THERMOCHRONOMETRIC CONSTRAINTS ON BASIN-FORMING IMPACT EVENTS FROM ^{40}AR/^{39}AR ANALYSES OF EXCAVATED CRUSTAL ROCKS. W. S. Cassata[1], L. E. Borg[1], C. A. Crow[1]
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Introduction: The history of lunar cratering is difficult to constrain as the Moon's surface has been reworked and surface rocks have been redistributed great distances by large, basin-forming impacts. Our understanding of the timing of basin-forming impacts is based largely on ^{40}Ar/^{39}Ar dating of impact melt rocks. A key observation that arose out of the Apollo era and years that followed is the majority of impact melt rocks and breccias in the Apollo collections formed ~3.8 – 4.0 billion years ago [e.g., 1-3]. One end-member explanation for this observation is the Late Heavy Bombardment Hypothesis [e.g., 4-5]. The other end-member explanation is a bias in the sample collection, either due to the geographically restricted sampling locations of the Apollo missions, stratigraphic superposition of younger impact ejecta in the regolith, and/or a lower probability of preserving older K-Ar ages during an era of intense cratering [e.g., 6].

There are many crustal rocks and clasts that yield ^{40}Ar/^{39}Ar ages that are significantly older than 3.8 – 4.0 Ga, but are younger than Sm-Nd, Rb-Sr, and U-Pb ages obtained on the same sample. While impact resetting is a plausible explanation for post-crystallization ^{40}Ar/^{39}Ar ages, other processes like conductive cooling in the crust and magmatic overprints cannot be dismissed. Moreover, the size of an impact event responsible for resetting the K-Ar system (i.e., basin forming or more local in scale) can be difficult to determine. Despite these challenges, crustal rocks can provide constraints on the bombardment history of the Moon prior to 3.8 – 4.0 Ga [1,6].

^{40}Ar/^{39}Ar thermochronometry offers means to assess (1) whether ages obtained from exhumed crustal rocks can be ascribed to impact events, and (2) the size of the impact event responsible for resetting the K-Ar system. Here, ^{40}Ar/^{39}Ar data obtained from anorthosite 60025 are coupled with thermal modeling to illustrate the limitations of the approach and to place constraints on a basin-forming event prior to 4.25 Ga.

Sample and methods: 60025 is a ferroan anorthosite from the Apollo 16 site. Modeling of exsolution development in pyroxene indicates a depth of formation of ~21 km [8]. Temperature-controlled diode laser heating with a co-aligned optical pyrometer was used to incrementally degas a mg-sized fragment of 60025. A **^{40}Ar/^{39}Ar age of 4285 ± 40 Ma** was obtained, consistent with results recently reported by [9]. The ^{40}Ar/^{39}Ar age is younger than both the Sm-Nd and Pb-Pb ages of 4367 ± 11 and 4359 ± 2 Ma [10], respectively, which suggests it may have been reset by an impact event. In the following section, diffusion data derived from the step-heating experiment are used along with petrographic constraints on the crystallization of 60025 to determine its thermal history.

Thermal modeling: *Question 1: Was the K-Ar system open or closed during long duration crustal residence (i.e., could the ^{40}Ar/^{39}Ar age reflect conductive cooling in the crust)?* Figure 1 shows the fractional retention of Ar during long duration residence in the lunar crust at different depths. At depths greater than ~30 km, Ar is effectively lost from 60025 at a rate faster than it is produced by radioactive decay. Radiogenic production exceeds loss by diffusion at a depth of ~24 km. At depths shallower than ~18 km, Ar is quantitatively retained. The inferred depth of formation of 60025 is 21 km [7], which is near the upper boundary of the Ar partial retention zone. As such, the post-crystallization ^{40}Ar/^{39}Ar age could plausibly reflect conductive cooling or a magmatic overprint that occurred in the crust. If the diffusive length scale was larger prior to exhumation (i.e., prior to shock) then it is even more likely that the K-Ar system was closed.

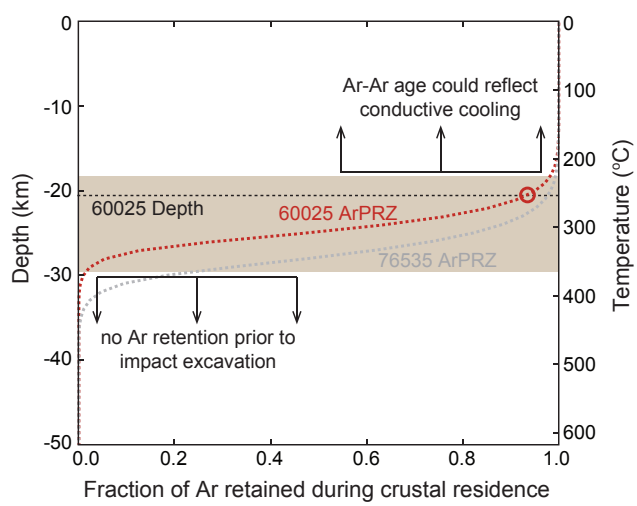

Figure 1: Ar partial retention in the lunar crust for sample 60025, calculated assuming a thermal gradient of 12.3 °C/km and using the diffusion parameters obtained from the incremental heating experiment (not shown).

Question 2: Would the K-Ar system be reset during excavation from 21 km depth? Figure 2 shows the fractional loss of Ar as a function of deposition location during the cooling of two ejecta blankets. The equation of [11] was used to obtain ejecta thickness as a function of crater size and distance from the transient crater

center. The models of [12] were used to constrain initial ejecta temperature (T_o) as a function of crater size. Given these starting conditions, the ejecta blankets were emplaced over cold regolith and allowed to conductively cool at the base and surface. Diffusive loss of Ar was modeled using diffusion parameters obtained from the incremental heating experiment (not shown).

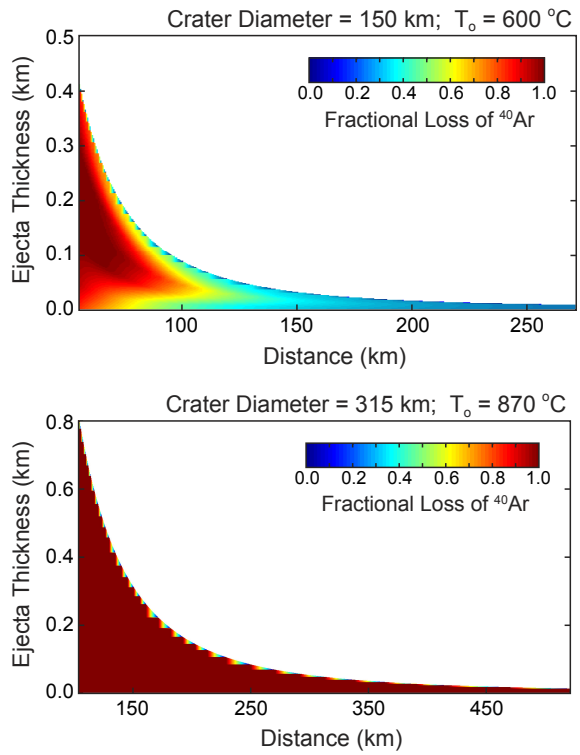

Figure 2: Predicted fractional loss of Ar from 60025 as a function of location within ejecta blankets associated with 150 km (a) and 315 km (b) craters.

For impact events generating a crater with diameter D = 150 km, mean ejecta blanket temperatures reach ~600 °C [12], and the fractional loss of Ar varies considerably with deposition location. Approximately 25 vol.-% of the ejecta blanket reaches sufficient t-T conditions to reset the K-Ar system (defined here as >90% loss). As the ejecta blanket thickens and the temperature increases in larger cratering events, the likelihood of resetting increases significantly. Assuming the excavation depth is 1/10th the diameter of the transient impact cavity and the ratio of observed crater rim diameter to transient cavity diameter is ~1.75 [13,14], excavation to 21 km depth requires a crater of D >315 km. Impact events of this size generate t-T conditions sufficient to reset plagioclase in 60025 in >95 vol.-% of the ejecta blanket. Thus the impact event that excavated 60025 is likely to have reset the K-Ar system. The probability of being reset in much smaller ejecta blankets associated with a crater of D = 100 km is also significant (>5%).

Impact history recorded by 60025: Collectively, the thermal modeling and inferred depth of crustal residence indicate (1) 60025 may have cooled below the Ar closure temperature during crustal residence, and *either* (2a) 60025 was excavated from ~21 km at 4285 ± 40 Ma resulting in a crater with D >315 km *or* (2b) 60025 was excavated from ~21 km depth prior to 4285 ± 40 Ma and was reset by a subsequent impact event at 4285 ± 40 Ma. Regardless, the modeling indicates that an impact event producing a crater with D >315 km occurred between 4285 ± 40 and the crystallization age of the sample at 4359 ± 2 Ma. Sample 60025 appears to have escaped heating associated with large (D > 100 km) impact events since 4285 Ma.

Conclusion: $^{40}Ar/^{39}Ar$ ages obtained from some exhumed crustal rocks, like 60025 and 76535 [6,14], constrain the timing of basin forming impact events prior to 3.8 – 4.0 Ga. In some instances, either because diffusion kinetics are not known or the depth of crustal residence is poorly constrained, it is not possible to determine whether a $^{40}Ar/^{39}Ar$ age reflects impact resetting or conductive cooling. For this reason, additional combined petrographic and thermochronometric studies would be useful in better understanding the relationship of crustal rocks to the bombardment history of the Moon. $^{40}Ar/^{39}Ar$ thermochronometry using pyroxenes may also help to distinguish conductive cooling from impact excavation/resetting, as this system has a significantly higher closure temperature (600-800 °C).

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