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TRACE ELEMENT STUDIES OF ROCKS AND SOILS FROM
OCEANUS PROCELLARUM AND MARE TRANQUILLITATIS

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ABSTRACT

Neutron-activation data on Zn, Ga, Ge, Cd, In and Ir are reported for six rocks and two soils from the Apollo-12 mission. A comparison of these data and similar data for Apollo-11 samples indicates extralunar components in the 12070 and 10084 soils of about 1.0 and 1.1% expressed in terms of an assumed composition which is the same as the water-free portion of C1 chondrites. A relationship between the integrated flux of extralunar material and the increase in concentration of such material in the fines portion of the lunar regolith is derived. According to this relationship, the integrated flux at the Apollo-11 site is 25% higher than that at the Apollo-12 site, which indicates that the influx of extralunar material has been decreasing with time. A reanalysis of the data on the Apollo-11 fines shows them to be consistent with a minor depletion in ilmenite and inclusion of a 20% component resembling Wood's anorthositic gabbro. Terrestrial basalts show relatively small variations in their contents of Zn, Ga, Ge, In and possibly Cd, independent of their geographical origin. Apollo-12 rocks have concentrations of Zn, Ge, Cd, In and possibly Ir which are lower by factors of 60 or more relative to terrestrial basalts. A mechanism is proposed for the late accretion of volatile-rich materials, including comets, in which a primitive terrestrial atmosphere is invoked to explain the significantly higher concentrations of such substances on the earth.

INTRODUCTION

Our previous work on Apollo-11 samples involved the determination of Ga, Ge, In, Ir and Au (Wasson and Baedeker, 1970). In the present study we have added Zn and Cd, which proved to have unexpectedly high concentrations in Apollo-11 fines and breccias, and we have dropped Au, which added little information in addition to that from our Ge and Ir determinations, and which was also measured by several other investigative teams.

Our six elements were chosen in order to provide a maximum of information about two different questions: (1) the nature of the extralunar component, and its usefulness in evaluating the formation of the lunar regolith; and (2) the distribution of elements of similar ionic properties in lunar crystalline rocks, and the interpretation of this data in terms of plausible geochemical and cosmochemical processes. Although our elements are not ideally suited for such questions, we have also attempted to assess (3) the nature of the extraneous lunar component in the soil.

Through the aid of the Manned Spacecraft Center curator and the Lunar Sample Analysis Planning Team, we have obtained a set of six rocks from the Apollo-12 mission which range from the Mg-rich, Al-poor to the Mg-poor, Al-rich extremes described in the LSPET (1970) report. In addition, we have investigated samples of two soils, the typical local (12070) and unusual light-gray (12033) fines.

Since a number of our geochemical interpretations depend on a comparison of data on lunar, terrestrial and meteoritic samples, we have followed the practice of Wasson and Baedeker (1970) and included the latter two types of samples in the same neutron activation runs as our lunar samples. In such a fashion we greatly reduce the magnitude of errors introduced in such comparisons. In this paper we have added an additional 14 terrestrial samples to the nine studied by Wasson and Baedeker (1970).

We have analyzed all samples in duplicate, in the belief that the resulting greater precision would reveal effects which would otherwise have been missed.

EXPERIMENTAL

Samples and Sample Preparation

Approximately one gram each of rock samples 12009, 12014 and 12038, and soil samples 12033 and 12070 were provided for analysis. Two 500-mg aliquots of a powder prepared by grinding a 10 g piece of rock 12063 were analyzed in this work, the remainder of the rock being used for Rb-Sr dating and lead isotope studies by Cliff et al. (1971). The procedures followed in preparing these samples for analysis were identical to those described in Wasson and Baedeker (1970).

In our studies on the Apollo-11 series of rock samples, large variations in the observed In content of the material which we analyzed indicated that the samples had been contaminated prior to their receipt in our laboratory, most probably by In-Ag seals on the rock boxes. Larger samples of rocks 12002 and 12022 weighing 3.5 and 3.2 g, respectively, were provided for our studies, in order that the surfaces of the rock chips could be removed prior to analysis, and thus minimize the possibility of surface contamination. An S.S. White airbrasive apparatus using 50- μ SiC as the abrasive powder was used for removing surface material. An analysis of the abrasive powder for the trace elements of interest showed concentrations of 0.48, 0.14, 0.00016, 0.00033, 1.89 and 0.027 ppm for Ga, Ge, In, Ir, Zn, and Cd respectively. Ge and Cd were present in higher concentrations in the SiC powder than was expected in the lunar material, Zn and Ir at approximately the same level of concentration, and Ga and In were significantly lower. If the lunar samples were contaminated by an 0.1% addition of SiC, this could cause as much as a 1% error in the results for Ge and Cd. We believe that the actual contamination is smaller. Approximately 280 mg of rock 12022 was abraded from the 3.2 g chip, and the sample washed in reagent-grade acetone using a sonic cleaner prior to crushing and splitting. Rock sample 12002 was split into one large fragment weighing 1.83 g and several smaller fragments. Two powdered samples of 500 mg each were prepared by crushing some of the uncleaned smaller fragments. The 1.83 g chip was treated with the same sandblasting

procedure as rock 12022, 480 mg being removed in the process. The remainder of the sample was crushed, and chips totaling 550 mg in weight were powdered and analyzed. As will be discussed in more detail later, our attempts to remove contamination were moderately successful.

Our sample of the light-gray fines, 12033, was packaged at the Lunar Receiving Laboratory in a stainless steel bolt-top container. Upon opening the container, the sample was found to be contaminated with metal turnings from the threads of the bolt-top. The fines were sieved through 176 μ screen to remove the turnings, and only the -176μ fraction was analyzed. The sieved fines were examined under a microscope; no metal turnings could be detected.

Sample packaging, flux monitor preparation and irradiation procedures were identical to those described in Wasson and Baedeker (1970). During one of our irradiations an experiment was performed to show whether systematic errors might arise from possible volatilization of the deposited salts from the flux-monitor foils as a result of heating during the irradiation. Each flux-monitor foil was wrapped with a second piece of high-purity aluminum foil; following irradiation the outer foils were processed in the same manner as the flux monitors. An irradiated blank sample of high-purity Al was also processed. The same level of activity for all six radionuclides was observed in the cover foils as was found in the blank. This indicates that any loss of activity from the flux monitor foils due to volatilization is negligible. Although one might argue that volatilized

monitor compounds would not necessarily recondense on the second foil during the irradiation, we believe that this is where the largest fraction should be found when any vapor condenses upon cooling at the end of the irradiation.

Analytical and Radiometric Procedures

The six elements Zn, Ga, Ge, Cd, In and Ir were determined simultaneously in each rock sample. Since description of the procedures used in the determination of Ga, Ge, In and Ir was provided in Wasson and Baedeker (1970), only our procedures for Zn and Cd will be discussed here.

Aliquots of carrier solutions of Zn and Cd (as well as the other four elements determined) were placed in zirconium crucibles, made basic with NH_4OH and evaporated to dryness. Following sample dissolution and chemical processing to extract In, Ga and Ge, the major amounts of Zn, Cd and Ir were recovered by passing an aqueous solution through a Dowex-1 anion-exchange column which completely removed the Ir. Zn and Cd were partially retained by the column and eluted with 6 N HNO_3 prior to the usual batch elution of Ir. The Zn- and Cd-containing eluate was mixed with the aqueous supernatant (processed for Au in our earlier procedure) from the first hydroxide precipitation (the supernatant contained minor amounts of Zn and Cd). This solution was evaporated to dryness, the resulting salt cake dissolved in H_2O , Cr and Sc carriers added, and Cr and Sc precipitated as hydroxides from a strongly ammoniacal

solution. The supernatant was boiled to remove excess NH_3 , the solution made slightly acidic with HCl , saturated with H_2S and the Zn and Cd sulfides precipitated by bringing the pH to about 8 with NaOH . The resulting precipitate was collected by centrifugation, dissolved in concentrated HCl , and the resulting solution evaporated to dryness. The residue was dissolved in an 0.12 M HCl solution which was 1.7 M in NaCl , and the solution passed through a Dowex-1 anion-exchange column. Following a procedure outlined by Kallmann et al. (1956) but modified slightly for our purposes, Zn was selectively removed from the column by eluting first with a solution 2 M in NaOH and 0.34 M in NaCl , and later with H_2O . The eluate was treated with H_2S to precipitate Zn as the sulfide. This precipitate was collected by centrifugation and dissolved in HCl . The resulting solution was counted and the chemical yield determined by atomic absorption.

The Cd was eluted from the ion-exchange column with 1 M HNO_3 , the eluate boiled to dryness, and the residue dissolved in 3 N HCl . Copper and Cr carriers were added, Cu precipitated as the sulfide with H_2S , and the precipitate removed by centrifugation and filtration. The clear solution was again saturated with H_2S , and CdS precipitated by lowering the acid concentration to 0.5 M by addition of NaOH . The precipitate was collected by centrifugation, dissolved in 3 N HCl , and the precipitation cycle repeated. The final CdS precipitate was washed twice with H_2O , dissolved in HCl , and the resulting solution counted. The chemical yield was measured by atomic absorption.

Gamma counting, using a 3 x 3 in NaI well detector was used to measure the Zn and Cd activities (the 0.335-MeV photopeak of the ^{115}In daughter of ^{115}Cd and the 1.114-MeV photopeak of ^{65}Zn).

Precision and Accuracy

In order to test the precision of our procedures we have analyzed several replicates of standard rocks BR and W-1. Based on the replicate analyses reported for W-1 in Table 2 and by Wasson and Baedeker (1970) we calculated relative standard deviations of ± 3.5 , 3.4, 20, 8, 7 and 25 respectively for the elements Zn, Ga, Ge, Cd, In and Ir. The precision for Zn and Ir was poorer by a factor of 2 in our analyses of BR, based on fewer determinations. The precision of our results for Ge appear to be poorer for these standard rocks than in most other terrestrial and lunar basalts we have investigated. We have analyzed 21 replicates of the Waianae, Hawaii basalt. If we assume that these all have the same Ge content, we can calculate a relative standard deviation of 7%, which should be an upper limit on the true value for our procedure. Ninety-five percent confidence limits on the means of duplicate determinations should be about 1.6 times greater than the above standard deviations.

The accuracy of our results can best be assessed by comparing them with the data obtained by other workers on the same rocks. A comparison of our data on the USGS series of standard

rocks has indicated no systematic differences for Ga, Ge, In and Ir which would point to systematic errors in our results. Our mean Zn value is about 10% higher than the average of some 20 analyses of W-1, which may indicate a systematic error in our Zn results. A check of our Zn flux monitor solution failed to explain this error. To date no reliable mean Cd value for W-1 is available for comparison.

RESULTS

The results of our analyses on the Apollo-12 returned lunar samples are presented in Table 1. Previously unpublished results for Ir in Apollo-11 rocks are included in an appendix to this paper.

In general the problem with In contamination encountered in the Apollo-11 samples appears to be less pervasive on the Apollo-12 samples. The In concentrations observed on the samples of 12022 and 12002 which were treated by the sandblasting procedure described above, and the untreated specimens of 12009, 12038 and 12063 appear to indicate an In concentration of 0.5 - 2 ppb in lunar crystalline rocks, slightly lower than the lower limit of 3 ppb obtained from the Apollo-11 data. However, rock 12014, the untreated specimen of 12002, and the soil samples 12033 and 12070 have very high In contents, which we believe indicates In contamination.

The amounts of Ge, Cd, and In detected in lunar crystalline rocks are near the limits of detection for our method.

Our Ge results for the lunar crystalline rocks are reported as upper limits. We are at present unable to account for the poor reproducibility of our Cd results. The high results reported in samples 12009, 12063, 12002, 12022 and 12070 are believed to be erroneous. The precision of our results for terrestrial rocks is considerably better. Because of the high U and low Cd abundances in lunar rocks, corrections for fission-produced ^{115}Cd have been applied to our data using the U values of Brunfelt et al. (1971), Cliff et al. (1971), Morrison et al. (1971), O'Kelley et al. (1971), Rancitelli et al. (1971), Rosholt and Tatsumoto (1971), Silver (1971) and Taylor et al. (1971). For rock 12014 no U data are as yet available, therefore no corrections have been carried out.

Our previous Ga results reported in Wasson and Baedeker (1970) should be increased by a factor of 1.05 to provide agreement with our latest Ga results, due to a recalibration of our monitor solution. Our most recent Ga results appear to be too high by 5 - 10 % when compared with these corrected results. We have applied a uniform correction factor of 0.95 to these values, and are searching for a discrepancy.

There appear to be some systematic differences between the Apollo-12 crystalline rocks and those recovered from Mare Tranquillitatis. A comparison of our data and that of Ganapathy et al. (1970b) with that of Ganapathy et al. (1970a) for the Apollo-11 rocks indicates that both Zn and Cd appear to be less abundant in the Apollo-12 samples. Whereas Ga was observed to be nearly constant in abundance in the Apollo-11 basalts (Wasson and Baedeker, 1970), a much wider range of Ga concentrations was found in the Apollo-12

rocks. This is consistent with the wider range in major element composition observed by LSPET (1970).

Data on a number of terrestrial rock samples which were analyzed simultaneously with the lunar rocks is presented in Table 2. We have extended the work reported by Wasson and Baedeker (1970) on the Waianae volcano to include three additional specimens from the lower member of this volcano, and have reanalyzed the three specimens from the lower member which were included in our previous study.

In addition to our work on the Waianae volcano, we have also analyzed 12 rock samples from the Skaergaard, Greenland intrusion. The results of our analyses on the chilled margin gabbro (which is believed to represent the parent liquid) are presented in Table 2. A detailed discussion and presentation of our data on the Skaergaard will be published elsewhere.

Two samples of peridotite were analyzed, PCC-1 and a rock from Salt Lake, Hawaii. Both rock samples were poorer in Zn, Ga, Ge, Cd and In than all basalt samples analyzed, and Ir was found to be markedly enriched in both rocks. Ga, Cd and In were found to be higher by a factor of 2 in the Salt Lake sample as compared to PCC-1.

The Brown Point pyroxene gabbro is a Ti-rich facies belonging to a mass of coarse gabbroic anorthosite from the Adirondack Mountains. The abundances of all elements are similar to those in other terrestrial igneous rocks, with the exception of Ga, which is lower.

THE EXTRALUNAR COMPONENT

The moon's cratered surface shows ample evidence of accretion of extralunar materials, and it is to be expected that evidence of this extralunar component should be found in the finely comminuted portion of the lunar regolith. Evidence for such a component in Apollo-11 soils and breccias was reported by the LSPET (1969) team, and later by several other investigators. The most comprehensive and conclusive study of Apollo-11 samples is that of Ganapathy et al. (1970a) who report that 12 elements are clearly enriched in the fines and breccias relative to the local rocks. They argued that the extra amounts of these siderophile and volatile elements were consistent with a 1.9% addition of material resembling C1 chondrites to a matrix composed of a roughly equal mixture of the two local rock types at the landing site.

We have studied two soils (but no breccias) from the Apollo-12 site. Sample 12033 is light-gray soil, and contains substantially smaller amounts of siderophilic and volatile elements than does 12070. The 12070 soil is similar in composition (Ganapathy et al., 1970b) and color (LSPET, 1970) to most of the fine material in the 12028 core, and is probably the more typical and older soil at the Apollo-12 landing site. We have attempted to define the magnitude of the extralunar component in 12070 and in the more typical portions of 12028 by a method similar to that used by Ganapathy et al. (1970a) but differing in important details. Like these authors, we have estimated

the concentrations of the elements of interest in the "lunar" matrix of the fines by taking an average of the available data on local rocks, and have subtracted these values from the average concentrations found in the 12070 and 12028 fines. Such data for 11 elements are summarized in Table 3. We have followed the practice of Wasson and Baedeker (1970) and normalized the results by dividing by C1 chondrite concentrations on a water-free basis. This has the important dividend that the resulting values are better estimates of the fraction of C1-like material actually present in the fines. In contrast to Ganapathy et al. (1970a) we list estimated errors in the resulting values (which mainly depend on the scatter in the data on the fines, since in most cases the additional contribution from the matrix is negligible), and have shown these errors as bars in Fig. 1a. This facilitates the assessment of the correctness of the assumption that the extralunar material closely resembles the composition of C1 chondrites.

We have also reevaluated the Apollo-11 data on fines and breccias in the manner described above. These are summarized in Table 3 and Figs. 1b and 1c.

As can be seen from the last three lines of Table 3, Ge and Ir are the first and second most precisely determined of the 11 elements listed. Interestingly enough, the magnitudes of the extralunar component estimated on the basis of Ir are about 1.6 times greater than those estimated on the basis of the Ge data. We have drawn bands across Figs. 1a, 1b and 1c which are just wide enough to include the Ge and Ir points.

If we take the midpoints of these bands to be the best estimates of the magnitudes of the extralunar component, and the widths of the bands as estimates of the errors, we arrive at values of 1.04 ± 0.24 , 1.14 ± 0.25 and $1.07 \pm 0.25\%$ for the Apollo-12 and 11 fines and the Apollo-11 breccias, respectively. The latter must be taken cum grano salis, since the breccias have not all originated in the same localized area of time and space.

The error bars of the other nine elements touch the defined bands, with the following exceptions: Cd in 12D; Zn and Cd in 11D and 11C. The significance of these exceptions will be discussed later. The remainder of the data support the conclusion of Ganapathy et al. (1970a) that the extralunar component has a mean composition similar to that of C1, C2 or E3-4 chondrites, and distinctly different from that expected from any other known class of meteorites. The agreement is not surprising, since C1 chondrites are believed to most nearly resemble the non-volatile portion of mean solar-system material (see Anders, 1971), and are probably the most abundant form of meteoritic material falling on the earth (Shoemaker and Lowery, 1967) or the moon. The composition of interplanetary dust is not known, but it is reasonable to assume that that portion which is produced by cometary attrition is similar in composition to the C1 chondrites.

That the observed Ir/Ge ratio in the extralunar component is 1.6 times greater than that found in C1 chondrites is an interesting and possibly important point. The errors we give in the last three lines of Table 3 are meant to be one standard

deviation, and do not include an estimate of the error in the C1 chondrite concentrations. Using the errors as given, and assuming errors of about 10% and 5% in the C1 chondrite concentrations of Ir and Ge respectively, we estimate a probability of less than 0.1 that the apparent difference in the Ir/Ge ratios between the extralunar component of the fines and the C1 chondrites reflects experimental error*. Thus it is fairly likely

*The concentration of Ir in C1 chondrites is an arithmetic average of one analysis of Ivuna and two analyses of Orgeuil reported by Ehmann et al. (1970). Atomic abundances (relative to Si) based on these data are very similar to those found by the same group for C2 and H chondrites. The data are based on neutron activation analysis. The data of Crockett et al. (1967), which are based on radiochemical neutron activation, show Ir abundances in C1 chondrites which are only about 60 % as large as those in C2 chondrites. It appears that the latter workers may have achieved incomplete exchange between carrier and the activated radionuclides, perhaps as a result of the highly reducing character of the C1 chondrites. Their data on the enstatite chondrites also appear to be too low, perhaps for similar reasons.

that the Ir/Ge ratio of the extralunar material is really higher than that of C1 chondrites. Germanium is one of the most volatile of the siderophile elements and the Ir/Ge ratio in the ordinary chondrites is about 4 times higher than the same ratio

for C1 chondrites. Also, this ratio varies by over five orders of magnitude in iron meteorites, and small additions of certain types of iron meteorite material to the soil could have a profound effect on the Ir/Ge ratio.

There is also an apparent tendency for volatile elements (other than Bi, to which we tend to assign less weight -- see the following paragraph) to be slightly higher in Figs. 1abc than would be predicted on the basis of the siderophilic-element contents of the soils and breccias. Concentrations of these elements are lower in "recrystallized" chondrites, and apparently in the bulk earth and bulk moon. If these "lost" volatiles condensed out in some colder portion of the solar system, it is possible that such a material enriched in these elements accounts for a measurable fraction of the extralunar component.

Although it may be ultimately possible to define a mean composition for the extralunar component which differs in some details from that of C1 chondrites, a number of other elements must be determined with precision comparable to those available for Ge and Ir, however, before detailed testing of hypothetical mixtures of different types of meteorites and/or other cosmic matter can be undertaken.

Anders et al. (1971) have made the novel suggestion that local variations (on a scale of tens or hundreds of meters) in the extralunar component can be found at the Apollo 11 and 12 sites. Such evidence, if correct, would be extremely important for evaluating the source of the extralunar component, since it would rule out a major contribution of interplanetary

dust to the extralunar component. Interplanetary dust, if present, should fall at comparable rates and have comparable compositions at all geographical locations on the moon. However, we can find no evidence in the tabulated data of Anders et al. (1971) to support their proposal, with the exception of their Bi concentrations. Their Bi values show large fluctuations both in soils and crystalline rocks, however, and we question whether an appreciable fraction of these variations may not result from experimental error.

As noted above, Cd is enriched in all soils and breccias by a factor of 3-5 relative to that expected on the basis of the estimated amounts of C1-like material; Zn is enriched by comparable factors in the Apollo-11 samples, but is not present in amounts in excess of that predicted in Apollo-12 samples. Ganapathy et al. (1970a) attribute the excess amounts of these elements in Apollo-11 samples to a hypothetical Cd- and Zn-rich lunar rock which was not included in the limited suite of rocks returned by the astronauts. Alternatively, one could attribute at least a portion of this material to the hypothetical volatile-rich extralunar material mentioned above. The great difference in the Cd/Zn ratios between the Tranquillitatis and Procellarum sites appears to lend support to the "unusual lunar rock" explanation.

The concentration of the extralunar component in the fines portion of the lunar regolith can be considered to be a function of the time elapsed since the regolith started to form.

In the discussion to follow, we will assume that the particle-size distribution of the extralunar material has remained constant during this period. This assumption is supported by the fact that the same cometary source is probably responsible for (1) the fine interplanetary dust, which Öpik (1969) believes provides most of the mass influx to the moon; and (2) the Apollo asteroid and live comet influx, which Öpik estimates causes most of the primary cratering. We shall also assume that the regolith turns over as a result of a bombardment which is a small fraction of that suffered by the surface since its formation. The lowest curve in Fig. 2 of Shoemaker et al. (1970) indicates that the bombardment of the surface at the Mare Tranquillitatis site has been sufficient to turn the surface over six times. The regolith thickness increases with increasing integrated flux of extralunar material. However, as the thickness increases, the lower limit of the projectile mass which can penetrate the regolith also increases, and the rate of regolith growth decreases. The relationship between the regolith thickness and integrated flux can be evaluated on the basis of data given by Oberbeck and Quaide (1968). They have estimated thicknesses of the lunar regolith at four different locations on the basis of Lunar Orbiter photographs. We have evaluated the extralunar flux from data given in their Fig. 7a (note that the points labelled 50 and 500 m on the abscissa of that figure should be 31.6 and 316 m respectively). Shoemaker et al. (1970) give a power law of the form $F = \chi c^\lambda$, where F is the cumulative number of craters/km² with diameters greater

than c (in meters), λ is -2.93 and X is a measure of the cumulative flux of extralunar material. In order to avoid the effects of crater obliteration by repetitive bombardment, we have fitted this equation to the large-crater-diameter portion of the Oberbeck and Quaide curves, where crater erosion processes have been relatively unimportant. In Table 4 are listed regolith thicknesses and relative cumulative fluxes for the four locations. The use of a value of -2.93 for λ in the power function is rather arbitrary in our treatment of the data of Oberbeck and Quaide. The actual slope of the lower portion of their curves appears to be somewhat higher. However, the actual value of λ is not of great importance providing that the actual projectile (and crater) size distributions at the four different locations were the same.

The relationship between these two variables is made clearer when their logarithms are plotted, as in Fig. 2. They appear to be related in exponential fashion, and a line is drawn in Fig. 2 which corresponds to the relationship

$$\log D_R = 0.585 \log X + 0.466 \quad (1)$$

where D_R is the regolith depth. According to this relationship, a factor of 10 increase in the integrated meteorite flux brings about an increase in the regolith depth by a factor of about 3.8.

Of more direct interest for us is the relationship between

the infall of extralunar material and the concentration of such material in the fines portion of the regolith. The concentration in the fines of a particular element resulting from the extralunar component will be proportional to the relative influx, and inversely proportional to the regolith depth: i.e., $[X] \propto X/D_R$, where $[X]$ is the net concentration of a "extralunar tracer" element such as Ir. Here we assume that the trace element is homogeneously distributed within the regolith. It then follows that

$$\log [X] = \log X - \log D_R + k' \quad (2)$$

where k' is a unitless constant. Combining equations (1) and (2) we obtain

$$\log [X] = 0.415 \log X + k \quad (3)$$

where k is a constant which can be calculated for each element of interest at each lunar landing site. For purposes of comparing the concentrations of a given element at two different landing sites (designated by subscripts 1 and 2) we can eliminate this constant and write the equation

$$2.41 \log \frac{[X]_2}{[X]_1} = \log \frac{X_2}{X_1} \quad (4)$$

We have calculated that the extralunar component at the Apollo-11 site corresponds to about an addition of 1.14% Cl

chondrite-like material, and that that at the Apollo-12 site is about 1.04%. The ratio of these two values is about 1.10. The error in this ratio is relatively small (1 σ of about 4%) since it involves only the experimental precision of our combined Ir and Ge measurements, and does not have to allow for systematic errors or the correctness of the assumption that the extralunar component has a composition resembling C1 chondrites. Using equation (4), we calculate that the extralunar influx at the Mare Tranquillitatis site was about 1.25 times greater than that at the Oceanus Procellarum site. (This factor would decrease to 1.17 if the value for λ used in the power function to fit the Oberbeck and Quaide curves is set equal to -4.3.

Rubidium-strontium ages at these sites are 3.65 Gyr (Papanastassiou et al., 1970) and 3.31 Gyr (Papanastassiou and Wasserburg, 1970), respectively. That the ratio of these two ages (1.10) is substantially smaller than the ratio of integrated extralunar fluxes at the two sites indicates that the flux of extralunar objects has been decreasing with the passage of time. A similar conclusion has recently been reached by Shoemaker and Soderblom (private communication) on the basis of crater counts and crater morphology.

Ganapathy et al. (1970a) have estimated an infall rate of $3.8 \times 10^{-9} \text{ g cm}^{-2} \text{ yr}^{-1}$ of C1 chondrite-like material (including H_2O) at the Apollo-11 site. Assuming the same regolith thickness at that site (4.5 m) and a thickness 1.14 times less at the Apollo-12 site (an estimate from equation (1)), the same

packed density (1.8 g/cm^3), fraction of large rocks (10%) and solar wind fraction (2%), we estimate influxes of $2.8 \times 10^{-9} \text{ g cm}^{-2} \text{ yr}^{-1}$ and $2.5 \times 10^{-9} \text{ g cm}^{-2} \text{ yr}^{-1}$ of material with C1-like composition (including H_2O). Because of the evidence for a flux decreasing with time, $2.5 \times 10^{-9} \text{ g cm}^{-2} \text{ yr}^{-1}$ must be regarded as an upper limit on the recent flux of such material on to the lunar surface.

Although we have estimated an average extralunar component of the Apollo-11 breccias, they should each be considered individually, since neither their time nor locus of formation is known. Our earlier data (Wasson and Baedeker, 1970) indicate that the extralunar components of breccia samples 10021, 10046 and 10048 are very similar to those of the 10084 fines, and are consistent with formation at a local site relatively late in the history of the regolith. Breccia 10060 has an extralunar component about 0.65 that of the other samples, and breccia 10056 contains less than 0.05 times the C1-like component of the more typical samples. In fact, Wasson and Baedeker (1970) pointed out that composition-wise, 10056 was a type-B crystalline rock.

McKay et al. (1970) have proposed that the lunar breccias are formed by sintering of material heated and transported during the base-surge phase of crater formation, and not by shock-lithification. They state that compositional evidence provides support for such an origin. We do not agree with the latter conclusion; rather, the compositional evidence defines conditions that must be met by any mechanism for breccia formation. The amount of extralunar component in 10021, 10046

and 10048 is too close to that of the local soil for this to be fortuitous; it is very likely that these consist of lithified (or sintered) preexisting regolith. Breccia 10060 is most likely a mixture of 65% regolith and 35% comminuted, deeper-lying bedrock. Breccia 10056 was investigated by McKay et al. (1970), who confirmed that it was petrologically a breccia. Assuming that the 10056 sample we investigated was not mislabelled, it should consist entirely of comminuted bedrock, and is probably a monomict breccia. Breccias 10056 and 10060 deserve further study, since they are likely to contain features contrasting with those of the more typical breccias.

The unusual light-gray fines 12033 has a meteoritic component which is about 0.59 times that in the 12070 and 12028 soils. Baedeker et al. (1971) proposed that this meteoritic component has resulted from either (1) a 59:41 mixture of local soil with an Al-, P- and K-rich, Fe- and Ti-poor exotic component from elsewhere on the moon; or (2) a fortuitous contribution of 0.59 C1 chondrite-like material to an Al-rich, Fe-poor rock type during the cratering event. They feel that the former possibility is more persuasive, and assuming such a picture they calculated the composition of an "exotic" component from major element data on both 12033 and the more typical soil 12070. Their calculated composition is similar in K and P content to the KREEP (high K, rare earth element, and P) component of Hubbard et al. (1971). These workers report that the 12033 soil contains 65% KREEP, and that 12070 contains 25% KREEP. If it assumed that the exotic component defined

by Baedeker et al. (1971) is KREEP, and amounts to 25% of the 12070 fines, the 59:41 12070/exotic ratio given by these authors indicates a 56% KREEP component in 12033, in relatively good agreement with the estimate of Hubbard et al. (1971). As indicated by Baedeker et al. (1971), the KREEP component is probably of more recent origin than the Copernican event, and we favor the Fra Mauro formation as the source of this material.

ANORTHOSITIC AND OTHER LUNAR COMPONENTS IN THE SOIL

Wood et al. (1970a) found that about 5% of the coarse fines from the Apollo-12 site were of anorthositic composition. Independently Wood et al. (1970b) and Wasson and Baedeker (1970) estimated from the composition of the "fine" fines and breccias that these materials contain 20% of material of anorthositic composition. That this fine material has more of this component than the coarser material is not surprising, since the finer material has higher ejection velocities in cratering events.

Goles (1970) and Goles et alia (1970) have argued for a "cryptic" component distinct from the anorthosite, chiefly on the basis of variation diagrams where Al, Si, Ca and Mg contents of crystalline rocks, soils and breccias are plotted versus their Ti contents. Although it seems quite certain that lunar materials other than anorthositic fragments and local rocks have contributed to the Apollo-11 soil, we fail to see the evidence for appreciable amounts of these materials in these authors' variation diagrams. Their choice of TiO_2 as the independent

variable was unfortunate, since the relatively dense ilmenite phase may be depleted in the lunar surface materials by settling (Compston et al., 1970). Thus, the Ti values of the soil and, to a lesser degree, the breccias, should be regarded as lower limits. When one allows for this by an error bar (+1% Ti in length) and adds appropriate error bars to the other plotted points, there is no difficulty in finding straight lines which intercept the crystalline rocks, breccias, soils and anorthositic gabbro on the variation diagrams of Goles (1970). We find no reason to believe that other compositionally distinct lunar materials account for as much as 5% of the mass of the Apollo-11 fines and breccias.

Like the Apollo-11 soils, the soil samples recovered from Oceanus Procellarum are chemically distinct from the crystalline rocks, being enriched in Al and depleted in Ti, and the existence of an exotic component in the lunar soil was indicated by the original LSPET (1970) data. As mentioned in the previous section, Hubbard et al. (1971) have characterized this component as being a high-K, -REE, -P rock which they call KREEP, and independently Wood et al. (1971) have characterized the same component, which they term "norite". It is difficult to estimate the abundance of most of our elements in the KREEP component due to their enrichment in the soil samples from the extralunar component. However, we can estimate the Ga concentration in the KREEP component, since the presence of the meteoritic component has only a minor effect on the observed Ga concentration of the soil. In Fig. 3 we have plotted Ga versus

the Al, Ca and Mg contents of the rocks and soils (the data on the latter elements are from Ansell et al., 1971; Compston et al., 1971; Engel and Engel, 1971; LSPET, 1970; Rose et al., 1971; and Willis et al., 1971). The regression lines are calculated using York's least squares fitting procedure, with a $\pm 5\%$ error assigned to all variables. The data point for 12014 was excluded from the fitting procedure, since the only major element data available on this sample was that of LSPET (1970), which we considered to be unreliable. The LSPET analyses generally yielded Al, Ca and Mg results which were systematically higher than the values obtained for the same rock samples by wet chemical or x-ray fluorescence techniques. The Ga contents of the crystalline rocks are positively correlated with Al and Ca and negatively correlated with Mg at confidence levels greater than 95%.

Hubbard et al. (1971) estimate that soils 12070 and 12033 contain 25 and 65% KREEP respectively, based on their Rb, Sr and rare earth-element data. The microprobe data of Wood et al. (1971) shows the KREEP component to be enriched in Al and Ca with respect to the local crystalline rocks, with a wide range of Mg concentrations (2.5 - 6.3% Mg). Fig. 3 points to an apparent enrichment of Ga in the KREEP component, and suggests a greater enrichment of Ga than Ca. From the available data on soils 12070 and 12033 and an assumed composition of 25% and 65% KREEP in 12033 and 12070 respectively, one can calculate average compositions for the crystalline rocks and KREEP components. The average crystalline rock Ga content calculated

is 3.9 ppm, 0.1 ppm higher than the mean Ga concentration of our suite of rocks, and the Ga content estimated for the KREEP component is 6.2 ppm.

The presence of the KREEP component in the fines and breccias is probably of minor import for the calculations used to evaluate the extralunar component in the previous section. In the following section we show that the concentrations of the elements we have studied show little variation between different types of mafic and acidic igneous rocks. Thus, a negligible error should be introduced in subtracting out a lunar contribution which is based only on the observed concentrations in the local rocks. A second potential problem has to do with the fact that the extralunar component is probably considerably greater in the highlands regolith than in the maria basins. Although the anorthositic and KREEP materials are probably of highlands origin, the material presently in the maria regolith has been ejected at high velocities from relatively large cratering explosions. The highest velocity ejecta is associated with the central "plume" during crater formation, and probably consists of material from well below the regolith. It should therefore contain nearly no extralunar material.

COMPOSITIONAL COMPARISON OF LUNAR AND TERRESTRIAL IGNEOUS ROCKS

We have attacked the problem of understanding the geochemical behavior of our selection of trace elements from two different standpoints. First, we have initiated studies of suites of cogenetic terrestrial samples, in order to observe whether fractionation trends are present, and hopefully, to gain some insights into the processes which have produced these fractionations. Second, it was thought that a comparison of bulk concentrations in similar materials from the earth and moon might reveal some trends associated with the separate origins of these two planets and assist in formulating and testing hypotheses regarding formational and planet-wide geochemical processes.

One of the first type of studies, an investigation of a rock suite from the Skaergaard Intrusion, will be published elsewhere. Data on another suite of related rocks from the Waianae, Hawaii, volcano, are given in Table 2. Wasson and Baedeker (1970) measured five trace elements in three rocks from the lower member and two rocks from the upper member of the Waianae Volcano, as these members are defined by Macdonald and Katsura (1964). Since rocks from the different members are probably not cogenetic, we have obtained an additional three specimens of the lower-member rocks originally collected by Macdonald. We have selected our samples to show as much chemical variation as possible while still plotting in the main trend on alkali-iron-magnesium triangular plots and alkali-titania two-parameter plots. Fig. 4 shows Zn, Ga, Ge and In data plotted

for these rocks versus the crystallization index (CI) of Poldervaart and Parker (1964). The CI is roughly proportional to the normative contents of Mg and Ca silicate minerals. Presumably, a lower CI indicates a later-forming differentiate. Also plotted are the In, Ge, and corrected Ga values reported by Wasson and Baedeker (1970) for the two rocks from the Waianae upper member.

The data on the six rocks are listed in order of increasing crystallization index in Table 2. As can be seen there, the Cd and Ir data tend to scatter, with no trends evident. For this reason we have not plotted them in Fig. 4. The scatter in Ir reflects relatively large experimental errors in our data, which are near the detection limit for this element. The Cd data are quite consistent between splits of the same powder, but vary widely from specimen to specimen. We attribute this variation to sampling problems. Apparently the Cd is concentrated in a minor phase which is very inhomogeneously distributed, and which contains only minor amounts of Zn, Ga, Ge and In.

Fig. 4 shows that Zn and Ge remain remarkably constant throughout the Waianae suite, although no Zn data are available on the rocks from the upper member. The trends of increasing Ga and In with decreasing CI which were observed in the mixed Waianae suite of Wasson and Baedeker (1970) are also observable in the suite from the lower member only. The increase in concentration of each element is small (10 - 15%) within this relatively small range in CI.

Although no trend in the Ir data is observed for the samples from the lower member of the Waianae suite, the average Ir content

appears to be higher than that in the rocks from the upper member (Wasson and Baedeker, 1970), perhaps reflecting a tendency for Ir to concentrate in early differentiates. The high Ir concentrations observed for the peridotite samples analyzed may reflect a still higher level of abundance of Ir in the mantle materials from which the basalts were derived.

In addition to the above samples we have investigated seven other rocks of basaltic composition. These include two basalts dredged from the Mid-Atlantic Ridge, the chilled marginal gabbro of the Skaergaard Intrusion, and four standard rocks: BCR-1 and W-1 of the USGS; BR of the CRPG, Nancy; and BM of the ZGI, Berlin. All elements (except Ir, for which the data are of lower quality) show remarkably small variations despite the fact that these basaltic rocks are from widely varying locations. Indium varies by a factor of 2.3, Zn by a factor of 2, and Ga and Ge by factors of 1.4. Cd varies by about a factor of 6; however, if we discard the highest and lowest values as reflecting sampling inhomogeneities, the spread is reduced to about a factor of 2. Most authorities now attribute the formation of basalts to partial melting and subsequent fractional crystallization processes occurring in the upper mantle (Ito and Kennedy, 1968; O'Hara, 1965). The limited concentration ranges observed above indicate that (1) the source material for these basalts shows very little variation in its composition (or, specifically, its content of these trace elements) independent of its geographical location; and (2) trace-element differentiation of the basaltic material during transport from its source, extrusion

and solidification is relatively minor (providing the final rock still has a "basaltic" major-element content).

Also listed in Table 2 are data on four other igneous rocks, the USGS standard granite, granodiorite and andesite, and Buddington's pyroxene gabbro, which has been studied by other lunar investigators. Inclusion of the first three together with the basalts causes the ranges in concentration to increase only in the cases of In (to 3.1) and Cd (to ~9) in excess over the elemental concentration ranges listed in the previous paragraph. Inclusion of the pyroxene gabbro increases the Ga and Ge ranges to 1.7 and 1.5 respectively. That basic and acidic igneous rocks are so similar clearly shows that they have originated in relatively simple processes (as opposed to multiple-plate fractionation processes, which should cause much larger concentration ranges to be observed).

The last three rocks listed in Table 2 are ultramafics -- the USGS standard dunite and peridotite, and a Hawaiian peridotite described by Kushiro and Kuno (1963). These yield concentrations of Zn and Cd which are lower by factors of about 2 than the lowest values in the mafic and acidic rocks. Gallium and In are lower in the ultramafics by factors of 5 to 30. Germanium is only slightly lower in the ultramafics. Iridium in the dunite is similar to the mafic values, but is about a factor of 10 higher in the two peridotite samples. The differences between the peridotites and the mafic rocks probably reflect the direction and general magnitude of the distribution of these elements between the basaltic magmas and the parent

material (i.e., during partial melting of upper mantle materials, Ir tends to remain in the unmelted solid).

The remarkably small range of Ge in terrestrial igneous rocks is illustrated in Fig. 5, a histogram showing all of our data on terrestrial rocks. Apparently Ge is distributed about equally between all major phases in these rocks, and for those rock types which have formed by partial melting about equally between the molten and solid phases. Germanium is clearly not a good "tracer" element for studying igneous fractionation processes.

As noted above, our Ir data show quite a bit of scatter on the terrestrial samples studied. In some cases, however, the data on one type of rock from a single locality are relatively consistent. For example, the Ir content of the Waianae basalts derived from the lower member is quite well defined at about 0.47 ± 0.10 ppb and those of the W-1 diabase at 0.32 ± 0.05 ppb. The BCR-1 and BR basalts gave values which are about an order of magnitude lower than the above values. Because of the large spread, it is not possible to speak of an average Ir content of terrestrial basaltic rocks. Other igneous rocks show Ir concentration ranges similar to those encountered in the basalts. Only the peridotites show relatively consistent and high values, as mentioned above, but our data are too few to establish a world-wide average for these rocks.

In Table 5 are listed average concentrations in terrestrial and lunar basalts for the five elements other than Ir. Gallium is 5.2 times lower in the Apollo-12 rocks than in average terrestrial basalts; the factor for Apollo-11 basalts is about

4.5, after correcting the Wasson and Baedeker (1970) data for the 5% systematic error noted in the experimental section. The other elements are depleted by factors of 60 or more, and could, within the scatter of the data, all be depleted by the same factor (of about 70 - 80). The difference between Ga and the other elements is striking. The lunar Ga abundances (relative to Si) are slightly lower than those in the ordinary chondrites (Tandon and Wasson, 1968; Fouche' and Smales, 1967), and slightly higher than the values for terrestrial peridotites (Table 2). In the LL-chondrites and mesosiderites, which have oxidation states comparable to lunar samples (as estimated by the Fe/Mg ratios of ferromagnesian silicates coexisting with metallic iron), Ga is about equally distributed between the magnetic (metallic) and non-magnetic (silicate) separates (Fouche' and Smales, 1967; Van Alstine et al., 1970). Thus, Ga should not be depleted by more than a factor of about 2 by the separation of a metallic phase. The Ga concentration in lunar "mantle" and separated metal surface rocks is more likely to reflect that of the bulk moon than Ge and Ir, which are more strongly concentrated into Fe-Ni.

The depletion of Ge may be ascribed to its extraction by a metal phase during the formation of the lunar basalts, since Ge is concentrated in the metal of LL-chondrites and mesosiderites (Fouche' and Smales, 1967; Van Alstine et al., 1970). The other siderophile elements such as Ir would also have been lost through such a process. The fact that Ge concentrations are higher in terrestrial rocks might be explained if the extraction process

on earth was not so efficient. The oxygen fugacity may also have been higher on the earth, such that the silicate/metal distribution ratio of Ge was higher.

Two different geochemical affinities can be invoked to explain the depletion of the other three elements in lunar rocks. Ringwood and Essene (1970) and Ganapathy et al. (1970a) classify them as volatile elements, and believe that their low abundances (together with those of Rb, Pb, etc.) may be characteristic of the (basaltic) source regions, and already established during the accretion of the moon. They and others (e.g., Tera et al., 1970) quite reasonably reject the possibility that these elements were boiled off during extrusion, since the $\text{Sr}^{87}/\text{Sr}^{86}$ intercept indicates that the Rb-Sr fractionation took place about 4.6 Gyr ago, 1.3 Gyr before the extrusion of the Oceanus Procellarum basalts. Alternatively, Zn, Cd and In may be chalcophile, and may have been extracted with a settling FeS phase (or, more likely, an FeS - Fe eutectic) during a molten stage in the moon's history. Unfortunately, we do not know either (1) how strong their chalcophile properties are, nor (2) how effectively they would be scavenged by an iron-troilite eutectic. It appears to be simplest to attribute the low abundance of the three elements to volatility, since Wasson (1971) has shown that this is the property which best explains the similar lunar/terrestrial and ordinary chondrite/C1 chondrite abundance patterns.

Two models have been proposed to explain the low volatile-element concentrations of lunar rocks relative to terrestrial basalts. Ringwood and Essene (1970) attribute the low abundance

of volatile elements on the moon to an incomplete recondensation of materials evaporated from a late, high-temperature stage during the formation of the earth. Ganapathy et al. (1970a) invoke a model developed by Anders (1968) and Turekian and Clark (1969) in which the accretion of the moon and earth occurred during a period of falling temperatures, with condensation and accretion of volatiles as a very late event during the formation of these planets. If, when this occurred, (1) the earth and moon had nearly their present masses, and (2) the relative capture velocities were low, it is reasonable that the earth should have swept up much more of the material than the moon. Ganapathy et al. (1970a) argue for an enhancement of this effect resulting from the geocentric orbital velocity of the moon.

Although Wasson (1971) has argued that the lunar/terrestrial concentration ratios are more in keeping with a "selective accretion" model of the latter type than with a "partial condensation" model as proposed by Ringwood and Essene (1970), Singer and Bandermann (1970) have shown that the maximum enrichment attainable by this mechanism is a factor of 22 more mass per unit area, or a factor of only 3.6 more mass per unit volume, if the volatile elements are distributed through masses on the earth and moon which are proportional to the total masses of these planets. Since four of the terrestrial lunar concentration ratios listed in Table 5 are ≥ 60 , it appears that the Ganapathy et al. (1970a) model must be rejected.

Wasson (1971) has proposed an alternative manner in which a late, volatile-rich material could be selectively accreted

by the earth relative to the moon, independent of the sign of the temperature-time differential. (Because of the inherently low probability for the capture of the moon had it formed elsewhere in the solar system (see, e.g., Wise, 1969), it is assumed that it and the earth formed at about the same distance from the sun.) It seems reasonable that the initial accretion of material to form the earth and the moon chiefly involved volatile-poor material in roughly circular heliocentric orbits, and that this process, once started, proceeded quite rapidly. At the end of this stage, the earth and moon had achieved masses very similar to their present ones. It is further assumed that the earth had formed and retained an atmosphere of mass comparable to or greater than its present one, whereas the moon had no appreciable atmosphere. Accretion continued, but at a much slower rate. If now, the major source of accreted material were comets which were occasionally perturbed by Jupiter into orbits which intersected those of the earth and moon, the selective accretion of volatile-rich material to the earth can be understood. According to Singer and Bander mann (1970), material with cometary geocentric velocities is captured at a higher rate (per unit mass) on the moon than on the earth (or at a rate slightly less than a factor of 2 in favor of the earth if expressed in units of mass per unit area. However, the retention efficiency for such material would have been much greater on the earth than on the moon. The earth's atmosphere would have given the earth a retentivity of nearly 1.0 for such material, whereas collisions of high-velocity, volatile-rich comets with

the moon may very well have resulted in the retention of only a very small fraction of the original mass. Loss would have occurred both because of the vaporization of the lost material to atomic velocities greater than the escape velocity, as well as the entrainment of unvaporized particles by the relatively dense cloud of vapor which would result from an explosively heated dirty-snowball-type comet (Whipple, 1950). It is not critical whether or not the moon was gravitationally attached to the earth when this process occurred. If the mixing volumes involved masses of the earth and moon which were proportional to their total masses, the retention efficiency would have to be 300 times lower on the moon than on the earth in order to explain a terrestrial/lunar concentration of 100 for volatile elements. More details are given by Wasson (1971).

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APPENDIX

Previously unpublished data on the concentrations of Ir in
Apollo 11 samples

Sample	Petrologic Class	Ir (ppb)	
		replicates	mean
10058,30	Crystalline B	0.09	0.09
10084,26	Fines	9.7,8.2*	9.8

*For previous replicates see Wasson and Baedeker (1970)

Table 1

Replicate and mean concentrations of six trace elements in six rocks and two soils from Oceanus Procellarum

Sample	Type	Zn(ppm)		Ga(ppm)		Ge(ppb)		Cd(ppb)		In(ppb)		Ir(ppb)	
		replicates	mean	replicates	mean	replicates	mean	replicates	mean	replicates	mean	replicates	mean
12009,24	A	2.0,1.3 [†]	1.8	3.2,3.2	3.2	≤42,≤40	≤41	2.2,≤38 [‡]	2.2	4.2 [‡] ,1.6	1.6	0.12,0.05	0.08
12038,24	A	1.3,34 [‡]	1.3	5.2,5.2	5.2	≤6,≤30	≤18	4.1,8.1	6.1	1.0,1.3	1.1	0.57 [‡] ,0.09	0.09
12063,51	A	1.9,2.7	2.3	4.1,4.6	4.3	≤30,≤10	≤20	1.1,≤21 [‡]	1.1	0.9,1.4	1.1	≤0.03,0.06	≤0.04
12002,121	B	1.1,1.2	1.2	2.9,3.0,2.8	2.9	86,10,≤18	≤57	50 [‡] ,2.0	2.0	5.0 [‡] ,14.2 [‡] ,0.62	0.6	0.04,0.06	0.05
12014,12		1.3	1.0,1.0	1.0	2.9,2.9	2.9	18,≤7	≤13	≤2.0	≤4.2 [*]	133,201 [‡]	≤133 [¶]	0.09,0.07
12022,51	B	1.2,3.3	2.0 [§]	3.8,3.9	3.9	≤10,50	≤30	≤50 [‡] ,6.4	6.4	1.1 ,2.4	1.6 [§]	0.09,0.49 [‡]	0.09
12033,94	D	7.2,7.0	7.1	5.4,5.4	5.4	221,225	223	56,72	64	56,38	≤47 [¶]	3.9,3.9	3.9
12070,84	D	9.0,8.8	8.9	4.4,4.6	4.5	349,361	355	456 [‡] ,195	195	131,248 [‡]	≤131 [¶]	8,1.7.4	7.7

*Not corrected for uranium fission contribution.

[†]Value of less than usual accuracy. Given one-half weight in the determination of the mean.

[‡]Value shows evidence of contamination. Not included in the determination of the mean.

[§]Geometric mean.

^{||}Surface of rock sample removed with SiC abrasive powder prior to analysis.

[¶]All analyses show evidence of contamination. Reported mean is probably too high.

Table 2

Zn, Ga, Ge, Cd, In and Ir concentrations in terrestrial basalts and other terrestrial rocks

Sample Location	Type	No.	So. §	Zn (ppm) repl.	mean	Ga (ppm) repl.	mean	Ge (ppm) repl.	mean	Cd (ppb) repl.	mean	In (ppb) repl.	mean	Ir (ppb) repl.	mean
Waianae [†] Hawaii	Thol. basalt	C-7	GAM	116,114	115	18.6,21.1 18.8	19.5	1.55,1.67 1.63	1.62	232,270	250	82,77 76	78	0.67,0.38 0.41	0.48
Waianae Hawaii	Thol. basalt	C-6	GAM	114,120	117	18.3,18.6*	18.2	1.53,1.61*	1.56	108,121	114	66,67*	72	0.53,0.48*	0.57
Waianae Hawaii	Thol. basalt	C-11	GAM	109,107	108	21.6,20.7*	21.0	1.66,1.67*	1.65	104,96	100	71,74*	80	0.43,0.41*	0.50
Waianae Hawaii	Thol. basalt	C-5	GAM	120,125	123	22.3,22.5 22.6	22.5	1.71,1.66 1.69	1.68	165,180	172	94,82 89	88	0.46,0.34 0.34	0.38
Waianae Hawaii	Thol. basalt	C-40	GAM	110,108	109	20.7,19.1	19.9	1.63,1.64 1.61	1.63	151,196	173	90,77 79	84	0.79,0.44 0.50	0.58
Waianae Hawaii	Thol. basalt	C-20	GAM	111	111	21.3,21.4*	21.8	1.81,1.80*	1.75	105,105	105	77,81*	89	<0.04,0.34*	
Mid-Atl. Ridge	Thol. basalt	1-3	FA	92,85	88	16.7,17.3	17.0	1.65,1.52	1.58	141,157	149	68,65	67	0.23,0.14	0.18
Mid-Atl. Ridge	Thol. basalt	56-2/3	FA	77,88	82	15.5,16.5	16.0	---,1.54	1.54	122,129	125	60,58	59	0.67,0.08	
Skaergaard	Marginal gabbro	EG4507	EAV	74,89	82	21.6,22.8	22.2	1.26,1.26	1.26	--,42	42	39,44	41	0.34,0.21	0.26
Bridal Veil Oregon	Basalt	BCR-1	USGS	129	129	22.4*	21.7	1.49*	1.55	82	82	93*	94	<0.11,0.07*	<0.12
Centerville Virginia	Diabase	W-1	USGS	91,93 89,96	92	17.4,17.3 18.1,18.4*	17.9	1.64,1.21 1.83,1.82*	1.74	---,142 157,167	155	66,64 60,60*	64	0.34,0.31 0.30,0.25*	0.32
Essey-la-Côte France	Basalt	BR	CRPG	180,156 165	167	18.9,19.6 18.4	19.0	2.26,1.38 1.63,1.39	1.67	125,122	124	55,75	65	0.04,0.03	0.04
Mellenbach Germany	Basalt	BM	ZGI	137,127	132	18.7,18.7	18.7	1.84,1.95	1.89	77,74	76	64,58	61	0.25,0.51	0.38
Westerly R.I.	Granite	G-2	USGS	93,93	93	23.6,23.8*	23.3	1.85,1.07*	1.36	26,27	27	28,26*	30	0.06,0.04*	0.07
Silver Plume Colorado	Grano- diorite	GSP-1	USGS	112	112	22.7,25.0 25.2	24.3	1.99,1.92 1.30	1.74	64	64	65,54 46	55	<0.25,<0.07	<0.16
Guano Valley Oregon	Andesite	AGV-1	USGS	100	100	20.7,22.1 21.3,22.6	21.7	1.64,1.17 1.26	1.36	75	75	48,46 42	45	0.29,<0.10	<0.20
Brown Point	Pyroxene gabbro	114	GG	154,144	149	13.7	13.7	2.05,1.92	1.98	201,162	181	--,87	87	2.8 [‡] ,0.07	0.07
Twin Sisters Washington	Dunite	DTS-1	USGS	51	51	0.43,0.44 [†]	0.45	0.83,0.84 [†]	0.97	10	10	2.7,2.2	2.5	0.52,0.43 0.72	0.56
Cozadero California	Peri- dotite	PCC-1	USGS	35	35	0.65 [†]	0.66	1.18,0.90 [†]	1.07	17	17	3.1,3.8 [†]	3.4	3.7,7.7	5.7
Salt Lake Honolulu	Peri- dotite	HK571- 01204d	IK	28,76		2.25,2.28	2.27	1.14,1.25 1.01	1.14	34,54	44	8.0,8.4	8.2	3.6,3.0	3.3

*For previous replicates and means see Wasson and Baedecker (1970).

†For previous replicates and means see Baedecker and Wasson (1970).

‡Waianae basalts listed in order of decreasing crystallization index.

§The sources are abbreviated as follows: FA -- F. Aumento, Dalhousie University; GG -- G. Gole, Univ. of Oregon (described by A.F. Buddington, 1939); IK -- I. Kushiro, Univ. of Tokyo; GAM -- G.A. Macdonald, Univ. of Hawaii; EAV -- E.A. Vincent, Oxford Univ.; CRPG, Centre des Recherches Petrographiques et Geochimiques, Nancy, France; USGS -- United States Geological Survey; ZGI -- Zentrales Geologisches Institut, Berlin, East Germany.

†Value shows evidence of contamination. Not included in the determination of the mean.

Table 3

Summary of data needed to estimate the magnitude of the meteoritic component at the Apollo 11 and 12 sites.*

	Ni (ppm)	Ir (ppb)	Pd (ppb)	Au (ppb)	Ge (ppb)	Ag (ppb)	Zn (ppm)	Cd (ppb)	Br (ppb)	Bi (ppb)	Tl (ppb)
11C	204±20	8.5±1.0	10±3	2.4±0.5	360±40	18±10	29±5	92±14	160±50	2.2±0.7	2.8±0.7
11D	190±20	8.9±1.0	9.8±3.0	2.0±0.5	390±40	10±5	21±5	44±10	107±30	1.6±0.7	2.2±0.7
11AB	5.8	0.1 [†]	3.3	0.071	20 [†]	1.5	1.64	4.6	22	0.33	0.67
12D	200±20	8.3±1.0		2.2±0.5	355±40	5.4±2.5	7.0±1.1	49±8	125±30	2.0±0.6	1.9±0.6
12AB	60±40	0.1		0.016	≤20	0.80	1.4±0.7	1.6	13.2	0.8±0.4	0.31
C1 [‡]	13300	634	625	208	42000	487	421	1260	6870	149	90
($\frac{11C-AB}{C1}$)	100 1.38±0.15	1.39±0.16	0.04±0.51	0.91±0.24	0.89±0.09	1.7±1.0	4.5±1.2	3.1±0.8	1.24±0.44	0.87±0.47	1.7±0.8
($\frac{11D-AB}{C1}$)	100 1.49±0.15	1.32±0.16	1.07±0.53	1.11±0.24	0.82±0.09	3.5±2.1	6.4±1.2	6.9±1.1	2.01±0.73	1.28±0.47	2.3±0.9
($\frac{12D-AB}{C1}$)	100 1.05±0.34	1.29±0.16		1.06±0.24	0.80±0.09	0.94±0.51	1.3±0.3	3.7±0.6	1.63±0.44	0.81±0.54	1.8±0.7

* Data taken from the following papers:

Apollo 11 -- Ansell and Helz (1970)[Ni]; Compston *et al.* (1970)[Ni]; Ganapathy *et al.* (1970a) [Ir, Pd, Au, Ag, Zn, Cd, Br, Bi, Tl]; Haskin *et al.* (1970)[Au, Ag, Br]; Morrison *et al.* (1970) [Zn]; Reed *et al.* (1970) [Br]; and Wasson and Baedeker (1970) [Ir, Au, Ge].

Apollo 12 -- Ganapathy *et al.* (1970b) [Ir, Au, Ag, Zn, Ce, Br, Bi, Tl]; and Laul *et al.* (1970a) [Ir, Au, Ag, Zn, Cd, Br, Bi, Tl]; this work [Ir, Ge, Zn, Cd].

Carbonaceous chondrites -- Crockett *et al.* (1967) [Pd, Au]; Ehmann *et al.* (1970) [Ir, Au]; Fouche' and Smales (1967) [Ge]; Goles *et al.* (1967) [Br]; Greenland (1967) [Pd, Ag, Zn, Cd]; Greenland and Lovering (1965) [Ni, Ge, Zn]; Laul *et al.* (1970b) [Tl]; Laul *et al.* (1970c) [Bi]; Lieberman and Ehmann (1967) [Br]; Reed and Allen (1966) [Br]; Reed *et al.* (1960) [Bi, Tl]; Schmitt *et al.* (1963) [Cd]; and Wiik (1969) [Ni, Zn].

[†] The 11AB Ge and Ir values of Wasson and Baedeker (1970) and Ganapathy *et al.* (1970a) showed some scatter. Since their higher values may include some contamination, we have used values for 11AB which are typical of the 12AB rocks.

[‡] C1 values raised by 1.25 to place them on water-free basis.

Table 4

Relationship between regolith thickness
and crater frequency for four regolith
types studied by Oberbeck and Quaide (1968)

Regolith type	Measured median thickness (m)	χ^* (km^{-2})
I	3.3	1.1×10^6
II	4.6	2.6×10^6
III	7.5	4.7×10^6
IV	16.0	1.8×10^7

* χ is the cumulative flux constant
in the equation $F = \chi c^\lambda$ (Shoemaker
et al., 1970) where F is the cumu-
lative number of craters with a dia-
meter (in m) greater than c in an area
of 1 km^2 , and $\lambda = -2.93$.

TABLE 5

A comparison of the trace-element composition of terrestrial and lunar (Oceanus Procellarum) rocks.

	Zn (ppm)	Ga (ppm)	Ge (ppb)	Cd (ppb)	In (ppb)
Terrestrial basaltic rocks	114	19.5	1620	118	68
Lunar basaltic rocks	1.6	3.7	≤20	≤2	1.2
Ratio terr/lun	71	5.2	≥80	≥60	57

FIGURE CAPTIONS

Fig. 1. Abundance pattern of 11 trace elements in the extralunar component obtained by subtracting average abundances for crystalline rocks from those of the soils and breccias and normalizing to C1 chondrites. The error bars represent estimates of one standard deviation based on the replicate analyses of the soil and breccia samples. a) Apollo-12 soils. The band represents an extralunar component amounting to a $1.04 \pm 0.24\%$ addition of a C1-chondrite-like material to soil derived from local crystalline rocks. b) Apollo-11 soils. The band represents an addition of $1.14 \pm 0.25\%$ of a C1-chondrite-like material to soil derived from local crystalline rocks. c) Apollo-11 breccias. The band represents an addition of $1.07 \pm 0.25\%$ of a C1-chondrite-like material to soil derived from local crystalline rocks.

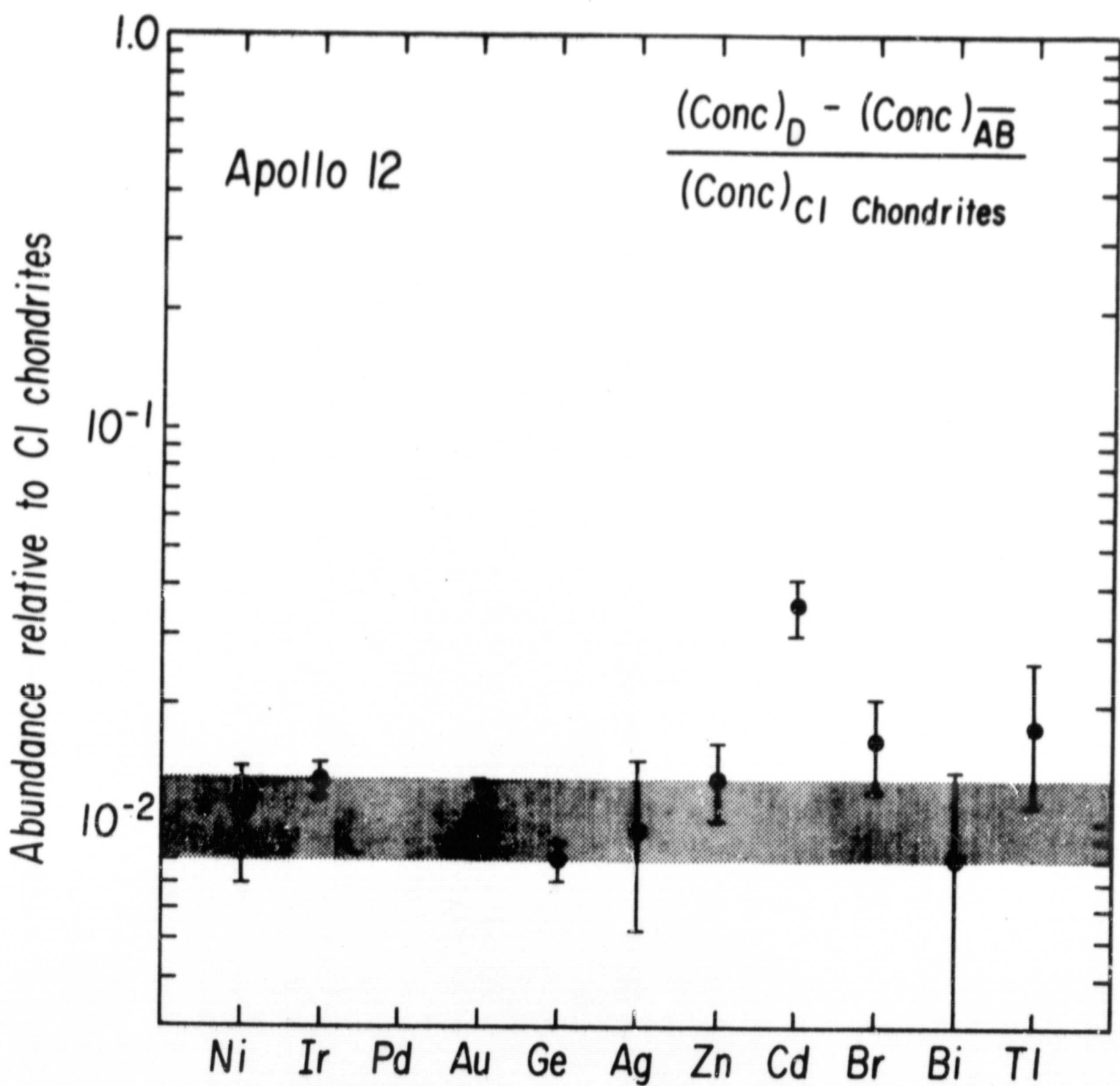
Fig. 2. A plot of the logarithm of the regolith depth (D_R) vs. the logarithm of the cumulative flux constant (χ) for the four regolith types studied by Oberbeck and Quaide (1968) using the data presented in Table 4.

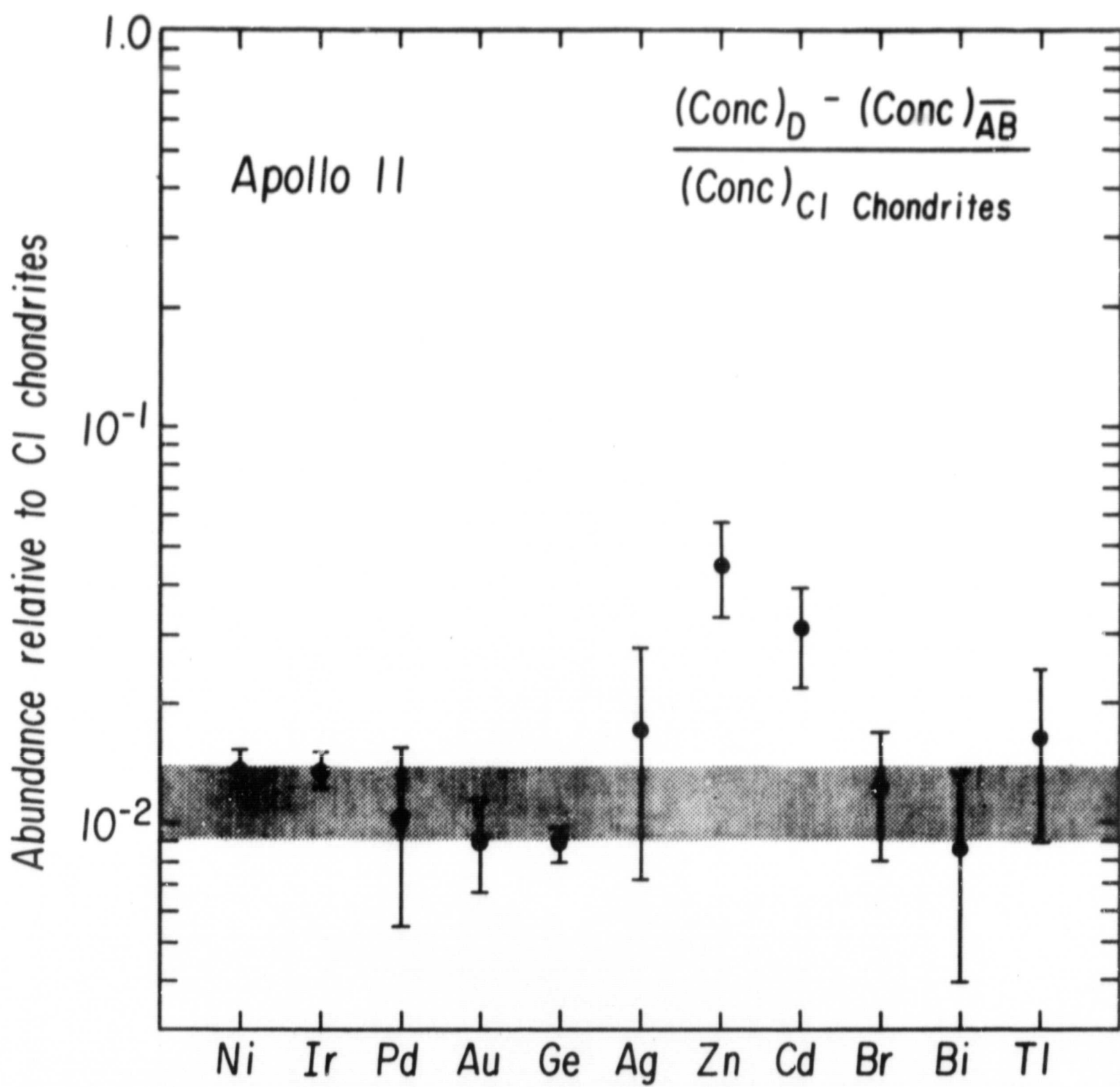
Fig. 3. Gallium data for six crystalline rocks and two soil samples plotted vs. their Al, Ca and Mg contents. The regression lines were based on the data points for the crystalline rocks represented by open circles (the open triangle is rock 12014, for which the major

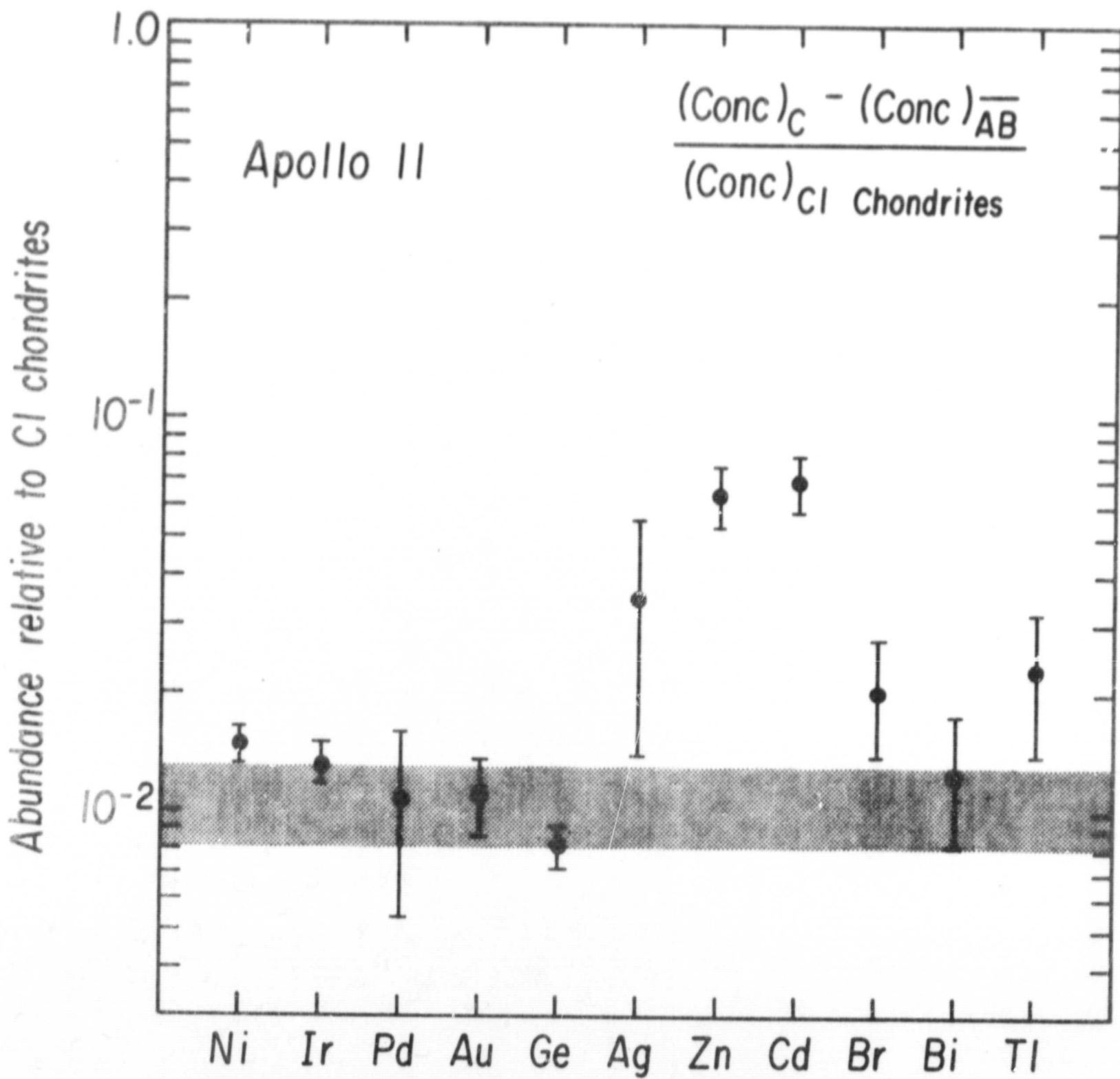
element data is of low quality). Correlations between each pair of variables are significant at confidence levels of $>95\%$.

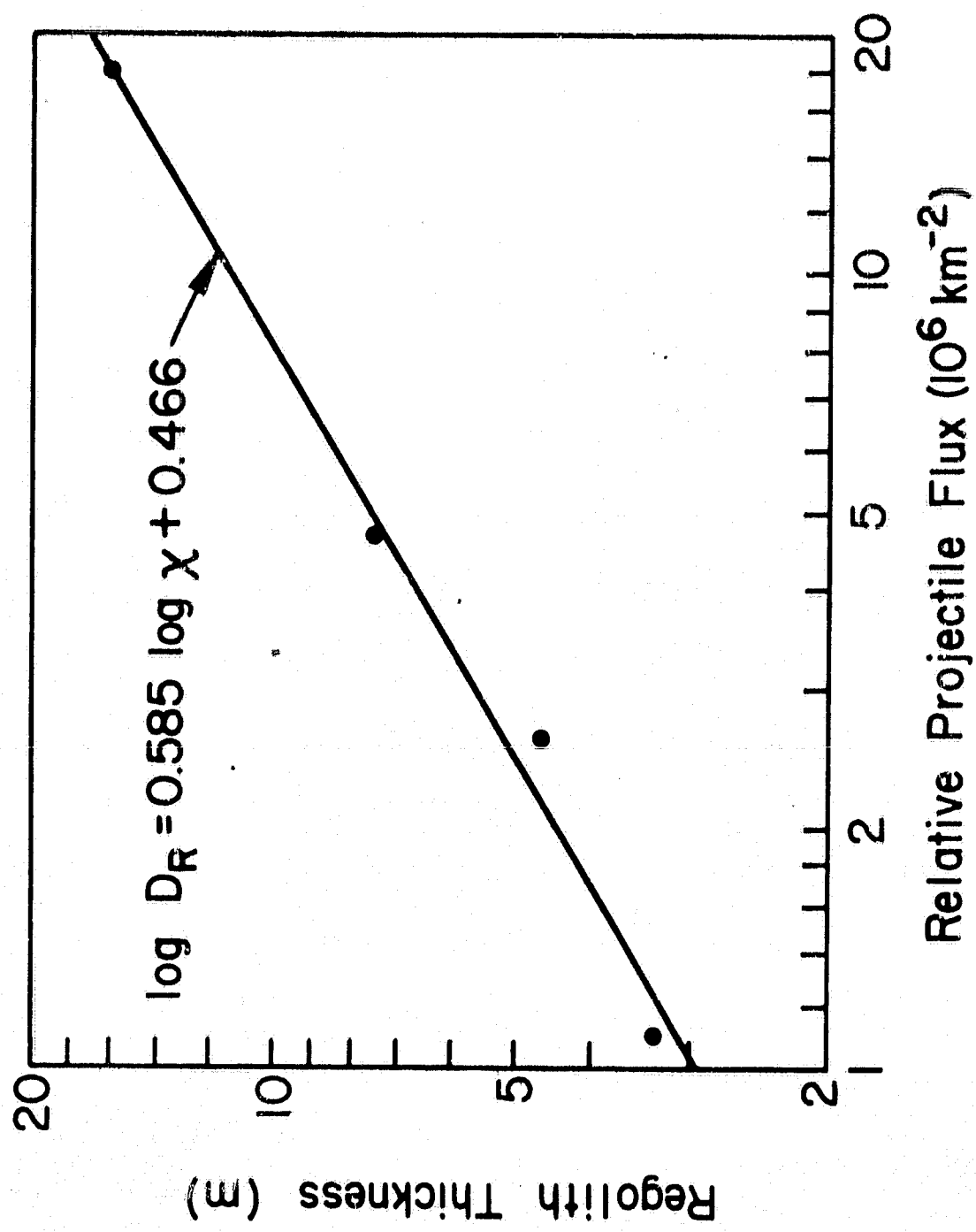
Fig. 4. Plots of our Zn, Ga, In and Ge data vs. crystallization index for a suite of Waianae, Hawaii, rocks. The error bars correspond to 95% confidence limits on the means of duplicate determinations.

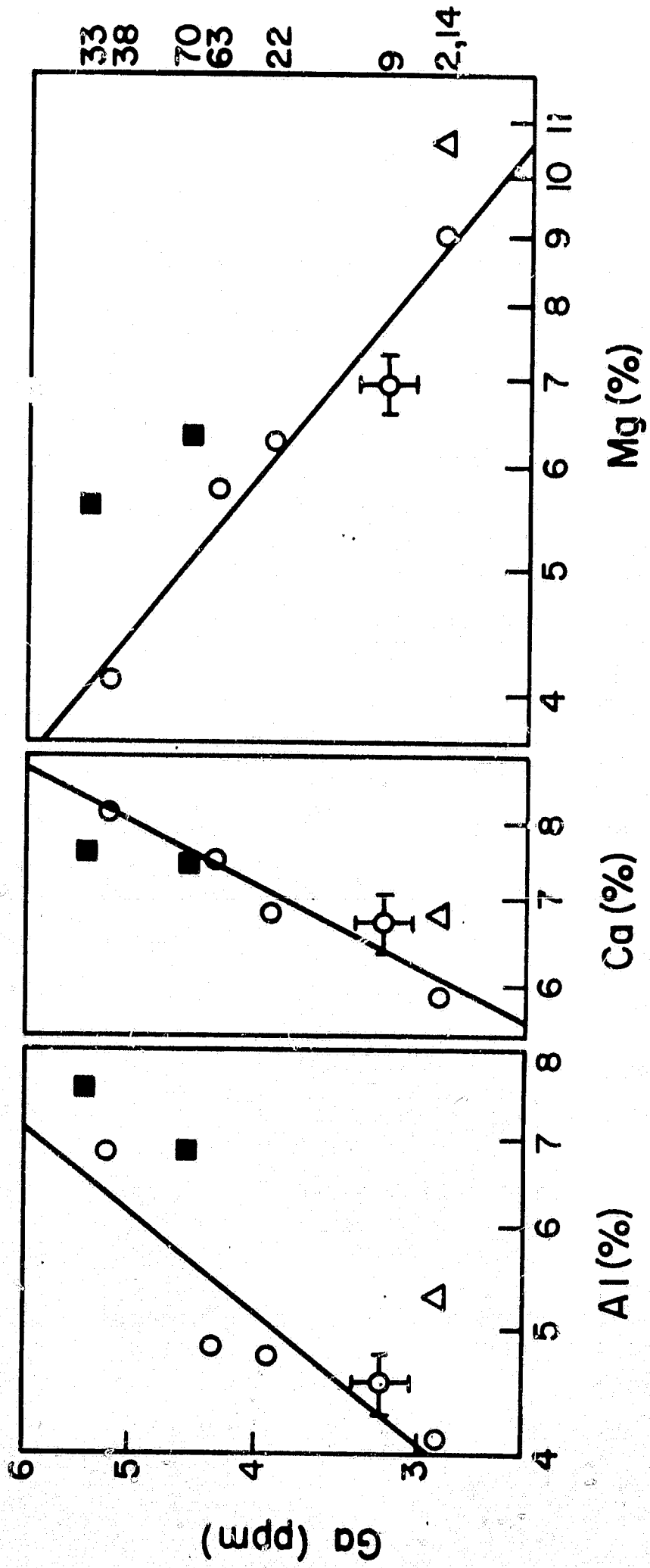
Fig. 5. A histogram of Ge data on terrestrial rocks (as presented in Table 2).











Rock suite, Waianae, Oahu, Hawaii

Zn (ppm)

Ga (ppm)

In (ppb)

Ge (ppm)

Crystallization Index

