this range [19] would result in the fragmentation and mixing of Imbrium ejecta with local material. Thus, any Imbrium materials in the vicinity of the Apollo 17 site should have had a mode of emplacement similar to Imbrium ejecta at the Apollo 14 or 16 site. In addition to Imbrium, several post-Serenitatis sub-basin-sized craters are seen in highlands east of Serenitatis [15]. Although none directly intersect the landing site, their presence indicates that ejecta from their formation may be emplaced in the site area. One enigmatic aspect of several of these craters, such as Littrow, is that although they appear to postdate the Serenitatis event, their rims are characterized by textures similar to the Sculptured Hills. This would imply that the formation of the Sculptured Hills texture postdates the Serenitatis event, and thus casts into doubt the interpretation of the Sculptured Hills as a Montes Rook analogue formed simultaneously with Serenitatis [20]. On the other hand, it may be that the seismic effects of the Imbrium Basin event, which must have been significant [21], caused movement that modified both the Sculptured Hills and superposed crater deposits.

Geological Setting of the Apollo 17 Site: Apollo 17 surface exploration provided an excellent sampling of the different massifs and deposits in the Taurus Littrow Valley [22]. The sample materials consist of fragment-laden impact melts showing a diversity of textures (poikilitic, ophitic, subophitic, aphanitic) and a rather narrow range of chemical composition [23,24]. These characteristics have been generally interpreted to indicate that the majority of the rocks were derived from cooled melt rocks containing lithic clasts (primarily granulite breccias and Mg-suite cumulates) and representing Serenitatis ejecta and impact melt. Although the characteristics of the sample suite support this view, the outline of events in the Serenitatis region in addition to the Serenitatis event itself (e.g., Procellarum, Tranquillitatis, Imbrium, post-Serenitatis craters, etc.) all suggest that there may be other candidates for many of the materials sampled [20]. In addition, recent work on the formation of impact basins and new multispectral imaging data on basins [11,24] have provided a basis for the reinterpretation of the Apollo 17 site region. In December 1992 the Galileo spacecraft will obtain multispectral images of the north polar nearside region, including the Serenitatis Basin, and these data will provide an excellent framework for reanalysis of the Serenitatis Basin. Among the important questions to be addressed are (1) What is the role of Procellarum and Tranquillitatis Basin ring deposits at the Apollo 17 site? (2) What are the correct ring assignments for the Serenitatis Basin? (3) How many impact events are recorded in breccias? (4) How can we provide a better link of sample data to impact crater environments? (5) What is the true age of the Sculptured Hills and what represents them in the sample collection?

References: [1] Hinners N. (1973) *NASA SP-330*, 1-1 to 1-5. [2] Head J. and Settle M. (1976) *Interdisciplinary Studies by the Imbrium Consortium, 1*, 5–14. [3] Schmitt H. (1973) *Science, 182*, 681–690. [4] Hartmann W. and Kuiper G. (1962) *LPL Comm., 1*, 12, 51–66. [5] Wilhelms D. (1987) *U.S. Geol. Surv. Prof. Paper 1348*. [6] Whitaker E. (1981) *Proc. LPS 12A*, 105–111. [7] Head J. (1974) *Moon, 9*, 355–395. [8] Schultz P. and Spudis P. (1979) *Proc. LPSC 10th*, 2899–2918. [9] Head J. and Wilson L. (1992) *GCA, 56*, 2155–2175. [10] Belton M. et al. (1992) *Science, 255*, 570–576. [11] Head J. et al. (1992) *JGR*, submitted. [12] Scott D. (1974) *Proc. LSC 5th*, 3025–3036. [13] Reed V. and Wolfe E. (1975) *Proc. LSC 6th*, 2443–2461. [14] Wolfe E. and Reed V. (1976) *U.S. Geol. Surv. J. Res., 4*, 171–180. [15] Head J. (1979) *Moon Planets, 21*, 439–462. [16] Lucchitta B. (1978) *U.S. Geol. Surv. Map I-1062*. [17] Head J. (1974) *Moon, 11*, 327–356. [18] McGetchin T. et al. (1973) *EPSL, 20*, 226–236. [19] Oberbeck V. (1975) *Rev. Geophys. Space Phys., 13*, 337–362. [20] Spudis P. and Ryder G. (1981) *Proc. LPS 12A*, 133–148. [21] Schultz P. and Gault D. (1975) *Moon, 12*, 159–177. [22] Wolfe E. et al. (1981) *U.S. Geol. Surv. Prof. Paper 1080*. [23] LSPET (1973) *Science, 182*, 659–672. [24] Heiken G. et al., eds. (1991) *Lunar Sourcebook*, Cambridge, New York.

N93-18795

THE ANCIENT LUNAR CRUST, APOLLO 17 REGION.

O. B. James, 959 National Center, U.S. Geological Survey, Reston VA 22092, USA.

P. 4

Introduction: The Apollo 17 highland collection is dominated by fragment-laden melt rocks, generally thought to represent impact melt from the Serenitatis basin-forming impact. Fortunately for our understanding of the lunar crust, the melt rocks contain unmelted clasts of preexisting rocks. Similar ancient rocks are also found in the regolith; most are probably clasts eroded out of melt rocks.

The ancient rocks can be divided into groups by age, composition, and history. Oldest are plutonic igneous rocks, representing the magmatic components of the ancient crust. Younger are granulitic breccias, which are thoroughly recrystallized rocks of diverse parentages. Youngest are KREEPy basalts and felsites, products of relatively evolved magmas. Some characteristics of each group are given below.

Plutonic Igneous Rocks: All large Apollo 17 samples of plutonic igneous rocks are members of the Mg-suite. They are troctolites, norites, one gabbronorite, and one dunite. Detailed investigations of smaller samples (breccia clasts, coarse fines, and rake samples) have identified additional samples of the first three of these rock types, plus several samples of ferroan anorthosites.

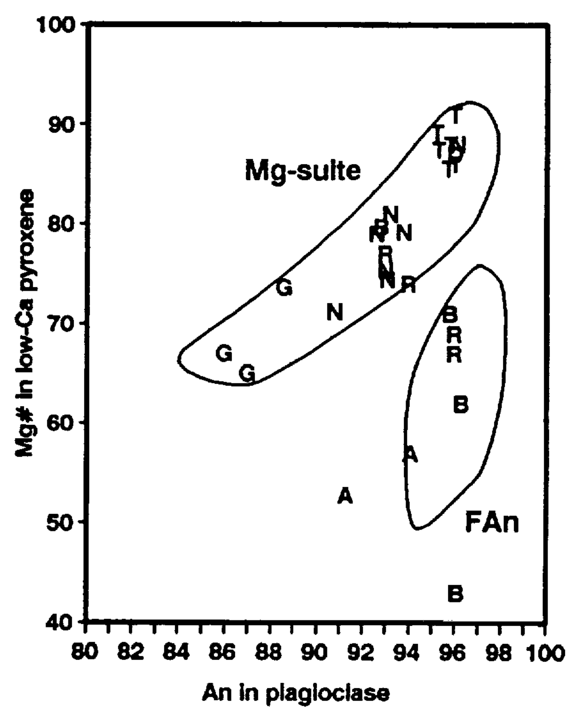

Fig. 1. Content of An in plagioclase vs. mg# in low-Ca pyroxene in some Apollo 17 ancient crustal rocks. Symbols are as follows: T, troctolite; D, dunite; N, norite; G, gabbronorite; A, anorthosite with relict igneous texture; B, anorthosite or norite with granulitic texture; R, granulitic breccia. Lines enclose the fields of compositional variation of minerals in all lunar Mg-suite rocks and in ferroan-anorthosite-suite rocks from sites other than Apollo 17.

Mg-suite rocks. The *troctolites* are mineralogically fairly simple. Most consist almost entirely of olivine and plagioclase, with only traces of other minerals (typically low-Ca pyroxene and chrome spinel) [1–8]. The largest troctolite, 76535 [1,2], also contains trace augite, whitlockite, apatite, baddeleyite, and K-feldspar. These rocks have high mg# (100 × molar Mg/[Mg + Fe]) in their mafic minerals and high An (100 × molar Ca/[Ca + Na + K]) in their plagioclases (Fig. 1). Mineral compositions are homogeneous in individual samples and fairly homogeneous in the group as a whole (Fig. 1). Most samples have cataclastic texture, but several have relict cumulate textures. Relict grains and monomineralic granulated patches are generally about 1–2 mm across. The coarse grain size and homogeneous mineral compositions suggest that these rocks formed deep within the crust; estimated depth of origin of 76535 is 10 to 30 km [2]. Estimates of crystallization age of 76535, from isotopic studies, range from 4.24 to 4.51 Ga [9–12].

The *norites* are mineralogically more diverse than the troctolites. Major minerals are plagioclase and low-Ca pyroxene, and most samples contain minor augite; many also have traces of a silica mineral, K-feldspar, phosphate minerals, chrome spinel, ilmenite, armalcolite, rutile, zircon, and baddeleyite [3,4,13–21]. Mineral compositions show a trend of increasing mg# in mafic minerals with increasing An in plagioclase (Fig. 1). In individual samples, mineral compositions are fairly homogeneous. All the norites have cataclastic texture, but some also have relict cumulate texture. Relict grain size is like that in the troctolites. One sample, 78527, has been granulated and recrystallized [14], but its mineralogy suggests that it is monomict. Depth of origin of the norites is difficult to determine quantitatively, but textures and mineral compositions indicate somewhat less reequilibration than in troctolite 76535, suggesting a shallower origin. Estimates of crystallization age, based on isotopic studies of several rocks, range from 4.33 to 4.43 Ga [22–26].

The *gabbronorites* (including one gabbro) contain more abundant augite relative to low-Ca pyroxene than the norites, and their assemblage of trace minerals is simpler [3,5,6,13,17]. Major minerals are plagioclase, low-Ca pyroxene, and augite; trace constituents are ilmenite, a silica mineral, K-feldspar, and chrome spinel. The pyroxenes average slightly more iron rich, and the plagioclases significantly more sodic, than in the norites (Fig. 1). The largest gabbronorite sample has cataclastic texture, but the smaller samples have relict igneous textures. Relict grain size is like that in troctolites and norites. As for the norites, quantitative determination of depth of origin is not possible, but textures and mineral compositions suggest a shallower origin than for troctolite 76535. Crystallization age of the only dated sample, 73255,27,45, is 4.23 Ga [26].

The *dunite* (72415-7) is mineralogically similar to troctolite 76535, except that olivine is by far the dominant mineral, K-feldspar and baddeleyite are absent, and a trace of armalcolite is present [1,27,28]. Compositions of the major minerals are similar to those in the troctolites (Fig. 1). The rock texture indicates a complex history [27]. The dunite was originally a cumulate, and vestiges of relict texture remain; zoning in the olivine has been interpreted as relict igneous variation [28]. After the dunite formed, it was subjected to multiple episodes of impact-induced shock and granulation, followed by recrystallization. The complex history of this rock obscures information on its depth of origin, but, if the olivine zoning is indeed primary, then the rock crystallized at a shallow depth [28]. Crystallization age, based on Rb-Sr studies, is 4.45 Ga [29].

Ferroan-anorthosite-suite rocks. The Apollo 17 anorthosites are fairly uniform modally but vary in compositions of their major minerals. Plagioclase makes up >93% by volume of most samples; low-Ca pyroxene, augite, and olivine are the most common minor constituents [5,21,30–32]. Mg# in mafic minerals spans a broad range and is lower than in most Mg-suite rocks (Fig. 1). One sample (76504,18) has unusually sodic plagioclase [5], and another (72275,350) has unusually iron-rich low-Ca pyroxene [31] (Fig. 1). Minerals in most samples are somewhat less homogeneous than in Mg-suite rocks. Textures of the rocks are also diverse. Three appear to be granulated coarse-grained plutonic rocks with little contamination from other rock types (74114,5; 72464,17; 76504,18) [5,30,32]. The others have had more complex histories: Two have been strongly shocked, three have been thoroughly recrystallized, and one is granulated and mixed with the surrounding breccia. In some of the complex samples, there may be unrecognized contamination by Mg-suite rocks, or the rocks may be genomict. Relict grain size, where present, is like that in Mg-suite rocks. There is no information on depth of origin or crystallization age, but data on Sr isotopic ratios in anorthosites from other sites indicate that such rocks formed very early in lunar history.

Granulitic Breccias: Numerous samples of granulitic breccias have been found at the Apollo 17 site. They range from relatively large samples, found in the regolith, to small clasts in breccias and fines. Typically, plagioclase content is 70–80% by volume and mafic minerals make up most of the remainder; commonly low-Ca pyroxene, olivine, and augite are all present. Mineral compositions in some granulitic breccias are like those in Mg-suite rocks; in others, they are like those in ferroan anorthosites (Fig. 1). The minerals are generally very homogeneous. Textures of the granulitic breccias range from coarse-grained poikiloblastic (e.g., 77017 [33]) to fine-grained granoblastic (e.g., 79215 [34], 78155 [35]) and indicate thorough recrystallization and reequilibration. Studies of granulitic breccias have established that such rocks can form by recrystallization of varied parents, including polymict breccias [34–36], fragment-laden melt rocks [37,38], and granulated monomict igneous rocks [39,40]. Samples 79215 and 78155 clearly formed from polymict precursors [34,35]. The precursor of 77017 may have been a monomict cumulate [33]; whether or not this interpretation is correct, mineral compositions in 77017 and 78155 indicate that the dominant components were ferroan-anorthosite-suite rocks, with little or no contribution from Mg-suite rocks. Recrystallization, in 78155 and three clasts from 73215, took place at about 4.25 Ga [41,42].

KREEPy Basalt: KREEPy basalt occurs at the Apollo 17 site only as granulated clasts making up the matrix of boulder 1 at station 2. Major minerals are plagioclase and clinopyroxene (ranging from magnesian pigeonite to ferroaugite); minor constituents are chrome spinel, a silica mineral, olivine, ilmenite, phosphate minerals, K-feldspar, zircon, and Si-rich glass [15,43,44]. The absence of meteoritic contamination and xenocrysts, and the presence of subophitic-intersertal texture and a broad compositional variation in the pyroxene, establish that these basalts are extrusive igneous rocks, not impact melts [43–45]. Crystallization age is about 4 Ga [25]. Chemically and mineralogically, these basalts are intermediate between high-alumina mare basalts and Apollo 15 KREEP basalts [43]. Possible explanations for this intermediate character are (1) mixing of a mare-type and a KREEP magma, (2) contamination of a mare magma with KREEP, or (3) partial melting of a distinctive source region with intermediate characteristics. The last of these alternatives may be the most likely [43].

Felsites: Felsites (fine-grained igneous rocks of granitic composition) are found as small clasts, in small numbers, at Apollo 17. Most are clasts in breccias from boulder 1 at station 2, South Massif [46], and from station 3 [36,47]. Major minerals in the felsites are quartz and Ba-K feldspar; minor minerals are Ca-K-rich ternary plagioclase, Ca-Fe-rich pyroxene, Fe-rich olivine, ilmenite, and

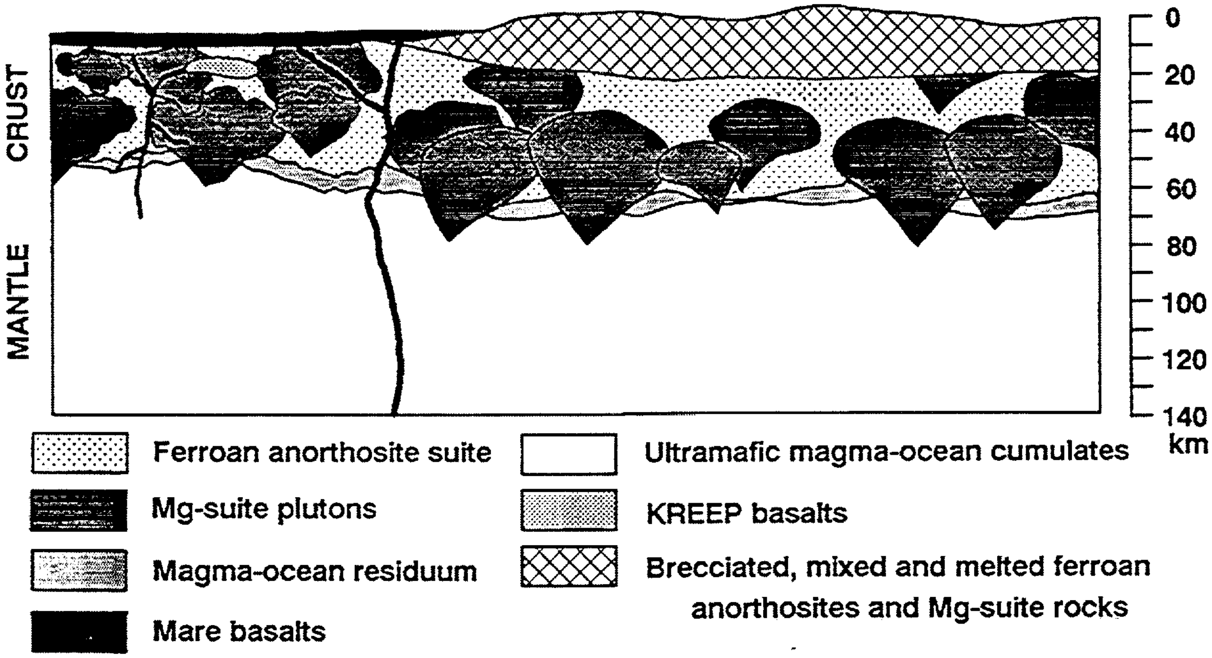

Fig. 2. Hypothetical cross section through the lunar crust and upper mantle.

phosphate minerals [36,46,47]. Many show relict igneous textures, most commonly micrographic or vermicular intergrowths of the quartz and feldspars. Many clasts have been partly melted. The relationship of felsites to other highland lithologies is not well understood. The evolved characteristics of these rocks suggest that their parent magmas could have formed by remelting within Mg-suite plutons, or by differentiation of KREEP basalt magmas; their relatively young crystallization age of about 4 Ga [48] indicates that they cannot be direct differentiates of Mg-suite parent magmas.

Origin and Evolution of the Lunar Crust: The results of lunar sample studies by numerous investigators have led to a hypothesis [49,50] for the evolution of the lunar crust and mantle. Figure 2 shows the structure of the outer 140 km of the Moon implied by this hypothesis, which is as follows. Differentiation of a primordial magma ocean several hundred kilometers deep formed a crust of ferroan anorthosite and related rocks (by plagioclase flotation) and an ultramafic mantle (by mafic-mineral sinking); the residual magma was rich in K, REE, and P and concentrated at or near the crust-mantle boundary. Partial melting near the end of magma-ocean crystallization, either in magma-ocean cumulates or in underlying undifferentiated mantle, produced mafic magmas that rose and intruded the primordial crust, forming large plutons of Mg-suite rocks. Impacts at the lunar surface granulated, mixed, and melted the crustal rocks, forming a near-surface layer of impact-modified rocks, including granulitic breccias. Later, more evolved magmas formed, perhaps by assimilation of the evolved parts of Mg-suite plutons by mare-basalt magmas, perhaps by remelting in Mg-suite plutons; these magmas formed KREEPy basalts and felsites. Locally, large impacts stripped away the top layers of the crust and formed multiring basins and fragment-laden impact melts. Subsequent melting in the magma-ocean cumulates produced mare-basalt magmas, which were extruded on the lunar surface and crystallized as mare basalts.

Ferroan anorthosites are rare at the Apollo 17 site, a fact noted many years ago. The most widely accepted explanation for the rarity of these rocks [4] is that the Serenitatis impact removed the upper part of the crust, which is dominated by ferroan anorthosite, from the region, and the samples are mostly rocks from the lower crust, where Mg-suite plutons probably dominate (see Fig. 2). It is also possible [32] that the layer of anorthosite formed by the primordial differentiation was thinner in the Apollo 17 region than elsewhere on the Moon. These two explanations are not mutually exclusive; both could be true.

Comparison of the Apollo 17 ferroan anorthosites to ferroan anorthosites found elsewhere on the Moon can potentially yield information on the nature of the primordial differentiation. The Apollo 17 anorthosites are modally and mineralogically (Fig. 1) much like anorthosites from elsewhere on the Moon. Differences are (1) relatively high contents of incompatible elements [30–32], (2) an unusually broad range of mineral compositions (Fig. 1), and (3) common development of granulitic texture. The significance of these differences is as yet unclear.

One of the greatest uncertainties in interpreting the history of the lunar crust concerns the nature of the precursors of granulitic breccias. Samples of these rocks have been variously shown to represent recrystallized polymict breccias, impact-melt rocks, and monomict igneous rocks. Those whose precursors are monomict igneous rocks are potentially the most important. Most large samples of granulitic breccias that have been exhaustively studied have been shown to be polymict, but small regions in such rocks are clearly monomict; thus, small granulitic breccia clasts could indeed represent monomict igneous rocks. Detailed studies of granulitic breccias to identify monomict samples could contribute significantly to our understanding of lunar magmatic processes.

References: [1] Dymek R. F. et al. (1975) *Proc. LSC 6th*, 301–341. [2] Gooley R. et al. (1974) *GCA, 38*, 1329–1339. [3] Warren P. H. and Wasson J. T. (1979) *Proc. LPSC 10th*, 583–610. [4] Warren P. H. et al. (1987) *Proc. LPSC 17th*, in *JGR, 92*, E303–E313. [5] Warren P. H. et al. (1986) *Proc. LPSC 16th*, in *JGR, 91*, D319–D330. [6] Warner J. L. et al. (1976) *Proc. LSC 7th*, 2233–2250. [7] Warren P. H. and Wasson J. T. (1977) *Proc. LSC 8th*, 2215–2255. [8] Winzer S. R. et al. (1974) *EPSL, 23*, 439–444. [9] Lugmair G. W. et al. (1976) *Proc. LSC 7th*, 2009–2033.

[10] Papanastassiou D. A. and Wasserburg G. J. (1976) *Proc. LSC 7th*, 2035–2054. [11] Bogard D. D. et al. (1975) *EPSL, 26*, 69–80. [12] Premo W. R. and Tatsumoto M. (1992) *Proc. LPS, Vol. 22*, 381–397. [13] James O. B. and Flohr M. K. (1983) *Proc. LPSC 13th*, in *JGR, 88*, A603–A614. [14] Nehru C. E. et al. (1978) *Proc. LSC 5th*, 773–788. [15] Stoeser D. B. et al. (1974) *Proc. LSC 5th*, 355–377. [16] Ryder G. et al. (1975) *Moon, 14*, 327–357. [17] James O. B. and McGee J. J. (1979) *Proc. LPSC 10th*, 713–743. [18] Warren P. H. and Wasson J. T. (1978) *Proc. LPSC 9th*, 185–217. [19] Chao E. C. T. et al. (1976) *Proc. LSC 7th*, 2287–2308. [20] McCallum I. S. and Mathez E. A. (1975) *Proc. LSC 6th*, 395–414. [21] Warren P. H. et al. (1983) *Proc. LPSC 13th*, in *JGR, 88*, A615–A630. [22] Premo W. R. and Tatsumoto M. (1991) *Proc. LPS, Vol. 21*, 89–100. [23] Nyquist L. E. et al. (1981) *Proc. LPS 12B*, 67–97. [24] Nakamura N. et al. (1976) *Proc. LSC 7th*, 2309–2333. [25] Compston W. et al. (1975) *Moon, 14*, 445–462. [26] Carlson R. W. and Lugmair G. W. (1981) *EPSL, 52*, 227–238. [27] Lally J. S. et al. (1976) *Proc. LSC 7th*, 1845–1863. [28] Ryder G. (1992) *Proc. LPS, Vol. 22*, 373–380. [29] Papanastassiou D. A. and Wasserburg G. J. (1975) *Proc. LSC 6th*, 1467–1489. [30] Laul J. C. et al. (1989) *Proc. LPSC 19th*, 85–97. [31] Salpas P. A. et al. (1988) *Proc. LPSC 18th*, 11–19. [32] Warren P. H. et al. (1991) *Proc. LPS, Vol. 21*, 51–61. [33] McCallum I. S. et al. (1974) *Proc. LSC 5th*, 287–302. [34] McGee J. J. et al. (1978) *Proc. LPSC 9th*, 743–772. [35] Bickel C. E. (1977) *Proc. LSC 8th*, 2007–2027. [36] James O. B. and Hammarstrom J. G. (1977) *Proc. LSC 8th*, 2459–2494. [37] Ostertag R. et al. (1987) *GCA, 51*, 131–142. [38] McGee J. J. (1989) *Proc. LPSC 19th*, 73–84. [39] James O. B. et al. (1989) *Proc. LPSC 19th*, 219–243. [40] Lindstrom M. M. and Lindstrom D. L. (1986) *Proc. LPSC 16th*, in *JGR, 91*, D263–D276. [41] Turner G. and Cadogan P. H. (1975) *Proc. LSC 6th*, 1509–1538. [42] Jessberger E. K. et al. (1976) *Proc. LSC 7th*, 2201–2215. [43] Ryder G. et al. (1977) *EPSL, 35*, 1–13. [44] Salpas P. A. et al. (1987) *Proc. LPSC 17th*, in *JGR, 92*, E340–E348. [45] Irving A. J. (1977) *Proc. LSC 8th*, 2433–2448. [46] Ryder G. et al. (1975) *Proc. LSC 6th*, 435–449. [47] Nord G. L. Jr. and James O. B. (1978) *Proc. LSC 9th*, 821–839. [48] Compston W. et al. (1977) *Proc. LSC 8th*, 2525–2549. [49] James O. B. (1980) *Proc. LPSC 11th*, 365–393. [50] Warren P. H. and Wasson J. T. (1980) *Proc. Conf. Lunar Highlands Crust*, 81–100.

N93-18790 p. 2

THE APOLLO 17 REGION: A COMPOSITIONAL OVERVIEW. R. Jaumann and G. Neukum, DLR, Institute for Planetary Exploration, Berlin/Oberpfaffenhofen, Germany.

Apollo 17 is located at a mare/highland boundary where the surface shows significant compositional heterogeneities. The composition of surface materials is estimated by analyzing their spectral/chemical correlations. Based on this spectral/chemical analysis the chemical and normative mineralogical composition of two highland units and three mare units has been estimated.

Introduction: The purpose of this investigation is to use spectral/chemical studies in order to determine the composition of Apollo 17 geologic units. The landing site itself and its surroundings are expected to expose different geologic settings ranging from basaltic materials over highland components to pyroclastic deposits [1–3]. Thus, Apollo 17 is an ideal place to study the mechanisms of mare volcanism, interaction of mare and highland materials, and pyroclastic activities. However, the understanding of such processes requires detailed knowledge about the chemical and mineralogical composition of the geologic units studied. Although the composition of the Apollo 17 landing site is known from the analysis of returned samples, only little compositional information can be provided from direct sample studies for the surrounding areas of this complex geologic structure. So far, data of the Apollo 17 Geochemical Orbiter Experiments provide information on the concentration of Fe, Ti, and radioactive elements as well as on Al/Si and Mg/Si concentration ratios [4,5]. This information, however, is restricted to low ground resolutions, which are not sufficient for detailed mineralogical analyses. On the other hand, when we consider spectroscopic measurements of the Apollo 17 area and combine these measurements with the spectral/chemical correlations of Apollo 17 samples, it is feasible to establish a compositional analysis of the Apollo 17 area that is calibrated to the spectral/chemical evidence of the Apollo 17 landing site [6].

Compositional Studies: Compositional information can be derived from an analysis of the spectral/chemical correlations of lunar surface materials. For this purpose the spectral variations of lunar samples are identified and compared with the variations of the concentration of chemical constituents [6]. Such a sample-based analysis provides parameters that can be used to interpret the spectral variations in remotely sensed spectroscopic data. By applying the technique to Apollo 17 soil samples, the concentration of some chemical key elements like Fe, Ti, Al, and Mg can be estimated from spectral measurements. When the technique is transposed to remotely

TABLE 1. Chemical composition of the materials exposed at the local surface points (Fig. 1).

Location	FeO	TiO_2	Al_2O_3	MgO
Ap17	16.8	7.9	12.9	11.0
Ap17 (L)	16.8	8.3	11.8	10.1
T5	19.8	8.9	9.8	8.8
S2	17.1	4.8	10.6	10.7
S3	17.1	7.4	11.7	11.0
S4	16.7	6.3	11.5	11.5
V2	11.2	1.0	20.6	6.3
L1	14.1	2.1	15.0	9.6
L2	14.0	2.2	14.2	11.2
σ	<1.7	<2.0	<2.5	<2.0

Ap17 (L) summarizes the results of the chemical analysis of Apollo 17 samples. All other data are derived from the spectral/chemical analysis of remote sensed measurements. The dimension is wt% and the error (σ) indicates an upper limit for accuracy of the analysis.

TABLE 2. Normative mineralogical composition of the materials exposed at the local surface points (Fig. 1).

Location	Ol	Px	Pl	Il
Ap17	13.8	34.1	36.5	15.6
T5	3.8	49.6	27.3	19.2
S2	8.4	50.1	30.0	11.4
S3	10.0	41.5	33.3	15.2
S4	11.2	43.1	32.5	13.2
V2	15.2	19.0	61.3	4.4
L1	13.6	35.4	44.3	6.6
L2	15.9	37.3	41.3	5.6
σ	<1.5	<2.0	<1.5	<2.0

The dimension is wt% and the error (σ) indicates an upper limit for the accuracy of the analysis. Ol = olivine, Px = pyroxene, Pl = plagioclase. Il = ilmenite.