sensed spectral data the accuracy is within 10% compared with the result of the chemical analysis of Apollo 17 samples (Table 1) or the results of the Apollo 17 Geochemical Orbiter Experiments [6]. Normative mineralogical information (Table 2) can be derived by comparing the calculated chemical concentrations with the concentration of Apollo 17 normative minerals (mineral phases of samples 74255 and 72395 [7]) [8]. Based on this approach the composition of the Apollo 17 surrounding areas can be determined relative to the ground-truth of the Apollo 17 landing site. The database consists of spectral/chemical parameters derived from the analysis of Apollo 17 soil samples, which transform spectral measurements into compositionally interpretable information, and of telescopic spectra obtained in the visible and infrared wavelength range [6,8]. The visible to near-infrared spectra (0.35–1.0 μm) are taken by using a CCD camera covering a regionally extended area, whereas the infrared spectra (0.7–2.5 μm) are taken at local surface points (Fig. 1) by using a CVF spectrometer. The ground resolution of the spectral measurements is 1–2 km. When the characteristics of spectra taken at the local surface points (Fig. 1) are used as end members, the distribution of compositional units can be estimated from the areally extended CCD spectral measurements.

Geological Interpretation: The geological interpretation of surface units is based on the chemical (Table 1) and mineralogical (Table 2) results, as well as on photogeological studies. Due to this interpretation, five geological units are identified (Fig. 1).

(1) Anorthositic highland materials (a) constitute an Al-rich unit that has an Fe/Al ratio < 0.6 and is characterized by its high content of plagioclase. The anorthositic highland materials are defined spectrally and chemically at the surface point V2 (Fig. 1). The unit is distributed in the form of small patches north and south of Apollo 17. Photogeological studies [1] indicate a pre-Nectarian age. (2) Mafic highland materials (n) are relatively Al-rich, but with an Fe/Al ratio ≈1 and a significant pyroxene component indicating the mafic characteristics of the unit. In terms of highland rocks these materials are comparable with gabbros (norites). The spectral/chemical reference for this unit is defined at the surface points L1 and L2 (Fig. 1).
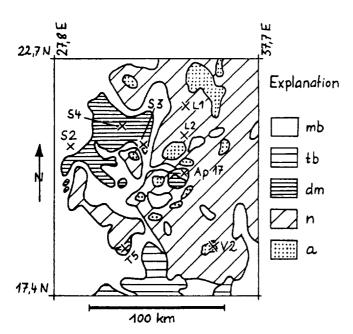
The mafic materials dominate the highlands surrounding Apollo 17. The age is assumed to be pre-Nectarian to Nectarian. The mare units of the Apollo 17 region are attributed to the Imbrian system [1,2]. In general, basalts are classified by their titanium content. Based on the TiO_2 concentrations normal mare basalts ($TiO_2 < 5$ wt%) can be distinguished from titanium basalts ($TiO_2 > 5$ wt%). (3) The spectral/chemical characteristics of a typical mare basalt (mb) as exposed at the surface point S2 (Fig. 1) can be compared with the characteristics of a pigeonite basalt defined by a relatively enriched pyroxene component. Mare basalts dominate the southwestern part of the mare/highland boundary between Mare Serenitatis and the Taurus Littrow mountains and also fill the valleys west of Apollo 17. (4) Titanium basalts (tb) as exposed at the surface points Ap17 and T5 (Fig. 1) are characterized by high TiO_2 concentrations and thus by a high ilmenite content. Titanium basalts are mainly exposed at Apollo 17 and south of the landing site toward Mare Tranquillitatis. (5) The materials exposed on the Mare Serenitatis shelf area west to the Apollo 17 landing site (S3 and S4 in Fig. 1) exhibit a spectral/chemical characteristic similar to that of the titanium basalts. However, this unit shows a fine texture, is smoother than the titanium basalt unit, and is enriched in the agglutinate and glass component [3]. The photogeologic interpretation indicates that this unit is mainly composed of pyroclastic materials (dm). The present study shows that the investigation of spectral properties is a useful tool to interpret the composition of lunar surface materials.

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N93-18793

NEOKREEP: A NEW LUNAR COMPONENT AT APOLLO 17.
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Introduction: The Apollo 11 (Mare Tranquillitatis) and Apollo 17 (Mare Serenitatis) landing sites are important as the only sources of high-Ti basalt visited by the Apollo missions. The lunar high-Ti basalts (>6% TiO_2; [1]) have no volumetrically comparable analogues among terrestrial basalts and require the presence of ilmenite in the source region, probably representing cumulates produced late in the crystallization of the lunar magma ocean [2].

Six principal groups of high-Ti basalts have been described, three from each of the two sites (see the review by [1]). The three groups of high-Ti basalts at the Apollo 17 site (termed "types") are all low-K (<2000 ppm K). These types are A, B, and C, with the type B basalts recently subdivided into varieties 1 and 2 (Fig. 1). It was shown by Neal et al. [3] that the type A basalts and both type B varieties can be generated through fractionation of observed phenocryst phases from a single magma. An evaluation of isotopic data [4] indicates an age of 3.75 ± 0.02 Ga for the type A basalts and 3.69 ± 0.02 Ga for the type B1/B2 basalts. The scarce type C basalts (only six examples) from Apollo 17 are more primitive than types A and B, having elevated MgO and Cr contents. However, the REE and alkali element (e.g., Rb) abundances in type C basalts are elevated relative to the type B basalts [5]. Neal et al. [5] suggest that the parent to the type C basalts was metasomatized by an alkali-rich fluid. Only two type C samples have been dated, yielding ages ranging from 3.63 ± 0.14 to 3.75 ± 0.08 Ga.
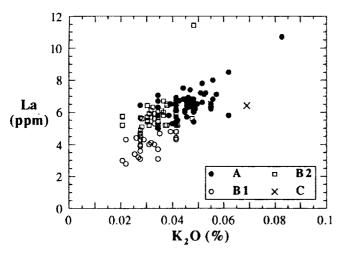

Fig. 1. Geologic map of the Apollo 17 area. (X) indicates the position of local surface points. For detailed description of the units see text.

Fig. 1.   Apollo 17 high-Ti basalts.

TABLE 1.   Fractionation calculations.

	A17 OG	30% xtallzd	avg A11-A
SiO_2	38.6	40.6	40.7
TiO_2	8.8	11.0	11.0
Al_2O_3	6.3	8.15	8.20
Cr_2O_3	0.75	0.12	0.33
FeO	22.0	21.7	20.2
MnO	0.27	0.27	0.23
MgO	14.4	7.06	8.01
CaO	7.7	10.4	10.5
Na_2O	0.36	0.50	0.52
K_2O	0.09	0.12	0.30
P_2O_5	0.06	0.08	0.19
Total	99.33	100.00	100.18
La	6.2	8.8	26.7
Ce	19.0	26.9	80.1
Nd	17.8	25.2	65.6
Sm	6.5	9.2	21.0
Eu	1.8	2.55	2.26
Gd	8.5	12.1	(26.8)
Tb	1.65	2.34	4.63
Dy	9.4	13.3	30.9
Er	5.1	7.2	18.9
Yb	4.43	6.13	17.0
Lu	0.61	0.86	2.46

Oxides are in %, REE in ppm.

Clearly these cannot be accurately placed into the stratigraphy represented by the types A and B basalts. A single example of another type of basalt was described by Ryder [6], who classified it as type D. This basalt has lower REE than type B basalts, but is richer in Mg and Cr and poorer in Ca and Ti.

Three groups of basalts also exist at the Apollo 11 site, divided into groups A, B, and D, with the low-K group B basalts subdivided into varieties 1, 2, and 3 (Fig. 2). None of the Apollo 11 basalt groups correspond to the types present at Apollo 17. Group B3 basalts are the most primitive at the Apollo 11 site and form a continuum of compositions with the B1 basalts (Fig. 2). The ages for these basalts, 3.71 ± 0.02 Ga and 3.67 ± 0.02 Ga for B3 and B1, respectively, can be considered coeval. Group B2 basalts are also low-K basalts, but are enriched in many incompatible trace elements, particularly the rare earth elements. The basalts of group B2 are the oldest at Apollo 11, with an age of 3.84 ± 0.02 Ga. Only three group D basalts (also low-K) have been described. Recent studies of the high-Ti basalts of Apollo 11 (e.g., [7–9]) have shown that all the basalts of groups B and D could have been generated through periodic melting of a single source region coupled with some fractional crystallization of the resulting magmas.

Apollo 11 Group A Basalts: An Anomalous High-Ti Group: The group A basalts of Apollo 11 differ in many respects from other
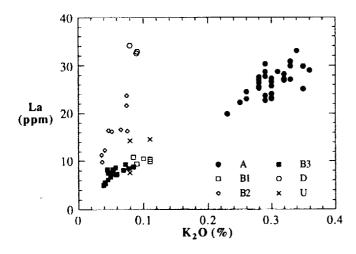

Fig. 2.   Apollo 11 high-Ti basalts.

high-Ti basalts of the region. Chemically, they are the only high-K (>2000 ppm K; [1]) variety of high-Ti basalt and are enriched in incompatible trace elements relative to other basalts from both the Apollo 11 and Apollo 17 sites (Fig. 2). In addition, group A basalts are the youngest of all high-Ti basalts, with an age of 3.56 ± 0.02 Ga [8]. The cluster of compositions is consistent with the Apollo 11 group A basalts representing a single flow (e.g., [7,10]). Papanastassiou et al. [11] have also indicated, based particularly on relatively young Rb-Sr model ages (3.8–3.9 Ga), the uniqueness of these basalts.

A model for the formation of the group A basalts was presented by Jerde et al. [7], wherein the Apollo 17 orange volcanic glass is the parent liquid. Fractionation of this composition, coupled with the assimilation of some incompatible-element-rich material, gave compositions akin to those of the Apollo 11 group A basalt population. It was postulated that this liquid was erupted near the Apollo 17 site and flowed into the Tranquillitatis Basin to the Apollo 11 site. Photographic and spectral evidence exists for a single flow unit spanning the distance between the two sites [12,13].

Evidence of KREEPy Contamination: Incompatible trace elements in group A basalts were modeled by Jerde et al. [7] using bulk partition coefficients calculated from the mineral assemblages indicated by fractionation calculations. It is evident from the results (Table 1) that the incompatible-element abundances in group A basalts are much too high (e.g., La = 26.5 ppm) to be explained by simple Rayleigh fractionation of Apollo 17 orange volcanic glass, which only leads to La = 8.8 ppm after 30% fractionation. The fact that K, P, and the REE are low in the calculated liquids suggests involvement of a KREEP component to obtain the basalts observed.
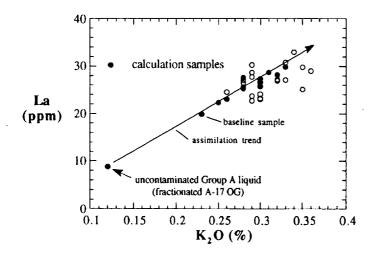
Fig. 3. Apollo 11 group A basalts.

In order to match the trace elements, ~20% of this component (the high-K KREEP of Warren [14]) must be added, which would imply that the Apollo 11 group A basalts are a young type of KREEP basalt. However, while this works chemically, isotopic considerations preclude it.

Addition of 20% of an old KREEP component (~4.4 Ga with ε_{Nd} = –3 at 3.56 Ga) would overwhelm the LILE in the magma, yielding basalts with ε_{Nd} values <0. However, ε_{Nd} values of the group A basalts of Apollo 11 fall in the range of +3 to +4. Therefore, any KREEPy assimilant could not have had an ε_{Nd} much less than +2 to +3. This KREEPy component must have been generated after complete crystallization of the LMO and must represent the late-stage crystallization products of a magma melted from the depleted "adcumulate" mantle. Assuming that this melt, indeed, was originally derived from the depleted cumulate mantle and that it crystallized with a $^{147}Sm/^{144}Nd$ greater than, but similar to, KREEP (=0.168), an age of 4.15 Ga can be calculated. This calculated model age is similar to the measured age (4.08 ± 0.07 Ga) of a KREEP basalt from Apollo 17 [15]. Furthermore, Shih et al. [15] suggested that other KREEPy rocks (including basalts, granites, troctolites, and norites) from various landing sites may also be cogenetic with this Apollo 17 KREEP basalt. This could further indicate that a Moon-wide melting event occurred at this time, which fused only the most evolved rocks This event would have occurred after crystallization of the lunar magma ocean but prior to initiation of high-Ti basalt volcanism. This post-LMO KREEPy
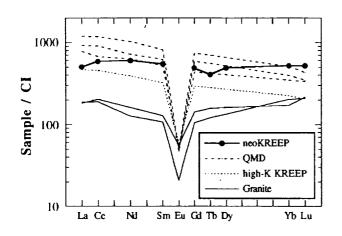
component has been dubbed neoKREEP because of its relatively young age compared to postulated urKREEP (dregs of the LMO).

An Incompatible Trace-Element Composition for NeoKREEP: To determine a composition for neoKREEP, a set of 12 samples was chosen from the suite of group A basalts. These samples fall on or near a line from the liquid composition generated from 30% fractionation of the Apollo 17 orange volcanic glass through the field of group A basalt compositions (Fig. 3). It is assumed that the array of sample compositions along this line represents varying degrees of assimilation of the neoKREEP component.

For each element i, the expression for the composition in a sample is given as

$$(1-x)L_i + xN_i = A_i$$

where x is the fractional amount of assimilation, L_i is the concentration of element i in the original liquid, N_i is the concentration of the same element in neoKREEP, and A_i is the concentration of this element in one of the group A basalt samples. This relation can be rearranged to

$$N_i = (A_i - L_i + xL_i)/x$$

providing an expression for the neoKREEP composition as a function of the amount of assimilation. For each sample and element, a series of N_i values were generated for various amounts of assimilation (x). A best-fit curve was then determined through the points. The actual amounts of assimilation, of course, are different for each sample, resulting in a new assimilation variable for each equation. To eliminate this problem, all the samples were related to a common baseline sample, showing the least assimilation (deviation from the original liquid). For the rare earth elements (REE), sample 10004,224 was defined as the baseline sample because it has the lowest concentration of REE among the 12 samples used. The assimilations in the other 11 samples were then given in terms relative to the baseline assimilation. This procedure provided 12 equations for the neoKREEP composition as a function of the baseline assimilation. Eleven pairs of simultaneous equations were then solved (each sample coupled with the baseline sample) for each element, yielding 11 values for the composition of neoKREEP and the baseline assimi-

Fig. 4. Evolved lunar rocks.

TABLE 2. neoKREEP composition.

		σ	x_{base}	σ
La	164	10	7.59	0.55
Ce	504	34	7.58	0.56
Nd	377	21	7.64	0.49
Sm	110	6	7.63	0.53
Eu				
Gd	134	8	6.59	1.10
Tb	21.0	1.1	7.47	0.51
Dy	167	8	7.54	0.40
Yb	114	7	7.98	0.56
Lu	17.6	0.9	8.20	0.54
Hf	96.0	6.5	9.06	0.71
Ba	1375	128	6.17	0.75
avg. x_{base}			7.59	0.18

All concentrations are in ppm.

lation. These 11 values were then averaged. The relations between the assimilation of each sample and the baseline assimilation were then used to determine the amounts of assimilation necessary in each sample. The results for some REE and other trace elements are given in Table 2. Approximately 7.5% assimilation of neoKREEP is required to produce the baseline group A basalt, and the entire array of group A compositions can be generated by 7.5–15% assimilation of neoKREEP with the composition in Table 2.

Comparison of NeoKREEP with Known Lunar Materials: The neoKREEP composition given in Table 2 and shown in Fig. 4 is more evolved than the postulated high-K KREEP composition given by Warren [14], but is not as evolved as materials such as the quartz monzodiorites from Apollo 15 (e.g., [16]). In addition, the chondrite-normalized REE slope is less than that of both high-K KREEP and the quartz monzodiorites, more akin to some lunar granites (e.g., 73215c, 14321,1027 in [17]). Warren et al. [17] noted the possibility that some lunar granites may be the late-stage remnants of younger, smaller intrusions formed through partial melting of the deep interior, which is essentially what we envisage as the origin of neoKREEP. Internal isochrons for granite 14321,1027 give ages of 4.09 ± 0.11 Ga and 4.11 ± 0.20 Ga from Rb-Sr and Sm-Nd, respectively [18], coinciding with our estimated age of neoKREEP of 4.15 Ga. Our neoKREEP composition is enriched in the REE over that of lunar granite. If neoKREEP represents some "pure" differentiate, perhaps the lunar granites reflect some dilution of similar material in the same way that KREEP basalts represent dilution of the "pure" differentiate urKREEP.

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N93-18792

POSSIBLE PETROGENETIC ASSOCIATIONS AMONG IGNEOUS COMPONENTS IN NORTH MASSIF SOILS: EVIDENCE IN 2–4-mm SOIL PARTICLES FROM 76503.

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Studies of Apollo 17 highland igneous rocks and clasts in breccias from the North and South Massifs have described magnesian troctolite, norite, anorthositic gabbro, dunite, spinel cataclasites (e.g., [1,2]), and granulitic lithologies that may have noritic anorthosite or anorthositic norite/gabbro as igneous precursors [3], and have speculated on possible petrogenetic relationships among these rock types. Mineral compositions and relative proportions of plagioclase and plagioclase-olivine particles in sample 76503 indicate that the precursor lithology of those particles was troctolitic anorthosite, not troctolite. Mineral and chemical compositions of more pyroxene-rich, magnesian breccias and granulites in 76503 indicate that their precursor lithology was anorthositic norite/gabbro. The combination of mineral compo-

sitions and whole-rock trace-element compositional trends supports a genetic relationship among these two groups as would result from differentiation of a single pluton.

View from the Soil: Although highland igneous lithologies in Apollo 17 materials have been described previously, the proportions of different igneous lithologies present in the massifs, their frequency of association, and how they are related are not well known. In this abstract, we consider the proportions of, and associations among, the igneous lithologies found in a North Massif soil (76503), which may represent those of the North Massif or a major part of it. Soil 76500 was collected far from any boulders [4] that might have added components to the soil. We assume for this exercise that the proportions of lithologies in the 2–4-mm fraction (76503) reasonably represent those of their igneous precursors. Some soil particles are polymict impact-melt breccias; we also seek to constrain the composition of igneous materials that were incorporated into them.

Based on a geochemical survey of 243 2–4-mm particles, soil at the base of North Massif comprises mainly impact melt breccias, regolith or glassy breccias, admixed high-Ti mare basalt and orange glass fragments, and surviving fragments of highland igneous lithologies [5]. Some 25 particles retain highland-igneous textures or have compositions indistinguishable from those that do. A subset of these igneous fragments comprises coarse single crystals or aggregates that have unshocked igneous or shocked, relict igneous textures. Some 40 particles are polymict breccias or are recrystallized and have granu-litic or polygonal textures. The polymict particles are dominated by igneous lithologies that are more pyroxene rich (anorthositic norite/gabbro) than those of the first group; they include two magnesian and one ferroan granulitic breccia groups (see [3]).

On the basis of a mixing-model calculation on the composition of <1-mm soil, [6] concluded that there must be some 15% of a high-Mg' (Mg/Mg + Fe) and low-ITE (incompatible trace element) component in addition to 36% of an anorthositic-norite component (roughly equivalent to gabbroic-anorthosite component of [7]). This is of special interest because the South Massif soils do not require the high-Mg' component in similar mixing models [6]. High-Mg' lithologies, however, are present in samples from both massifs.

The proportions of highland lithologies after removal of mare basalt and orange glass components and of regolith breccias and agglutinates whose compositions reflect mare-basalt admixture are as follows: 39 wt% noritic impact melt breccias, 18% magnesian granulitic breccias, 5% ferroan granulitic breccias, 9% other breccia lithologies that have generally anorthositic-norite compositions, and 29% troctolitic and noritic anorthosite fragments, including ~5% coarse, unshocked plagioclase crystals.

Olivine-Plagioclase Lithology: Trace-element and mineral compositions indicate that the coarse-grained plagioclase crystals found in the 2–4-mm size fraction and the fragments of coarse plagioclase and olivine ± orthopyroxene are closely related to those of 76535 troctolite [8,9], 76335 troctolitic anorthosite [10], and a troctolitic clast in 76255 [11] (Fig. 1). The mass weighted mean composition of this group of fragments, however, is substantially more anorthositic (~85–90% plagioclase) than bulk 76535, suggesting that these derive from a more anorthositic body of rock, such as 76335. We have not found coarse olivine fragments in troctolitic proportion, nor is olivine present in sufficient abundance in finer soil fractions [12] to indicate a parent of troctolitic bulk composition. The proportion of troctolitic-anorthosite component in 76503 (19% on a regolith-breccia-free basis) is greater than expected from previous investigations. This component is part of the high-Mg' component of [6] and appears to be more prevalent in the local North Massif soil than in the breccia boulders from higher on the Massif and more prevalent than in the