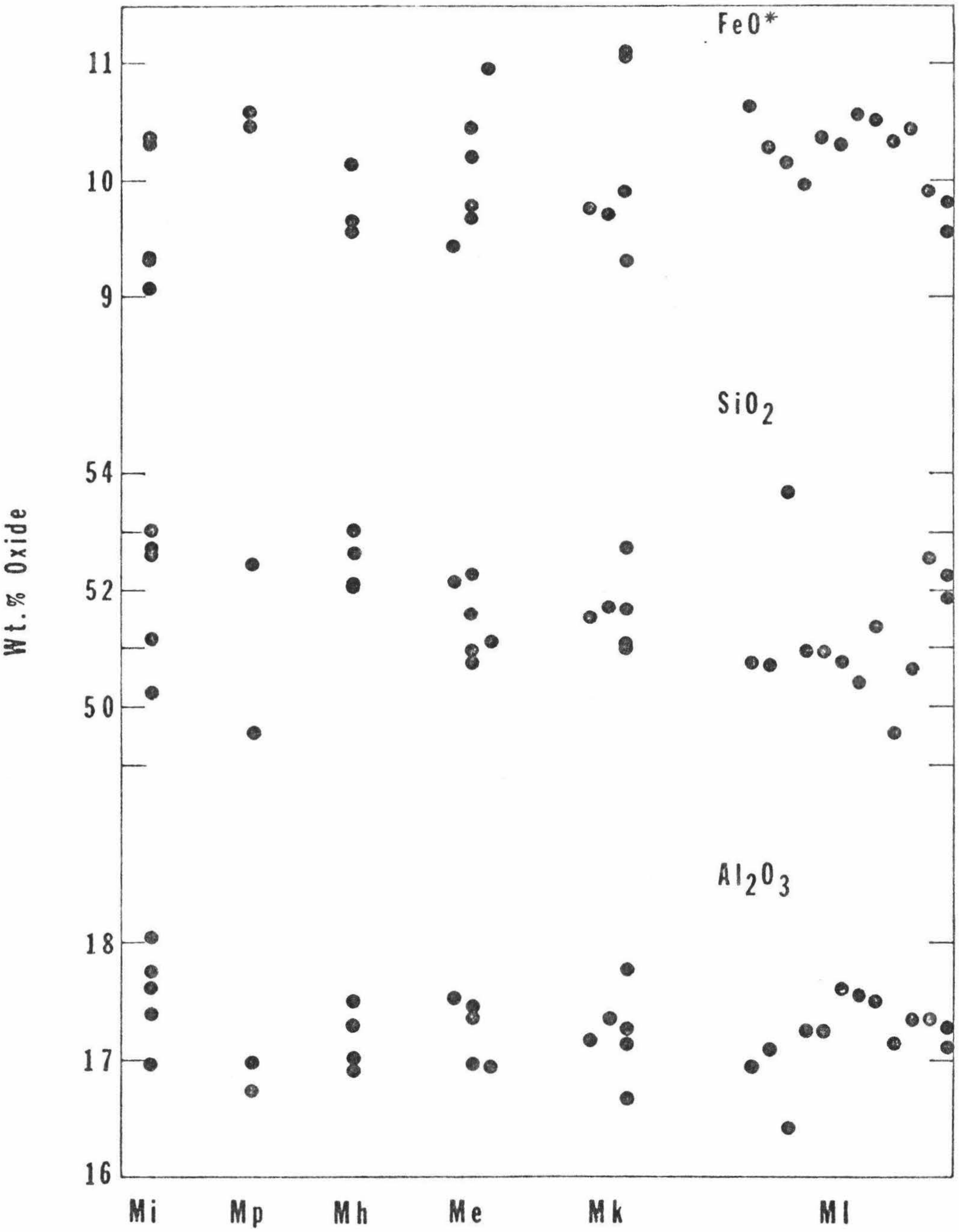

FIGURE 26

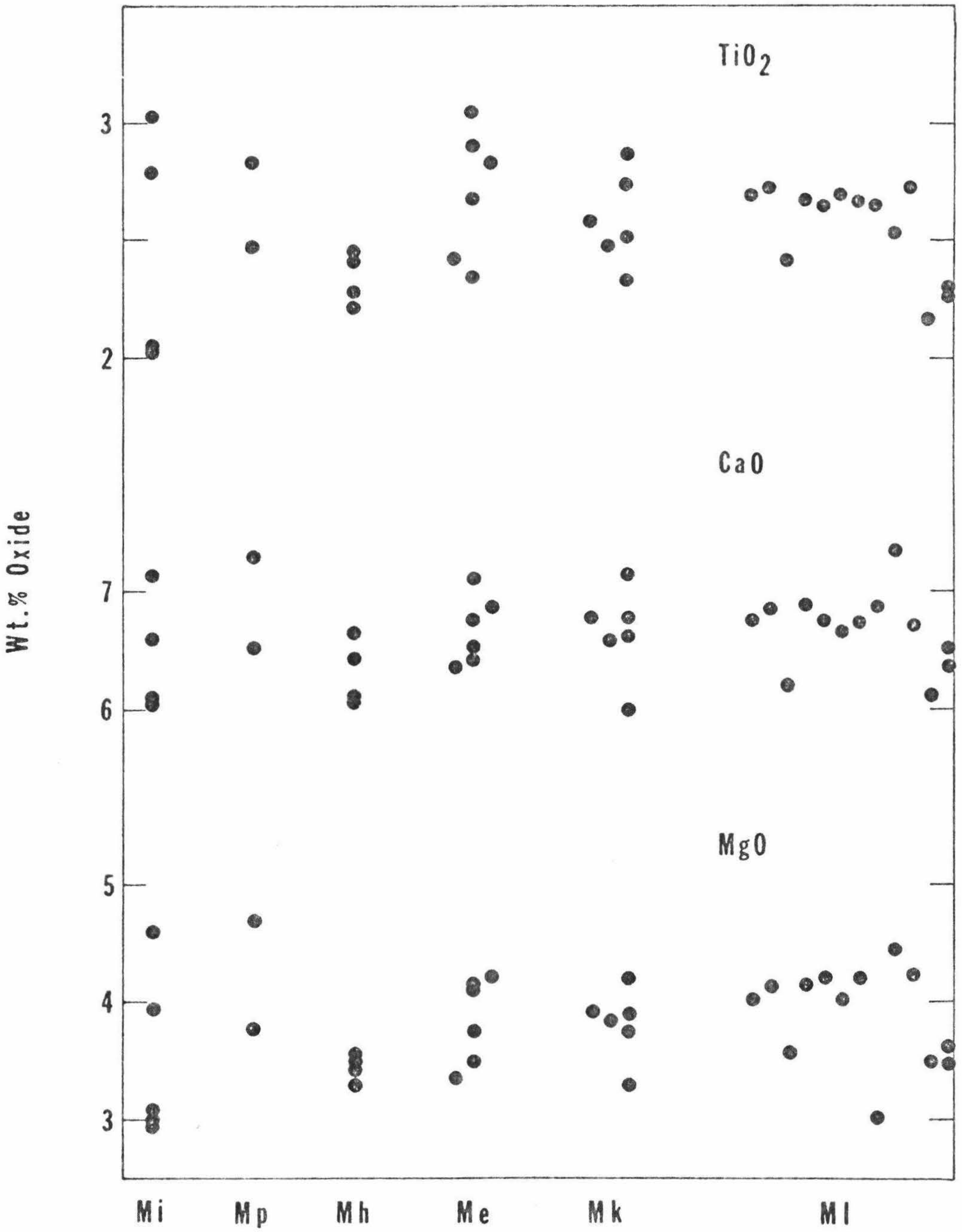

FIGURE 27

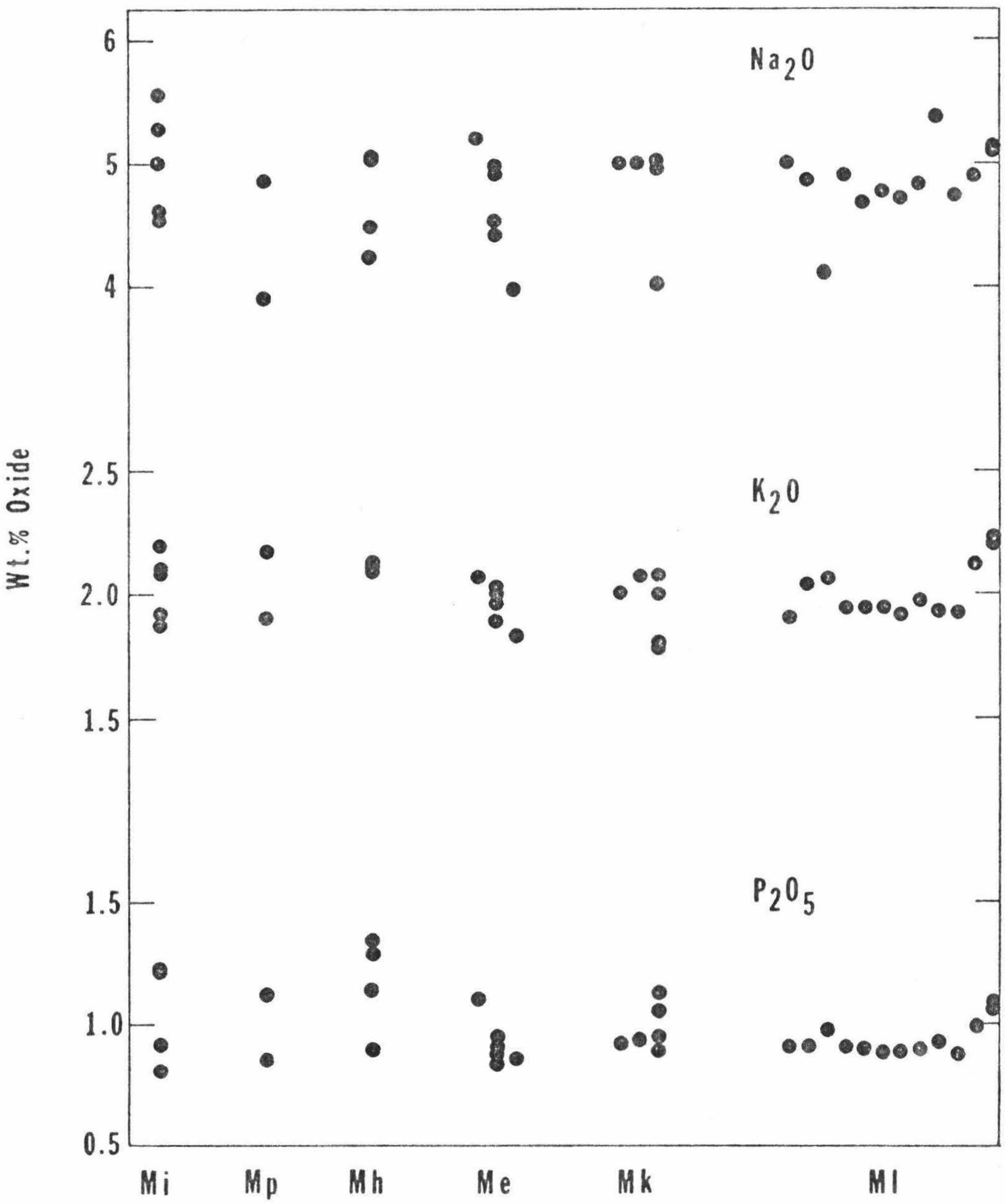

FIGURE 28

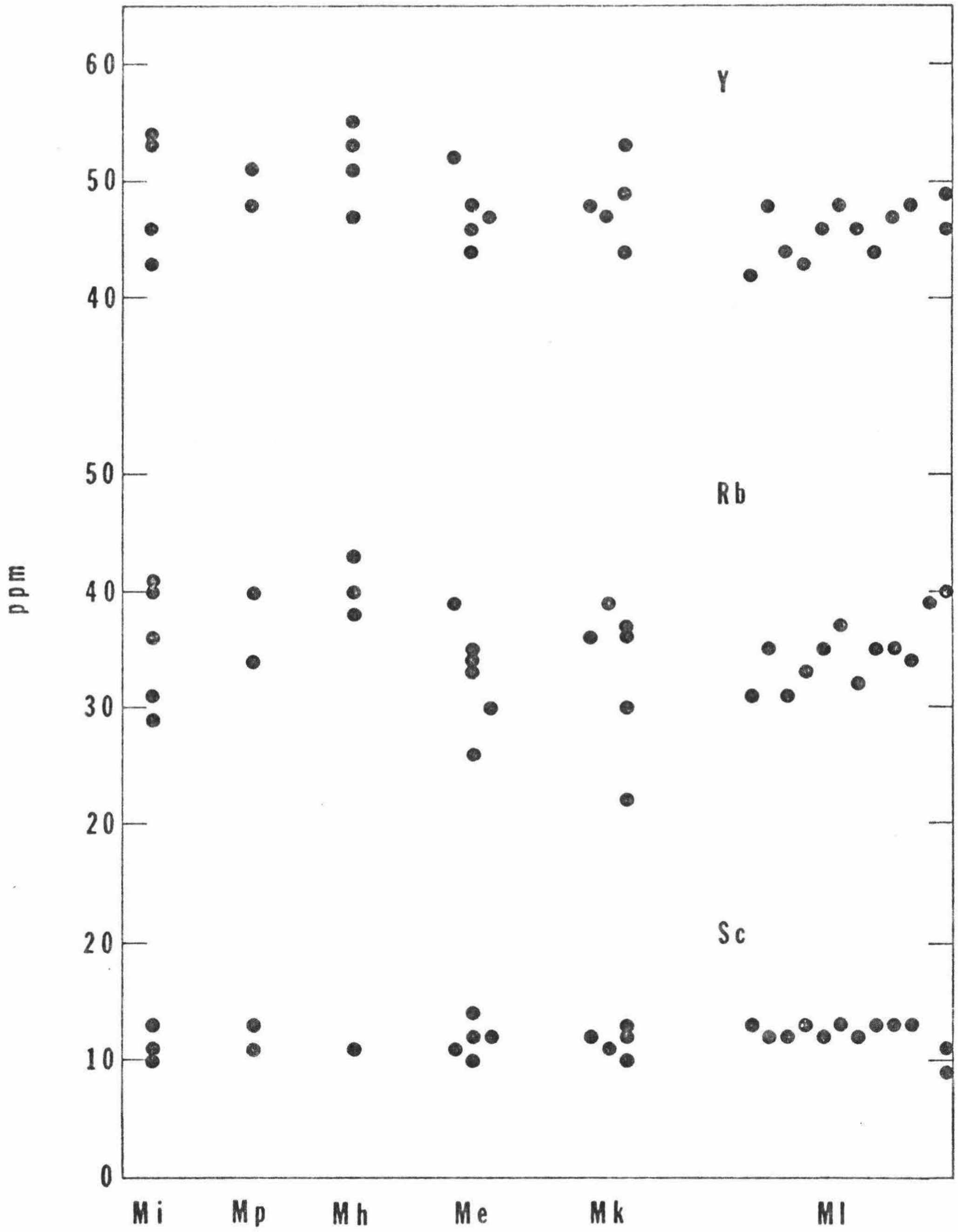

FIGURE 29

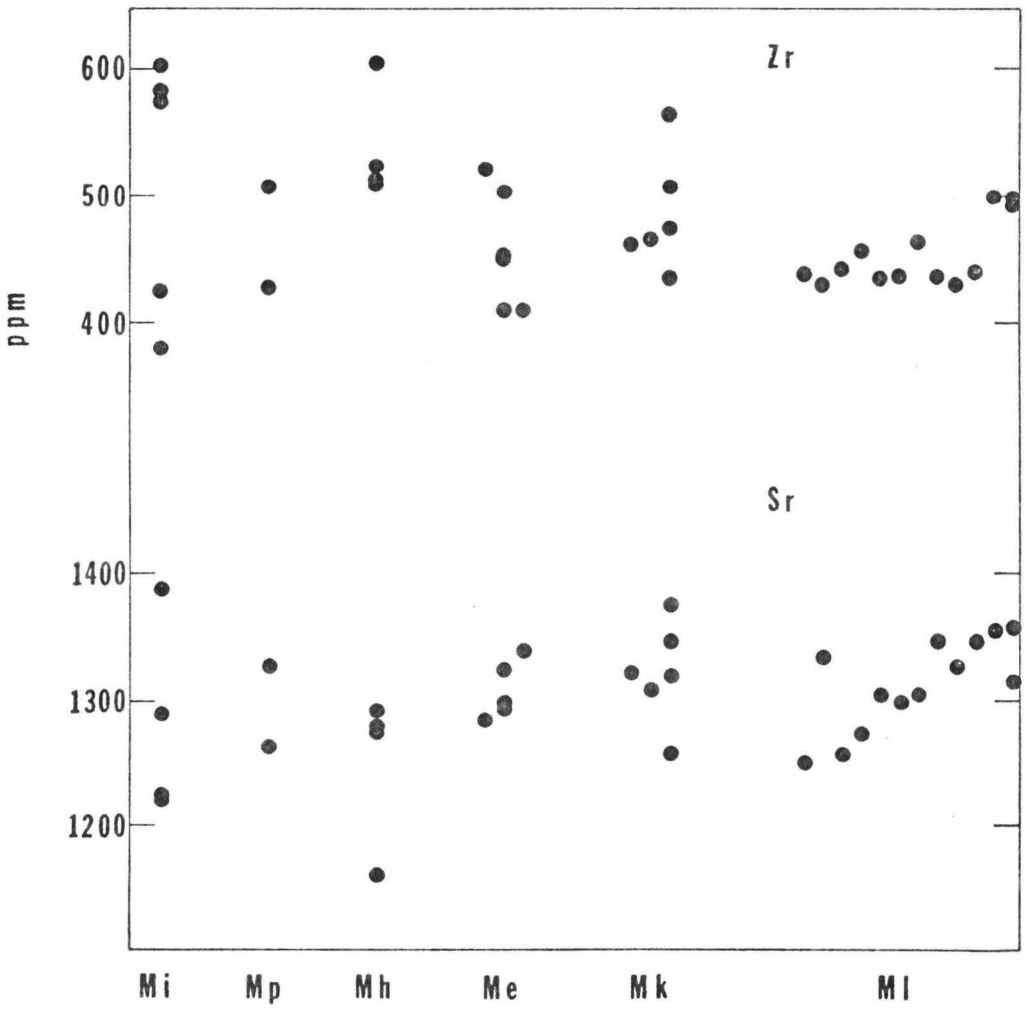

FIGURE 30

Figure 31: Zr/Nb versus Zr for south rift lavas. The increase in ratios suggests either fractionation of a phase capable of preferentially incorporating Nb (probably magnetite) or successively decreasing amounts of partial melting. Magnetite fractionation is probably indicated since these south rift lavas are probably not primary.

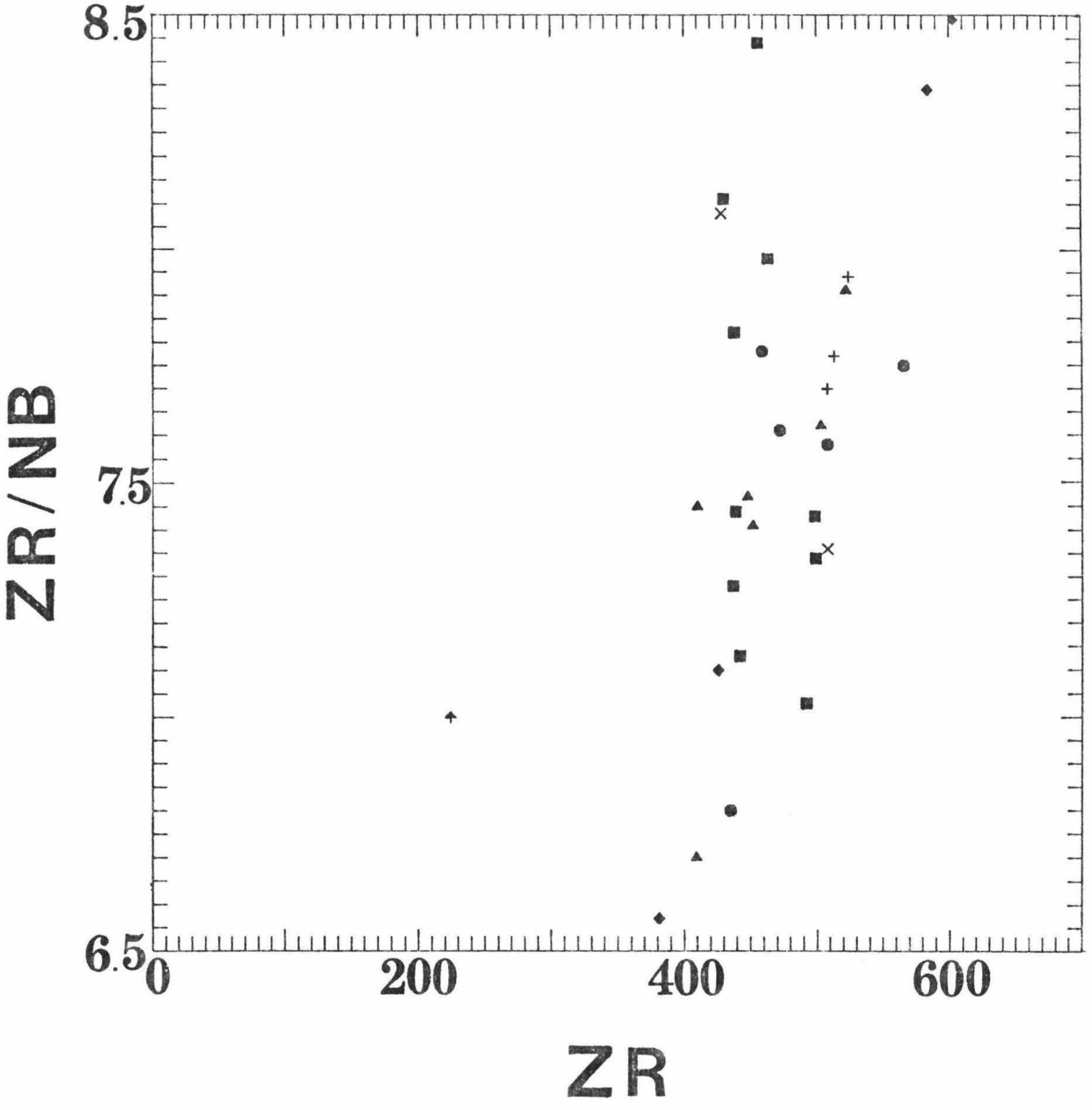

FIGURE 31

Figure 32: The Loaloan Stage rift. Lavas are exposed primarily near the saddle between Mauna Kea and Mauna Loa. Numbers refer to the sequence of eruption (Porter, 1973) with #13 representing the most recent lavas.

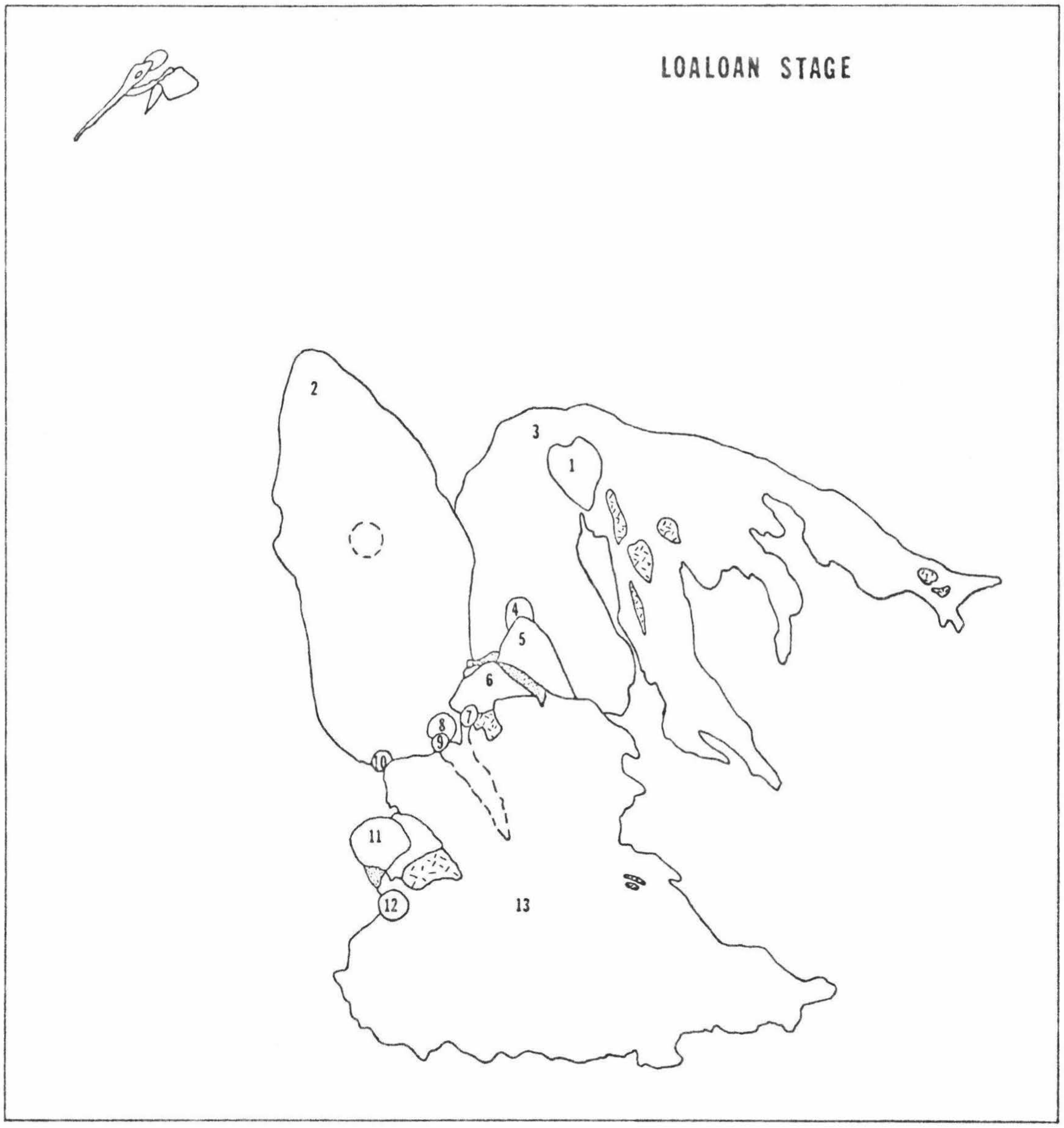

FIGURE 32

Figures 33-37: Oxide and trace element variation plotted against position along the Loaloan Stage rift from northeast to southwest (left to right). Thus, age also decreases from left to right. Note the systematic increase in Sr with position (Figure 37).

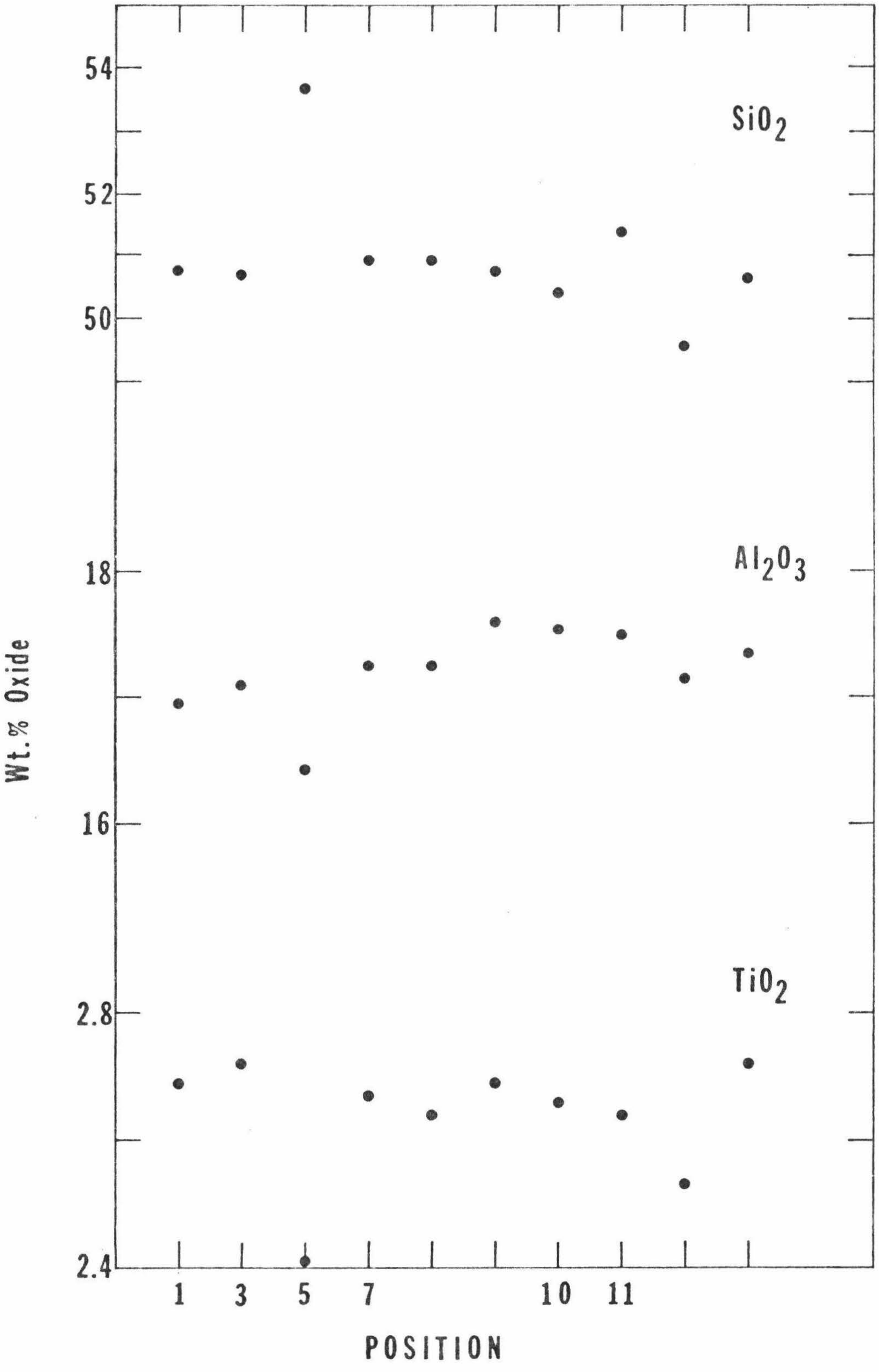
54
52
50
18
16
2.8
2.4
SiO_2
Al_2O_3
TiO_2
Wt.% Oxide
1
3
5
7
10
11
POSITION

FIGURE 33

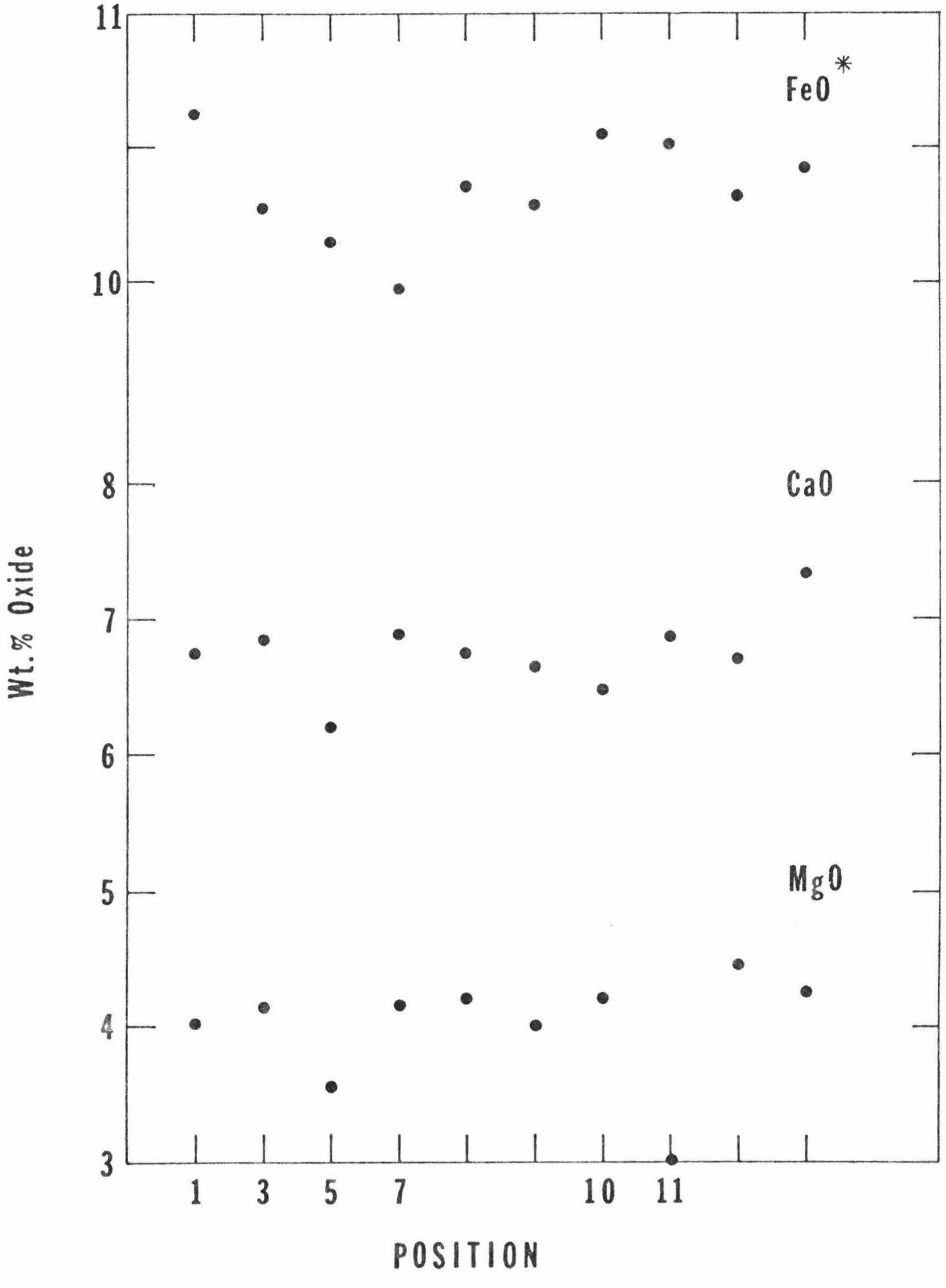

FIGURE 34

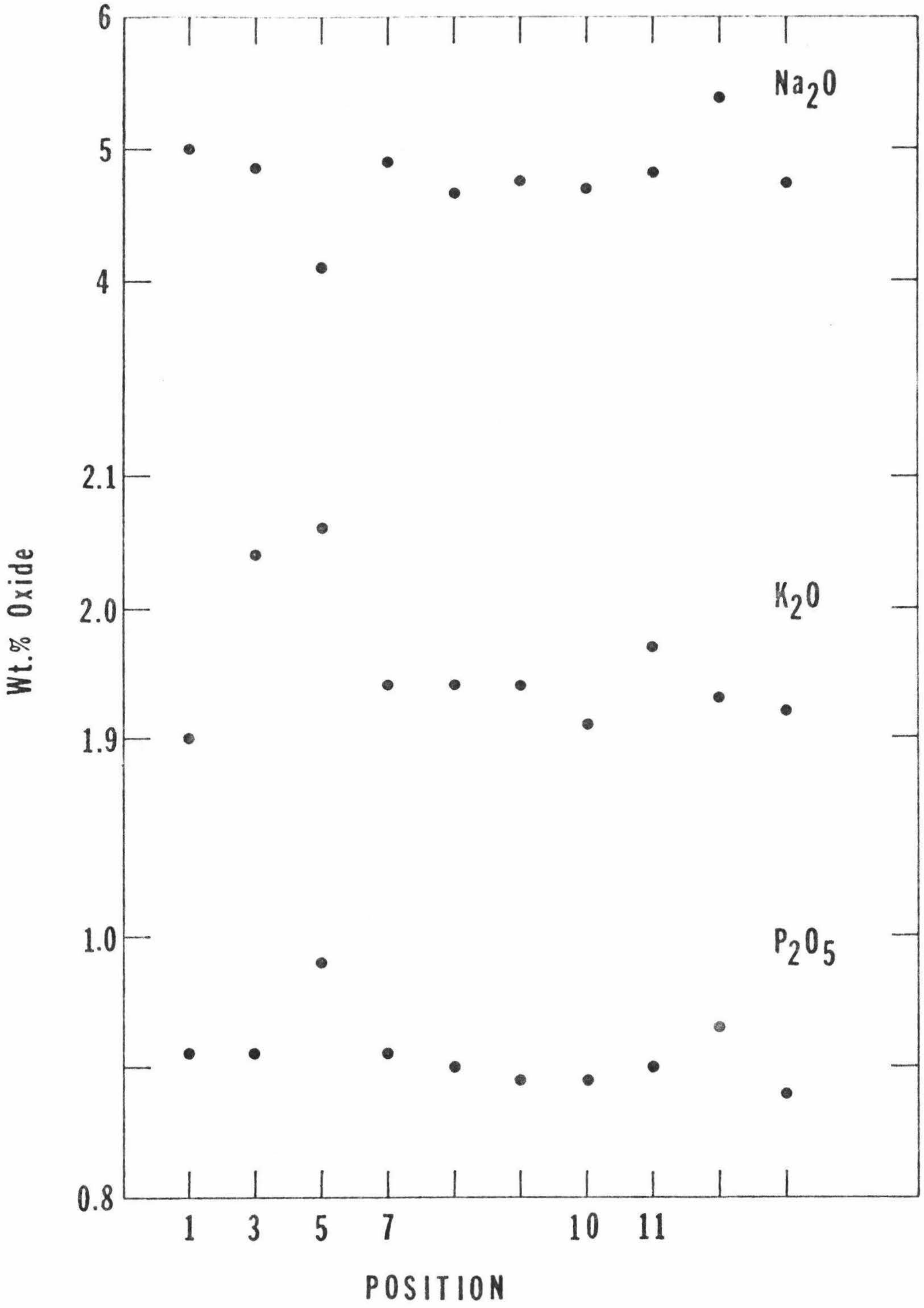

FIGURE 35

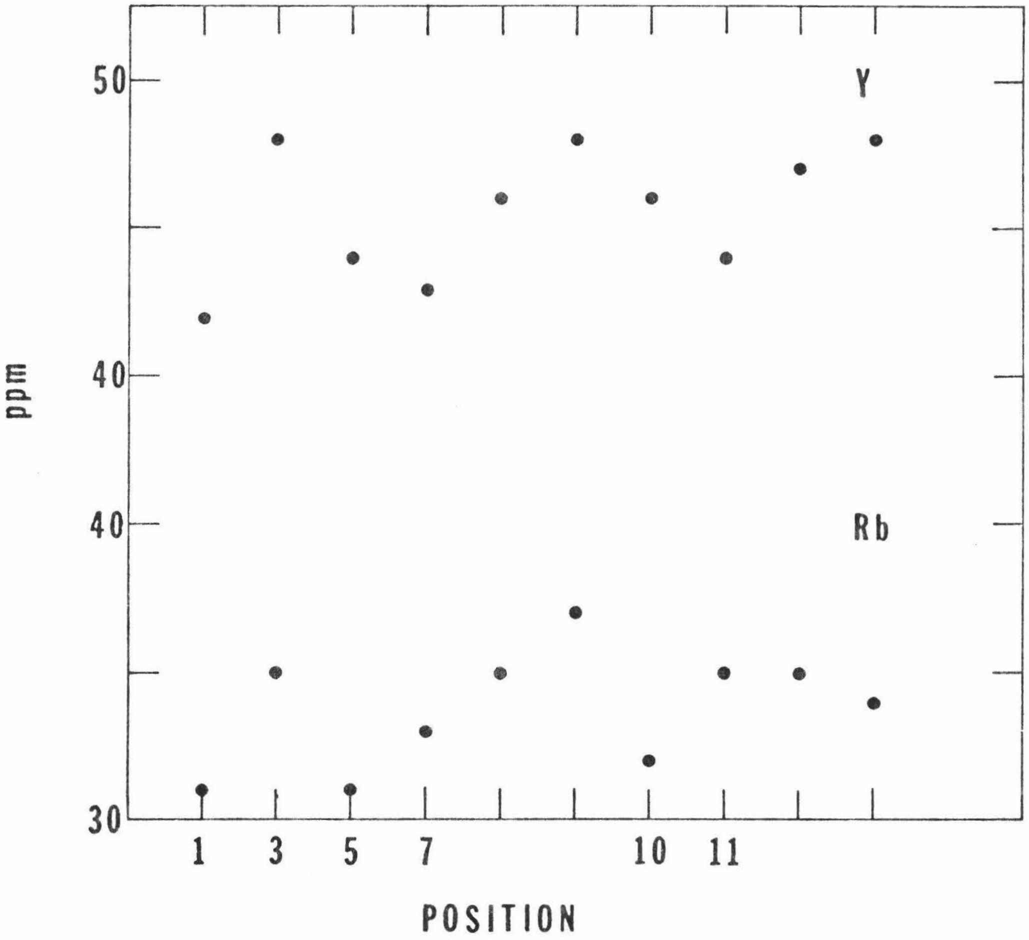

FIGURE 36

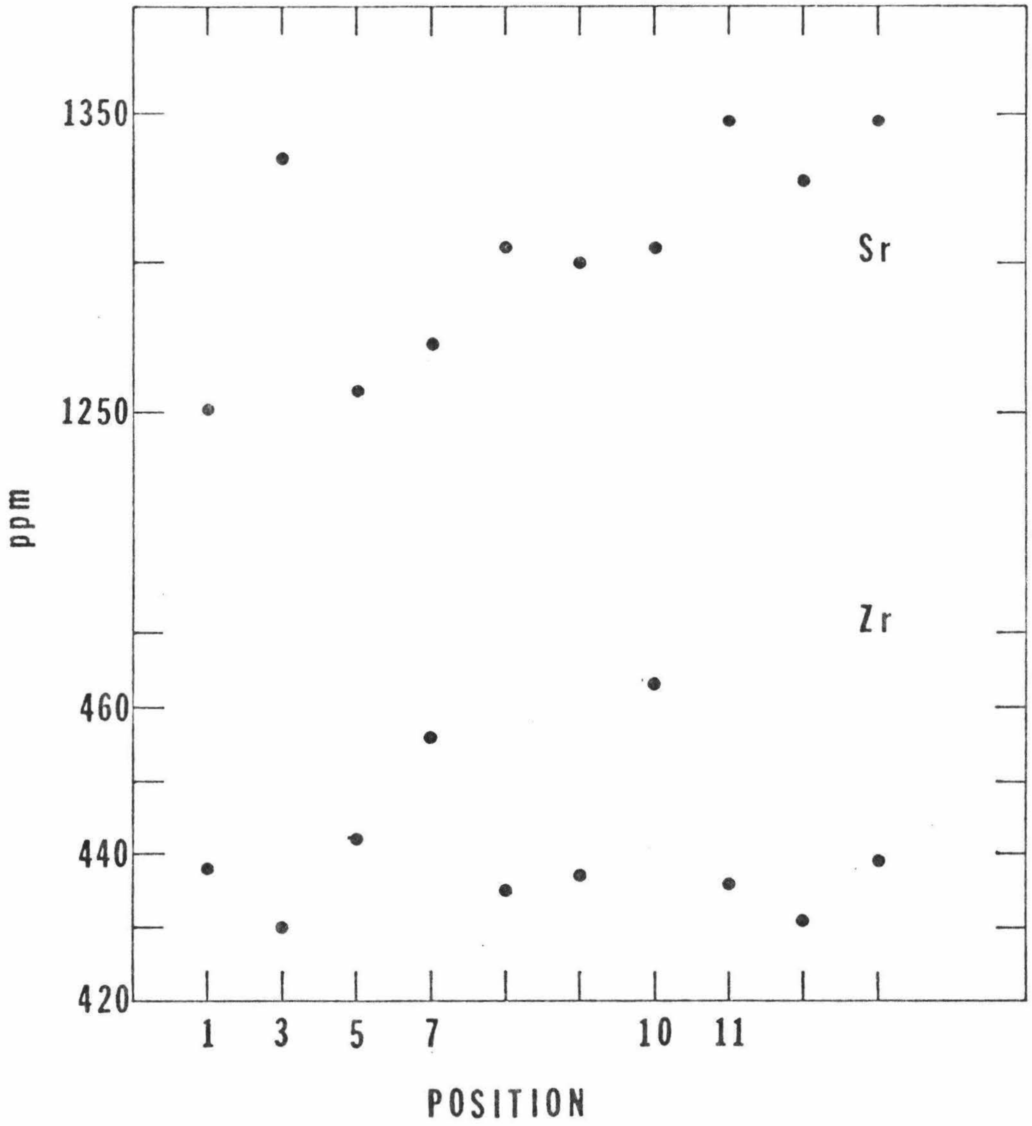

FIGURE 37

Figures 38-42: Oxide and trace element variation plotted against distance from the summit (0-14 km.).

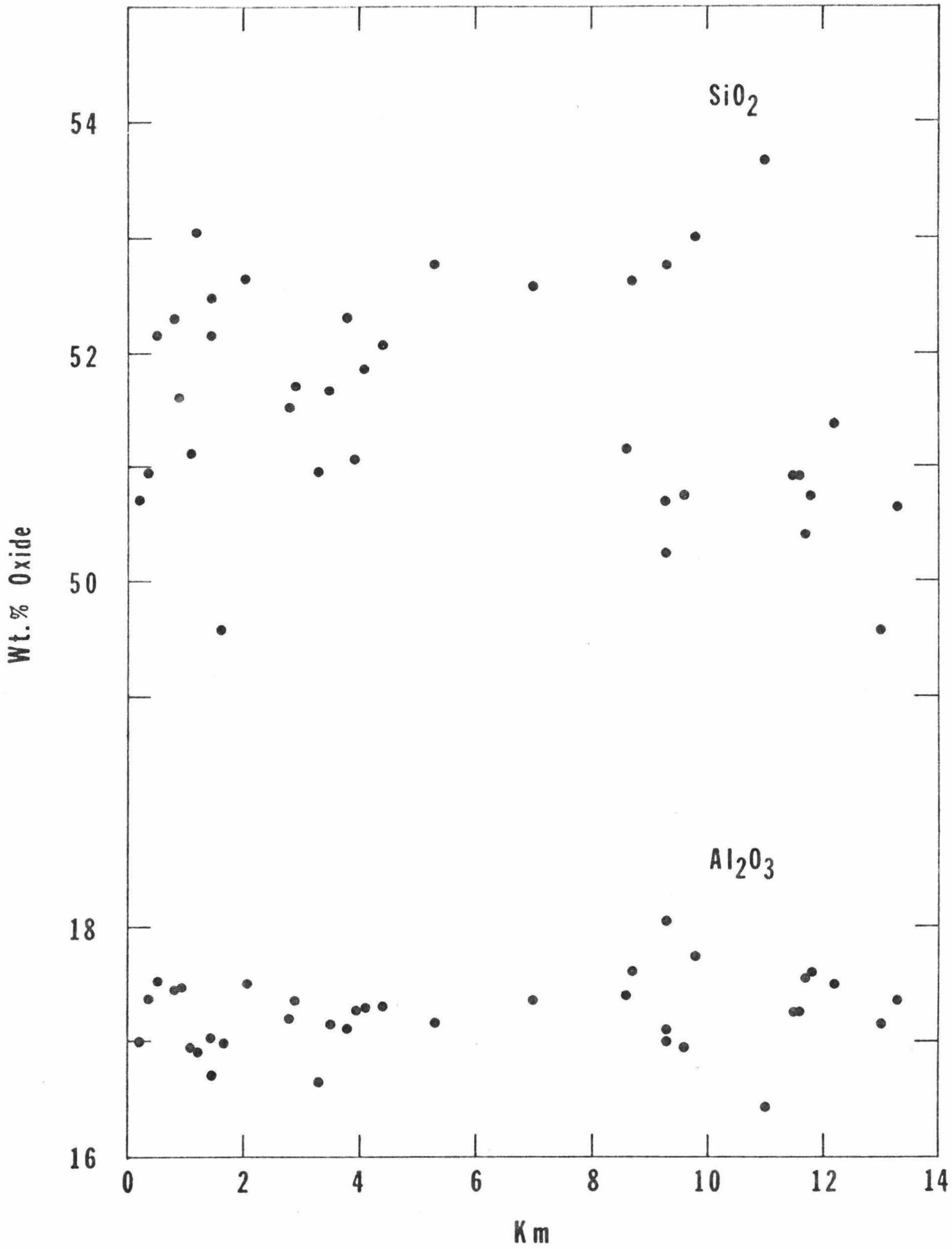

FIGURE 38

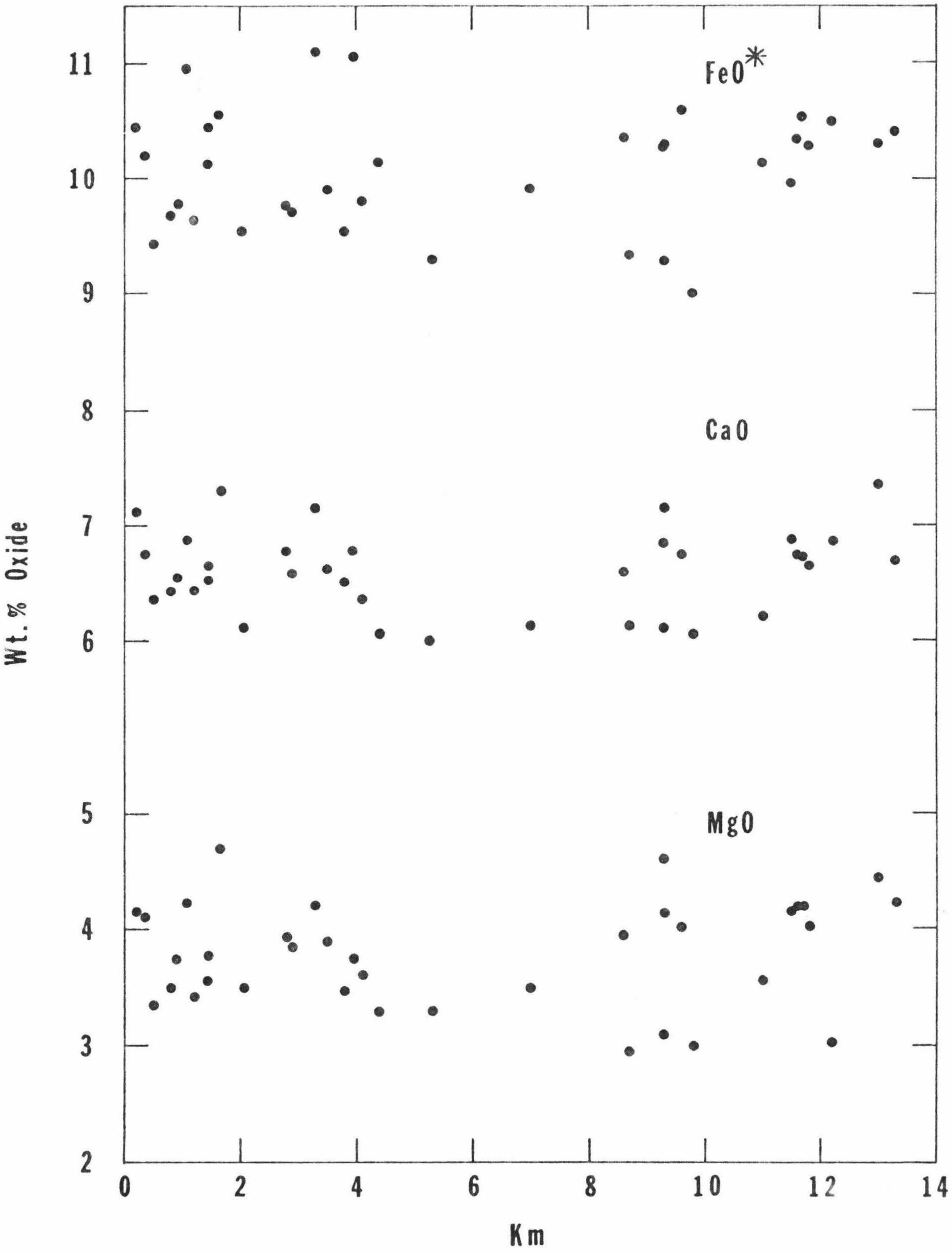

FIGURE 39

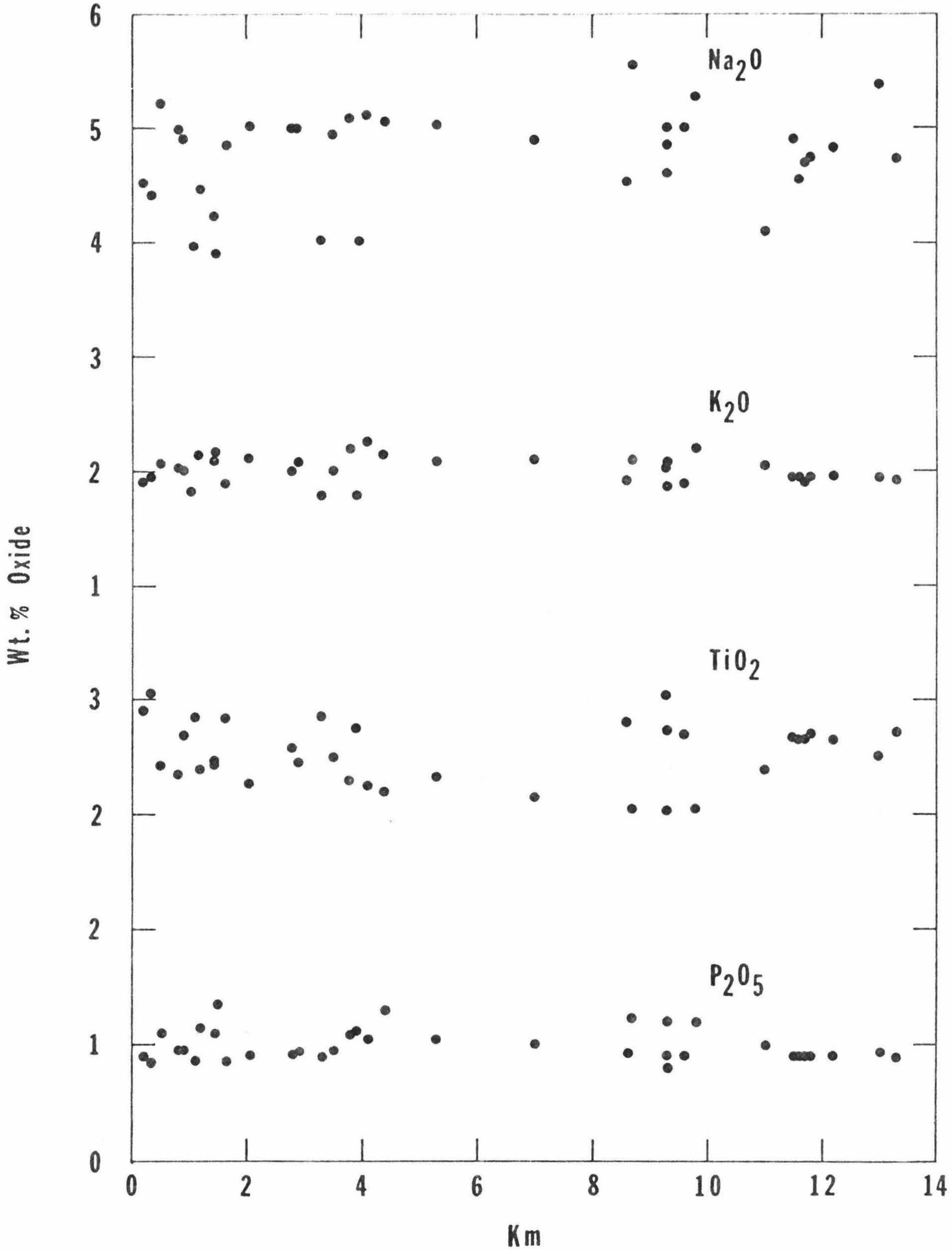

FIGURE 40

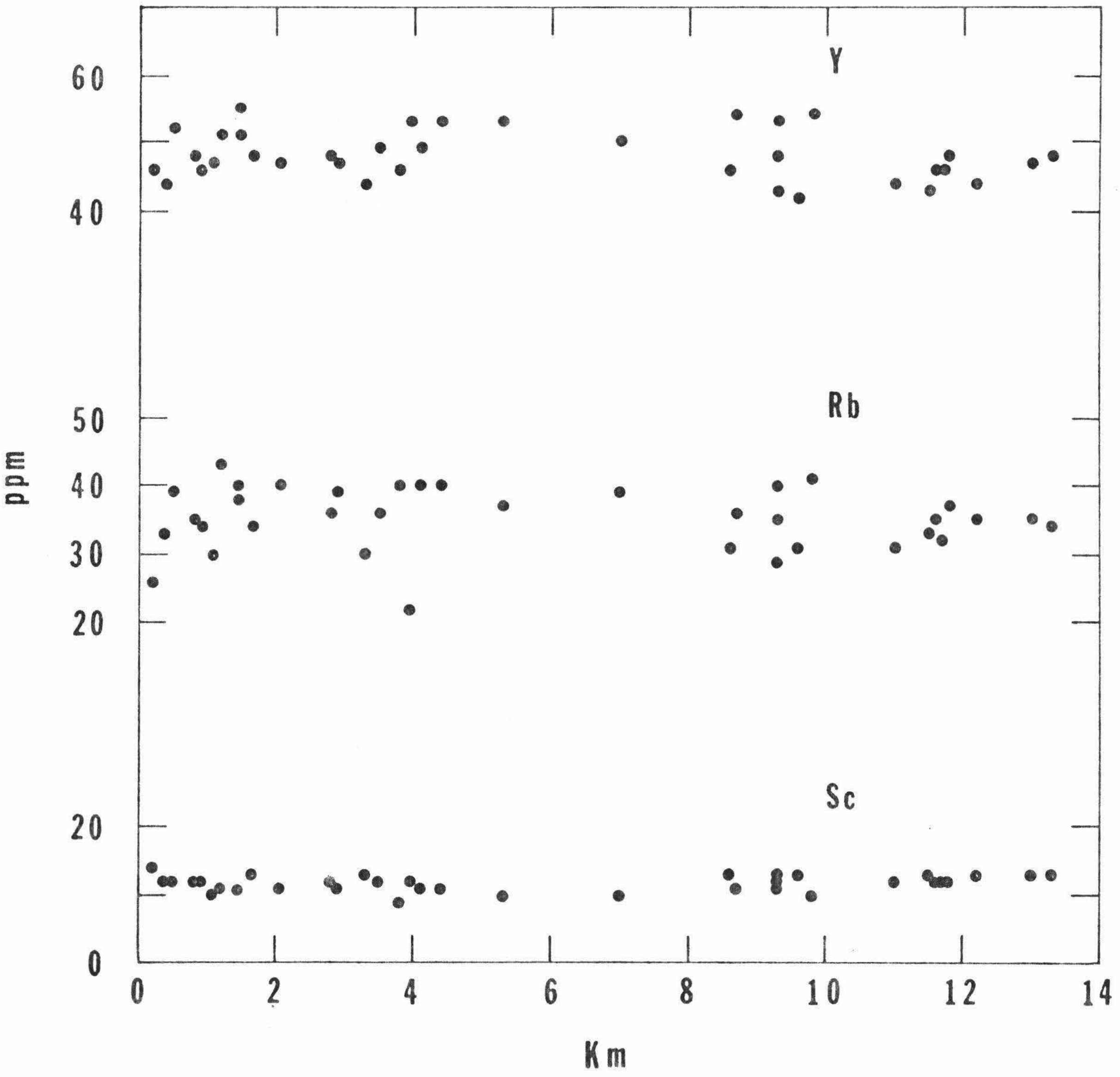

FIGURE 41

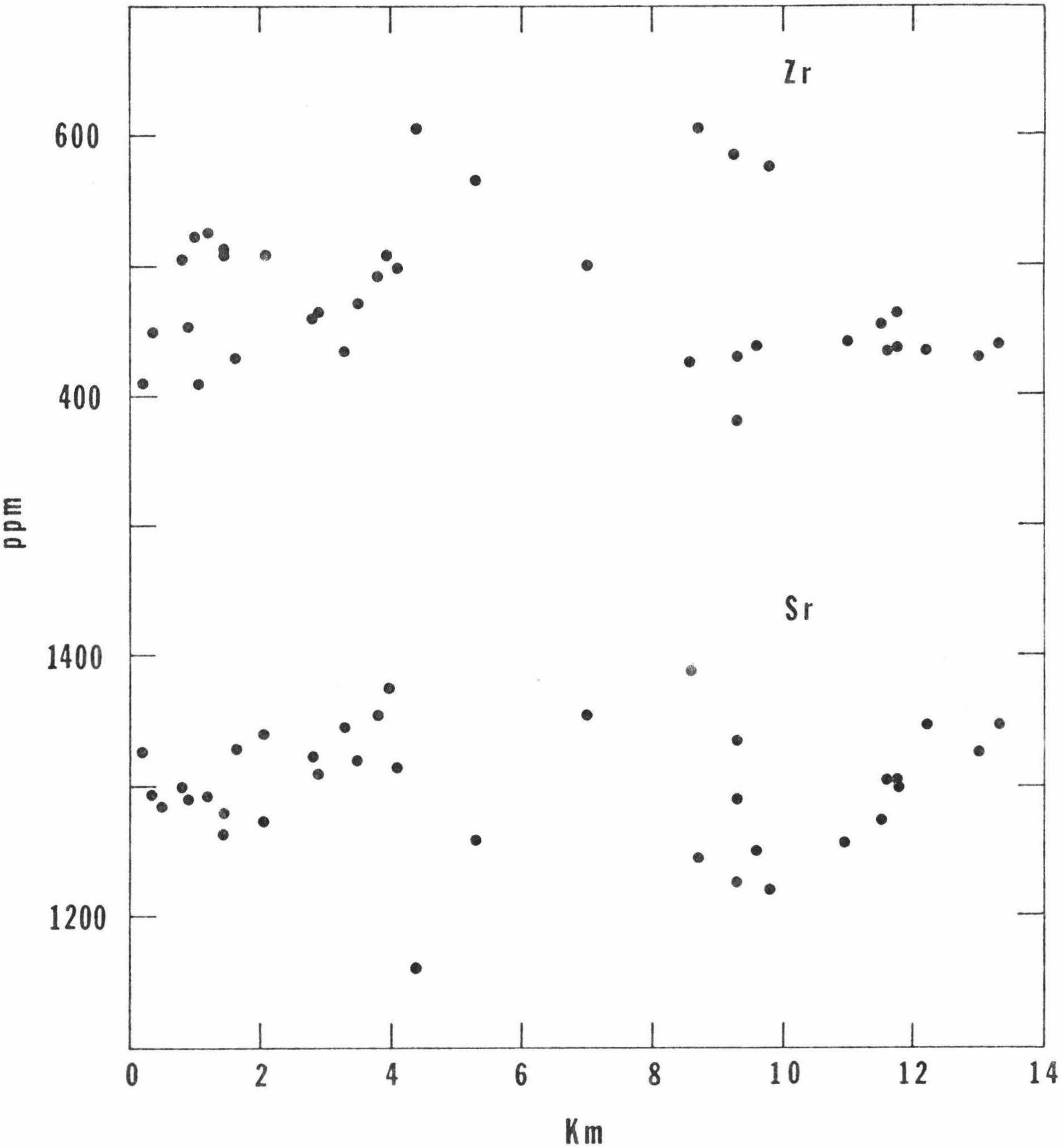

FIGURE 42

Figure 43: Ne-Ab-Fo-Q ternary diagram. Symbols same as in Figure 5. Note the separation of Hamakua and Laupahoehoe group lavas.

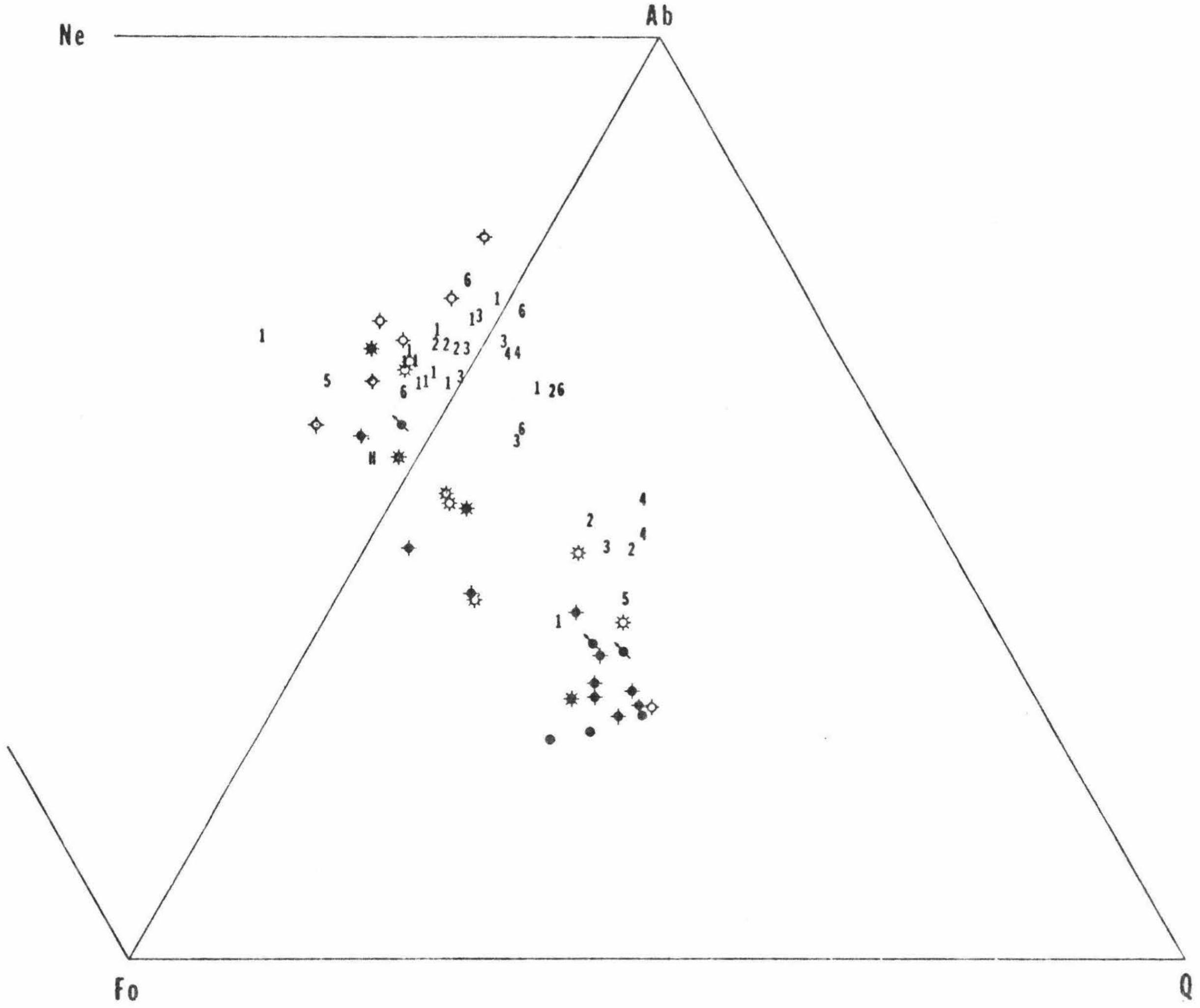

FIGURE 43

Figures 44-45: D.I. versus normative nepheline and normative hypersthene diagrams. Symbols same as in Figure 6, except the following: hourglass = Hamakua Group lavas (Macdonald and Katsura, 1964; Macdonald, 1968), asterisk = Hamakua Coast lavas from the Hamakua Group (Garcia, unpub.), picnic table = Laupahoehoe Group lavas (Macdonald and Katsura, 1964; Macdonald, 1968), Y = lavas from the west slope of Mauna Kea (Wise, unpub.), Z = lavas from both groups collected from the Hamakua Coast (Murata, unpub.). Note the distinct separation of Hamakua and Laupahoehoe group lavas (Laupahoehoe Group lavas lie on the upper trend).

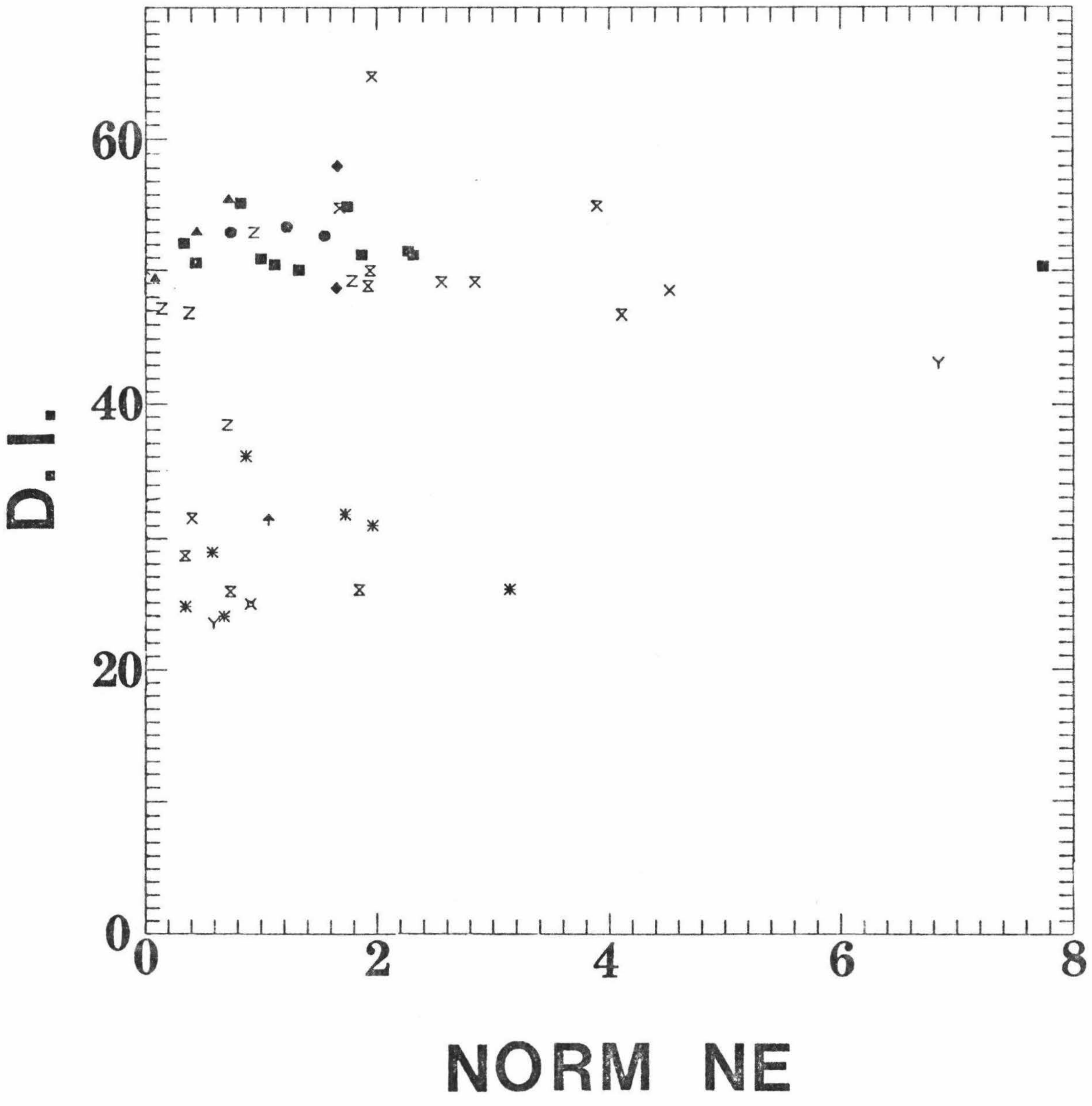

FIGURE 44

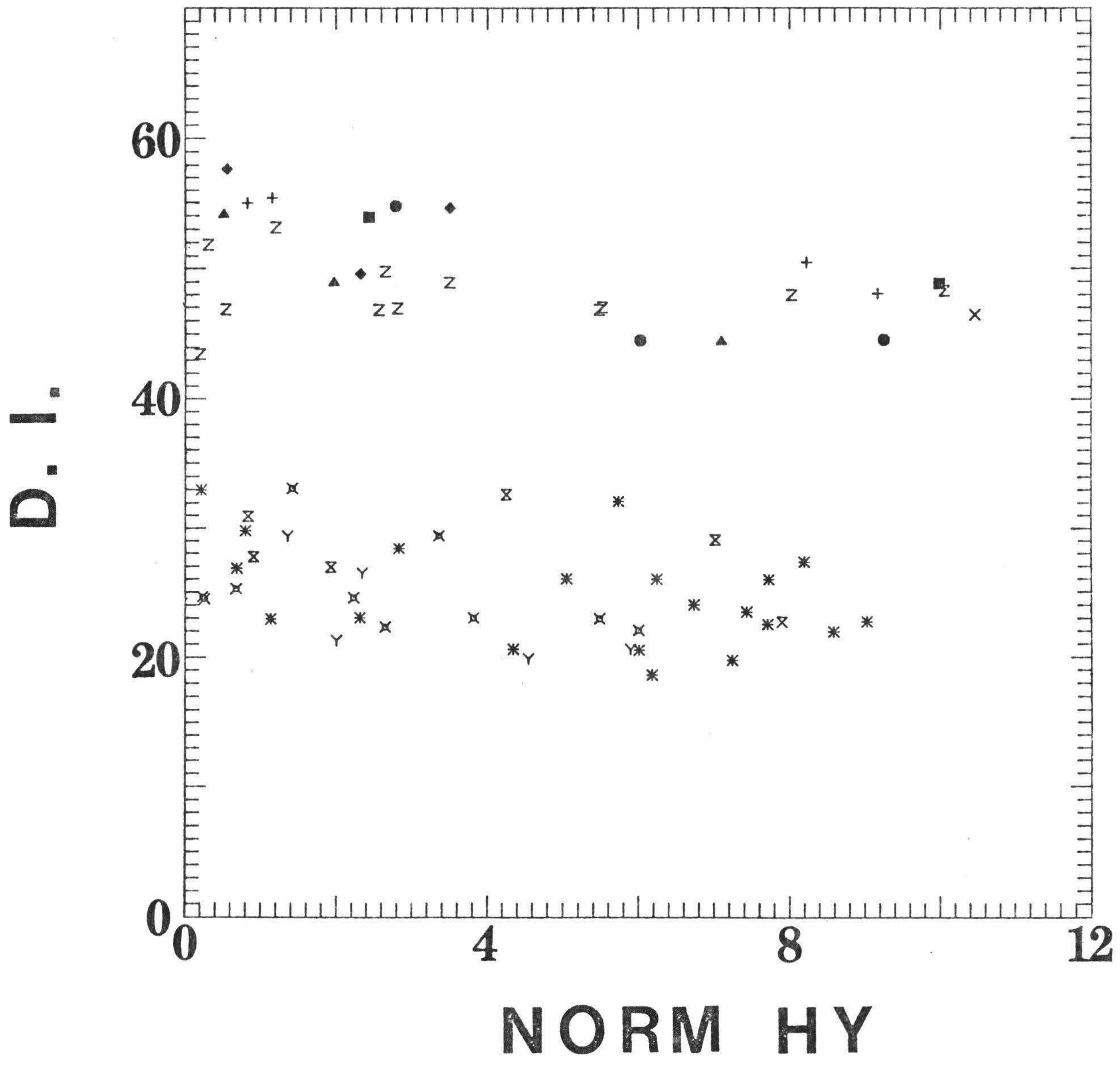

FIGURE 45

Figure 46: D.I. versus % An diagram for Laupahoehoe and Hamakua group lavas. Symbols same as in Figures 44 and 45. Note the slightly different slopes exhibited by the two groups (Laupahoehoe Group lavas lie on the upper trend) and the "kink" which separates them. Such a kink is inconsistent with the two groups being related by fractional crystallization.

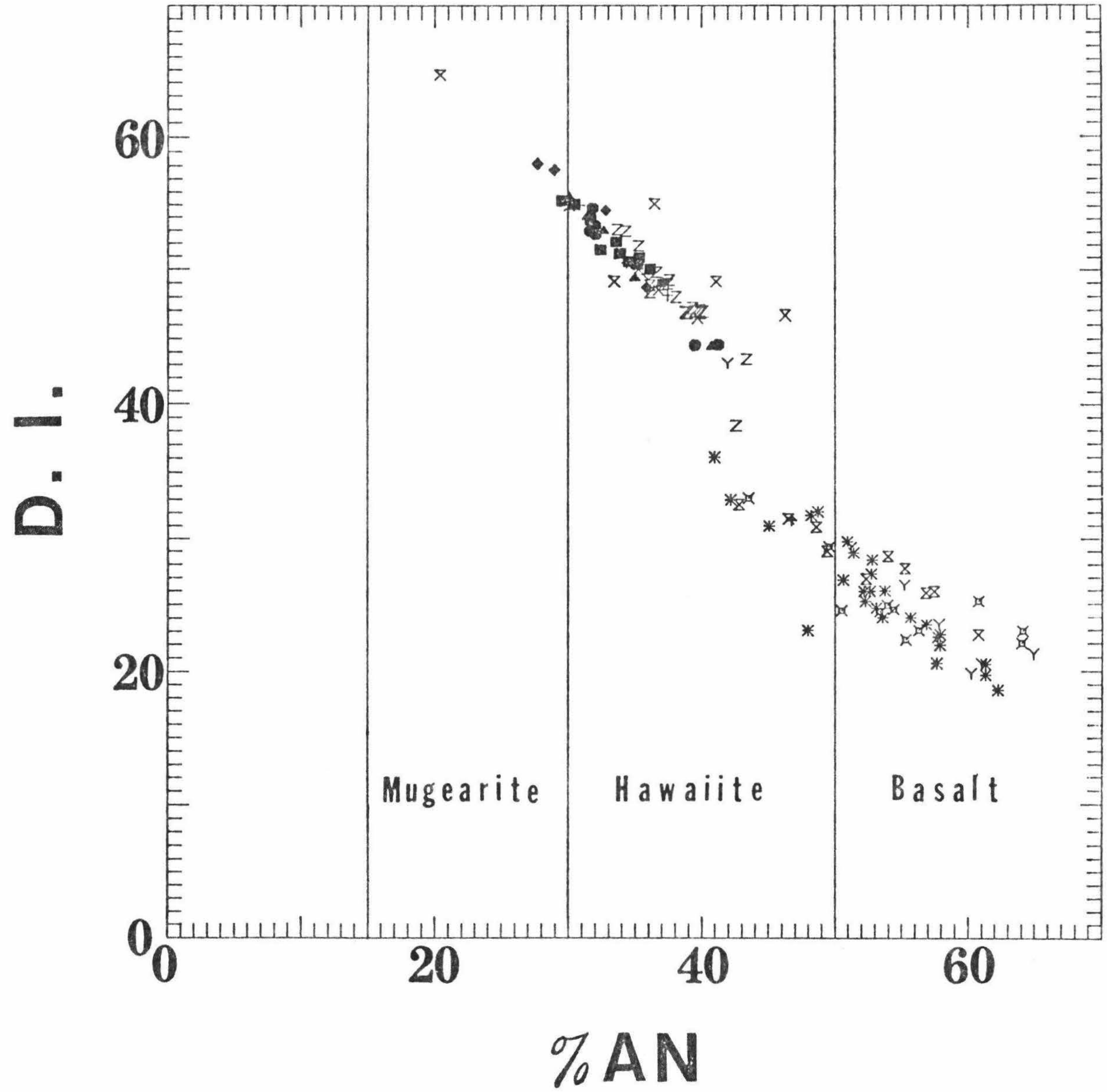

FIGURE 46

APPENDIX 1

Summary of Methods Used in the Analytical Laboratory Department of Earth Sciences, University of Manitoba*

Element	Method
Si Al Fe (Total) Mg (High) Ca K Ti Mn Zr	X-ray Fluorescence Spectrometry. Sample plus $Li_2B_4O_7$ and La_2O_3 are heated in a graphite crucible at about 1100 degrees celsius for one half hour. Resulting glass bead with H_3BO_3 (total weight, 2.1000 grams) ground to -200 mesh and then compressed to 50,000 p.s.i. Elements then simultaneously analyzed on multi-channel ARL X-ray Spectrometer.
Na_2O, MgO (Low)	Atomic Absorption Spectrophotometry. Rock dissolved with HF, H_2SO_4, and HNO_3 in platinum crucibles. Perkin-Elmer 303 A.A.S. used for determinations.
P_2O_5	Colorimetry. Solution as for Na_2O above. The absorption at 430 m of molybdivanadophosphoric acid complex. Unicam sp 500 spectrophotometer.
FeO	Rock decomposed with HF and 1:4 H_2SO_4 solution titrated with $K_2Cr_2O_7$ using Sodium Diphenylamine sulfonate as indicator.

APPENDIX 1 (continued)

Summary of Methods Used in the Analytical Laboratory
Department of Earth Sciences, University of Manitoba*

Element	Method
H_2O^-	Determined by heating sample to constant weight at 1100 degrees celsius.
H_2O (Total)	Determined by heating sample in a stream of dry oxygen in an induction furnace (Temperature 1100 degrees celsius H_2O collected on Anhydrone and weighed.
H_2O^+	H_2O (Total) $-H_2O^-$.
CO_2	Sample decomposed by HCl and heat. Evolved CO_2 passed through drying train and collected on Ascarite.
CO_2 (Low S samples)	Determined simultaneously with H_2O (Total). CO_2 collected on Ascarite; small amounts of SO_2 removed on MnO (act).

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*written communication by K. Ramlal (1979)

APPENDIX 1 (continued)

Summary of Analytical Methods

Constituents, Concentration, Precision, and Accurary
of Major Element Determinations,
Department of Earth Sciences, University of Manitoba*

Constituent	Concentration Percent	Instrument Precision	Accuracy of Replicates
SiO_2	59.60	.12	.20
Al_2O_3	9.34	.05	.13
Fe_2O_3 (Total)	10.08	.017	.03
MgO	.404	.04	.10
CaO	10.22	.02	.07
K_2O	2.69	.01	.01
MnO	.41	.01	.01
TiO_2	.48	.02	.02
Na_2O	4.20	.01	.05
H_2O (Total)	1.60	.03	.06
CO_2	1.15	.05	.12
P_2O_5	0.20	.01	.01
FeO	10.92	--	.04

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*written communication by K. Ramlal (1979)

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