# ANORTHOSITE EXPOSURES IN THE INNER ROOK MOUNTAINS OF THE LUNAR ORIENTALE BASIN.

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**Introduction:** The Orientale Basin is a 930 km multi-ringed impact structure on the western limb of the Moon. It is the youngest and most well-preserved major lunar basin, having been only partially filled by post-impact mare basalts. As a result of its great size and excellent preservation, the basin's composition and structure provide unique insight into both the Moon's crustal stratigraphy and basin-forming processes in general [e.g. 1-3]. The Inner Rook Mountains (IRM) form Orientale's innermost ring. They represent upper crustal material that was uplifted and exposed at the surface during the basin-forming event, providing a window into the interior of the lunar highland crust [e.g. 1-8].

From earth-based telescopic data [6, 7] and later Clementine multispectral data [8], the IRM were thought to be comprised of anorthosite based on their high albedo and the absence of spectral evidence for mafic minerals. Anorthosite is a rock type consisting of >90% plagioclase that is believed to have formed the upper crust of the Moon when buoyant plagioclase crystals accumulated at the top of a solidifying magma ocean [9-11]. The mineralogy and distribution of anorthosite on the lunar surface may therefore provide key constraints for questions surrounding processes of magma ocean crystallization, such as melt-solid segregation [12] and the scale of lateral homogeneity (e.g. serial magmatism vs global crystallization).

Recent observations by the the SELENE Multiband Imager (MI) and the Chandrayaan-1 Moon Mineralogy Mapper ($M^3$) orbital near-IR spectrometers have confirmed the anorthositic character of the IRM by identifying the diagnostic plagioclase absorption feature (~1250 nm) in a few isolated exposures [13, 14]. The presence of this absorption feature, which is caused by electronic transitions of trace $Fe^{2+}$ in plagioclase [15-17], requires that the mineral is in its crystalline form, constraining peak shock pressures experienced by the mineral to <25 GPa [18, 19]. Further, the strength of the feature is related to the abundance of plagioclase vs. mafic minerals in the host rock. With these new datasets, it is therefore possible to elucidate important information regarding the purity of anorthosite in this region and aspects of the pressure regime during the basin-forming process.

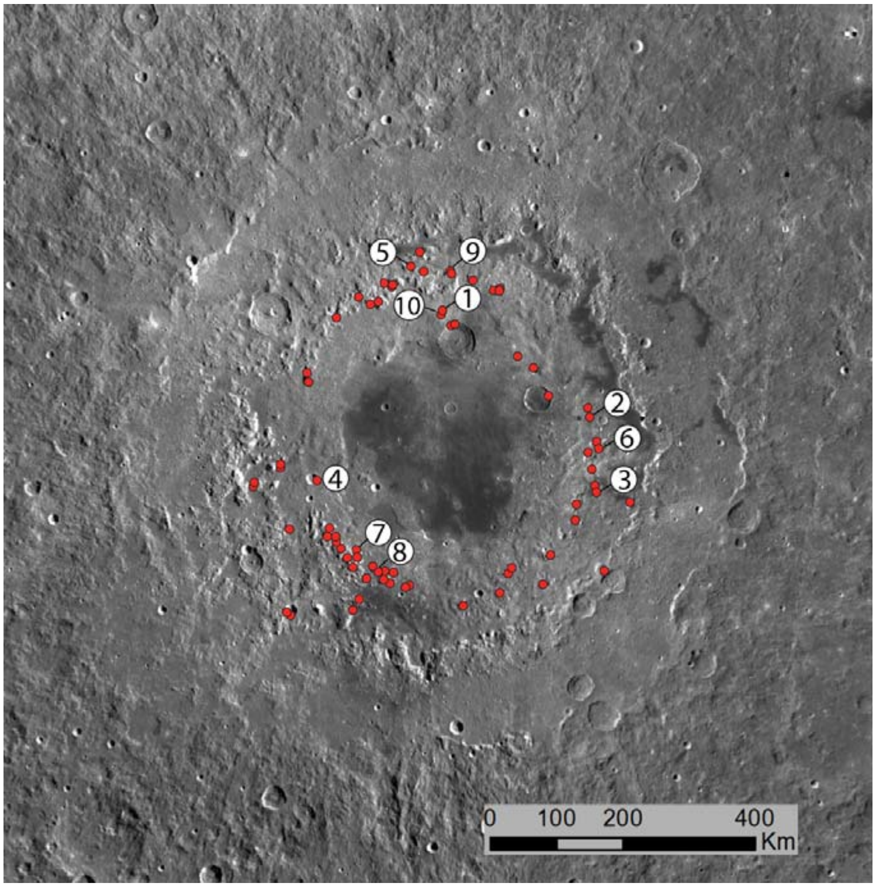

Figure 1: $M^3$ data showing pure crystalline plagioclase detections (red) overlaid on an LROC WAC mosaic. Numbered locations correspond to the spectra in Fig. 2.

**$M^3$ data and methods:** Here we present $M^3$ data describing the spatial distribution and spectral character of crystalline plagioclase detections in Orientale's IRM. These results are part of a larger global survey of $M^3$ plagioclase identifications, details of which can be found in [20]. The $M^3$ instrument is a hyperspectral imaging spectrometer that acquired images of the lunar surface in 85 wavelengths simultaneously at a spatial resolution of 140 m/pixel [21]. It therefore not only provides the high spectral resolution necessary to characterize the plagioclase absorption, but also places this mineralogic information in spatial context.

We have developed a spectral parameter specifically to aide in the identification of crystalline plagioclase in $M^3$ data:

$$\sum_{n=0}^{26} 1 - R(1029+20n)/Rc(1029+20n)$$

Here, a continuum is removed between 1030 and 1700 nm and band depths are summed over all 27 channels between 1030 and 1580 nm. Using this parameter we located pixels within the Orientale Basin that have a plagioclase absorption and have verified these detections by evaluating their spectra.

**Results:** Crystalline plagioclase exposures in Orientale are almost exclusively confined to the IRM (Fig. 1). They occur primarily on the crests and steepest slopes of massifs and in certain small (<30 km) craters. Non-IRM occurrences include a few massifs in the Outer Rook Mountains, as well as the walls of Maunder, Kopff, and other small craters in the basin floor. Crystalline plagioclase is not identified in any of the central peaks within the Orientale Basin, consistent with the interpretation that lunar anorthosite is derived from the upper portions of the crust [e.g. 8].

The IRM anorthosites are remarkably pure. Nearly all spectra with a plagioclase absorption show the ~1250 nm feature exclusively (Fig. 2a, b); the lack of mafic signatures likely constrains the abundance of plagioclase in these rocks to 98% or greater [13, 22]. However, spectra from a few isolated locations do show a weak pyroxene absorption in addition to the plagioclase feature (Fig. 2c). Comparing the shapes of these spectra with preliminary nonlinear mixing calculations [22] suggests that the crystalline components of these particular “impure anorthosite” exposures contain closer to only 95% plagioclase (here, ‘impure anorthosite’ refers to spectral purity: all spectra discussed are likely anorthosite by definition, but those with spectral evidence for mafic minerals are referred to as ‘impure’).

**Conclusions:** Crystalline plagioclase is pervasive throughout Orientale’s Inner Rook Mountains, indicating that an extensive, coherent layer of anorthosite was sampled by the impact. Further, the IRM anorthosites are comprised nearly exclusively of plagioclase (likely >98%) that in many instances did not experience shock pressures >25 GPa [18, 19]. The purity, crystallinity, and geographic extent of these anorthosites should place valuable constraints for models of magma ocean crystallization and impact basin formation.

**References:** [1] Head J. W. (1974) *Moon, 11,* 327-356. [2] Moore H. J. et al. (1974) *PLPSC 5th*, 71-100. [3] Wilhelms D. E. (1987) *USGS* 1348. [4] Grieve R. F. et al. (1981) *PLPSC 12A*, 37-57. [5] Croft S. K. (1981) *PLPSC 12A*, 227-257. [6] Spudis P. D. et al. (1984) *PLPSC 15th*, C197-C210. [7] Hawke B. R., et al. (2003) *JGR, 108 (E6)*, 5050. [8] Head J. W. et al. (2010) *LPS XLI*, Abstract #1030. [8] Bussey B. J. and P. D. Spudis (2000) *JGR, 105(E2),* 4235-4243. [9] Wood J. A. et al. (1970) *PLPSC, 11,* 965–988. [10] Dowty E. M. et al. (1974) *EPSL, 24,* 15-25. [11] Warren P. H. (1990) *AM, 75,* 46-58. [12] Parmentier E. M. and Y. Liang (2010) *LPS XLI,* Abstract #1824. [13] Ohtake M. et al. (2009) *Nature, 461,* 236-240. [14] Pieters C. M. et al. (2009) *LPS XL,* Abstract # 2052. [15] Adams J. B. and T. B. McCord (1971) *PLPSC 3,* 2183-2195. [16] Bell P. M. and H. K. Mao (1973) *GCA, 37,* 755-759. [17] Adams J. B. and L. H. Goulland (1978) *PLPSC 9,* 2901-2909. [18] Adams J. B. et al. (1979) *LPS X,* Abstract # 1. [19] Johnson J. R. and F. Horz (2003) *JGR 108(E6),* 5120. [20] Donaldson Hanna K. L. et al. *these abstracts*. [21] Green R. O. (2011) *JGR, 116,* E00G19. [22] Cheek L. C. and C. M. Pieters, *these abstracts*.

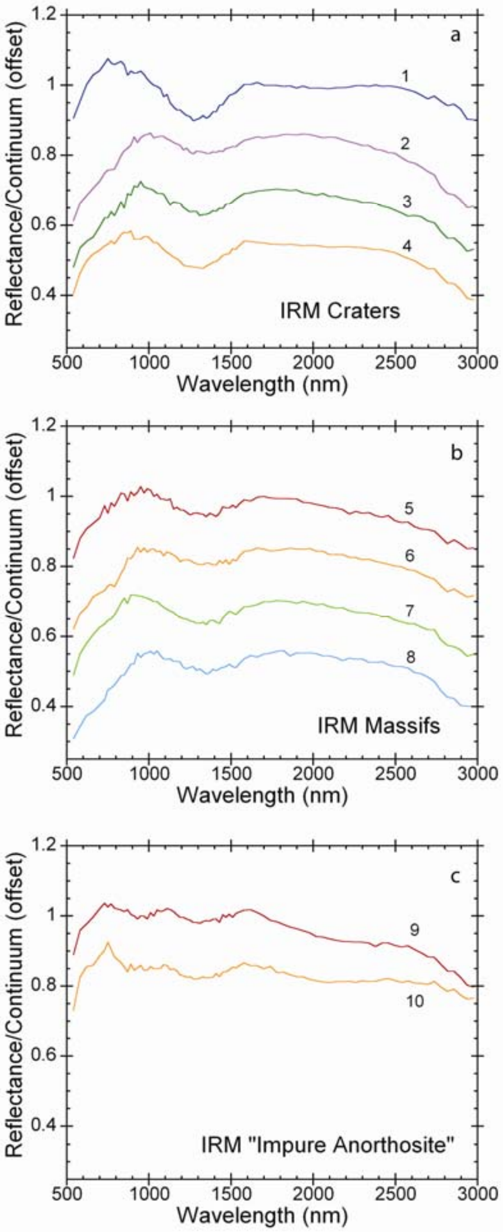

Figure 2. Representative spectra from the IRM. a) pure anorthosite in craters b) pure anorthosite on massifs c) anorthosite spectra with greater pyroxene abundance (plagioclase <98%). A continuum has been removed around 1030-1700 nm, and spectra are offset for clarity. Numbers correspond to locations in Fig. 1.