CONSTRAINTS ON THE TEMPERATURE AND MINERALOGY OF THE MOON FROM A JOINT INVERSION OF APOLLO SEISMIC, GEODETIC DATA AND LP-CLEMENTINE GRAVITY DATA

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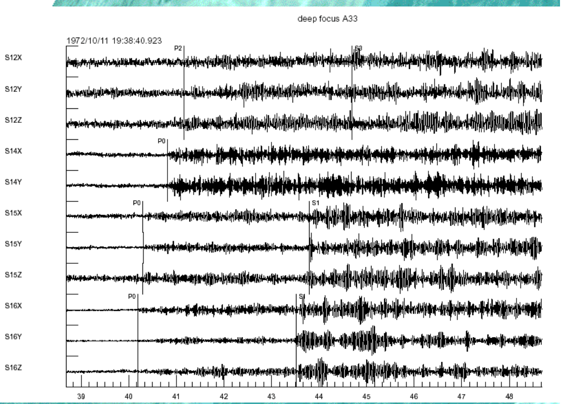
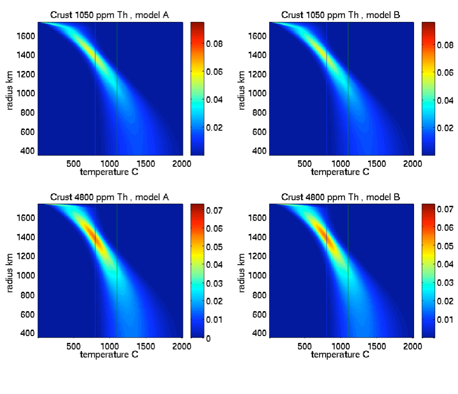
Abstract: We determine seismic profiles of the Moon interior using travel times obtained from a re-analysis of the Apollo data. Due to the small amount of informations, care has been taken in associating all these secondary seismic data with errors and modeling possible lateral variations of the crust. Our model confirms a mean crust of about 40 km, an upper mantle and a more primordial lower mantle. Mean values in the upper mantle are rather well constrain in the range 7.6-7.8 km s^{-1} and 4.4-4.5 km s^{-1} for P and S respectively. Velocity variations in the middle mantle are more cautious. In the zone of deep Moonquakes, velocities slightly increase to 4.6-4.7 ± 0.15 km s^{-1} and 8.2-8.4 ± 0.3 km s^{-1} for P and S waves respectively. We then perform an aposteriori analysis of the seismic profile in term of mineralogy and temperature profile by using the gravity and geodetic constrains. Our preferred models are a pyroxenite model for the upper mantle and a magnesian model for the lower mantle, with increasing Mg# with depth. We find temperature of 1073K (elastic lithosphere limit) and 1473K (thermal lithosphere limit) for radius of about 1400 km and 1000 km and show that the lunar mantle is probably depleted by about 70% compared to an Earth reference of 25.7 ppb. Taking a mean crustal thickness of 40 km with 1010 ppb in Th and our mantle value for the depletion, we then find a bulk Th and U abundance comparable to the Earth values within the error bars, and even possibly smaller

Step 1: Arrival times determinations

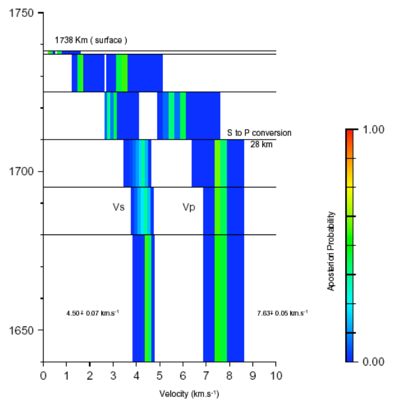


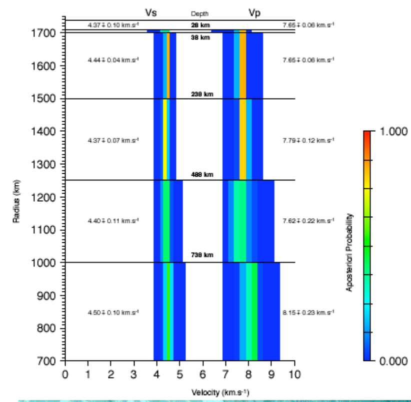
Importance of data weight and of model parameters: Errors bars ranging from 1sec to 30 sec are associated to arrival times. The mean error in the data set is about 2sec. Very large differences are observed, up to 25 sec between the S arrival times picking listed in Lognonné et al. [2003] and other data sets published in the literature, showing the importance of data's weight with respect to their apriori quality for velocity profile determinations. The quality of our error determination has been for example checked with respect to the arrival times of Nakamura [2004] for all deep events. For the 142 common arrival times, we have a root mean square difference of 6.8 sec if the differences are not weighted by errors (or if they are all equally weighted). By using our errors values as weights in the variance computation, we found a root mean square difference of 2.8 sec between the two data sets, still greater than the statistical mean error (1.7 sec) but nevertheless much smaller than the unweighted root mean square difference. The choice for model parameters is done with respect to the mean velocities in a few layers representative of the crust, upper mantle and lower mantle. If such representation is probably not optimized with respect to possible discontinuities, it nevertheless induce a theoretical error in the ray tracing and arrival time determination much smaller than 2sec.



Moon Temperature profiles

Step 2: seismic inversion





Top: Obtained seismic models for the crust (left) and the mantle (right) for a mean spherical model, in term of a posteriori probability. The bimodal distribution reflects probably the signature of lateral variations and a crust of about 30 km is found. Left: crustal thickness determination for inversion based only on the impact taking into account lateral variation in the crust. The crust ranges from 34 to 41 km. The mean crust can be estimated, due to the highlands, to about 40 km.

Site	Lat	Lon	R_{topo}	R^{med}_{Moho}	H_{seis}	σ_{seis}	n_{data}
#1 (A12)	-3.04	-23.42	1736.0	1701.7	34.3	3.7	24
#2 (A14)	-3.65	-17.48	1736.2	1702.6	33.7	4.0	21
#3 (A15)	26.08	3.66	1736.1	1701.8	34.2	4.8	21
#4 (A16)	-8.97	15.51	1737.6	1696.8	40.8	4.1	19

Step 3: Mineralogical determination with geodetic constrains

Due to the limited resolution, we test a set of proposed models with our obtained seismic velocities (See Table 1) instead inverting for their composition, choosing the temperature as new parameter for the inversion. Two models fit better than all other ones and are considered (model 5 of table 1 and model 7). Constraints from inertia and mean density data can now be used, together with the crustal determination. They lead to the rejection of model 7, for which too high densities are found. The same procedure is done for the lower mantle, and leads to selection of model IV. The last procedure can be started, leading to temperature estimations

Step 4: Temperature inversion

Our temperature model is primarily defined by the temperature at the base of the crust and by the temperature gradients in the upper and lower mantle. Although complex models can be used for more detailed studies, we choose a rather simple thermal model, due to the low quality of the seismic constraints. We took 4 layers, two for the crust and two for the mantle. For each layer n, the temperature is provided by the steady state equation in spherical coordinates,

$$\frac{1}{r^2}\frac{\partial}{\partial r}(r^2 k_n \frac{\partial T}{\partial r}) + \rho_n H_n = 0 ,$$

where k_n, ρ_n and H_n are the thermal conductivity, density and heating constant per unit of mass of the layer n. Let us note T_n the temperature at the top of each layer. In addition to these equations, let us note that temperature and heat flux are continuous at each boundary and let us neglect the heat flux at the bottom layer, i.e. at the interface between the mantle and core of our model. In all the following tests, the surface temperature will be taken as 250 K, the thermal conductivity of the deep crust and mantle will be taken equal to 2 and 3.3 $Wm^{-1}K^{-1}$ respectively. The other parameters of our model are the heating constant H_n in all layers and the thickness and conductivity of the top most regolith layer in the crust, acting as an insulator. Compared to a reference of 25.7 ng/g of U, equivalent to a H value of 6.18 10^{-12} J/kg^3 with a Th/U ratio of 3.67, the crust is strongly enriched and the mantle are probably strongly depleted.

Upper mantle ⇩ Lower mantle ⇩

SiO_2	43.69	42.30	46.10	44.78	**54.13**	52.3	**50.2**	54.0	44.3	47.5	48.9	**42.8**	44.54	44.6
Al_2O_3	7.65	3.62	3.51	4.32	**5.10**	4.0	**4.0**	4.0	0.6	3.39	3.6	**3.8**	2.80	4.3
FeO	13.12	16.62	12.62	9.14	**13.76**	20.7	**17.6**	23.8	9.9	13.0	10.5	**12.8**	10.24	9.1
MgO	29.36	34.54	34.97	38.25	**22.94**	20.0	**25.2**	15.2	44.7	32.9	34.2	**37.6**	37.94	38.1
CaO	6.18	2.92	2.80	3.51	**4.07**	3.0	**3.0**	3.0	0.7	2.6	2.7	**3.0**	3.32	3.5
Mg#	80	78.7	83.2	88.2	**74.8**	63.3	**71.9**	53.2	88.9	81.9	85.3	**84**	86.9	88.2

Table 1: Mineralogical models tested in the study for the upper-middle mantle and the lower mantle. For the upper-middle mantle, Model 1 is an Al and Ca-rich composition[*Morgan et al*, 1978]. 2 is a Fe-rich composition and 3 an intermediate model with orthopyroxene [*Jones and Delano*, 1989]. 4 a pyrolite composition[*Ringwood and Irifune*, 1988]. 5 a model of lunar pyroxenite constrained by the source of mare basalt at depths of 200-500 km[*Ringwood and E. Essene*, 1970]. 6 to 8 are pyroxenite models satisfying the mean velocity the upper velocity bound and lower velocity bound of *Nakamura* [1983] model obtained by *Kusvov* [1995]. For the lower mantle, model 1 is from *Taylor and Jakes* [1977], model 2 results from and impactor model [*Taylor*, 1987], Model 3 is with supplementary constrains on FeO to accomodate density and magnetic requirements [*Taylor and Jakes*, 1977] and model 4 is from *Jones and Delano* [1989].

Results: The temperature gradient in the mantle is mainly constrained by the depletion in U in both the upper and lower mantle and the left Figure shows the space of acceptable values for these temperature. We generally find temperature of 1073K (elastic lithosphere limit) and 1473K (thermal lithosphere limit) for radius of about 1400 km and 1000 km, comparable to the depth found in thermal evolution models [Spohn et al., 2001]. We find a peak probability at about 70% depletion for the upper and lower mantle in the PKT case compared to the Earth reference of 25.7 ppb and retrieve for the PKT crust a thickness of about 30 km from temperature constrains. This yields to about 8.2 ppb in U and 30 ppb in Th in the mantle, values proposed for the abundance given by Waenke et al. [1977] and Taylor [1982]. In the mean case, due to the smaller heating of the crust, we find a depletion in the range of 60-65%, providing abundance by about 15% larger. Taking a mean crustal thickness of 40 km with 1010 ppb in Th and this mantle value for the depletion, we find a bulk Th and U abundance comparable to the Earth values within the error bars, and even possibly smaller.