Introduction: The bulk density of the porous lunar crust has been mapped using high-resolution gravity provided by the Gravity Recovery and Interior Laboratory (GRAIL) mission [1, 2, 3]. However, the vertical structure of the crust, which is key to understanding its thermal [4, 5] and seismic [6] characteristics, and its origin and subsequent modification, is currently poorly known [2, 3]. Here, we analyze GRAIL data using an admittance approach to determine the vertical density structure of the lunar crust.

Approach: We used spherical harmonic coefficients of the Moon’s gravity [1], topography [9], and topography-induced gravity [2] fields up to spherical harmonic degree $\ell = 550$. The gravity data were derived from the GRAIL nominal and extended missions’ tracking data; the topography data were derived from the principal axis referenced Lunar Orbiter Laser Altimeter (LOLA) data [9]. A localized, multitaper spherical harmonic analysis was performed [7, 8] at each position on a grid of 400 nodes distributed in a quasi-equal area fashion [2], yielding a node spacing of $\sim$290 km. Optimal windows (tapers) of large spectral bandwidth ($L = 58$) were used to localize the gravity and topography-induced gravity data in spherical cap regions $15^\circ$ in radius. The resulting fields were expanded in spherical harmonics, from which the effective density $\rho_{\text{eff}}$ (which compares the observed gravity to the predicted gravity) as a function of $\ell$ was calculated:

$$\rho_{\text{eff}}(\ell) = \frac{S_{gb}(\ell)}{S_{fg}(\ell)},$$

where $g$ and $b$ refer to the observed gravity and to the gravity predicted from topography assuming unit density, respectively, and $S_{fg}$ is the cross-power spectrum of two functions $f$ and $g$ on the sphere.

The resulting localized effective density spectra typically show an increase in density with decreasing $\ell$ (e.g., Figure 1), implying that density increases with depth [3, 10]. These spectra were fit to two kinds of theoretical spectra assuming either a linear or an exponential density profile in the crust. The fits were performed for the high-degree part of the data ($250 \leq \ell \leq 550$), in order to avoid the effects of flexure and/or crustal thickness variations at lower degrees. Our method is therefore mostly sensitive to the shallow density structure of the crust (typically depths $z < 10-15$ km). The best-fit parameters describing the theoretical spectra were derived using a grid search.

Linear density model: Figure 2 shows the spatial distribution of the best-fitting linear density gradient $a$. The mare regions exhibit high surface densities and a distinct decrease in density with depth, as expected if (high density) mare basalts over-
Exponential density model: We now focus on regions characterized by a density increase with depth, in particular the farside and the SP-A basin region. In the following, regions with gradients less than $+5\,\text{kg}\,\text{m}^{-3}\,\text{km}^{-1}$ (our finite grid-search step) were excluded (purple dashed line in Figure 2). We assume a more realistic density distribution consisting of an exponential profile given by $\rho(z) = \rho_{\text{surf}} + \Delta \rho \left(1-e^{-z/d}\right)$, where $d$ is an e-folding depth and $\Delta \rho$ is the total density contrast across the crust. We assume that $\rho_0 = \rho_{\text{surf}} + \Delta \rho$ is that of the intact surface rock (i.e., grain) density. For each location, the local, window-averaged value of $\rho_0$ was calculated from a grain density model [12], although our results did not change if a single, constant value for $\rho_0$ was used. For the farside highlands as a whole, the best-fit density profile characteristics are: $d = 10 \pm 2\,\text{km}$ and $\rho_{\text{surf}} = 2261^{+34}_{-42}\,\text{kg}\,\text{m}^{-3}$, corresponding to a typical surface porosity of 21-24% (the average grain density being 2917 $\text{kg}\,\text{m}^{-3}$); the equivalent density gradient with a linear model is $a \approx 30\,\text{kg}\,\text{m}^{-3}\,\text{km}^{-1}$. Such high inferred surface porosity values are compatible with Apollo samples and lunar meteorites [13]. Figure 3 shows the spatial distribution of the e-folding lengthscale $d$. The most striking result is that the SP-A impact basin region has a significantly shallower low-density (porous) layer than the rest of the farside, with $d = 5-10\,\text{km}$, instead of 15-25 km.

The SP-A region appears to possess a shallower porous layer than the rest of the farside. This could be the result of: impact-induced removal (excavation) of pre-existing fractured material; annealing of pre-existing fractures [2] within SP-A; generation of a thick, pore-free impact melt sheet [15] within SP-A; intrusive processes; redistribution of thick (porous) SP-A ejecta deposits over the rest of the farside; or a combination of these factors. The SP-A impact may have caused widespread deposition of up to a few kilometers of ejecta [16, 17].

Conclusion: Mare regions are characterized by a distinct decrease in density with depth, while the farside is characterized by an increase in density with depth at an average rate of $\sim 30\,\text{kg}\,\text{m}^{-3}\,\text{km}^{-1}$ and typical surface porosities of 20%. The Apollo 12 & 14 landing site region has a similar density structure to the farside, permitting a comparison with seismic velocity profiles [18, 19]. The SP-A impact basin region appears distinct with a near-surface low-density (porous) layer 2-3 times shallower than the rest of the farside. This result suggests that redistribution of material during the large SP-A impact likely played a major role in sculpting the lunar crust. Mapping the spatial distribution of shallow porosity, as we have attempted here, will allow comparison with other data sets (e.g., seismic velocities, surface composition).