COMPARISON OF SURFACE GRAVITY DATA
WITH SATELLITE DATA

I — Introduction

A new estimate of the earth's gravity field has recently been obtained by Gaposchkin and Lambeck [1], [2] and designated as the 1969 Smithsonian Standard Earth II (SE II). This solution is based on some 100,000 camera and laser range observations to numerous satellites collected from a world wide tracking network. The solution included input from deep space probes provided by the Jet Propulsion Laboratory [3] as well as surface gravity data. The total field is represented by spherical harmonic coefficients complete to degree and order 16 plus a number of higher terms. The accuracy of this global solution has been estimated as ± 3 m in geoid height or about 9 mgal.

Figure 1 gives the free-air gravity field corresponding to this solution referred to an ellipsoid of flattening 1/298.255. Its most significant features, not shown up in earlier solutions, are the new detail found in southern latitudes and that almost all of the areas of tectonic activity are associated with positive gravity anomalies; both the areas where new crust is being generated, such as the ocean ridges and areas where the crust is being destroyed, such as the trenches. Earth models for explaining this apparent paradox have already been proposed by Kaula [4], [5].

In the present paper those aspects of the Smithsonian Standard Earth II solution relating to surface gravity data, that is the comparison and combination of the satellite based solution with these data, are discussed and some requirements for future solutions explored. Details of the data used and methods of analyses developed are given in [1].

It should be noted that the level of accuracy of this solution is such that questions concerning, for example, the validity of mixing data collected on the earth's surface with measurements of the geopotential in space, or the presence of "forbidden harmonics" in the surface gravity data, do not present serious problems yet and that more or less classical methods can be used for the combinations.

II. The data used

The satellite solution comprises both dynamic and geometric aspects of satellite geodesy and the solution is for coefficients describing the geopotential and for corrections to the locations of the tracking stations. The dynamic solution provides the main source of information for the gravity field although the geometric solution is important for reducing some of the correlations between the two types of parameters. All $C/S (\ell, m)$ terms complete through $\ell = 12$, $m = 12$ except for $11,7, 12,8$ and $12,9$ were determined as well as some selected higher order terms; $C/S (\ell, 1) \leq \ell \leq 16$; $C/S (\ell, 2) \leq \ell \leq 15$; $C/S (14,2); C/S (\ell, 12) \leq \ell \leq 19$; $C/S (\ell, 13) \leq \ell \leq 21$; $C/S (\ell, 14) \leq \ell \leq 22$.

These higher order terms with order (i.e. $m$) 1, 2 and 3 are weakly determined because the orbital perturbations caused by them are small compared to the accuracy of the observational data and because they are difficult to separate as the satellites that are sensitive to them are in low orbits and consequently infrequently observed. The terms with order 12, 13 and 14 have been determined from resonance effects and some of these may not truly reflect the earth's gravity field because of truncation errors. These aspects are discussed further by Gaposchkin in [6] as well as in [1].

The surface gravity data used for the combination solution has been prepared by Kaula [7]. His basic data consists of 1' x 1' mean free air anomalies computed essentially by the techniques described by Uotila [8]. These anomalies were combined to form mean values for approximately 60 x 60 nautical mile (n.m) squares to form a statistically uniform sample. To obtain 300 x 300 n.m area means Kaula estimated 60 n.m area anomalies for the unsurveyed areas using linear regression methods [9] to all the 60 n.m squares within the 300 n.m. area. Finally, he computed the 300 n.m means as the arithmetic mean of all the observed and extrapolated 60 n.m squares within each larger area. The results yielded 935 values covering 56.6% of the globe and are listed in [1] table 12.

The anomalies $\Delta g_i$ can be related to the harmonic coefficients $C$ and $S$ according to

\[
\Delta g_i = \gamma \sum_{\ell=3}^{\infty} (\ell-1) \sum_{m=0}^{\infty} \left( \frac{2}{\ell} \right) (C_{\ell m} \cos m \lambda_i + S_{\ell m} \sin m \lambda_i) P_{\ell m} (\sin \varphi_i)
\]

and normal equations constructed whose inversion would provide estimates of the $C_{\ell m}$ and $S_{\ell m}$. Such a solution would however, be very poor due to the incomplete coverage of the data. The summation has been carried out to $\ell = 16$ and $m = 16$.

Instead, these normal equations are combined directly with the normal equations from the satellite solution (the combined geometric and dynamic
solutions) and inverted. The procedure is described in more detail in [1]. Two important points that have to be considered in such a combination are:

1)- What accuracy estimates have to be given to the two solutions?

and

2)- What has to be done with the 43.5% of the globe that has no surveyed data at all?

For the satellite solution it was assumed that the computed correlations between the various parameters were realistic but that the scale of the computed covariance matrix was underestimated. A scaling factor was consequently obtained from a comparison of station coordinates (and their estimated accuracies) determined by the geometric and dynamic methods independently and from a comparison of the values for the parameters estimated in the successive iterations towards the final adopted solution. In the last iteration the derived scale factor by which the covariance matrix was multiplied was 4.5.

The accuracy with which each surface gravity anomaly was estimated was assumed to be $\sigma_{\Delta g} = \frac{33}{(n + 1)}$ where $n$ is the number of 60 x 60 n.m. squares with some surface gravity data. Thus a 300 n.m. area with $n = 25$ gives $\sigma_{\Delta g} = 6.4$ mgal whereas one with $n = 1$ gives $\sigma_{\Delta g} = 24$ mgal. These standard deviations are higher by a factor of 2 than those used by Kaula in his 1966 combination solution with the same gravity data [7], but tests of the combination solution have indicated the validity of the adopted choice: In any one iteration orbits computed from the combination solution geopotential coefficients yielded better residuals than those determined solely from the satellite solution rather than cause a deterioration of orbital residuals as has been the case for all previous combinations made by other investigators.

This important characteristic of the SE II solution is a direct result of the iterative procedure adopted for the computations. In any one iteration, the satellite solution determined is tested against the surface gravity data and against satellite orbits as well as the geometric solution results and other data types. Combinations with the surface gravity data are then made and the tests repeated using, if necessary, different plausible weighing factors. When a satisfactory solution is found, i.e. the one that gives the best comparisons for all the tests, it is used as the basis for the next iteration.

For the unsurveyed areas three alternative hypotheses were made:

(i)- Only the observed $\Delta g$ were used to compute the normal equations,

(ii)- Model anomalies generated by Kaula [10] from a linear regression of the 935 observed squares were used,

(iii)- The anomalies were set to zero.
The same standard deviations were assigned to all the predicted anomalies in cases (ii) and (iii); namely $\sigma_{\Delta g} = 33\text{ mgal}$.

A comparison of results obtained for the last two hypotheses showed no fundamental difference ($< 1\text{ m}$ in geoid height) and this may have been expected since the root mean square of the predicted model anomalies was only $11\text{ mgal}$, considerably less than the $\sigma_{\Delta g}$ associated with them. The results of these two hypothesis did however show a difference when compared with the first hypothesis for those extensive areas in the southern latitudes where there were no surface data. The combination solution based on (i) introduced some rather large amplitudes in some of the major geoid features in these areas whereas the combination solution based on (ii) or (iii) reduced these by about $5\text{ m}$ and brought them closer to the satellite solution. Elsewhere the difference in geoid heights between (i) on the one hand and (ii) and (iii) on the other did not exceed $2\text{ m}$. For all subsequent iterations the model anomalies (ii) have been used.

These tests have indicated that some form of model anomalies must be adopted for the large unsurveyed areas though the question as to what type is best is unresolved and will be important for future analyses. That the linear regression method gives estimates that are too small is shown by the above estimate of $11\text{ mgal}$ for the r.m.s. of the predicted anomalies as compared to $16.5\text{ mgal}$ for the surveyed areas. Even the latter value is bound to be too small as it is also based on linear regression over the $60 \times 60\text{ n.m.}$ areas. Kohnlein [11] used gravity anomalies obtained from the satellite solutions for the unsurveyed areas, but such an approach means that the two data sets are no longer independant.

A comparison of the combination solution against surface gravity data is of course not entirely valid if the same gravity information is inherent to both data sets. Thus, in order, to obtain some independant tests some recent surface data compiled by Le Pichon and Talwani for the Atlantic Ocean [12] and for the Indian Ocean [13] has been used for further comparisons.

III — Method of comparison

In making the comparisons of the two data types, we follow the approach of Kaula [7] the essentials of which are outlined below:

For each $300\text{ n.m.}$ free air anomaly $\Delta g$ we have two independant estimates, one $g_T$ based on the terrestrial data

$$g_T = g_H + \delta g + \delta_T$$

(2a)

and the other $g_S$ based on the orbital perturbations of satellites,

$$g_S = g_H + \epsilon_S$$

(2b)

where $g_H$ is the true value of the contribution to $\Delta g$ from the $C_{km}$ and $S_{km}$.
estimated from the satellite theory; that is equation (1).

The $\epsilon_T$ and $\epsilon_S$ are the errors in the estimates of $g_T$ and $g_S$, while $\delta g$ is the difference between the true gravity anomaly and $g_H$. As the quantities $g_H$, $\delta g$, $\epsilon_T$ and $\epsilon_S$ are independent, we obtain the following relationships from equations 2.

$$E \{ g_H^2 \} = [ g_T g_S ]$$

$$E \{ \epsilon_S^2 \} = [ g_S^2 ] - [ g_T g_S ]$$

$$E \{ (g_T - g_S)^2 \} = [ (g_T - g_S)^2 ] = [ g_T^2 ] - 2 [ g_T g_S ] + [ g_S^2 ]$$

$$E \{ \delta g^2 \} = E \{ g_T^2 \} - E \{ \epsilon_T^2 \} - E \{ g_H^2 \}$$

Where the $E \{ \} \ldots$ denote estimates of the mean of the quantity contained in the brackets and the $[ \ldots ]$ denote the mean value of the enclosed quantity. Thus for sets of $g_T$ and $g_S$, these various estimates can be evaluated. For an estimate of the mean square of $\epsilon_T$, Kaula has adopted.

$$E \{ \epsilon_T^2 \} \approx [ [ g_T^2 ] / n ]$$

where $n$ is the number of observed 60 n.m. squares used to compute the 300 n.m. anomaly.

If the satellite solution gave a perfect estimate of $C$ and $S$, that is:

$$[ g_S^2 ] = [ g_H^2 ]$$

then $E \{ \epsilon_S^2 \} = 0$, even though $g_S$ would not contain all the information necessary to describe the total field. The information not contained in the satellite field—the errors of omission $\delta g$—then consists of the neglected higher order coefficients.

The quantity $[ (g_T - g_S)^2 ]$ provides a measure of the agreement between the two estimates $g_T$ and $g_S$ of the gravity field and is equal to the sum of the three types of errors. Thus,

$$[ (g_T - g_S)^2 ] = E \{ \epsilon_S^2 \} + E \{ \epsilon_T^2 \} + E \{ \delta g^2 \} .$$

Another estimate $D$ of $g_H$ can be obtained from the gravimetric estimates of the degree variances $V_k^2$ defined by Kaula [9]. Namely,
\[
E \left\{ g_H^2 \right\} = D = \sum_{\ell} \frac{n_\ell}{2\ell + 1} V_\ell^2
\] (5a)

where \(n_\ell\) is the number of coefficients of degree \(\ell\) included in \(g_H\).

The degree variances can also be estimated from the satellite and combination solutions according to:

\[
(V_\ell')^2 = \gamma^2 (\ell - 1)^2 \sum_m (C_{\ell m}^2 + S_{\ell m}^2)
\] (5b)

and a further estimate \(D'\) of \(g_{H-1}\) is obtained by replacing the \(V_\ell^2\) in equation (5a) by \((V_\ell')^2\).

An analogous form of this expression but using the accuracy estimates of the \(C\) and \(S\) instead of these values yields a further estimate \(F\) for \(E \left\{ \varepsilon_S^2 \right\}\); that is

\[
v_\ell^2 = \gamma^2 (\ell - 1)^2 \sum_m (\sigma_{C\ell m}^2 + \sigma_{S\ell m}^2)
\] (6a)

and

\[
F = \sum_{\ell} \frac{n_\ell}{2\ell + 1} v_\ell^2
\] (6b)

Then for both a "perfect" satellite and for "perfect" gravity data

\[
\begin{bmatrix} g_S^2 \\ g_T^2 \end{bmatrix} = \begin{bmatrix} g_T & g_S \end{bmatrix} = D = D'
\]

and

\[
E \left\{ \varepsilon_S^2 \right\} = F = 0
\]

and

\[
-E \left\{ \varepsilon_T^2 \right\} = 0
\]

IV. — Results of the comparisons

Figure 1 gives the degree variances obtained from the three data sets. Those based on surface gravity have been estimated from the world wide covariance analysis given by Kaula [7]. The estimates from the satellite solution have been computed according to equation (5b) but beyond \(\ell = 12\) the calculations are based on the assumption that for a given \(\ell\) the undetermined coefficients have the same mean square value as the determined coefficients.

Generally the estimates from the satellite solution are larger than the corresponding values from the surface gravity data (the reverse trend was noted by
SURFACE GRAVITY DATA AND SATELLITE DATA

Kaula [9] using earlier data sets for both data types. This tendency is probably the result of the linear regression methods causing estimates that are too small as already noted in the preceding section. A further explanation for some of the discrepancies is that the surface gravity degree variances have been estimated from a data set that does not cover the entire globe as does the satellite solution.

In the important range of degree 8 to 16 where we expect the two data types to be perhaps of comparable importance in contributing to the overall gravity field, the two degree variance functions show very similar tendencies, indicating that the two methods are providing estimates of the same phenomena despite the fact that the satellite terms for \( l \geq 13 \) are perhaps too large.

The \( v_k^2 \) are also given in figure 1 for the satellite and combination solutions.

The decrease of the \( v_k^2 \) with increasing \( l \) can be adequately approximated by the empirical rule [14] \( v_k^2 \approx 120/\lambda \text{ mgal}^2 \) though a constant of \( 100 \text{ mgal}^2 \) gives a somewhat better fit. The increase of the \( v_k^2 \) with increasing \( l \) can be approximated by \( v_k^2 \approx 1.8 \times 10^{-3} \lambda^3 \) and the two curves crossover in the region \( \lambda = 16 \) to \( \lambda = 18 \) indicating that, with the present data sets, there is little information to be gained in extending the field beyond 16.

Table 1 summarizes the various estimates obtained from equations (3) through (6) using only those 300 n.m. square anomalies in which the number of observed 60 n.m. squares was \( \geq 20 \) (a total of 136). For the new satellite solution the comparisons are made for the field truncated at different degrees. Also shown are the comparisons of this surface data with the 1966 SAO Standard Earth solution [15], as well as with the new combination solution truncated at 8,8, 10,10 and for the complete 16,16 field. These last comparisons are of course not entirely valid as mentioned in the preceding sections but they do serve to illustrate several points.

The new satellite solution truncated at 8,8 and the 1966 8,8 solution give equivalent comparisons. For both the \( E, e_s^2 \) are small and there is good agreement between the four estimates \( \{ [g_T g_S. [g_S^2]. D. D'] \} \) of \( E, g_h^2 \) indicating that the 8,8 field contains almost all the information in a "correct" 8,8 field. A comparison with the new combination solution truncated at 8,8 further indicates that this solution is virtually equivalent to the two satellite solutions and that the surface gravity data does not contribute significantly to the low order field. At 10,10 the new satellite solution shows a marked improvement for \( \{ (g_T - g_S)^2 \} \) and \( E, g_s^2 \) and this field is also about as good as can be expected. The 10,10 comparison with the combined solution indicates that the surface gravity data is beginning to exert influence in the solution for the coefficients. (The negative value obtained for \( E, e_s^2 \) is a direct consequence of ignoring the correlation that exists between the solution and the data against which the tests are made). Beyond about the 11th order the comparisons between the new satellite solution and the surface gravity data begin to deteriorate due to
the difficulty of estimating some of these terms as discussed in section 2. These higher order comparisons, however, do not invalidate the earlier conclusions that this satellite information is representative of the Earth's gravity field since the expected uncertainty in \[ ((g_T - g_S)^2) \] is of the order \( E \{ \epsilon_T^2 \} + E \{ \epsilon_S^2 \} \) or, for the total satellite field,

\[ ((g_T - g_S)^2) = (177 \pm 54) \text{ mgal}^2 \]

The good agreement between the two estimates \( E \{ \epsilon^2 \} \) and \( F \) up to about degree 12 indicates that the modified accuracy estimates for the \( C \) and \( S \) of the satellite solution are of the right magnitude. The discrepancies that began to occur in the satellite field beyond this are due to the fact that some of these terms are determined from resonance effects and that typically the precision estimates derived for such terms are smaller than those of the non–resonant terms, in this solution, by a factor of about 10. As these estimates may be misleading under certain conditions \([1]\) this suggests that in future combination solutions different scaling factors may have to be applied to the precision estimates derived from the two types of terms. This has not been done in the present solution. The comparisons made between the total combination solution field and the surface gravity data indicate that the errors of omission \( E \{ \delta g^2 \} \) are still quite large when compared with the \( E \{ \epsilon_T^2 \} \) and \( E \{ \epsilon_S^2 \} \) so that the surface gravity data contains additional information.

Such a conclusion, however, assumes the independence of the various quantities used in estimating these errors. Instead, the \( F \) offers a better estimate of the error in the combination solution coefficients. Also, the \( E \{ \epsilon_T^2 \} \) is likely to be too small since the estimates for \( g_T^2 \) appear too small. Thus, in reality, there appears to be little further information left in the surface gravity data.

Consequently an extension of the combination solution beyond 16,16 with the present data has not been attempted.

Additional tests with surface gravity were made for the combination solution using the recent compilations by Le Pichon and Talwani for the Atlantic \([11]\) and Indian \([12]\) Oceans. Gravity anomaly profiles (referenced to an ellipsoid of \( 1/297.0 \)) from \( 5^\circ \times 5^\circ \) area means for both data sets were computed for selected profiles and are shown in figure 5. The first profile, along latitude \( 32^\circ 5 \) in the North Atlantic, lies midway between two parallel ship cruises but the remainder are taken along the ships tracks where continuous gravity recordings were made. Table (2) summarizes the results of the comparisons. The accuracy of the surface data has been taken as \( 5 \text{ mgal} \) \([11]\), \([12]\) except for the South Atlantic profile for which \( \sigma_{5g} = 7 \text{ mgal} \). If for \( E \{ \epsilon_S^2 \} \) the value \( F \) for the total combination field (table 1) is taken, then \( E \{ \delta g^2 \} \) follows directly and its average value is \( 50 \text{ mgal}^2 \). Thus this surface data does contain additional information, at least in the areas where the geoid features are of considerable magnitude, but of short wavelength. If it is assumed that the rule \( V_\ell \approx 120 / \ell \) is valid beyond
SURFACE GRAVITY DATA AND SATELLITE DATA

\[ \ell = 16 \] then a simple calculation yields the approximate limit to which a solution can be extended if global \(5^\circ \times 5^\circ\) (or \(300 \text{ n.m.} \times 300 \text{ n.m.}\)) coverage with an accuracy of 5 mgal was available. That is

\[
\ell_{\text{max}} \approx 25
\]

\[
\sum_{\ell=17}^{120} \frac{1}{\ell} = 50
\]

V. Conclusions and Remarks

Surface gravity data is a valuable source of information for testing the satellite solutions for the geo-potential. The tests described here have indicated that the two types of information yield estimates of the same parameters. That is, a separation of the various perturbing functions has been achieved in the satellite method. Consequently, a combination of the two data sets is valid and subsequent tests have shown that such a solution yields better estimates for the parameters than either data set individually [1], [2].

With the satellite data presently available it will be difficult to determine the earth's gravity field completely beyond about \(12,12\). The presently available surface gravity data can be used to extend the field up to \(16,16\) but contains little information on a global basis beyond this.

With very precise laser tracking, with better orbital coverage and with lower perigee satellites with a good inclination distribution the present methods can provide a complete field to perhaps \(20,20\) [6] plus resonant terms.

At this resolution, considerable improvements in both the accuracy and distribution of the surface data will be required if it is to make a significant contribution to the global solution. It is unlikely that global coverage will become available in the near future so that some prediction for unsurveyed areas will still be necessary. Methods for such prediction need further investigations as do methods for assessing the accuracies of the mean gravity anomalies. In this respect the method proposed by Matheron [16], [17] may be of value if extended to the prediction of gravity anomalies to unsurveyed areas.

The data from measurements at sea indicates that complete coverage of the oceans with the present instruments will enable the field to be determined to perhaps order and degree 25.
REFERENCES


SURFACE GRAVITY DATA AND SATELLITE DATA.


Table 1
Comparison of Satellite and global solutions
with surface Gravity Measurements (mgal^2).

<table>
<thead>
<tr>
<th></th>
<th>(g_T - g_S)^2</th>
<th>[g_T g_S]</th>
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<th>D</th>
<th>D'</th>
<th>[g_T^2]</th>
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<th>E {δg^2}</th>
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Table 2
Summary of comparisons between surface data and combination solution for the profiles in Fig. 3.

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<th>PROFILE</th>
<th>([\langle y_s - y_T \rangle^2 \text{ mgal}^2])</th>
<th>(E { \epsilon_T^2 } \text{ mgal}^2)</th>
<th>(E { \epsilon_S^2 } \text{ mgal}^2)</th>
<th>(E { \delta g^2 } \text{ mgal}^2)</th>
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<td>AVERAGE</td>
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Fig. 1 - Free Air anomalies (mgal) referred to ellipsoid of flattening 1/298.255. Based on the GAPOSCHKIN and LAMBECK solution [1]. Contour interval is 10 mgal. Shaded areas correspond to negative anomalies.
Fig. 2 – Degree Variances $V_q^2$ and $v_q^2$

Fig. 3 – Comparisons of continuous gravity profiles from shipboard measurements compiled by Talwani and Le Pichon (broken lines) with profiles computed from the combination solution (solid lines).
Indian Ocean

Fig. 3 – (Cont.)

LONGITUDE

Indian Ocean

Fig. 3 – (Cont.)

JUNCTION OF INDIAN AND ATLANTIC RIDGE SYSTEMS

Fig. 3 – (Cont.)
North Atlantic
Northwest-Southeast
Profile from Halifax to Dakar
Fig. 3 – (Cont.)

South Atlantic
Northwest-Southeast
Profile from Equator to Capetown
Fig. 3 – (Cont.)