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Determination of regional geoid models by combining local and global datasets

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The geoid as an equipotential surface of the Earth's gravity field plays an essential role in geosciences and engineering. A geoid model can be determined by formulating a boundary value problem with the boundary (the geoid) being itself a function of the disturbing potential.

In principle, the solution of this boundary value problem by Stokes's (1849) method requires the global distribution of the gravity anomalies. Yet the application of the Stokes integral formula remains impractical, due to incomplete geographical coverage of the (terrestrial) gravity data. Broad characteristics of the global gravity field can be defined from the accurate tracking of Low Earth Orbit satellites. The recent advances of satellite technology (e.g. the gravimetric GRACE mission) have resolved the long-wavelength component of the global geoid with an accuracy of a few cm. However, the spatial resolution of such information is limited to about 200 km. Further improvements to the Earth gravity models at medium and short wavelengths (regional scale) should come from the use of terrestrial surveys and satellite altimetry (over the oceans). In particular, modifications of the original Stokes integral formula combine local terrestrial gravity anomalies and the GGM-derived long-wavelength component (i.e. the global trend) of the geoid. This combination is very effective since some recent advances in technology and geodetic theory have created necessary conditions for achieving 1-cm accuracy in high-resolution regional geoid modeling.

The Stokesian integration requires gravity observations referred to the geoid as a boundary surface while in reality gravity measurements are taken at the topographic surface. Thus, to satisfy the boundary condition, the gravity anomalies need to be downward continued to the geoid level. Downward continuation requires harmonicity of the quantities to be downward continued; thus a number of different corrections related to the existence of the masses that are external to the geoid (such as topography and atmosphere) need to be accounted for very carefully. One way of estimating the effect of external masses is to use Helmert's (1884) second condensation model. According to this model, the masses are replaced by an infinitesimal condensation layer on the geoid. To use this model, the observed gravity anomalies are first transformed into Helmert anomalies, which will then be used for downward continuation from the Earth's surface to the geoid.

Accordingly, this contribution focuses on Stokes-Helmert's approach to geoid determination and various important aspects that are often overlooked in regional geoid modeling.