Geophys. J. Int. (2018) **XXX**, 1–2

Gravity disturbance or gravity anomaly?

Vanderlei C. Oliveira Jr[1*], Valéria C. F. Barbosa[1] and Leonardo Uieda[2]
[1] *Department of Geophysics, Observatório Nacional, Rio de Janeiro, Brazil*
[2] *Department of Geology and Geophysics, University of Hawaii, Manoa, USA*

Received 2018 Month XX; in original form 2018 Month XX

SUMMARY
Gravity anomalies have long been used by geophysicists for the purpose of determining density distributions in subsurface.
However, gravity anomalies
In this paper, we discuss the fundamental concepts
URGENTE: (Marussi et al. 1974), (Torge & Müller 2012), section 4.2.1

Key words: potential fields – gravity disturbance – gravity anomaly – gravity modeling.

1 INTRODUCTION

The resultant of gravitational force and centrifugal force acting on a body at rest on the Earth's surface is called *gravity vector* and its intensity is called simply *gravity* (Hofmann-Wellenhof & Moritz 2005). In the case of gravimetry on moving platforms (e.g., airplanes, helicopters, marine vessels), there are additional non-gravitational accelerations due to the vehicle motion, such as Coriolis acceleration and high-frequency vibrations (Glennie et al. 2000; Nabighian et al. 2005; Baumann et al. 2012). Geophysicists use gravity for estimating the Earth's internal density distribution whereas geodesists use gravity to estimate the geoid (Li & Götze 2001). Hence, geophysicists are usually interested in the gravitational component of the observed gravity, which is produced by the Earth's internal density distribution. The first step of the procedure for isolating this gravitational component consists in removing the non-gravitational effects due to the vehicle motion and also the time variations such as Earth tides, instrumental drift and barometric pressure changes, for example. If these effects are properly removed, the resultant gravity data can be considered as the sum of a centrifugal component due to the Earth's rotation and a gravitational component produced by the whole Earth's internal density distribution. The isolation of this particular gravitational component and its subsequent use for estimating density distributions related to geological structures in subsurface are the main goals in applied geophysics (Blakely 1996).

Based on well-established concepts of the literature, we present a discussion aiming at bringing some light to the following question: in geophysical applications, should we use the gravity disturbance or gravity anomaly? It seems that this theoretical issue has been debated within the scientific community from a more geodetic than geophysical point of view (LaFehr 1991; Chapin 1996; Li & Götze 2001; Fairhead et al. 2003; Hackney & Featherstone 2003; Hinze et al. 2005). Our reasoning suggests that the gravity disturbance is more appropriated than gravity anomalies for approximating the gravitational effect produced by the Earth's internal density distribution.

* Sei la, Nacional, Brazil

2 NORMAL EARTH AND NORMAL GRAVITY

Traditionally, the Earth's gravity field is approximated by the gravity field produced by a geocentric and rigid ellipsoid of revolution, which has the minor axis b coincident with the mean rotating axis of the Earth Z, the same total mass (including the atmosphere) and also the same angular velocity of the Earth (Figure A1). Another characteristic of this model is that its limiting surface coincides with a particular equipotential of its own gravity field. This model is called as *normal Earth* (Heiskanen & Moritz 1967; Vaníček & Krakiwsky 1987; Hofmann-Wellenhof & Moritz 2005; Torge & Müller 2012).

It is worth noting that, although the normal Earth has the same total mass (including the atmosphere) of the Earth, its internal density distribution is unknown.

The only condition imposed on its internal density distribution is that it produces a gravity field having a particular equipotential which coincides with its limiting surface.

For convenience, we denote any density distribution satisfying this condition as a *normal density distribution*.

Similarly to the gravity vector and gravity, the resultant of the virtual gravitational and centrifugal forces exerted by the normal Earth on a body at rest at a point P is called *normal gravity vector* and its intensity is called simply *normal gravity*.

The normal Earth gives rise to the *geodetic coordinate system*. In this coordinate system, the position of a point P is defined by the *geometric height* h, *geodetic latitude* φ and *longitude* λ (Figure A1). Geodetic coordinates (h, φ, λ) can be easily converted into geocentric Cartesian coordinates (X, Y, Z) (Figure A1).

The plane containing the point P, the axis Z and the origin O of this *geocentric Cartesian coordinate system* is called meridian plane (gray plane in Figure A1).

ACKNOWLEDGMENTS

The authors would like to thank the editor and all the reviewers for their criticisms and corrections.

REFERENCES

Baumann, H., Klingelé, E., & Marson, I., 2012. Absolute airborne gravimetry: a feasibility study, *Geophysical Prospecting*, **60**(2), 361–372.

Blakely, R. J., 1996. *Potential Theory in Gravity and Magnetic Applications*, Cambridge University Press.

Chapin, D. A., 1996. The theory of the bouguer gravity anomaly: A tutorial, *The Leading Edge*, **15**(5), 361–363.

Fairhead, J. D., Green, C. M., & Blitzkow, D., 2003. The use of gps in gravity surveys, *The Leading Edge*, **22**(10), 954–959.

Glennie, C. L., Schwarz, K. P., Bruton, A. M., Forsberg, R., Olesen, A. V., & Keller, K., 2000. A comparison of stable platform and strapdown airborne gravity, *Journal of Geodesy*, **74**(5), 383–389.

Hackney, R. I. & Featherstone, W. E., 2003. Geodetic versus geophysical perspectives of the gravity anomaly, *Geophysical Journal International*, **154**(1), 35–43.

Heiskanen, W. A. & Moritz, H., 1967. *Physical Geodesy*, W.H. Freeman and Company.

Hinze, W. J., Aiken, C., Brozena, J., Coakley, B., Dater, D., Flanagan, G., Forsberg, R., Hildenbrand, T., Keller, G. R., Kellogg, J., Kucks, R., Li, X., Mainville, A., Morin, R., Pilkington, M., Plouff, D., Ravat, D., Roman, D., Urrutia-Fucugauchi, J., Véronneau, M., Webring, M., & Winester, D., 2005. New standards for reducing gravity data: The north american gravity database, *Geophysics*, **70**(4), J25–J32.

Hofmann-Wellenhof, B. & Moritz, H., 2005. *Physical Geodesy*, Springer.

LaFehr, T. R., 1991. Standardization in gravity reduction, *Geophysics*, **56**(8), 1170–1178.

Li, X. & Götze, H.-J., 2001. Ellipsoid, geoid, gravity, geodesy, and geophysics, *Geophysics*, **66**(6), 1660–1668.

Marussi, A., Moritz, H., Rapp, R. H., & Vicente, R. O., 1974. Ellipsoidal density models and hydrostatic equilibrium: Interim report, *Physics of the Earth and Planetary Interiors*, **9**(1), 4–6.

Nabighian, M. N., Ander, M. E., Grauch, V. J. S., Hansen, R. O., LaFehr, T. R., Li, Y., Pearson, W. C., Peirce, J. W., Phillips, J. D., & Ruder, M. E., 2005. Historical development of the gravity method in exploration, *GEOPHYSICS*, **70**(6), 63ND–89ND.

Torge, W. & Müller, J., 2012. *Geodesy*, de Gruyter, 4th edn.

Vaníček, P. & Krakiwsky, E. J., 1987. *Geodesy: The Concepts, Second Edition*, Elsevier Science.

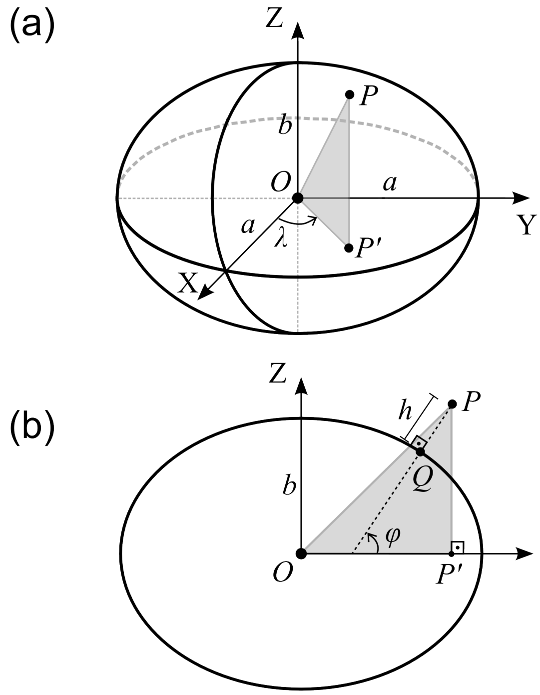

Figure A1. Schematic representation of the geodetic coordinate system defined by an oblate ellipsoid with semi-minor axis b, coincident with the mean Earth's rotation axis, and a semi-major axis a. In this coordinate system, the position of a point is determined by the geometric height h, the geodetic latitude φ and longitude λ. The Earth's origin is represented by O, P represents a point (h, φ, λ) and P' represents the projection of P on the plane XY (Equatorial plane). The plane containing O, P and P' is represented by the gray triangle in (a) and (b). In (b), the point Q represents the projection of P on the ellipsoid surface.