The Deep Structure of Lunar Basins:
Implications for Basin Formation and Modification

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We present models for the structure of the crust and upper mantle beneath lunar impact basins from an inversion of gravity and topographic data from the nearside of the moon. All basin models display a thinner crust and an elevated Moho beneath the central basin region compared to surrounding areas, a signature of the processes of basin excavation and mantle uplift during collapse of the transient cavity. There is a general decrease in the magnitude of apparent uplift of mantle material with increasing basin age; we attribute this relation primarily to enhanced rates of ductile flow of crustal material early in lunar history when crustal temperatures were relatively high and the effective elastic lithosphere was thin. The more relaxed topographic and Moho relief associated with older basins on the central nearside may, in particular, be at least partly a consequence of the extensive subsurface heating associated with the formation of the large Procellarum basin. The deep structure of the youngest basins constrains the geometry of the cavity of excavation and the amount of crustal material ejected beyond the basin rim.

From the volumes of the topographic basin, of mare basalt fill, and of uplifted mantle material, the volume of crustal material ejected beyond the basin rim for an Orientale-sized event was of the order of \(10^7 \text{ km}^3\). A near-constant thickness of nonmare crustal material beneath the central regions of young basins of various diameters and preimpact crustal thicknesses suggests that the transient cavity excavated to at least the base of the crust for the largest basins; significant excavation into the mantle may have been impeded by an abrupt increase in strength at the lunar Moho.

INTRODUCTION

The formation of multiring impact basins has played a major role in the geological evolution of the moon. During the first billion years of lunar history the impact of large projectiles onto the lunar surface resulted in the excavation of basin cavities hundreds of kilometers in diameter [Wood and Head, 1976] and the implantation of large quantities of heat into the lunar interior [O'Keefe and Ahrens, 1977]. Impact basins also became the focus for volcanic and tectonic activity over a considerable time period following the basin formation events [e.g., Head, 1976; Solomon and Head, 1979; Solomon et al., 1982]. Important constraints on the processes of basin formation and modification are provided by the present volumes of the topographic basin, of material ejected during basin formation, and of mare basalt fill as well as by the degree of involvement of the mantle in isostatic compensation of basin relief. The topographic volumes of the basins are reasonably well known. Estimates for the volumes of mare basalt and basin ejecta deposits have also been obtained from photogeological studies [e.g., Moore et al., 1974; Head et al., 1975; DeHon and Waskom, 1976; Head, 1982], but such techniques are generally limited in their ability to resolve the thickness of the deposits. The depth of excavation of several basin-forming events on the moon has also been estimated from the chemistry and mineralogy of ejecta deposits inferred from remote sensing data [Spudis, 1982, 1983; Spudis et al., 1984], but such estimates depend critically on the accurate identification of primary ejecta and on assumptions about chemical layering of the lunar crust. In this paper, we apply gravity and topographic data to infer the three-dimensional structure of the crust and upper mantle beneath impact basins on the lunar nearside. With the derived structural models we constrain several of the important geometrical and physical parameters related to the processes of basin formation and modification, and we assess their variations with basin age and size.

Muller and Sjogren [1968] were the first to recognize that the youngest nearside mare basins are characterized by positive gravity anomalies, which they attributed to "mascons." Since that discovery, a number of efforts have been made to model the gravity anomalies over mascon basins with contributions to the anomalous mass placed at the surface [Conel and Holstrom, 1968; Baldwin, 1968; Booker et al., 1970], at the lunar Moho [e.g., Wise and Yates, 1970], or at both locations [e.g., Hulme, 1972; Wood, 1972; Bowin et al., 1975; Sjogren and Smith, 1976]. At least some portion of the anomalous mass contribution to mascon anomalies resides near the surface [Phillips et al., 1972], but models where mare basalt fill is the sole source of anomalous mass require an unreasonable thickness of mare basalt to fit the measured gravity field [Thurber and Solomon, 1978]. While both mare fill and an elevated Moho likely contribute to the observed gravity, the solution for the distribution of anomalous mass between the two locations given only gravity and topographic data requires additional assumptions.

Structural models consistent with gravity and topographic
data have been constructed for a number of individual basins [e.g., Bowin et al., 1975; Sjogren and Smith, 1976; Phillips and Dvorak, 1981; Janle, 1981a, b]. Most of these models, however, were developed under dissimilar sets of assumptions and constraints, thus hindering a comparison of the inferred structures beneath different basins. Also, the interpretation of gravity data over a single basin requires that the investigator make subjective judgements about regional trends in the gravity data arising from structures outside the area of interest or occurring over wavelengths greater than the scale of the basin. In contrast, global models for lunar crustal structure [Wood, 1973; Bills and Ferrari, 1977a; Thurber and Solomon, 1978] calculated under uniform sets of constraints and assumptions permit an internally consistent assessment of structural variability on a regional scale. These global models were developed to address crustal structure at scales greater than the dimensions of most lunar basins, however, and with the exception of the study of Thurber and Solomon [1978], none considered mare basalt as a significant contributor to the observed gravity field.

In this paper we determine models for the crustal structure in the vicinity of nine impact basins on the lunar nearside (Figure 1). The models are derived as part of a simultaneous inversion of nearside gravity and topographic data, using a procedure similar to that of Thurber and Solomon [1978]. The moon is divided into a grid of blocks, each 5° x 5° in horizontal extent (Figure 2). These block dimensions represent the approximate limits of resolution of the available gravity data, as discussed further below. The disturbing gravitational potential at any point over the nearside can be approximated as the sum of the potentials due to the distribution of anomalous mass within each block. In this section we describe the adopted procedure for calculation of gravity anomalies, as a forward problem, given a distribution of topography or anomalous mass that is uniform within each block. We then discuss the constraining assumptions that we have chosen to make the inverse problem well posed. Finally, we describe an iterative, linearized inversion procedure to determine the crustal structure within each block from gravity and topography, subject to the adopted constraints.

**Computation of Gravity Anomalies**

In previous analyses of the gravity anomaly fields of planets using spherical shell segments [Morrison, 1976; Thurber and Solomon, 1978], contributions to the anomalous mass within each segment have been approximated by uniform surface masses located at a constant radius from the center of the planet. While this method simplifies the computation of the gravitational potential, it may provide a poor approximation to the actual potential when the thickness of the block is similar in magnitude to the distance between the block and the observation point.

As an improved approximation, we represent contributions to the anomalous mass within a spherical shell segment as
arising from a finite volume of thickness $b_r$ and of latitudinal and longitudinal extent $b_{\phi}$ and $b_{\lambda}$, respectively. In general, the disturbing potential at a point $r$ above the lunar surface due to a block of anomalous mass within a spherical shell segment is

$$u(r) = \int \int \int F \, d\theta' \, d\phi' \, dr'$$

where

$$F = G \rho(r') \left( \frac{r'}{r} \right)^2 \cos \theta'$$

and $G$ is the gravitational constant, $\rho$ is the density (or density contrast), $\theta$ is the latitude, $\phi$ is the longitude, and primed variables denote coordinates of anomalous mass within the block. All radius vectors are with respect to the lunar center of mass. If $F$ is expanded to second order in a Taylor series about the center of the block, equation (1) simplifies to

$$u(r) = b_0 b_{\phi} b_{\lambda} \left[ F + \frac{1}{2} (F_{00} b_{\phi}^2 + F_{0\phi} b_{\phi}^2 + F_{\phi\phi} b_{\phi}^2) \right]$$

where the double subscripts designate second derivatives of $F$ with respect to the primed coordinates evaluated at the center of each block.

The gravitational potential due to the anomalous mass within all $M$ blocks in the grid is obtained by summation:

$$U(r) = \sum_{j=1}^{M} u_j(r)$$

and the free air gravity anomaly field $g(r)$ is obtained by the simple finite difference relation:

$$g(r) = \frac{U[r + \Delta r(r/r)] - U(r)}{\Delta r}$$

Equation (3) is used to compute the contributions from blocks beneath and near the observation point. For a block more than 900 km distant from the observation point the distribution of anomalous mass in the right-hand side of (1) for that block is represented by four point masses; for a block more than 1200 km distant, a single point mass is used. These simple approximations provide accuracy comparable to that of (3) with a considerable savings of computation time.

### Constraints and Assumptions

Following Thurber and Solomon [1978], we assume that Airy compensation dominates on the moon and that there are three principal contributions from each block to the anomalous gravity field (Figure 2): (1) surface topography, measured relative to a datum $D_r$ and contributing to gravity in proportion to the density $\rho_c$ of the upper crust, (2) Moho relief, measured relative to a datum $D_m$ and contributing to gravity in proportion to the density contrast $\Delta \rho_m = \rho_m - \rho_c$ between the densities $\rho_c$ and $\rho_m$ of the crust and mantle, respectively, and (3) mare basalt fill with a density contrast $\Delta \rho_b = \rho_b - \rho_c$. We adopt the values $\rho_c = 2.9 \text{ g/cm}^3$, $\rho_m = 3.4 \text{ g/cm}^3$, and $\rho_b = 3.4 \text{ g/cm}^3$, estimates that are uncertain by about 0.1 g/cm$^3$ [Solomon, 1975]. Thus $\Delta \rho_m = \Delta \rho_b = 0.5 \text{ g/cm}^3 = \Delta \rho$ in this problem. The contribution of topography is evaluated using (3)-(5) and is subtracted from the free air anomaly field to yield the Bouguer anomaly field.

To invert formally the Bouguer anomaly field for crustal structure, it is necessary to impose two additional constraints [Thurber and Solomon, 1978]. The first constraint is that by assumption, (1) topography in mare areas was isostatically compensated by an Airy mechanism before the emplacement of mare basalts and (2) no compensation of mare units has occurred since their emplacement. These two assumptions, for which we use the term "premare isostasy constraint" below, permit a unique decomposition of the gravity anomaly into contributions from Moho relief and mare fill. The first assumption receives support from the argument that the crust surrounding a recently formed impact basin was likely to have been hot and incapable of supporting the large deviatoric stresses associated with a deep basin depression [e.g., Bratt et al., 1981]. The second assumption, however, has little influence on the estimated thickness of nonmare crustal material. The effects of both assumptions on the derived structural models are discussed further below. Under the premare isostasy constraint, the basalt thickness $t_j$ and the depth to the Moho $d_j$ within any mare block $j$ are related by the equation

$$(t_j + h_j) \rho_c = (D_j - d_j) (\rho_m - \rho_c)$$

where $h_j$ is the topography. Both $t_j$ and $d_j$ are measured relative to the surface of a standard section which is assumed to be in isostatic equilibrium and which has a known depth to Moho $D_j$. This relationship is illustrated in Figure 3.

The second constraint on the derived crustal models follows from the fact that the crust in the general vicinity of the Apollo 12 and 14 landing sites ($\sim 3^\circ S, 20^\circ W$) is known from seismic refraction measurements to be 55 km thick [Toksoz et al., 1974]. Because the Apollo 12 and 14 sites are located in
regions of ancient premare structure and thin mare fill in the Oceanus Procellarum–Mare Cognitum area, it is reasonable to assume that the degree of isostasy in the region is nearly complete. We therefore adopt a 55-km-thick layer of nonmare crustal material of density $\rho_c$ as the standard section for (6). With this assumption the model for the nearside crustal structure developed in this study automatically matches the crustal thickness in the vicinity of the Apollo 12 and 14 landing sites.

Inversion for Crustal Structure

To determine the Moho configuration and thickness of mare basalt on the lunar nearside, we invert the Bouguer anomaly data, subject to the adopted assumptions and constraints. The inversion scheme iteratively improves on an initial estimate of the crustal structure. The criterion used to determine a best fit model is the minimization of the root-mean-square residual gravity anomaly:

$$\text{rms residual} = \left[ \frac{1}{N} \sum_{k=1}^{N} (g_k^{\text{obs}} - g_k^{\text{calc}})^2 \right]^{1/2}$$

where $N$ is the number of gravity anomaly observations, $g_k^{\text{obs}}$ is the $k$th observed gravity anomaly, and $g_k^{\text{calc}}$ is the $k$th gravity anomaly calculated from the block model using the nonlinear formulations given in (3)-(5).

At each iteration in the inversion scheme, (3)-(5) are approximated by a linear relation between the Bouguer gravity anomaly and the thickness of anomalous mass within a block:

$$g_k^{\text{calc}} = -G\Delta\rho \sum_{j=1}^{N} \frac{A_j b_j \cos \alpha_{kj}}{r_{kj}^2} \quad k = 1, 2, \ldots, N$$

where $b_j$ is the total thickness of the anomalous mass (Moho uplift plus mare fill) in the $j$th block, $\Delta\rho$ is the density contrast between the block and the surrounding crust (0.5 g/cm³), $A_j$ is the surface area of the $j$th block, $r_{kj}$ is the vector between the $k$th observation point and the center of mass of all anomalous mass in the $j$th block, and $\alpha_{kj}$ is the angle at the observation point between $r_{kj}$ and the downward vertical. From the linear relation (8) between $g_k^{\text{calc}}$ and $b_j$, it follows that $\Delta g_k = g_k^{\text{calc}} - g_k^{\text{obs}}$ can be linked to perturbations in the thickness of anomalous mass by the relationship

$$\Delta g_k = \sum_{j=1}^{N} \frac{\partial g_k^{\text{calc}}}{\partial b_j} \Delta b_j$$

or

$$\Delta g = P \Delta b$$

where $\Delta g$ is an $N \times 1$ column vector of residual gravity anomalies, $P$ is an $N \times M$ matrix of partial derivatives ($P_{kj} = -G \Delta\rho A_j \cos \alpha_{kj}/r_{kj}^2$), and $\Delta b$ is an $M \times 1$ vector of thickness corrections to the anomalous mass.

For $N > M$, (10) can be treated as a least squares problem and inverted by any number of methods [e.g., Lawson and Hanson, 1974] to obtain the corrections to the thickness of anomalous mass in each block. These corrections are then added to the mass thicknesses in the crustal model from the prior iteration, subject to the constraint of premare isostasy, imposed where applicable, to separate the contributions from Moho relief and mare fill. After each iteration the global datum $D_M$ for Moho relief is adjusted to meet the constraint on crustal thickness inferred from seismic measurements for the region of the Apollo 12 and 14 landing sites. Finally, a new Bouguer anomaly field $g^{\text{calc}}$ is computed from the adjusted model using (3)-(5), the rms residual gravity anomaly is determined from (7), and the inversion process is repeated until the rms residual converges to a minimum.

We impose one additional constraint on blocks in mare regions. If the Bouguer anomaly over a mare block is sufficiently small, the premare isostasy constraint may not be consistent with a finite thickness of basalt. Therefore, if the adjustment to the basalt thickness within a mare block forces the total mare thickness to fall below 250 m, the constraint of premare isostasy for that block is relaxed and the basalt thickness is held constant at 250 m for the remaining iterations.

A substantial savings in computation time and core storage was achieved by filling the matrix $P$ only with those partial derivatives associated with the block directly below the observation point and the eight surrounding blocks. This approximation is equivalent to assuming that the residual anomaly above a given block is caused only by errors in the adopted thicknesses of the anomalous mass in the nine nearest blocks. The resulting partial derivative matrix is then sparse and can be easily manipulated into a diagonally dominant form. We solve for the vector $\Delta b$ of thickness corrections by converting the matrix $P$ to upper triangular form by using a series of Householder transformations applied to sequentially accumulated rows [Lawson and Hanson, 1974, pp. 207–311].

Gravity and Topographic Data

The inversion procedure described above to determine crustal and upper mantle structure would be straightforward if the free air gravity anomaly and topography were known everywhere on the moon within a well-prescribed uncertainty. Such, unfortunately, is not the case. Information on the lunar gravity field is derived from measurements of Doppler shifts in the frequency of radio transmissions along a line-of-sight (LOS) direction between the earth and a satellite in orbit about the moon. High-frequency variations in these Doppler observations contain a signature of gravity anomalies arising from near-surface heterogeneities in density. To extract this signature, the effect of orbital parameters on the LOS Doppler data must be removed or modeled. The estimation of a representation of the free air gravity field over a significant area of
the lunar surface requires the simultaneous inversion of Doppler data from a large number of orbits for the parameters of both the spacecraft orbits and the gravity field. The first such representation for a large fraction of the lunar nearside was described by Wong et al. [1971]. They presented both point mass and disk mass representations of lunar gravity from Lunar orbiter and early Apollo tracking data as well as a full discussion of the procedures used in data inversion.

For this study we employ an improved representation of the low-latitude nearside gravity field obtained by Wong et al. [1975] from an inversion of tracking data from low-altitude Apollo spacecraft. Their representation consists of 350 near-surface disk masses distributed between ±30° latitude and ±100° longitude and superimposed on the triaxial model for lunar gravity of Liu and Laing [1971]. The location and size of each disk were assigned a priori, either on the basis of geological information or in an otherwise regular spacing, generally 5° in latitude and longitude. This spacing is a measure of the horizontal resolution of the gravity field model. Disk radii range from 75 to 300 km; the largest disks represent major impact basins. The mass of each disk was derived from the simultaneous inversion of a large quantity of Doppler tracking data from Apollo 14, 15, and 16 spacecraft. Orbits used in the inversion ranged from 15 to 200 km in altitude and ±30° in latitude.

The free air gravity anomaly at an altitude of 100 km (or a radius of 1836 km from the lunar center of mass) may be readily calculated from the disk model of Wong et al. [1975] and is shown in Figure 4 (a similar figure is given by Sjogren [1974]). Unfortunately, the error in this free air anomaly field is not as readily determined. Wong et al. [1975] did not determine formal errors in the mass of each disk. An estimate of the typical uncertainty in the free air gravity field may be obtained, however, from the time derivatives of the differences between the LOS Doppler observations and the predictions of the disk mass model of Wong et al. [1975]. For five orbits over the central nearside with Doppler residual data displayed by Wong et al. [1975] and with spacecraft altitudes between 80 and 160 km, the residual LOS acceleration is typically within ±7 mgal. For instance, the LOS acceleration residual for the Apollo 15 subsatellite at 100-120 km altitude over Serenitatis, Crisium, and Smythii for orbit 1516 are 6, 4, and 6 mgal, respectively. While these LOS acceleration residuals over the central nearside (e.g., Serenitatis) provide a measure of the error in the derived free air gravity anomaly, the residuals at greater longitudes (e.g., Smythii) reflect more nearly horizontal spacecraft accelerations and are less simply related to errors in the gravity field. For the purpose of estimating the error in the derived crustal structure in the next section of this paper, we estimate that the free air anomalies at 100 km elevation shown in Figure 4 have an associated error of ±10 mgal over the central nearside and ±20 mgal at the corners of the area represented by the disk mass solution.

Information on lunar topography comes from a variety of sources, including Apollo laser altimetry [Roberson and Kaula, 1972; Wollenhaupt and Sjogren, 1972; Wollenhaupt et al., 1974; Kaula et al., 1972, 1973, 1974], landmark tracking [Wollenhaupt et al., 1974], limb profiling [Watts, 1963], and photogrammetry [Hopmann, 1967; Mills and Sudbury, 1968; Arthur and Bates, 1968]. Bills and Ferrari [1977b] synthesized a large number of these topographic measurements from all sources and placed them in a consistent center-of-mass coordinate system.

In this paper we use averages, in blocks of 5° latitude by 5° longitude, of 15,887 observations of nearside topography compiled by B. G. Bills (personal communication, 1982). There are a total of 672 blocks spanning the latitude range ±40° and the longitude range ±105°. In regions where the topography is poorly resolved, values from the harmonic representation of Bills and Ferrari [1977b] supplement the block averages. The topography of the eastern half of the Orientale basin, the youngest of the nearside basins, has recently been reevaluated by Head et al. [1981] using earth-based telescopic measurements of limb heights [Watts, 1963]. Apollo laser altimetry over the northern edge of Orientale suggests that while these limb height measurements provide an accurate representation of the basin relief, they require a downward adjustment of about 2 km to fit smoothly with the other topographic data.
Fig. 5. Topography of the lunar nearside, relative to a datum at 1730 km radius. The map is obtained from 5° × 5° averages of topography compiled by Bills and Ferrari [1977b] and also incorporates limb height observations of the Orientale basin summarized by Head et al. [1981]. The contour interval is 1 km. The elevation minima within several basins are also indicated.

The accuracy of the block-averaged topography depends on both the technique employed in making individual measurements and the number of measurements in each block. For the individual determinations of topographic height from Apollo laser altimetry and orbital photogrammetry, Bills and Ferrari [1977b] have estimated an error of ±0.3 km. The majority of these measurements were made between 15°S and 40°N latitude on the nearside of the moon. Error estimates for topography measured over the remainder of the nearside using landmark tracking, earth-based photogrammetry, and limb profiling range from 0.4 to 0.9 km.

The error in the average topography over a 5° × 5° block can be estimated from the uncertainties in the individual determinations within that block and from the assumption that all individual measurements are independent. The estimated error in the average topography over Orientale, Humorum, and Nubium for some blocks approaches the estimates for individual observations because these basins are poorly covered by spacecraft tracking data. As a result, errors in topography over these basins represent a large source of uncertainty in the analysis of associated basin structure. For example, the contribution to the Bouguer correction from an 0.5-km error...

Fig. 6. Bouguer gravity anomaly for the lunar nearside at 100 km elevation. A datum D_r at a radius of 1736 km was used to make the topographic correction. The contour interval is 50 mgal. Local maxima over several basins are also indicated.
in topography within a $5\degree \times 5\degree$ block of crust of density 2.9 g/cm$^3$ near the lunar equator is about 20 mgal at 100 km altitude, or up to twice the estimated uncertainty in the free air anomaly derived from the mass disk gravity model. Because the total error in the Bouguer correction is a function of errors in topography everywhere, a baseline error in the topography over a larger region would produce a still greater error in the calculated gravity anomaly.

Fortunately, the large quantity of topographic data collected by orbiting spacecraft from the central nearside results in greatly improved estimates of the average topography for $5\degree \times 5\degree$ blocks in that region. Uncertainty in the mean topography of blocks covered by Apollo orbital tracking, some of which are sampled by as many as 135 observations, may be less than 0.1 km. Thus the contribution of uncertainty in topography to errors in the Bouguer anomaly over the central nearside is small.

The "observed" Bouguer anomalies are calculated above the center of each block at two altitudes, 100 and 200 km, which bracket the altitudes of the majority of Apollo 14, 15, and 16 orbits used to derive the disk gravity model [Wong et al., 1975]. We found that using 1344, rather than 672, gravity "observations" greatly improves the stability of the inversion and acts to smooth the resulting structural model. Of course, the 1344 "observations" are not fully independent. The mass disk model is described by 1400 parameters, but many of these parameters are strongly correlated [Wong et al., 1975].

The Bouguer gravity anomaly field obtained from the application of (3)-(5) to the observed topography (Figure 5) and free air gravity anomalies (Figure 4) is shown in Figure 6. The topographic datum $D_T$ assumed in calculating the topographic correction is a sphere of radius of 1736 km, selected to be representative of the typical elevation of central nearside regions. The mascon basins Orientale, Humorum, Imbrium, Serenitatis, Nectaris, Crisium, and Smythii display the most prominent Bouguer anomalies on the figure. Orientale, Serenitatis, and Crisium are sites of the largest anomalies (> 200 mgal) relative to surrounding regions.

The estimated error $(\sigma_A)_h$ in the Bouguer anomaly field over the $k$th block due to uncertainty in both the gravity and topographic data sets may be estimated from

$$
(\sigma_A)_h^2 = (\sigma_t)_h^2 + \sum_{j=1}^{N} (\sigma_t)_j^2 \quad k = 1, 2, \cdots, N
$$

where $(\sigma_t)_h$ is the error in the free air gravity anomaly over the $k$th block, and $(\sigma_t)_j$ is the error in the Bouguer correction for the $k$th block due to the uncertainty in mean topography for the $j$th block. The area of most reliable Bouguer gravity lies beneath the orbital tracks of the Apollo spacecraft. This area includes the Serenitatis, Tranquillitatis, Crisium, Fecunditatis, Nectaris, and Smythii basins (Figure 1). The Bouguer gravity anomalies over Orientale, Humorum, and Nubium are less accurately represented, but these basins have nonetheless been included in our study for completeness. Signatures of the Imbrium and Australe basins are visible in the gravity and topographic data, but the structures determined in this study for these regions, located at the edge of the area of good coverage, should be regarded as highly uncertain.

**Basin Structural Models**

We have inverted the Bouguer gravity anomaly field of Figure 6, following the inversion procedure described above, to obtain values for the thicknesses of the nonmare and mare basalt components of the crust in each of the 672 blocks on the lunar nearside. Maps of these quantities are shown in Figures 7 and 8, respectively. The crustal thicknesses in blocks containing the Apollo 12 and 14 landing sites are within 1.5 km of the seismically measured value of 55 km. After the final iteration, the rms residual Bouguer anomaly was 6.5 mgal.

The uncertainty $(\sigma_t)_h$ in the thickness $t$ of nonmare crustal material in the $k$th block can be estimated approximately with the assumption that it is primarily the result of the uncertainty in the Bouguer anomaly for that block. Under this assumption,

$$
(\sigma_t)_h = \frac{s_k^2(\sigma_A)_h}{GA_k \Delta \rho}
$$
where $s_k$ is the distance from the observation point (100 km elevation) to a point located at the centroid of the block of nonmare crustal material. We use $s_k = 127.5$ km for all $k$. The estimated uncertainties in the values of $t_c$ shown in Figure 7 are displayed in map form in Figure 9. The estimated error in $t_c$ over most of the nearside ranges from about 2 to 3 km in areas beneath the tracks of Apollo spacecraft (between 15°S and 40°N latitude) to about 5 km at high latitudes. Errors in $t_c$ derived for the western limb, including the Orientale basin, are generally larger (up to 12 km) because of the uncertainty in the correct baseline for the local topography. Near the edges of the grid area, additional errors may arise from neglect of structural variations beyond the grid and from the assumption of a uniform topographic datum. We repeated the inversion using Bouguer anomaly data obtained with $D_T$ larger by up to 3 km than the value assumed above and found the resulting structure similar to that of Figures 7 and 8; even in blocks at the edge of the grid the crustal thicknesses differ by only a few percent for different values of $D_T$. Another source of error in the computed structure is the uncertainty ($\pm 0.1$ g/cm³) in the assigned densities. An underestimate of the density contrast between crustal and mare or mantle material of 20%, for instance, results in an overestimate of both the Moho relief and mare basalt thickness by about the same percentage.

It is important to recognize that relaxation of the premare isostasy constraint would not alter significantly the values of crustal thickness depicted in Figure 7. If isostatic compensation of the premare basin topography was less than complete or if some percentage of the mare fill has been compensated, the actual thickness of mare basalt would be greater than in our model by some amount. Because mantle material and mare basalt are similar in density, however, the Moho relief

Fig. 8. Mare basalt thickness on the lunar nearside. See Figure 7 (and text) for model assumptions. The dashed line indicates the boundary of mare regions. Contour interval is 1 km.

Fig. 9. Estimated error in crustal thickness (exclusive of mare basalt fill) resulting from uncertainties in gravity and topographic data as calculated using equation (12). The contour interval is 2 km.
would be less by an approximately equal amount. This approximately even trade-off of anomalous masses occurs because the Bouguer anomaly, when measured 100–200 km above the sources of anomalous mass, is much more sensitive to the total anomalous mass than to the distribution of that mass between the surface and Moho. If high-density basaltic to gabbroic intrusions (dikes, sills) permeate the crust beneath the basalt fill in large basins [e.g., Janle, 1981a, b], then lesser amounts of Moho relief and mare fill than shown in Figures 7 and 8 would be necessary to match the observed gravity. The equivalent thickness of low-density crustal material, however, would not differ greatly from the values shown in Figure 7. Though some system of magma conduits clearly must have existed at depth to feed the extensive mare units within most basins, the volume of such structures is not known.

The crustal thickness model of Figure 7 indicates that the crust is thinner beneath each of the major nearside basins than in immediately surrounding areas. Such a result, of course, would follow at least qualitatively from the assumption of premare isostasy. We interpret this thinner crust to be principally the result of the basin formation process, including formation of the transient cavity and collapse of the cavity by some combination of inward and upward flow of crustal and mantle material [e.g., Melosh and McKinnon, 1978]. By this interpretation, the Moho relief beneath a basin region is a measure of the amount of uplift of mantle material during cavity collapse, an important parameter in describing the thermal evolution of a basin region [Bratt et al., 1981]. Of course, the crustal model of Figure 7 reflects only the present Moho relief beneath basins; the relief immediately following basin formation may have been substantially modified by the effects of ductile flow in the crust.

Cross sections through the basin structural models are shown in Figure 10. The dashed lines show the Moho depth according to the block model along the profile lines shown in Figure 1. The solid lines show smoothed cross sections of both the Moho and the base of mare basalt deposits. The smoothed profiles were obtained from the contour maps in Figures 7 and 8 and represent azimuthally averaged cross sections. From the smoothed Moho profiles and the assumption of cylindrical symmetry, we have estimated the magnitude of Moho relief and the apparent volume \( V \) of uplifted mantle...
beneath each basin. These quantities are listed in Table 1, in order of increasing basin age [Wilhelms, 1981].

Table 1 and Figure 10 suggest a general decrease in the apparent volume of uplifted mantle with increasing basin age. The Moho beneath the central mare units of Orientale (Figure 10a) is raised by 66 km relative to the ambient depth of about 80 km. Serenitatis (Figure 10b), a basin about the same diameter as Orientale and of somewhat greater age, is surrounded by crust with an average thickness of 55 km; mantle relief beneath Serenitatis is 30 km, but the uplifted volume is broader and thus somewhat larger (7.5 x 10⁶ km³ versus 6.0 x 10⁶ km³) than that beneath Orientale. The volume of apparent mantle uplift beneath Orientale is also more uncertain, however, because of poor data coverage over the western portion of the basin. The difference in the magnitude of Moho relief between the two basins nonetheless exceeds the sum of the errors in these two quantities (Figure 9). The oldest basin examined is Tranquillitatis (Figure 10). Though slightly larger in diameter than the nearby Serenitatis basin, Moho relief beneath Tranquillitatis is only 12 km, also a significant difference in relation to the errors shown in Figure 9. We interpret the decrease in the apparent mantle uplift with increasing basin age (Table 1 and Figure 10) to be the result of an age dependence of the basin formation or modification processes. We consider the implications of this inference in a later section.

Figure 7 shows a generally thinner crust over the central nearside of the moon than near the limb regions, a result also noted in several earlier studies of lunar crustal structure [e.g., Kaula et al., 1974]. This large region of thinner crust coincides approximately with the location of the ancient Procellarum basin [Whitaker, 1981], a 3200-km-diameter structure centered near 26°N and 15°W. This coincidence suggests that some record of mantle uplift during basin formation is still apparent even for the largest and one of the oldest of the preserved nearside basins [Wilhelms, 1981].

The thicknesses of mare basalt, shown in map view in Figure 8, are deserving of several comments. The inversion procedure preserved the constraint of premare isostasy within all major basins for the complete set of iterations. Equivalently, the Bouguer gravity anomaly over these basins is sufficiently large so as to require both an elevated Moho and more than 250 m of high density mare basalt. Basalt thicknesses beneath the centers of the irregular mare basins (Nubium, Fecunditatis, and Tranquillitatis) range from 0.5 to 1.3 km, while the younger circular mare basins (Orientale, Serenitatis, Crisium, Humorum, Nectaris, and Smythii) contain as much as 3.5-4.5 km of basalt fill within the central depression region. That these thicknesses in the central portions of the circular basins are greater by up to 2 km than the values derived by Thurber and Solomon [1978] under a similar set of assumptions can be attributed to the improved accuracy of the topographic data used here and to the method adopted to evaluate the gravitational potential due to anomalous mass (equation (3)). It should be noted, however, that the region of thickest mare fill in each of the Orientale and Smythii basins constitutes only a single block in the model depicted in Figure 8. In some irregular mare regions (e.g., Oceanus Procellarum) the constraint of premare isostasy was relaxed during the inversion because the basalt thickness within one or more blocks dropped below the 250-m limit after an iteration. For such regions, either the surficial mare basalt unit is very thin or the crust is somewhat thicker than would be predicted from complete isostatic compensation of premare relief.

As noted earlier, the mare basalt thicknesses depicted in Figure 8 should probably be regarded only as lower bounds on the actual thicknesses within the central regions of most basins [Thurber and Solomon, 1978]. If the lunar lithosphere maintained finite strength during the formation of at least the younger basins, then the premare basin would have had incompletely compensated relief and a greater thickness of mare basalt would be required to match the Bouguer gravity field. Further, some portion of the mare basalt load in the maria has likely been compensated by lithospheric flexure [e.g., Solomon and Head, 1979, 1980], which also would lead to mare units somewhat thicker and Moho relief somewhat less than indicated by the structural model presented here. It is interesting to note that these thicknesses estimated by DeHon and Waskom [1976] and DeHon [1978] from rim heights of craters buried or partially buried by mare deposits are generally comparable to the values in Figure 8 for many mare regions. This similarity between the two determinations is probably coincidental, however, because the rim height technique tends to underestimate mare thickness by as much as a factor of 2 [Head, 1982].

It is also possible that the mare basalt thicknesses shown in Figure 8 for some regions may be overestimates because of a significant contribution to the gravity field from subsurface intrusions of high-density mare basaltic material. The thickness of mare basalt within the central block of the Orientale

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**TABLE 1. Estimates of Moho Relief, Apparent Volume of Mantle Uplift, Basin Volume, and Ejecta Volume for Major Basins on the Lunar Nearside**

<table>
<thead>
<tr>
<th>Basin</th>
<th>Radius, km</th>
<th>Moho Relief, km</th>
<th>Ambient Crustal Thickness, km</th>
<th>V_m 10⁶ km³</th>
<th>V_e 10⁶ km³</th>
<th>V_o 10⁶ km³</th>
</tr>
</thead>
<tbody>
<tr>
<td>Orientale</td>
<td>310</td>
<td>66</td>
<td>85</td>
<td>6</td>
<td>0.7</td>
<td>7</td>
</tr>
<tr>
<td>Serenitatis</td>
<td>305</td>
<td>30</td>
<td>60</td>
<td>8</td>
<td>1</td>
<td>9</td>
</tr>
<tr>
<td>Crisium</td>
<td>225</td>
<td>31</td>
<td>55</td>
<td>6</td>
<td>1</td>
<td>7</td>
</tr>
<tr>
<td>Humorum</td>
<td>205</td>
<td>27</td>
<td>60</td>
<td>3</td>
<td>0.5</td>
<td>4</td>
</tr>
<tr>
<td>Nectaris</td>
<td>300</td>
<td>28</td>
<td>60</td>
<td>4</td>
<td>0.7</td>
<td>5</td>
</tr>
<tr>
<td>Smythii</td>
<td>300</td>
<td>33</td>
<td>65</td>
<td>4</td>
<td>0.6</td>
<td>5</td>
</tr>
<tr>
<td>Nubium</td>
<td>340</td>
<td>10</td>
<td>60</td>
<td>2</td>
<td>0.3</td>
<td>2</td>
</tr>
<tr>
<td>Fecunditatis</td>
<td>345</td>
<td>14</td>
<td>60</td>
<td>2</td>
<td>0.2</td>
<td>2</td>
</tr>
<tr>
<td>Tranquillitatis</td>
<td>340</td>
<td>12</td>
<td>60</td>
<td>2</td>
<td>0.2</td>
<td>2</td>
</tr>
</tbody>
</table>

Basins are listed in order of increasing age [Wilhelms, 1981]. The indicated radius corresponds to the second ring [Head, 1977; Wilhelms, 1981].
basin, for instance, is 3–4 km in Figure 8. On photogeological grounds, however, the basalt thickness has been estimated to be no greater than 1 km [Head, 1974]. A significant excess mass within the submare crust is a possible explanation for this discrepancy. The large uncertainty in the topography and gravity in the Orientale region may be at least an equal contributor, however.

**Geological Implications**

The structural models derived in this paper have a number of implications for the processes that accompanied the formation and evolution of large impact basins on the moon. We divide the discussion into basin formation processes, including transient cavity collapse and emplacement of ejecta deposits, and basin modification processes, including volcanism and lithospheric deformation on time scales longer than those normally associated with the basin formation event. We recognize that this division is somewhat arbitrary and that the present characteristics of each basin may be the product of a combination of both types of processes.

*Early Cavity Geometry*

The volume of material ejected from a basin during impact is an important quantity, for several reasons. Its relation to basin diameter provides a strong constraint on the cratering process for large impacts, particularly on the early cavity geometry. The ejecta volume also provides information on the sampling depths of large impacts and thus on the interpretation of returned lunar samples for crustal composition. Head et al. [1975] have defined the volume $V_C$ of "primary ejecta" as the volume of material excavated from a crater and thrown beyond the crater rim. The quantity $V_C$ is thus approximately equal to the volume of the early transient cavity minus the volume of material which falls back in to the collapsing cavity. This equality ignores the effects of density changes due to compression of target material, brecciation of fallback material, or other inelastic effects. The volume of primary ejecta residing outside a major basin, however, is a difficult quantity to determine from photogeologic observations of ejecta deposits, among other reasons because of the difficulty in separating primary ejecta from the locally derived, secondary ejecta surrounding the basin [Head et al., 1975]. From a variety of observations, models, and scaling relations, Head et al. [1975] estimated that $V_C$ for a basin the size of Orientale lies between $0.4 \times 10^6$ and $12 \times 10^6$ km$^3$. This wide range of possible ejecta volumes is equivalent to a large uncertainty in the thickness of ejecta deposits expected from an Orientale-sized impact. For instance, if the ejecta volume were spread evenly over the lunar surface, the thickness could be as little as 10 m or as large as 300 m.

More recently, Spudis [1982, 1983] and Spudis et al. [1984] have estimated the depth of excavation for lunar basins from remote geochronological observations and scaling relations under the assumption that the lunar crust is layered, with an anorthosite layer overlying a more noritic layer [Ryder and Wood, 1977]. Spudis [1982, 1983] and Spudis et al. [1984] concluded that the chemistry and mineralogy of the ejecta deposits of five nearside basins provide support for an excavation model in which the ratio of the depth of excavation to basin diameter is about 0.1 for basin-sized impacts. On the basis of such a model the excavation depth for the younger nearside basins ranged from 40 to 80 km [Spudis, 1981], and the volume $V_E$ of primary ejecta excavated during the Orientale basin impact is estimated to be $5 - 8 \times 10^6$ km$^3$ [Spudis et al., 1984]. These estimates have considerable uncertainty, however, associated with the reality of the assumed simple model of crustal layering and with the ability to separate primary from secondary ejecta.

The structural models for lunar nearside basins depicted in Figure 10 provide a straightforward and internally consistent framework with which to address cavity and ejecta volumes. Let $V_p$ be the volume of the present basin, including the volumes of the topographic depression and of mare fill. We make the assumption that the crustal thickness of the preimpact basin region was similar to the present thickness at 2–3 basin radii from the basin center (excluding other nearby basins). Then the quantity $V_p = V_C + V_e$, where $V_C$ is the volume of apparently uplifted mantle, provides a measure of preimpact crustal material now external to the basin rim. The quantity $V_e$ should be somewhat less than $V_p$. This is because inward transport of crustal material during basin modification as well as deposition of ejecta from younger basins or of nonmare volcanic material will act to reduce one or both of $V_C$ and $V_e$. Therefore $V_e$ can be regarded as the difference between the volume of crustal material ejected beyond the basin rim and the volume of crust returned to the basin region by these later processes. For the youngest nearside basins, which may have formed at a time when near-surface temperatures were comparatively low and the elastic lithosphere relatively thick (e.g., Orientale, Serenitatis, and Crisium), $V_e$ may approach $V_p$.

Values for $V_e$ derived from the smoothed cross sections of Figure 10 are given in Table 1. The values range from 2 to $9 \times 10^6$ km$^3$. It should be repeated that these values are insensitive to the assumption of premare isostasy, because to first order the Bouguer anomaly data yield the total anomalous mass (mantle uplift plus mare basalts), and it is the sum of the two components that contributes to $V_p$. By analogous reasoning, these values of $V_e$ are also insensitive to the presence of high-density intrusions within the low-density crust underlying the basin. The uncertainties in gravity and topography and in the assumed densities and model simplifications result in an uncertainty in $V_e$ of about 30% for Orientale and 20% for the other basins studied. The values of $V_e$ in Table 1 fall within the range of estimates for $V_e$ for basins the size of Orientale or Serenitatis [Head et al., 1975; Spudis et al., 1984].

An interesting relationship between apparent mantle uplift and crustal thickness is suggested by the data from the six youngest basins in Table 1. The thickness $t_c$ of the nonmare crust beneath the central region of each of these six basins is remarkably uniform, 20–30 km, despite differences in basin size, age, and preimpact crustal thickness. The Moho relief for five basins in crust 55–65 thick is nearly constant at about 30 km, despite a variation of 200 km in basin diameter. While the Moho relief beneath Orientale is greater than that of the other five basins by 30–40 km, the surrounding crust is also thicker by 20–30 km.

We suggest that this near uniformity of nonmare crustal thickness beneath the younger basins can be understood only
Fig. 11. Three scenarios for the formation of young impact basins formed when the elastic lithosphere is comparable or greater in thickness than the crust. In all cases, uplift of the sublithospheric mantle dominates the late-stage collapse of the transient cavity. (a) The excavated cavity is confined to the crust. (b) With a larger ratio of crater diameter to crustal thickness the excavated cavity is prevented from extending significantly deeper than the Moho by the relatively greater strength of mantle material. (c) With a still larger ratio of crater diameter to crustal thickness the transient cavity may extend into the uppermost mantle. If the preserved nonmare crust beneath the basin center is dominated by fallback of ejected crustal material in the scenarios depicted in Figures 11b and 11c, it may be similar in thickness for the two cases. Discovery of mantle material among the returned lunar samples would strengthen the possibility of the scenario in Figure 11c.

If excavation of the transient cavity for basins 400-600 km in diameter extended to the base of the crust or to the uppermost mantle. If the excavated volumes did not extend at least to near the base of the crust for these basins (Figure 11a), then the value of \( t_c \) for a basin of a given diameter should be greater for a thicker preimpact crust. The quantity \( t_c \) should also decrease with increasing diameter for basins formed in crust of similar thickness. Neither relation is indicated by the values in Table 1. Of course, the youngest six basins in Table 1 may have been affected to varying degrees by long-term viscous relaxation and other modification processes, but it is hard to imagine how such processes would have led coincidentally to the similar present values of \( t_c \). As noted above, the conclusion that excavation of the youngest nearside basins extended to the lower crust has been suggested on independent geochemical and photogeological grounds by Spudis [1982, 1983] and Spudis et al. [1984].

If we accept that the excavated volumes for the basins in Table 1 extended in depth at least to the lowermost crust, we still must seek a mechanism for the near constancy of \( t_c \). One explanation is motivated by the observation that mantle material is significantly stronger than crustal material at laboratory strain rates [e.g., Brace and Kohlstedt, 1980]. If this difference in strength can be extrapolated to the high strain rates thought to accompany cavity formation [e.g., O'Keefe and Ahrens, 1977], the Moho may have acted more or less as a barrier to the deepening of cavity excavation (Figure 11b). The 20- to 30-km thickness of nonmare crust beneath the central regions of large young impact basins, by this hypothesis, would be a combination of ejecta fallback and crustal material transported laterally during cavity collapse. These two components would be difficult to distinguish.

An alternative explanation for the uniformity of \( t_c \) is that the transient cavity for the largest basins excavated through the crust and into the mantle (Figure 11c). If the equivalent thickness of preimpact crustal material that fell back into the central basin region was only weakly dependent on basin diameter, then the indicated value of \( t_c \) would be similar for basins that had excavated to different mantle depths. Of course, the actual ejecta would have consisted of a mixture of crustal and mantle material, but the inversion procedure we have followed cannot distinguish a thin nonmare crust from a thicker layer with the same volume of low-density crustal rock admixed with mantle material. The paucity of candidates in the Apollo sample collection for material from the lunar mantle [see Head et al., 1975], however, suggests that exca-
Fig. 12. Two possible scenarios to account for the structure of older lunar basins (e.g., Figures 10g–10i). (a) Because of the high crustal temperatures and thin elastic lithosphere at the time of impact, flow of both the crust and the mantle accompany transient cavity collapse, leading to a structure similar to that presently observed. (b) The collapse of the transient cavity at left is accomplished primarily by uplift of mantle material contemporaneously with the return of impact ejecta and melt to the basin, yielding an initial structure similar to that of younger basins. Because of high near-surface temperatures this initial structure undergoes long-term viscous relaxation of both topographic and Moho relief [Solomon et al., 1982].
portant [Wilhelms, 1983]. Nubium and Tranquillitatis lie just within, and Fecunditatis just outside, the third ring of the Procellarum basin [Whitaker, 1981] and are thus within a radial range where significant thermal effects can be expected [Bratt et al., 1981]. As noted earlier, the Procellarum basin may have contributed to the generally lower elevations and thinner crust on the central nearside than elsewhere on the moon. That the Procellarum basin did not itself experience complete viscous relaxation of surface and Moho topography may be ascribed to the insensitivity of the controlling time constant to spatial scale [Solomon et al., 1982] and perhaps as well to the constraints on lateral flow imposed by spherical geometry when the basin diameter approaches the diameter of the moon.

**Nonmare volcanism.** An additional modification process that can act to reduce basin relief and that would be enhanced by higher temperatures is nonmare volcanism, i.e., deposition of nonmare basalts with a density similar to the typical density of the crust and therefore not resolvable from gravity anomalies. The volumetric significance of nonmare volcanism for lunar crustal evolution in general and for old lunar basins in particular is not well known. To the extent that the crust, as opposed to the upper mantle, is the source of magmas for this volcanic activity, this process involves lateral transport of crustal material to fill the basin depression and is not readily distinguishable from ductile flow (Figure 12). If temperatures exceed the solubus in crustal material adjacent to a newly formed basin, in fact, the effective viscosity might be sufficiently low so that bulk flow would dominate magma transport as a mechanism for reducing relief.

**Mare volcanism.** The source regions of mare basalt magmas are thought to lie at depths in excess of 100 km [e.g., Wyllie et al., 1981]. At such depths the deposition of kinetic energy as heat, the release of hydrostatic pressure, and the uplift of mantle isotherms during basin formation [Bratt et al., 1981] are likely to have had only a small effect on magma genesis and then only if the impact was very large [Hubbard and Andre, 1983] and the ambient temperature was already near the solubus. That the thickness of mare basalts appears to be less within the older basins (Nubium, Fecunditatis, Tranquillitatis) than nearby younger basins (Figure 8) suggests that mare volcanism probably has not contributed substantially to the age dependence of preserved basin structure.

Of the basin modification processes considered above, lateral flow of lunar crustal material, perhaps augmented by nonmare volcanism, is the most likely explanation of the general decrease of preserved mantle uplift with increasing basin age. Because both of these processes involve the radially inward movement of crustal material, the two are difficult to separate on the basis of gravity and topographic data alone. The poor degree of preservation of relief of the older basins on the central nearside may be a consequence of both the generally high temperatures in the early lunar crust and the extensive additional crustal and subcrustal heating associated with the formation of the ancient Procellarum basin.

**Summary and Conclusions**

We have presented models for the crustal and upper mantle structure beneath major impact basins on the nearside of the moon. The models are derived from an inversion of nearside gravity and topography, subject to the assumptions that pre-mare topography was isostatically compensated by an Airy mechanism and that compensation of mare basalt units may be neglected. An additional constraint is the crustal thickness inferred beneath the Apollo 12 and 14 landing sites from seismic observations.

The basin structural models are characterized by thinned crust and an elevated Moho beneath the central basin regions, compared to surrounding areas, presumably the preserved effects of basin excavation and mantle uplift during collapse of the transient cavity. Orientale exhibits over 60 km of apparent mantle uplift, while the Moho relief beneath Serenitatis, Crisium, Humorum, Nectaris, and Smythii is about 30 km. The three oldest basins included in this study, Nubium, Fecunditatis, and Tranquillitatis, display only a modest (~10 km) amount of preserved Moho relief.

The general decrease in the amount of preserved mantle uplift with increasing basin age is attributed primarily to enhanced rates of lateral flow of crustal material early in lunar history when crustal temperatures were relatively high and the elastic lithosphere comparatively thin. Nonmare volcanic activity may also have contributed to the greater extent of modification of older basins. The relaxed topographic and Moho relief for older basins on the central nearside, in particular, may be at least partly a consequence of the extensive crustal and subcrustal heating associated with the formation of the large Procellarum basin.

If we make the assumption that the preimpact crustal thickness beneath the basin region was similar to the present thickness beneath surrounding regions, then the sum of the volume of uplifted mantle, the volume of mare basalt fill, and the volume of the topographic basin provides a lower bound to the volume of crustal material ejected beyond the basin rim during basin formation. This bound may be similar to the true ejected volume for the youngest basins formed when the elastic lithosphere may have been significant in thickness and when any modification subsequent to cavity collapse may have been relatively minor. On this basis, we estimate the volume of crustal material ejected from basins the size of Orientale to be of the order of 10^{7} km^{3}.

The thickness of nonmare crustal material beneath the central regions of the six youngest basins considered in this study is remarkably uniform (20-30 km) despite large differences in basin size and in the preimpact thickness of the crust. These results suggest the hypothesis that basin excavation extended in depth at least to the lowermost crust for these impacts and that significant deepening of cavity excavation for the largest basins was most likely impeded by an abrupt increase in strength at the crust-mantle boundary. The preserved layer of nonmare crust beneath the basin centers, by this view, may be some combination of fallback and crustal material transported laterally during cavity collapse.

The structural models described in this paper provide important constraints on the thermal budget of the basin-forming process and on the subsequent thermal and tectonic evolution of the basin region [Bratt et al., 1981]. While these models are of necessity simple, they provide a basis for testing the ideas presented here with data of higher spatial resolution in individual basin regions on the moon and the other terrestrial planets.

**Acknowledgments.** We thank Bruce Bills for providing us with a tabulated version of his compilation of lunar topographic data, Bill Siigren for several constructive discussions on lunar gravity, Paul Spudis for a copy of his paper prior to publication, and Jan Nattier-Barbaro for help with manuscript preparation. Critical reviews by J. Dvorak, an additional reviewer, and an Associate Editor resulted in significant improvements to this paper. This research was supported by NASA grants NSG-7081 and NSG-7297.


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(Received June 11, 1984; revised September 25, 1984; accepted November 12, 1984.)