Formulation of the boundary-value problem for geoid determination with a higher-degree reference field

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Accepted 1996 March 4. Received 1996 March 4; in original form 1995 March 7

SUMMARY
In this paper we formulate the boundary-value problem for the determination of the gravimetric geoid considering a satellite gravitational model as a reference. We show that the long-wavelength part of the gravitational field generated by topographical masses must be added to the satellite model in order to prescribe a reference gravitational potential for a partly internal and partly external problem for geoid determination. We choose a reference potential that does not depend on the way topographical masses are compensated or condensed, but only on the satellite reference model and on the difference of gravitational potentials induced by topographical masses in the spaces outside the Earth and below the geoid. The latter contribution to the reference potential is expressed in the form of an ellipsoidal harmonic series, and the expansion coefficients are tabulated numerically up to degree 20.

Key words: boundary-value problem, ellipsoidal harmonics, geoid determination, gravitational potential.

1 INTRODUCTION
Two techniques have recently been employed for the determination of the geoid over continental areas. Combining GPS positioning with orthometric heights results in the geometrical geoid, whose undulations with respect to the level ellipsoid are given as the ellipsoidal (GPS-determined) heights minus the orthometric heights (e.g. Hofmann-Wellenhof, Lichtenegger & Collins 1992). On the other hand, surface gravity observations supplemented by geodetic levelling can be used to construct the so-called gravimetric geoid (e.g. Vaníček et al. 1987). As a matter of fact, these techniques are not independent, since both make use of a density hypothesis within the Earth. Only the ellipsoidal heights resulting from GPS positioning on the one hand and gravimetric data on the other hand are independently determined. The geoid, therefore, can be determined in two ways, both of which depend on the density distribution within the Earth. Hence, there is hope that in the near future these two techniques can be combined, which will lead to an improvement of density distribution modelling in the uppermost part of the Earth. In order to reach this goal, the geoid should be determined with an accuracy of 1 dm (or better), since a change in the density model for the uppermost part of the Earth from a commonly accepted constant value of 2.67 g cm\textsuperscript{-3} to a more realistic 3-D density model changes the geoid by a few decimetres only (Martinec 1993).

The magic accuracy of 1 dm in the determination of the gravimetric geoid—not yet realizable in mountainous terrain—requires not only highly accurate surface gravity observations but also accurate theories and corresponding numerical codes for geoidal height computations. The last requirement has not yet been resolved satisfactorily since existing theories for geoid computations still contain some assumptions which do not allow the desired accuracy of 1 dm to be reached.

In this paper, our aim is to formulate the boundary-value problem for the determination of the gravimetric geoid with an accuracy of 1 dm. Throughout the paper, we will call this problem the boundary-value problem for geoid determination (BVPGD). Besides the usual data of surface gravity measurements and heights of the Earth's surface above the geoid, we shall assume that the low-degree potential harmonic expansion obtained from analysis of satellite orbit perturbations, truncated approximately at degree 20, is known a priori, so that it can serve as the reference potential. Let us note that it is the spherical harmonic coefficients of the Earth's gravitational field that are derived from the analysis of satellite orbit perturbations; to obtain the ellipsoidal harmonic coefficients of this field, the transformation relations between spherical and ellipsoidal harmonics derived by Jakeli (1988) are used. Formulating the BVPGD for a higher-degree reference potential has several advantages. For instance, the truncation error of Stokes' integral applied to observed gravity data reduced to the reference gravity field is significantly smaller than in the case where the classical Stokes theory is applied to original gravity data that contain low- as well as high-frequency components (Vaníček & Sjöberg 1991). This reduced truncation
error can be evaluated numerically using a global gravity model truncated at degree 120 at most (Martinec 1993).

However, the formulation of the BVPGD with a reference potential given a priori may encounter some difficulties, since it is not as easy as in, for instance, the case of Molodensky's boundary-value problems (Heck 1991). Whereas Molodensky's problems are governed by the Laplace equation in the external space, and a reference satellite potential represents low-degree components of the solution in the whole space of interest, the reference satellite harmonics used in a partly internal and partly external BVPGD represent a solution only in the external space. The gravitational potential induced by topographical masses must be added to the satellite gravity model in order to construct the low-degree part of the solution within the topographical masses.

Vaníček et al. (1995) have made a first attempt to use satellite potential harmonics in the BVPGD as the reference. They defined the reference potential for Helmert's disturbing potential as the difference between the satellite model and the low-degree components of the direct topographical effect on the potential. This means that the reference harmonics of the sought potential depend on the way the topographical masses are condensed. In this paper, we will show that the reference potential for the BVPGD can be introduced differently, such that it does not depend on the way the topographical masses are condensed or condensed. This model better reflects the physical and mathematical background of the BVPGD, because the long-wavelength part of its solution is uniquely determined by the boundary conditions on the geoid and the Earth's surface and, of course, by the Laplace-Poisson equation. Note that the uniqueness and stability of the short-wavelength part of the solution of the BVPGD is influenced by the way the topographical masses are condensed (Engels et al 1993). Another question not answered by Vaníček et al. (1995) is how to reduce the observed surface gravity to the reference field. Again, we will show that such a reduction can be performed without specifying the model of compensation of topographical masses.

2 FORMULATION OF THE BVPGD

The BVPGD will be formulated in ellipsoidal coordinates, as, later on, after linearization, the geoid will be approximated by a level ellipsoid—the ellipsoidal coordinates are most suitable to introduce this approximation. The 3-D ellipsoidal coordinates \((u, \beta, \lambda)\), can, for instance, be introduced by their relation to Cartesian coordinates \((x, y, z)\) (e.g. Heiskanen & Moritz 1967, Sect. 1-19; Thung & Grafarend 1989).

\[
\begin{align*}
\sqrt{u^2 + E^2 \sin^2 \beta \cos \lambda}, \\
\sqrt{u^2 + E^2 \sin^2 \beta \sin \lambda}, \\
\sqrt{u \cos \beta},
\end{align*}
\]

where the parameter \(E\) is constant and defines the common focal distance of the family of confocal ellipsoidal coordinate surfaces \(u = \text{const.}\).

Let the geoid \(S_g\) be described by a function \(u = u(\Omega, \Omega)\), where \(\Omega\) stands for the pair of angular coordinates \((\beta, \lambda)\), i.e. \((u(\Omega, \Omega), \Omega)\) are points on the geoid. We will assume that the function \(u(\Omega, \Omega)\) is not known. Let \(H(\Omega) (\geq 0)\) be the height of the Earth's surface \(S_g\) above the geoid reckoned along a coordinate line \(\beta = \text{const.}, \lambda = \text{const.}\). We will assume that \(H(\Omega)\) is a known function. Finally, let the following quantities be given: the gravity \(g(\Omega)\) measured on the Earth's surface, the density \(\rho(\Omega, \Omega)\) of the topographical masses (the masses between the geoid and the Earth's surface), and the gauge value \(W_0\) of the gravity potential on the geoid.

Since it is our intention to deal with the gravity field generated by the Earth's internal masses, we make a few simplifying assumptions. First, we assume that the observations of \(g\) are corrected for the attraction of the atmosphere and the direct gravitational effect of the other bodies, mainly the moon and the sun. Second, we assume the Earth is a rigid, undeformable body, uniformly rotating (with a constant angular frequency \(\omega\)) around a fixed axis passing through its centre of mass. This assumption excludes consideration of the indirect gravitational effect of other celestial bodies, such as the tidal deformation of the Earth. Third, as already mentioned, we assume the height \(H(\Omega)\) of the Earth's surface \(S_g\) above the geoid reckoned along a coordinate line \(\Omega = \text{const.}\). It can be defined analogously to the usual orthometric heights, \(H = C/g_\Omega\), where \(C\) is the geopotential number and \(g_\Omega\) is the mean value of the \(\omega\)-component of gravity along a coordinate line \(\Omega = \text{const.}\). between the geoid and the Earth's surface. Values of \(g_\Omega\) can be estimated by a procedure similar to that used for \(g\) (Heiskanen & Moritz 1967, Sect. 4-4). However, \(g\) determined in this way is a rough estimate of the actual value. Only after finding the geoid with a high accuracy (better than 1 dm) will we be able to improve both \(g_\Omega\) and \(g\).

The question we pose is: how do we determine the gravity potential \(W(u, \Omega)\) inside and outside the topographical masses and the radius \(u(\Omega)\) of the geoid? The problem is governed by Poisson's equation with the boundary conditions given on the free boundaries \(S_g\) and \(S_e\) coupled by means of height \(H(\Omega)\):

\[
\begin{align*}
\nabla^2 W &= -4\pi G \rho + 2\omega^2 \\
|\text{grad} W| &= g \\
W &= W_0
\end{align*}
\]

outside \(S_g\), on \(S_t\), on \(S_e\),

\[
W = \frac{1}{2} \omega^2 (x^2 + y^2) + \frac{GM}{r} + O\left(\frac{1}{r^2}\right), \quad r \to \infty,
\]

where \(GM\) is the geocentric gravitational constant, \(g\) is equal to zero outside the Earth, and \(r\) is the distance from the geocentre, \(r = (x^2 + y^2 + z^2)^{1/2}\). The first-degree harmonics are left out of the potential \(W\) because of the geocentric coordinate system.

The gravity potential \(W\) can now be split into the normal (known) gravity potential \(U\) and a disturbing (unknown) gravitational potential \(T\):

\[
W = U + T,
\]

where the normal potential \(U\) is generated by a level ellipsoid (of minor semi-axis \(b_s\), say) spinning with the same angular velocity as the Earth (Heiskanen & Moritz 1967, Sect. 2-7). Throughout the paper, we will assume that the mass of the level ellipsoid is equal to the mass of the Earth, and that the mass-centre of the level ellipsoid coincides with that of the Earth. Then the zero- and first-degree harmonics of the potential \(U\) are equal to those of the potential \(W\), and thus they vanish in the disturbing potential \(T\). Moreover, we will assume that the normal gravity potential \(U\) on the level ellipsoid is equal to the actual gravity potential \(W_0\) on the
geoid. The free, non-linear boundary-value problem (2)-(5) can be reformulated for the disturbing potential \( T \):
\[
\nabla^2 T = -4\pi GQ \quad \text{outside } S_g, \\
\left| \nabla T \right| = \frac{g}{\rho} \quad \text{on } S_g, \\
U + T = W_0 \quad \text{on } S_g, \\
T \sim O\left( \frac{1}{r^2} \right) \quad r \to \infty.
\]

The unknowns to be determined by solving the problem (7)-(10) are the disturbing potential \( T \) in the space outside the geoid, and the ellipsoidal a-coordinate of \( S_g \). Clearly, the boundary value of potential \( U \) cannot be subtracted from \( W_0 \) in boundary condition (9) because the surface \( S_g \) is not known, and thus the normal potential \( U \) cannot be directly evaluated on \( S_g \). The asymptotic condition (10) imposed on \( T \) at infinity follows from (5), the fact that zero- and first-degree harmonics of the normal potential \( U \) are equal to those of the gravitational potential \( W \), and that both fields, \( W \) and \( U \), are related to the same angular velocity.

The non-linear boundary-value problem (7)-(10) with a free boundary will be treated in a linearized form. Let us define points \( P, P_g \) and \( Q \) on the Earth's surface, the geoid, and the level ellipsoid \( u = b_0 \), respectively, such that they lie on the same coordinate line \( \Omega = \text{const} \). Then the boundary condition (8) can be linearized and written in the form
\[
\left. \frac{\partial T}{\partial r} \right|_r + \frac{2}{r} \left. T \right|_r - \left. \varepsilon u(T_0) - \varepsilon u(T_p) \right|_r = -\Delta g^F,
\]
(Martinec 1990; Heck 1991; Martinec et al. 1993), where we have introduced the free-air gravity anomaly \( \Delta g^F \),
\[
\Delta g^F = g_P - g_Q + F,
\]
where \( F \) is the free-air reduction (Heiskanen & Moritz 1967), \( \varepsilon u(T_0) \) and \( \varepsilon u(T_p) \) are the ellipsoidal corrections,
\[
\varepsilon u(T_0) = -\frac{e^2}{2} \sin^2 \beta \left. \frac{\partial T}{\partial u} \right|_P,
\]
\[
\varepsilon u(T_p) = \frac{e^2}{2} \sin^2 \beta \frac{T_p}{b_0},
\]
\( e \) is the first numerical eccentricity,
\[
e = \frac{E}{\sqrt{u^2 + E^2}},
\]
and \( E_0 \) is the first eccentricity of the level ellipsoid \( u = b_0 \),
\[
E_0 = E/\sqrt{h_0^2 + E^2}
\]
It can be simply shown that leaving out the non-linear terms in eq. (11) results in a relative error of \( 10^{-7} \) and an absolute error of the order of 0.1 mgal. The bias in the geoidal heights induced by this linearization error is at most 15 mm (Seitz, Schramm & Heck 1994).

By the above linearization, the free, non-linear boundary-value problem (7)-(10) can be reduced to a fixed, linear boundary-value problem [described by eqs (7), (10) and (11)] for determining the disturbing potential \( T \) outside the surface \( S_g \). In contrast to in the problem (7)-(10), \( S_g \) is now considered to be known and fixed. The easiest and most often used way to approximate the geoid is by a mean sphere. The relative error introduced by this spherical approximation is of the order of \( 3 \times 10^{-3} \) (Heiskanen & Moritz 1967, Sect. 2-14), which then causes an error of at most 0.5 m in the geoidal heights.

To reach a better accuracy in the determination of the disturbing potential \( T \), we will approximate the geoid in the problem (7), (10) and (11) by a level ellipsoid, i.e. we put
\[
u_0(\Omega) = b_0.
\]
In fact, the actual shape of the geoid deviates from a level ellipsoid by 100 m at most. Therefore, if we treat the geoid as the level ellipsoid in the formulae relating to the disturbing potential \( T \), this causes a relative error of up to \( 1.5 \times 10^{-5} \), the absolute error in geoidal heights then does not exceed 2 mm.

After solving the fixed, linear boundary-value problem described by eqs (7), (10) and (11), and finding the disturbing potential \( T \) outside the geoid, the height \( N \) of the geoid above the level ellipsoid is obtained from boundary condition (9); its linearized form turns out to be the well-known Bruns' formula for height \( N \) (Heiskanen & Moritz 1967, eq. 2-144):
\[
N = \frac{T_{e,0}}{Q_{e,0}}.
\]

### 3 COMPENSATION OF TOPOGRAPHICAL MASSES

Now, let us look for a particular solution to the Laplace-Poisson equation (7). The gravitational potential \( V'(u, \Omega) \) induced by the topographical masses,
\[
V'(u, \Omega) = \Delta g \left( \int_{\Omega} \frac{Q'(u', \Omega')}{L(u, u', \Omega', \Omega')} w(u', \beta') du' d\Omega', \right)
\]
where
\[
w(u, \beta) = u^2 + E^2 \cos^2 \beta
\]
\( \Omega \) is the full solid angle, \( L(u, u', \Omega, \Omega') \) is the distance between the computation point \( u, \Omega \) and an integration point \( u', \Omega' \), and \( d\Omega' = \sin \beta' d\beta' d\lambda' \), is a quantity that satisfies eq. (7). [Note that the ellipsoidal approximation (16) of the geoid has already been used in the formula for the potential \( V'(u, \Omega) \).]

However, it is a well-known fact that the equipotential surfaces of \( V' \) undulate by several hundreds of metres with respect to a level ellipsoid. Thus, it is not very advantageous to consider only \( V' \) as a particular solution of eq. (7); another solution of this equation must be added to the potential \( V' \) in order to reduce its large magnitude.

The fact that the known undulations of the geoid are significantly smaller than those induced by the potential \( V' \) indicates that a compensation mechanism must exist which reduces the gravitational effect of topographical masses. This mechanism is probably mainly connected with lateral mass heterogeneities of the crust (Martinec 1994a) but also partly with deep dynamical processes (Martinec 1994b; Matyusa 1994). To describe the compensation mathematically, a number of more or less idealized compensation models have been proposed. For the purpose of geoid computation, we can, in principle, employ any compensation model generating a harmonic gravitational field outside the geoid. For instance, the topographic-harmonic compensation models (e.g. Rummet et al. 1988; Moritz 1990) are based on compensation by the anomalies of density distribution \( q_0(u, \Omega) \) in a shell between the geoid and the compensation level \( b_0 = D(\Omega), D(\Omega) > 0 \); i.e.;
the gravitational potential

\[ \psi_{\text{const}}(u, \Omega) = G \int_{\Omega} \frac{q_0(u, \Omega)}{L(u, \Omega, b_0, \Omega)} w(b_0, \Omega) \, db_0 \, d\Omega, \]

(20)

reduces the gravitational effect of topographical masses.

In the limiting case, the topographical masses can be compensated by a mass layer located on the geoid. This kind of compensation, called the Helmert second condensation (Helmert 1884), produces a potential described by the surface Newton integral:

\[ \psi_{\text{conden}}(u, \Omega) = G \int_{\Omega} \frac{\sigma(\Omega)}{L(u, \Omega, b_0, \Omega)} w(b_0, \Omega) \, db_0 \, d\Omega, \]

(21)

where \( \sigma(\Omega) \) is the density of the condensation layer. The 2-D condensation density \( \sigma(\Omega) \) can be chosen in various ways depending on the approximation used for fitting the topographical potential \( \psi \) with the condensation potential \( \psi_{\text{conden}} \).

Martinec (1993).

Having introduced a compensation mechanism for the topographical masses, the associated compensation potential \( \psi_{\text{c}} \) 'approximating' the topographical potential \( \psi \) reads

\[ \psi_{\text{c}} = \psi - \psi_{\text{const}} \text{ or } \psi = \psi_{\text{c}} + \psi_{\text{const}}. \]

(22)

for the isostatic compensation and Helmert's condensation of topographical masses respectively. Finally, a particular solution of the Laplace-Poisson equation (7) can be chosen as

\[ \delta \psi = \psi - \psi_{\text{c}}, \]

(23)

where \( \delta \psi \) is the so-called residual topographical potential, i.e. the misfit of \( \psi_{\text{c}} \) and \( \psi \).

4 A HIGHER-DEGREE REFERENCE GRAVITATIONAL POTENTIAL

Now, let us assume that some low-degree harmonics of the gravitational potential \( T \) have been determined from satellite orbit analyses. The question arises of how to reformulate the fixed boundary-value problem described by eqs (7), (10) and (11) so that a low-degree satellite gravity model can be considered as a reference gravitational potential.

Since the disturbing potential \( T \) is harmonic outside the Earth and it vanishes at infinity, it can be represented as a series of spherical harmonics \( e^j|m| q_m(e) Y_m(\Omega), j = 0, 1, \ldots, \) and \( |m| \leq j \), which all vanish at infinity (Heiskanen & Moritz 1967, p. 43, eq. 1-111b):

\[ T(u, \Omega) = \sum_{|m| = 0}^{\infty} \sum_{j = 1}^{\infty} T_{jm}(e) \left( \frac{e}{e_0} \right)^{j+1} \frac{q_{jm}(e)}{q_{jm}(e_0)} Y_{jm}(\Omega), \]

(24)

where \( q_{jm}(e) \) are defined by eq. (A4) in the Appendix. This series is convergent outside the bounding ellipsoid \( u = b_1 \), but it may be divergent in the space between the Earth's surface and this bounding ellipsoid (Sjöberg 1977; Jekeli 1983; Grafarend & Engels 1994). To define a higher-degree reference potential for the above boundary-value problem, however, we are interested only in the low-degree part of the potential \( T \). Let us thus split the disturbing potential \( T \) into the (known) low-degree reference potential \( T_{\text{r}} \) and a (unknown) higher-degree gravitational potential \( T' \):

\[ T = T_{\text{r}} + T'. \]

(25)

The set of potential coefficients \( T_{jm} \) (where superscript \( \text{r} \) stands for external) determined by analyses of perturbations of satellite orbits will be taken as the reference. The reference gravitational potential \( T_{\text{r}} \) in the space outside the bounding ellipsoid \( u = b_1 \) is then represented as

\[ T_{\text{r}}(u, \Omega) = \sum_{j = 1}^{\infty} \sum_{|m| = 0}^{\infty} T_{jm}(e) \left( \frac{e}{e_0} \right)^{j+1} \frac{q_{jm}(e)}{q_{jm}(e_0)} Y_{jm}(\Omega), \]

(26)

where \( \ell \) is the cut-off degree of the reference potential coefficients, say \( \ell = 20 \). Moreover, since \( |T_{jm}| < \infty \), the finite series (26) has finite values not only outside the bounding ellipsoid \( u = b_1 \), but also in the space between the Earth's surface and the ellipsoid \( u = b_1 \). Therefore, \( T_{\text{r}} \) represented by the finite series (26) can also be considered as the reference gravitational potential for gravity observations performed on the Earth's surface.

Analogously, the residual gravitational potential \( \delta \psi \) can also be split into low-degree and high-degree parts:

\[ \delta \psi = \delta \psi_{\text{r}} + \delta \psi' \]

(27)

The low-degree part \( \delta \psi_{\text{r}} \) is given by the low-degree parts of expansions (A16) derived in the Appendix:

\[ \delta \psi_{\text{r}}(u, \Omega) = \begin{cases} \sum_{j = 0}^{\ell} \sum_{|m| = 0}^{\infty} (V_{jm} - V_{jm}) \left( \frac{e}{e_0} \right)^{j+1} \frac{q_{jm}(e)}{q_{jm}(e_0)} Y_{jm}(\Omega) & \text{for } u \geq b_0 + H(\Omega), \\ \sum_{j = 0}^{\ell} \sum_{|m| = 0}^{\infty} (V_{jm} - V_{jm}) Y_{jm}(\Omega) & \text{for } u < b_0, \end{cases} \]

(28)

Applying the same argument as in the preceding paragraph, we represent \( \delta \psi_{\text{r}} \) by the first series not only above \( u = b_1 \), but also between the Earth's surface and the bounding ellipsoid \( u = b_1 \). Note that formula (28) does not determine the potential \( \delta \psi_{\text{r}}(u, \Omega) \) within the topographical masses.

5 REFERENCE GRAVITY ANOMALY

The crucial point of the problem of a higher-degree reference field is determining the (low-frequency) part \( \Delta \phi \) of the free-air gravity anomaly \( \Delta \phi \) that is generated by the reference potential \( T_{\text{r}} \). Obviously, \( \Delta \phi \) is given by boundary condition (11) applied to \( T_{\text{r}} \):

\[ \Delta \phi = - \frac{\partial T_{\text{r}}}{\partial u} \bigg|_{u = b_0 + H(\Omega)} - 2 \frac{\partial T_{\text{r}}}{\partial u} \bigg|_{u = b_0} + \epsilon(\Omega) \epsilon(\Omega), \]

(29)

The first and third terms of \( \Delta \phi \) can easily be computed employing the representation (26) of \( T_{\text{r}} \). Unfortunately, the same representation cannot be used for evaluating the second and fourth terms of \( \Delta \phi \), because formula (26) is valid only outside the Earth (not on the geoid). Hence, the next step will be devoted to deriving the ellipsoidal harmonic expansion of \( T_{\text{r}} \) at a point on the geoid.

The gravity potential \( W \) can be considered to be a sum of the gravitational potential \( \psi \) generated by the masses below the geoid, the topographical potential \( \psi' \), and the centrifugal

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potential $V^\mu$:
\[ W = V^\mu + V^\nu + V^\omega. \]  
(30)
The two different decompositions (6) and (30) of the gravity potential $W$ can now be put together so that the disturbing potential $T$ reads
\[ T = V^\nu + V^\omega - U. \]  
(31)
This equation is valid everywhere inside and outside the Earth. Outside the geoid, in particular, the gravitational potential $V^\nu + V^\omega - U$ is harmonic, and it can be represented by an ellipsoidal harmonic series of the form (valid also for $u = b_0$)
\[ V^\nu + V^\omega - U = \sum_{\ell=0}^{\infty} \sum_{m=-\ell}^{\ell} \frac{q_{\ell m}(u)}{q_{\ell m}(b_0)} Y_{\ell m}(\Omega) \quad \text{for} \quad u \geq b_0, \]  
where $(V^\nu + V^\omega - U)_{\ell m}$ are expansion coefficients. Substituting the last formula together with the expansions (A9) and (26) into eq. (31), comparing the coefficients by ellipsoidal harmonics $e^{\ell j} q_{\ell m}(b_0) Y_{\ell m}(\Omega)$ up to degree $\ell$, and considering the continuation property of harmonic functions, in particular the unique extension of the region of definition of a harmonic function (Kellogg 1935, Theorem V, Chap. X), we obtain
\[ (V^\nu + V^\omega - U)_{\ell m} = T_{\ell m} - V_{\ell m}^{\omega}, \]  
where $j = 0, 1, \ldots, \ell$, $|m| \leq j$, and $V_{\ell m}^{\omega}$ are given by integrals (A10).
On the geoid, $u = b_0$, formula (31) together with expansions (A11) and (32), yields
\[ T(b_0, \Omega) = \sum_{\ell=0}^{\infty} \sum_{m=-\ell}^{\ell} \left[ V_{\ell m}^{\omega} + (V^\nu + V^\omega - U)_{\ell m} \right] Y_{\ell m}(\Omega). \]  
(34)
The low-degree part $T_\ell$ of potential $T$ [see decomposition (25)] at a point on the geoid then reads
\[ T_\ell(b_0, \Omega) = \sum_{j=0}^{\ell} \sum_{m=-j}^{j} \left[ V_{\ell m}^{\omega} + (V^\nu + V^\omega - U)_{\ell m} \right] Y_{\ell m}(\Omega), \]  
(35)
or, on substituting for $(V^\nu + V^\omega - U)_{\ell m}$ from eq. (33), we have
\[ T_\ell(b_0, \Omega) = \sum_{j=0}^{\ell} \sum_{m=-j}^{j} T_{\ell m} Y_{\ell m}(\Omega), \]  
(36)
where
\[ T_{\ell m} = T_{\ell m}^{b_0} - V_{\ell m}^{\omega}. \]  
(37)
\[ j = 0, 1, \ldots, \ell, \quad \text{and} \quad |m| \leq j. \]
Finally, we are ready to evaluate the reference free-air gravity anomaly $\Delta g_f$. Substituting eqs (26) and (36) into (29), we obtain
\[ \begin{align*}
\Delta g_f(\Omega) &= \left( 1 + \frac{\ell^2}{2} \sin^2 \beta \right) b_0 \sum_{\ell=0}^{\infty} \sum_{m=-\ell}^{\ell} \left( \frac{\ell}{\ell} \right)^{\ell+1} \times \left[ \frac{1}{u^{\ell+1}} \right] \left( \frac{q_{\ell m}(u)}{q_{\ell m}(b_0)} \right) Y_{\ell m}(\Omega) \\
&\times T_{\ell m} Y_{\ell m}(\Omega) + \left( \frac{1}{u^{\ell+1}} \right) \sum_{\ell=0}^{\infty} \sum_{m=-\ell}^{\ell} \left( \frac{\ell}{\ell} \right)^{\ell+1} \times \left[ \frac{1}{u^{\ell+1}} \right] \left( \frac{q_{\ell m}(u)}{q_{\ell m}(b_0)} \right) Y_{\ell m}(\Omega), \end{align*} \]  
(38)
It is important that the reference free-air gravity anomaly $\Delta g_f$ does not depend on the way the topographical masses are compensated or condensed, but only on the reference satellite harmonics $T_{\ell m}$, the density distribution of topographical masses via differences, $V_{\ell m}^{\omega} - V_{\ell m}$, and on the topographical height $H(\Omega)$.}

6 RVPGD WITH A HIGHER-DEGREE REFERENCE FIELD

Subtracting eq. (29) from eq. (11) and using the decomposition (25), we have
\[ \frac{\partial T}{\partial u} \bigg|_{u=b_0 + \delta H} + \frac{\partial T}{\partial b_0} \bigg|_{u=b_0} = - \Delta g_f, \]  
(39)
where $\Delta g_f$ is the high-degree part of the potential $T$, and $\Delta g_f^{\omega}$ is the high-degree part of the free-air gravity anomaly,
\[ \Delta g_f^{\omega} = \Delta g_f - \Delta g_f. \]  
(40)
On the strength of assumption (16), the residual topographical potential $\delta V$, and thus also its high-degree part $\delta V$, can be considered as known quantities at points on the Earth's surface and the geoid. The latter quantity can readily be determined from formula (27), where $\delta V$ is given by the ellipsoidal harmonic expansion (28). This makes it possible to introduce a new unknown potential $T^{h}$:
\[ T^{h} = T - \delta V. \]  
(41)
By noting that the function $T - \delta V$ is harmonic outside the geoid, its high-degree part $T^{h}$, the high-degree part of Helmert's so-called disturbing gravitational potential (Martinec et al. 1993) when the topographical masses are compensated according to Helmert's second condensation technique, satisfies the boundary-value problem of the form
\[ \nabla^2 T^{h} = 0 \quad \text{for} \quad u > b_0, \]  
(42)
\[ \frac{\partial T^{h}}{\partial u} \bigg|_{u=b_0 + \delta H} = \frac{\partial T^{h}}{\partial b_0} \bigg|_{u=b_0}, \]  
(43)
\[ - \epsilon_f (T^{h}) \bigg|_{u=b_0 + \delta H} - \epsilon_f (\delta V) \bigg|_{u=b_0} = - \Delta g_f^{\omega} - \Delta g_f, \]  
(44)
where the boundary condition (43) follows immediately from the substitution of eq. (41) into eq. (39). The high-degree parts of the direct and secondary indirect topographical effects on gravity (Martinec & Vanicek 1994a, b) read
\[ \begin{align*}
\delta A^{f} &= \left. \frac{\partial \delta V(u, \Omega)}{\partial u} \right|_{u=b_0 + \delta H} - \sum_{\ell=0}^{\infty} \sum_{m=-\ell}^{\ell} \left( V_{\ell m}^{\omega} - V_{\ell m} \right) \left( \frac{\ell}{\ell} \right)^{\ell+1} \times \left[ \frac{1}{u^{\ell+1}} \right] \left( \frac{q_{\ell m}(u)}{q_{\ell m}(b_0)} \right) Y_{\ell m}(\Omega), \end{align*} \]  
(45)
eqs (25) and (41) into Bruns' formula (17), we obtain

\[ N = N_N + \frac{1}{\gamma_0} (\mathbf{T}^{a,b} + \mathbf{\Delta}^{a,b}) \mathbf{I}_{N}, \]

where we have introduced low-degree geoidal undulations \( N_N \) as

\[ N_N = \frac{1}{\gamma_0} T_{L} \mathbf{I}_{N}. \]

From eqs (36) and (37), we obtain

\[ N_N = \frac{1}{\gamma_0} \sum_{j=0}^{\infty} \sum_{m=-j}^{j} (T_{\mathbf{I}}^{a,b} + V_{\mathbf{I}}^{a,b} - V_{\mathbf{I}}^{b,a}) \mathbf{I}_{N,0}(\Omega). \]

It should be emphasized that neither \( \Delta g^{a,b} \) nor \( N_N \) depends on the way topographical masses are compensated or condensed, but only on the reference harmonics \( T_{\mathbf{I}}^{a,b} \) and on the differences \( V_{\mathbf{I}}^{a,b} - V_{\mathbf{I}}^{b,a} \) of ellipsoidal harmonics induced by topographical masses. On the other hand, the boundary-value problem (42)-(44) for \( T^{a,b} \) depends on the way the topographical masses are compensated. There is an open question, not addressed here, whether Helmert's now popular condensation technique (e.g. Martinec et al. 1993) is the best way to compensate the gravitational effect of topographical masses when the problem (42)-(44) is to be solved.

### 7 NUMERICAL RESULTS FOR \( V_{j}^{a,b} - V_{j}^{b,a} \)

Let us try now to estimate the effect of the gravitational field generated by topographical masses on the reference free-air gravity anomaly \( \Delta g^{a,b} \) and the reference geoidal undulations \( N_N \). To do this, we have evaluated the differences \( V_{j}^{a,b} - V_{j}^{b,a} \) for low degrees, \( j = 0, 1, \ldots, 20 \), by a numerical quadrature applied to integrals (A10) and (A12) found in the Appendix. As a first approximation of \( V_{j}^{a,b} - V_{j}^{b,a} \), we have assumed that the density \( \rho(u, \Omega) \) of topographical masses is constant and equal to the mean crustal density of \( q_0 = 2.67 \text{ g cm}^{-2} \). The actual density of topographical masses is expected to vary around \( q_0 \) by 10 to 20 per cent. Later on, we will estimate the effect of such topographical density variations on the differences \( V_{j}^{a,b} - V_{j}^{b,a} \).

Table 1 gives the differences \( V_{j}^{a,b} - V_{j}^{b,a} \), \( j = 0, 1, \ldots, 20 \), \( |m| \leq j \), for the IUGS/7 global spherical harmonic terrain model (Wieser 1987) complete up to degree and order 180. We have found that the contributions of \( V_{j}^{a,b} \) and \( V_{j}^{b,a} \), \( j = 0, 1, \ldots, 20 \), \( |m| \leq j \), to geoidal heights lie within the interval \((-2.80, 0) \text{ m})\); the minimum \(-2.80 \text{ m}\) is located in the Himalayas. Note that this result is in agreement with that obtained by Sjöberg (1994). A plot of this effect for the territory of Canada is shown in Fig. 1. As expected, the minimum value, which reaches \(-0.43 \text{ m}\), is connected with the highest part of the Canadian Rocky Mountains. Since the density of topographical masses enters Newton's integral linearly, in order to achieve the 1 dm accuracy of geoidal heights, the regional density of the Rocky Mountains massif should be known with a relative accuracy of better than 25 per cent.

Inspecting eq. (38), we can observe that the part of the reference free-air gravity anomaly \( \Delta g^{a,b}(\Omega) \) that originates from differences \( V_{j}^{a,b} - V_{j}^{b,a} \) is

\[ \Delta g_{a,b,\text{topo}}(\Omega) \approx -\frac{2}{b_i} \sum_{j=0}^{\infty} \sum_{m=-j}^{j} (V_{j}^{a,b} - V_{j}^{b,a}). \]

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Table 1. Spherical harmonic coefficients $V_{lm}^{12} - V_{lm}^{13}$ (in $m^2 s^{-2}$) for the TUG87 global terrain model (Wieser 1987) and for $s_{0} = 2.67 g cm^{-3}$.

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Boundary-value problem for geoid determination
We have evaluated $\Delta gr_{\text{e.g.}}(Q)$ globally for $\ell = 20$ and found that it reaches at most 0.09 mgal.

8 CONCLUSION

To employ a reference potential in the computation of the gravimetric geoid has an evident advantage: it reduces the magnitude of quantities we work with and thus enables us to linearize the originally non-linear boundary-value problem for geoid determination. There are certainly a lot of possible ways to bring the reference potential into the geoid computation. Here, a reference potential has been considered to consist of the harmonics derived from the analyses of satellite orbits.

We have assumed that such a priori information on long-wavelength components of the gravitational field is fixed and should not be corrected from surface gravity observations. This fact is expressed by an asymptotic condition (44), which says that only the short-wavelength part of the gravitational potential is sought from surface gravity data. On the other hand, introducing a reference potential into the problem of geoid determination requires that the reference potential harmonics be accurate enough and that they contain meaningful information. Therefore, we have assumed that reference satellite harmonics are taken up to degree about 20.

It should be emphasized that satellite potential harmonics define the reference gravitational potential in the space external to the Earth. In order to construct a reference potential for a partly internal and partly external boundary-value problem of geoid determination, the low-degree part of the gravitational potential induced by topographical masses must be taken into account. Vanicek et al. (1995) have already formulated the BVPGD for the case where a satellite reference potential is taken as the reference. They confined themselves to the so-called Stokes-Helmert technique for geoid computation, and introduced the reference potential for Helmert's disturbing potential as a satellite gravitational potential minus the direct topographical effect on the potential. Evidently, such a reference field depends on the way topographical masses are condensed.

Here, we were motivated by whether the BVPGD could be formulated in such a way that a higher-degree reference potential would be independent of the way the topographical masses are compensated. We have shown that such a formulation exists; the reference free-air gravity anomaly as well as the reference geoidal height are determined by a satellite gravitational model and by the differences of the external and internal gravitational fields generated by topographical masses.

The reference potential for geoid determination is insensitive to the way the topographical masses are compensated. Numerically, the magnitude of that part of the reference geoidal heights which comes from low-degree components of the topographical potential is approximately three times larger than the corresponding direct topographical effect on the potential in the Stokes-Helmert technique (Vanicek et al. 1995).

ACKNOWLEDGMENTS

The authors are grateful to B. Heck and an anonymous reviewer for their comments on the submitted version of the manuscript. They have helped to clarify some points and improve the presentation. Ms W. Wells has done an excellent job of correcting the English language of this paper. The research has been partly sponsored by the Grant Agency of the Czech Republic through Grant No. 205/94/0500.

REFERENCES

Bjerhammar, A. 1987. Discrete Physical geodesy, Rep. 380, Dept. of Geodetic Science and Surveying, The Ohio State University, Columbus, OH.


Helmert, F.R., 1884. Die mathematischen und physikalischen Theorien


Moritz, H., 1990. The Figure of the Earth: Theoretical Geodesy and the Earth's Inertia, Wicksman, Karlskrona.

Rummel, R., Rapp, R.H., Stükel, H. & Tscherning, C.C., 1988. Comparisons of global topographic/isostatic models to the Earth's observed gravity field, Rep. 388, Dept. of Geodetic Science and Surveying, The Ohio State University, Columbus, OH.


Sjöberg, L., 1977. On the errors of spherical harmonic developments of gravity at the surface of the Earth, Rep. 257, Dept. of Geodetic Science and Surveying, The Ohio State University, Columbus, OH.


APPENDIX A: ELLIPTOIDAL HARMONIC REPRESENTATION OF $\delta V$

In this Appendix, we express the residual topographical potential $\delta V$ in a spectral form. Because of the ellipsoidal approximation (16) of the geoid, it is advantageous to represent $\delta V$ in terms of ellipsoidal harmonics. The expansion of the reciprocal distance in terms of ellipsoidal coordinates can be found in Hobson (1965, pp. 424–430; for $u > u'$,

$$\frac{1}{L(u, \Omega, u', \Omega')} = 4\pi \sum_{j=0}^{\infty} \sum_{m=-j}^{j} (-1)^j \frac{(j+m)!}{(j^2 - m^2)} P_j^m(u') \times P_j^m(u) R_{jm}(\Omega) R_{jm}(\Omega'),$$

(A1)

where $i = \sqrt{-1}$, $P_j^m(u'/u)$ and $Q_j^m(u'/u)$ are the Legendre functions of the first and second kinds normalized according to Hobson (1965). $Y_{km}(\Omega)$ are fully normalized spherical harmonics normalized according to Varshalovich, Moskalev & Khersonskii (1989), and the asterisk denotes the complex conjugate.

Thong (1993) has shown that Legendre's functions $P_j^m(u'/u)$ and $Q_j^m(u'/u)$ can be written as infinite power series of the first eccentricity $e$, defined by eq. (15), as

$$P_j^m(u'/u) = (-1)^{j/2} e^{- j} p_{jm}(e),$$

(A2)

$$Q_j^m(u'/u) = (-1)^{(j+1)/2} e^{j+1} q_{jm}(e),$$

(A3)

where

$$p_{jm}(e) = \sum_{k=0}^{\infty} d_{jm,k} e^{2k},$$

$$q_{jm}(e) = \sum_{k=0}^{\infty} b_{jm,k} e^{2k}.$$  

(A4)

The coefficients $d_{jm,k}$ and $b_{jm,k}$ can be defined, for instance, by the following recurrence relations:

$$d_{jm,k} = \frac{(-j + 2k - 2)^2 - m^2}{2k(2k - 2 - 2k - 1)} d_{jm,k-1}$$

for $k \geq 1$  

(A5)

with $d_{jm,0} = 1$, and

$$b_{jm,k} = \frac{(j + 2k - 1)^2 - m^2}{2k(2k + 1)} b_{jm,k-1}$$

for $k \geq 1$.  

(A6)

again with $b_{jm,0} = 1$. For points lying on the Earth's surface or close to the Earth's surface, the eccentricities $e$ and $e'$ are very close and smaller than $10^{-3}$ at most, and the series over $k$ in eq. (A4) quickly converge for low degrees $j$. Later on, we will restrict ourselves to ellipsoidal harmonics of degrees $j \leq 20$; in such a case, it will be sufficient to sum the series in eq. (A4) only up to $k = 2$ and still keep the relative accuracy of the order of 227

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with coefficients

\[
V_{jm}^{\delta u} = \frac{4\pi G}{2j+1} \left( \frac{c_0}{\epsilon_0} \right) \left( \frac{c}{\epsilon} \right)^{j+1} \sum_{\nu=0}^{\infty} \int_{\Omega'} \frac{q_{jm}(\epsilon')}{q_{jm}(c_0)} Y_{jm}(\epsilon') \frac{d\Omega'}{d\Omega}.
\]

(A12)

As we have discussed, the compensation of topographical masses plays an important role in geoid determination. Employing expansion (A7) of the Newton kernel, the compensation potential \( V^c \), eq. (22), at points outside/on the geoid can be expressed as a series of ellipsoidal harmonics:

\[
V^c(u, \Omega) = \sum_{j=0}^{\infty} \sum_{m=-j}^{j} V^c_{jm} \left( \frac{c}{c_0} \right)^{j+1} \frac{q_{jm}(c)}{q_{jm}(c_0)} Y_{jm}(\Omega) \quad \text{for } u \geq b_0,
\]

(A13)

where the coefficients \( V^c_{jm} \) are equal to

\[
V^c_{jm} = \frac{4\pi G}{2j+1} \left( \frac{c_0}{\epsilon_0} \right) \left( \frac{c}{\epsilon} \right)^j \sum_{\nu=0}^{\infty} \int_{\Omega'} q_{jm}(\epsilon') Y_{jm}(\Omega') w(\epsilon', \beta') d\epsilon' d\Omega'.
\]

(A14)

for the isostatic compensation of topographical masses, and

\[
V^s_{jm} = \frac{4\pi G}{2j+1} \left( \frac{c_0}{\epsilon_0} \right) \left( \frac{c}{\epsilon} \right)^j \sum_{\nu=0}^{\infty} \int_{\Omega'} q_{jm}(\epsilon') Y_{jm}(\Omega') w(b_0, \beta') d\epsilon' d\Omega',
\]

(A15)

for Helmert’s second condensation.

Finally, we are ready to write the ellipsoidal harmonic representation of the residual topographical potential \( \delta V = V^s - V^c \). Using expansions (A9), (A11), and (A13), the potential \( \delta V \) reads

\[
\delta V(u, \Omega) = \sum_{j=0}^{\infty} \sum_{m=-j}^{j} \left( V^s_{jm} - V^c_{jm} \right) \left( \frac{c}{c_0} \right)^{j+1} \frac{q_{jm}(c)}{q_{jm}(c_0)} Y_{jm}(\Omega).
\]

(A16)

for \( u \geq b_0 \),

\[
\delta V(u, \Omega) = \sum_{j=0}^{\infty} \sum_{m=-j}^{j} \left( V^s_{jm} - V^c_{jm} \right) Y_{jm}(\Omega)
\]

for \( u = b_0 \),

where the coefficients \( V^s_{jm} \), \( V^c_{jm} \), and \( V_{jm}^{\delta u} \) are given by eqs (A10), (A12), and (A14), respectively.