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Title: Insights from recent gravity satellite missions in the density structure of continental margins - with focus on the passive margins of the South Atlantic

Article Type: SI: Rifting to Passive Margins

Keywords: continental margins; satellite gravity missions; spatial resolution; omission error; interpretation gravity effects; interpretation gravity gradients.

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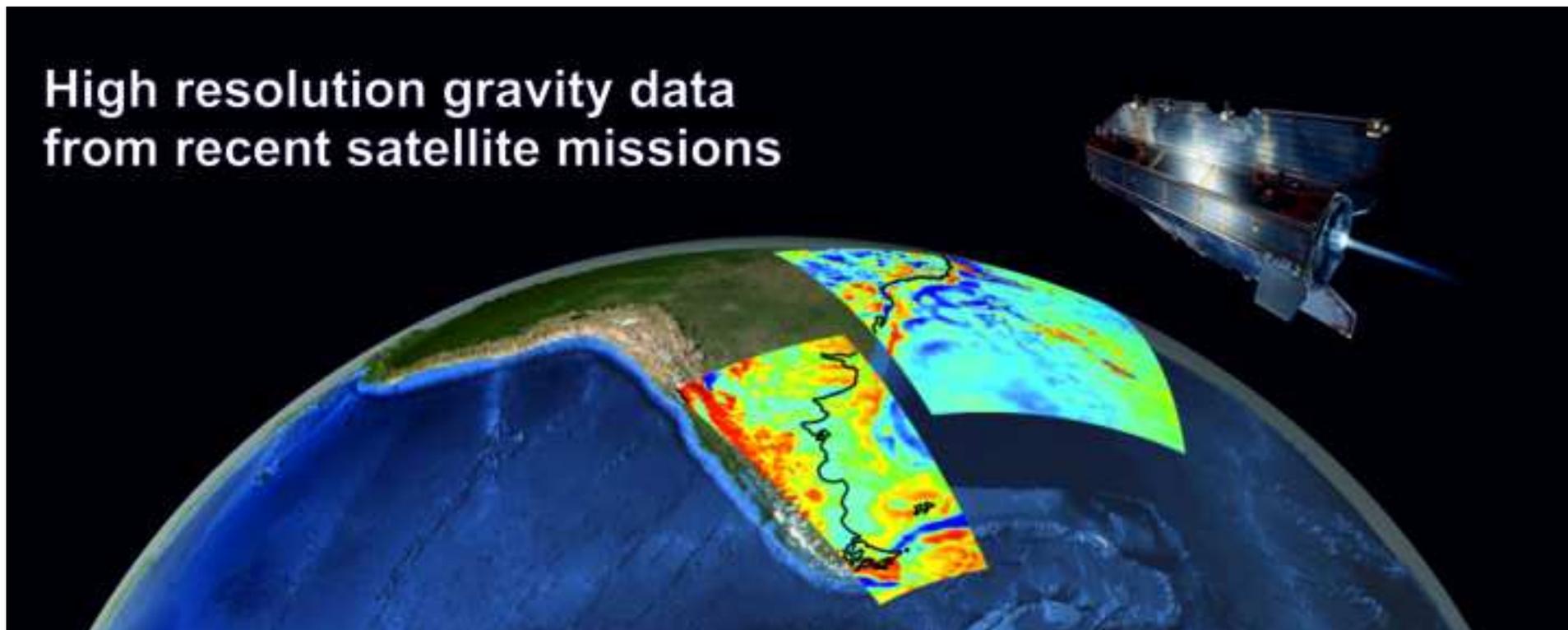
First Author: Hans-Juergen Goetze, Prof. Dr.

Order of Authors: Hans-Juergen Goetze, Prof. Dr.

Abstract: We focus on new gravity and gravity gradient data sets from modern satellite missions GOCE, GRACE and CHAMP, and their geophysical interpretation at passive continental margins of the South Atlantic. Both sides, South Africa and South America, have been targets of hydrocarbon exploration and academic research of the German Priority Program SAMPLE (South Atlantic Margin Processes and Links with onshore Evolution). The achievable spatial resolution, driven by GOCE, is 70 - 80 km. Therefore, most of the geological structures, which cause a significant gravity effect (by both size and density contrast), can be resolved. However, one of the most important aspects is the evaluation of the omission error, which is not always in the focus of interpreters. It results from high-frequency signals of very rough topographic and bathymetric structures, which cannot be resolved by satellite gravimetry due to the exponential signal attenuation with altitude. The omission error is estimated from the difference of the combined gravity model EIGEN-6C4 and the satellite-only model GOCO05S. It can be significantly reduced by topographic reductions. Simple 2D density models and their related mathematical formulas provide insights in the magnitude of the gravity effect of masses that form a passive continental margin. They are contrasted with results from satellite-only and combined gravity models. Example geophysical interpretations are given for the western and eastern margin of the South Atlantic Ocean, where standard deviations vary from 25 - 16 mGal and 21 - 11 mGal, respectively. It could be demonstrated, that modern satellite gravity data provide significant added value in the geophysical gravity data processing domain and in the validation of heterogeneous terrestrial data bases. Combined models derived from high-resolution terrestrial gravity and homogeneous satellite data will lead to more detailed and better constrained lithospheric density models, and hence will improve our knowledge about structure, evolution and state of stress in the lithosphere.

Response to Reviewers:

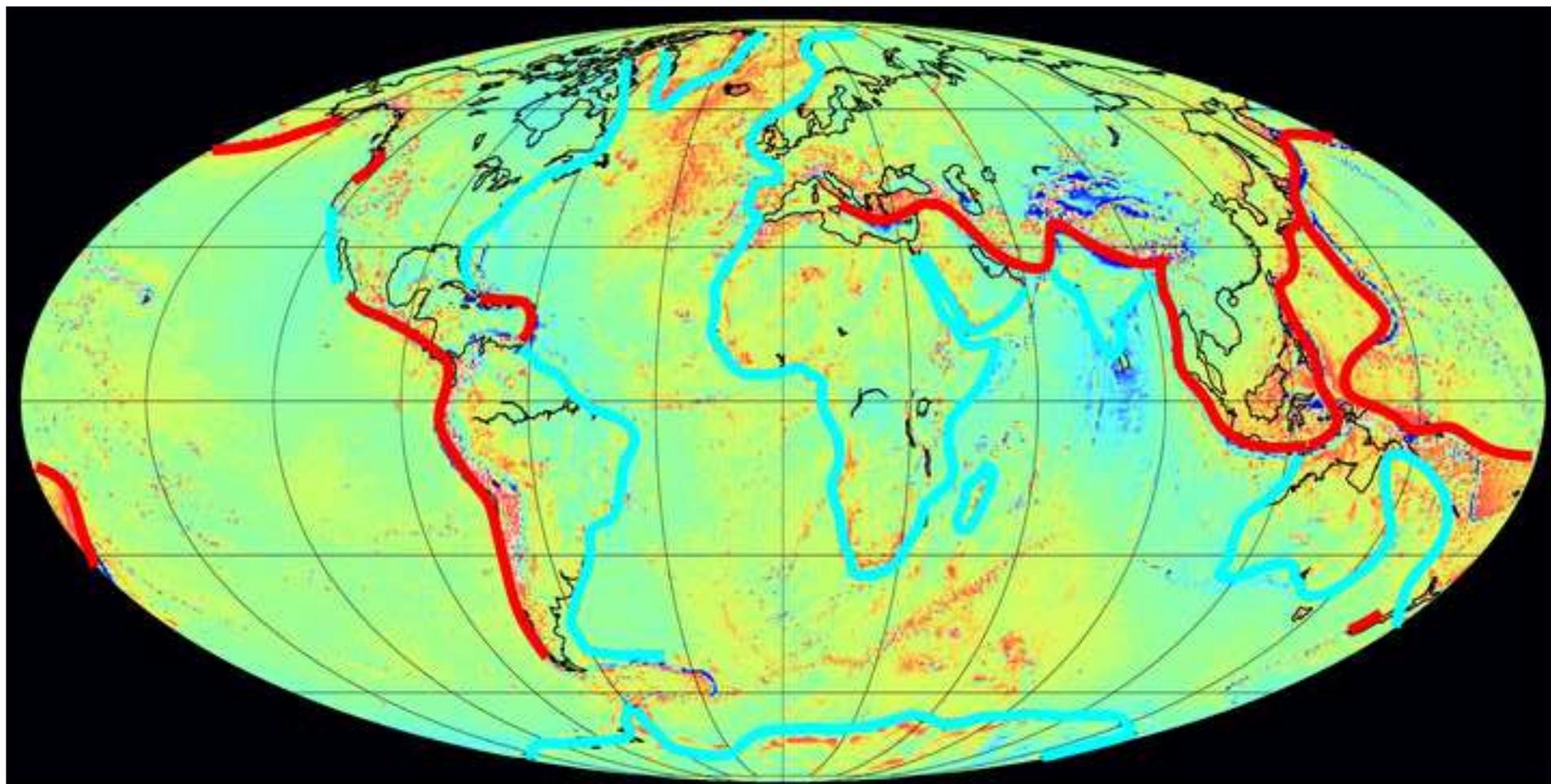
# High resolution gravity data from recent satellite missions



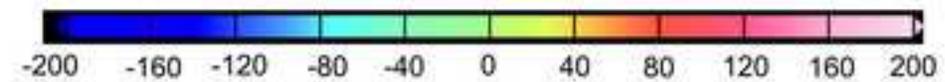
## \*Highlights (for review)

- Combine terrestrial and satellite gravity to provide insights into passive margins.
- Satellite gravity and gradients were used to validate terrestrial gravity databases.
- 2nd derivations of satellite gravity can re-examine the continent-ocean boundaries.

Figure  
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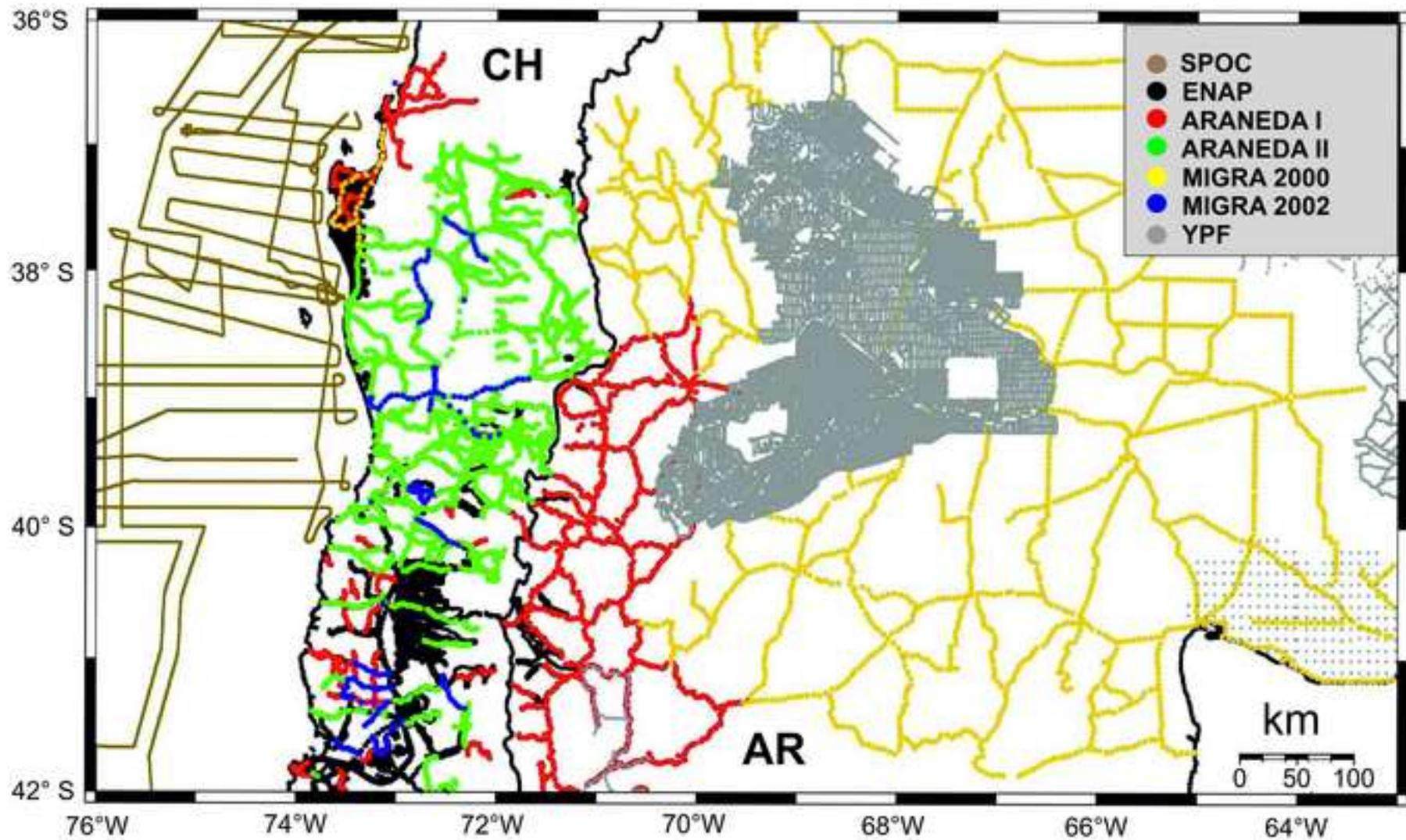


 Passive continental margins  
 Active continental margins



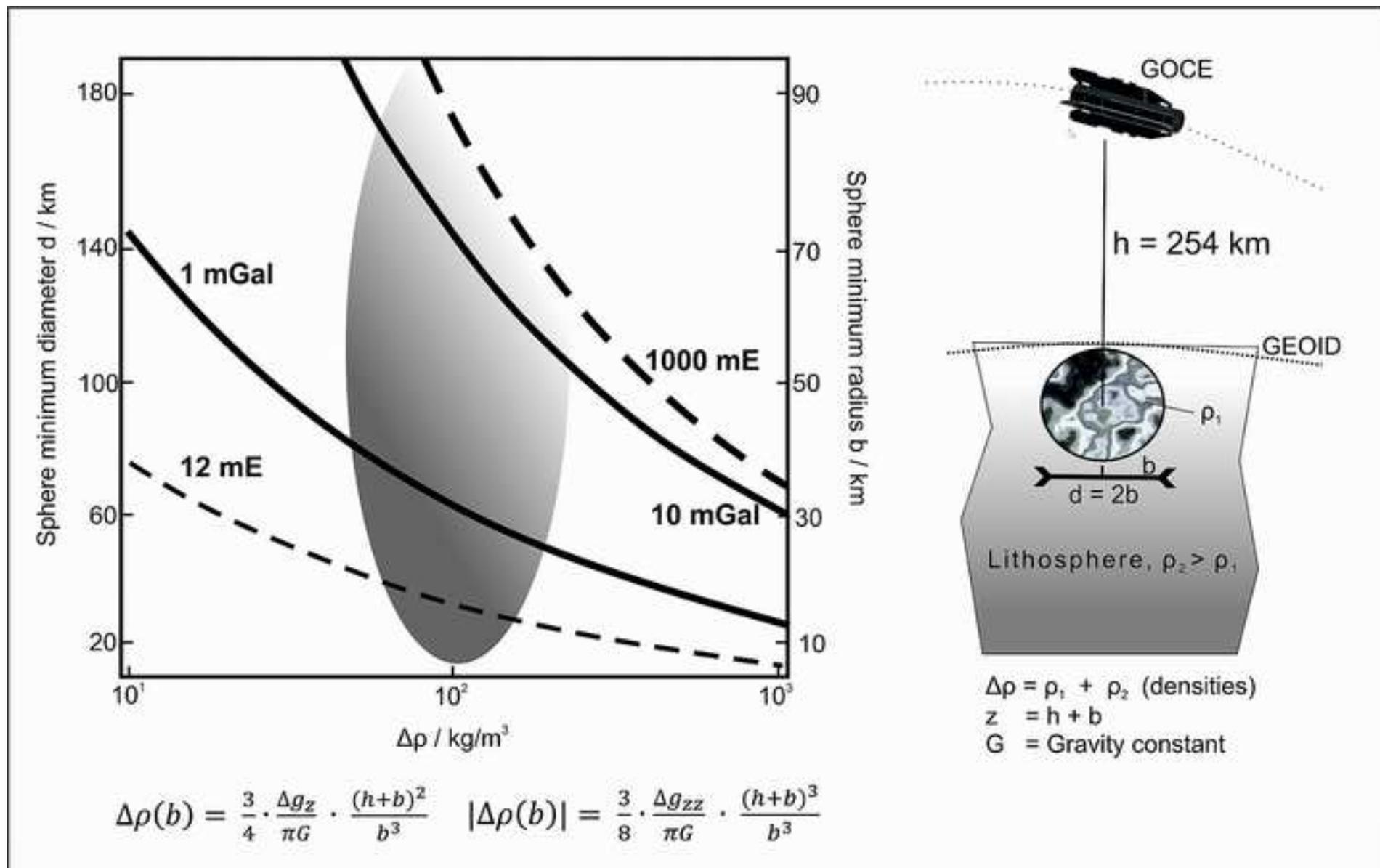
Global field of gravity disturbance, units:  $10^{-5} \text{ m/s}^2$

### Distribution of observations



Figure

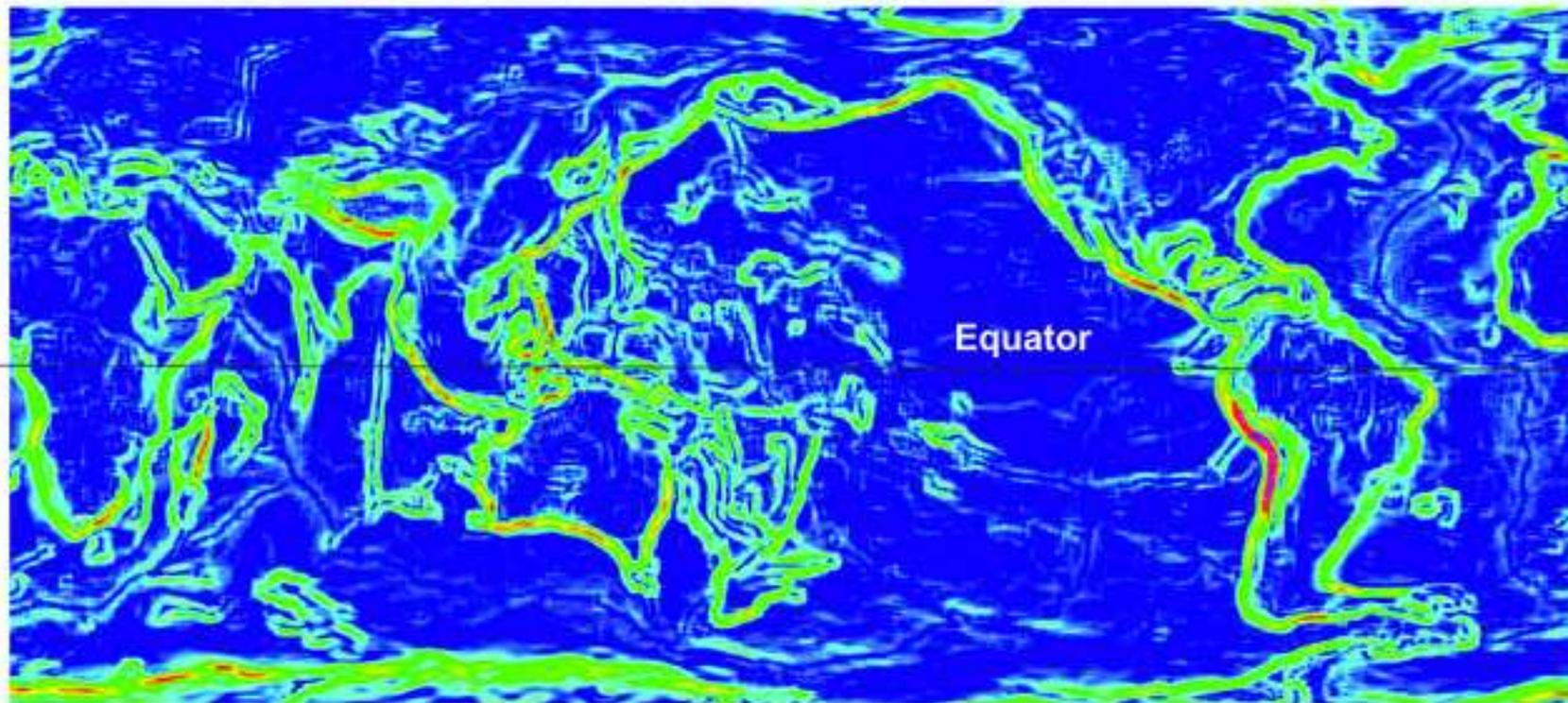
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Figure

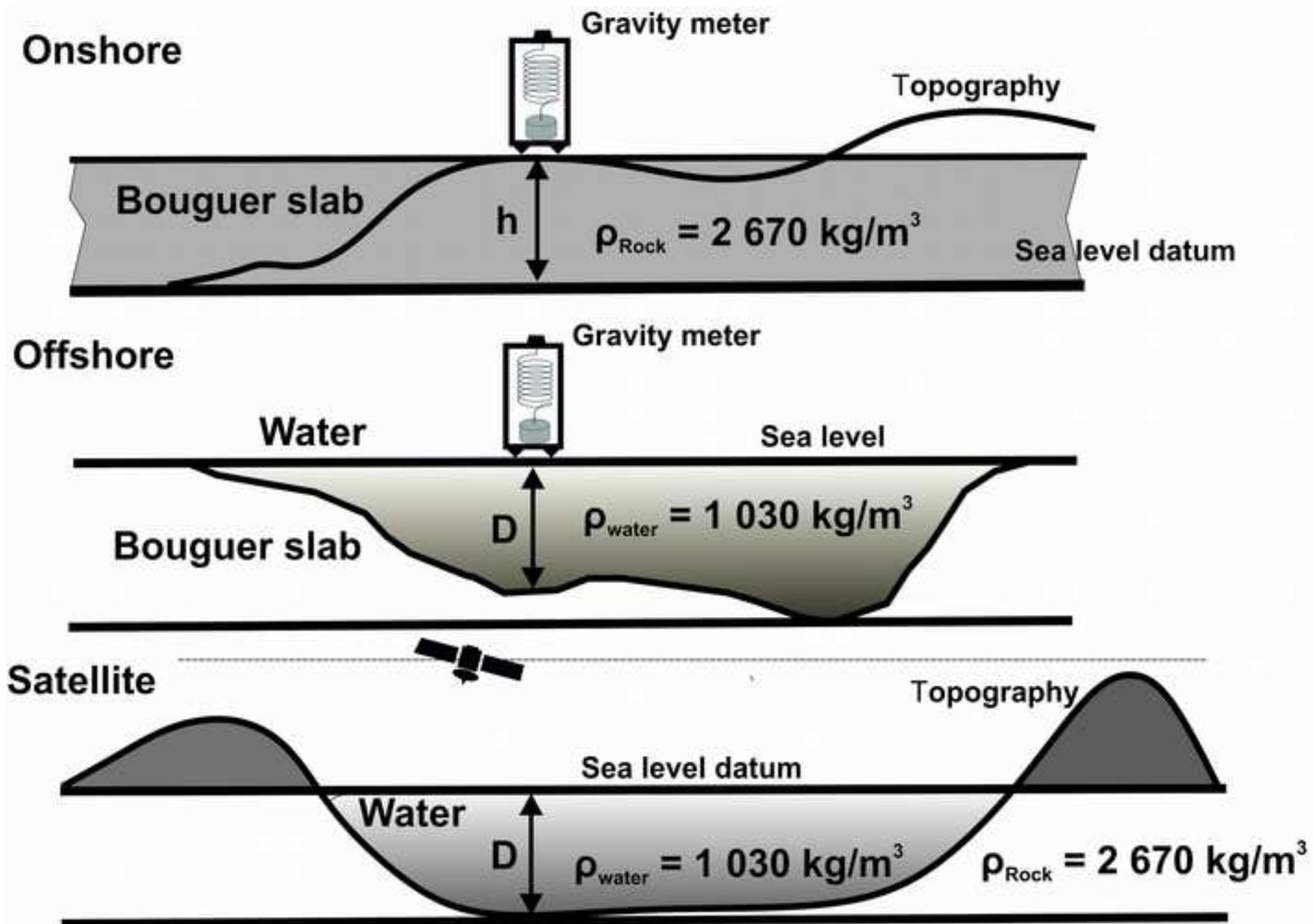
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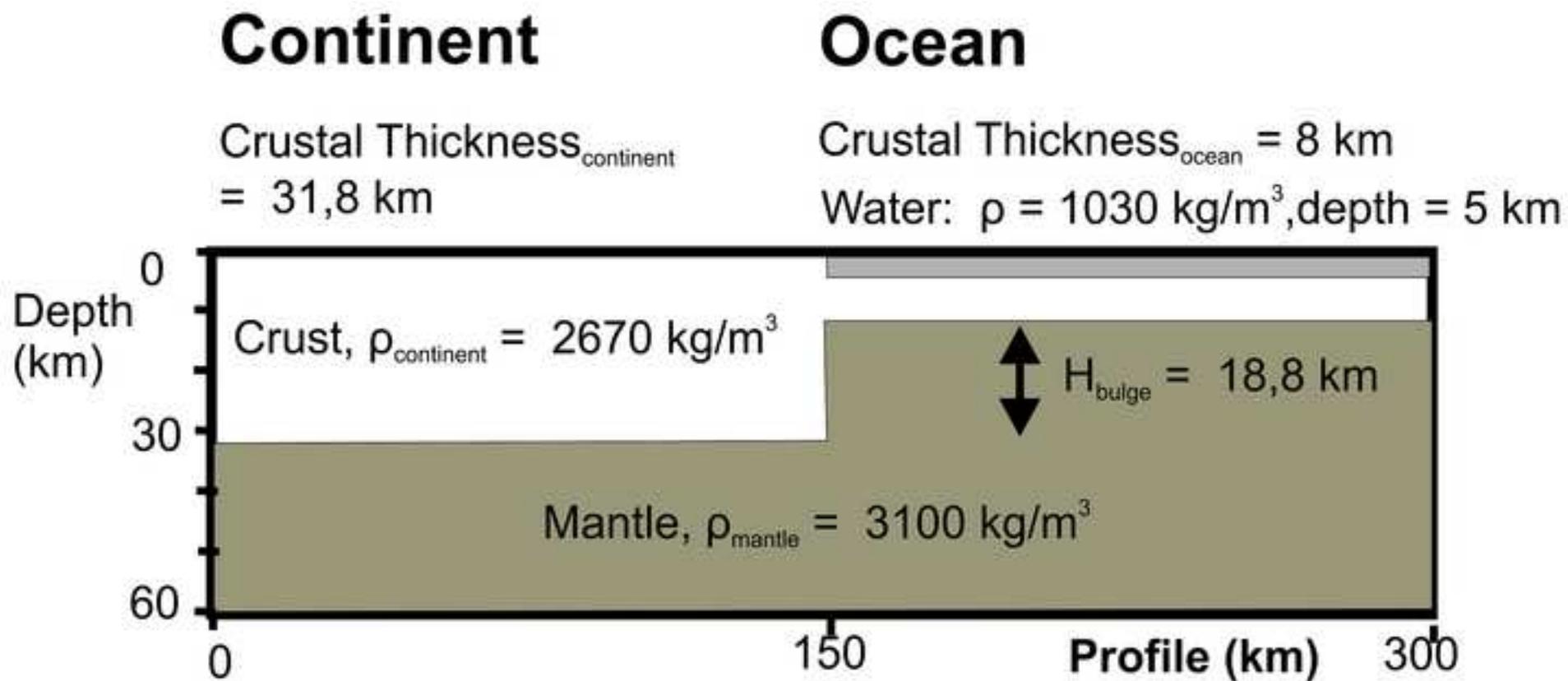
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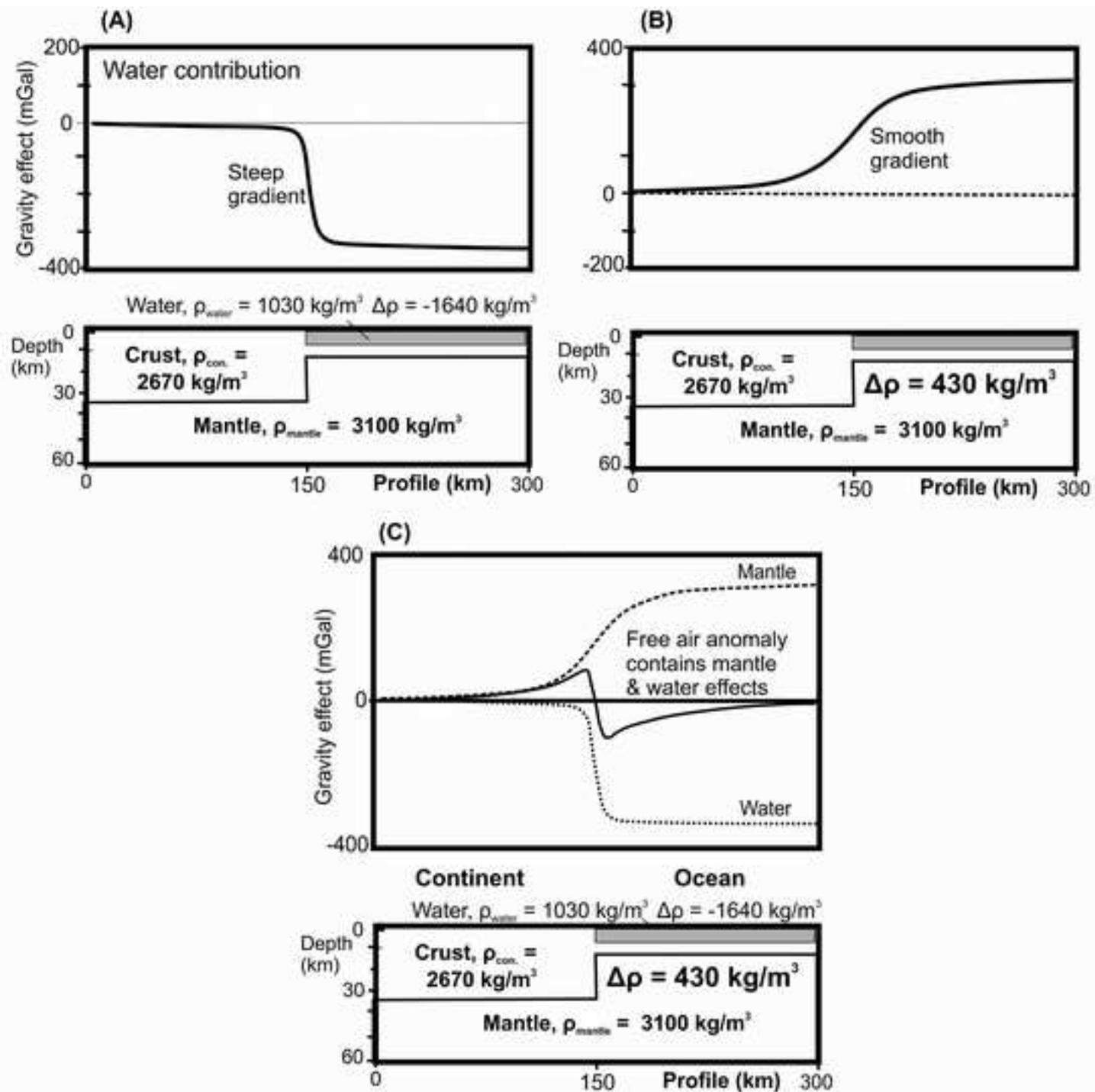
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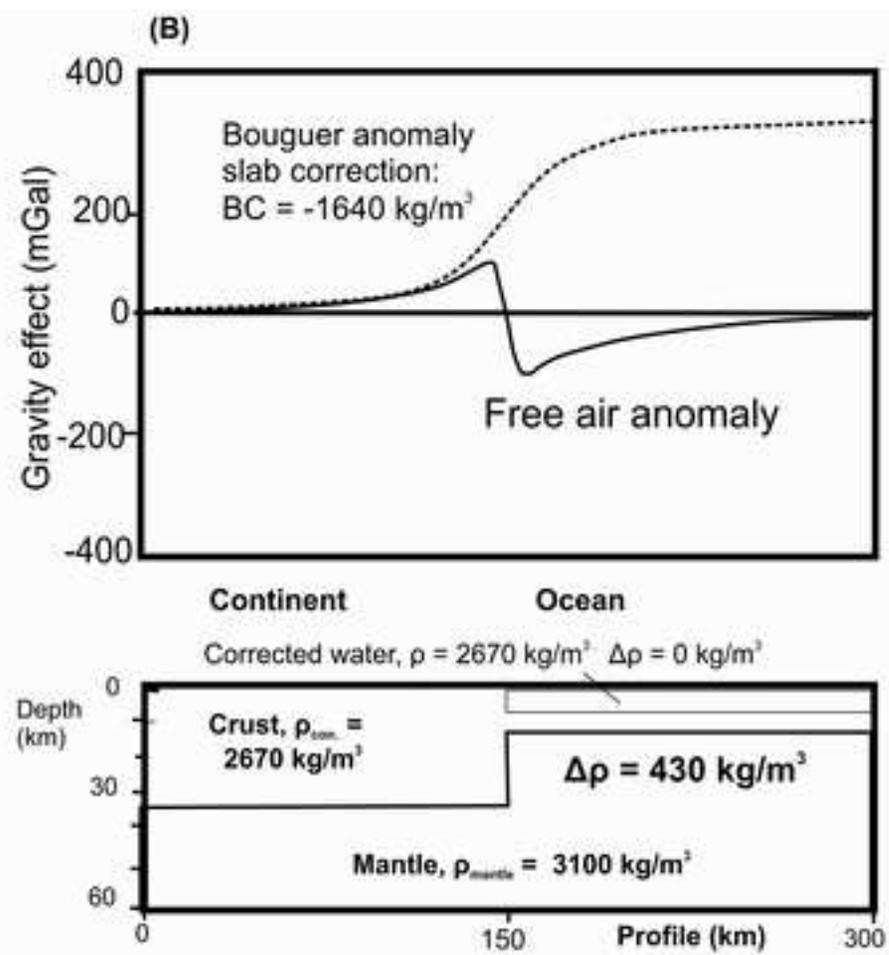
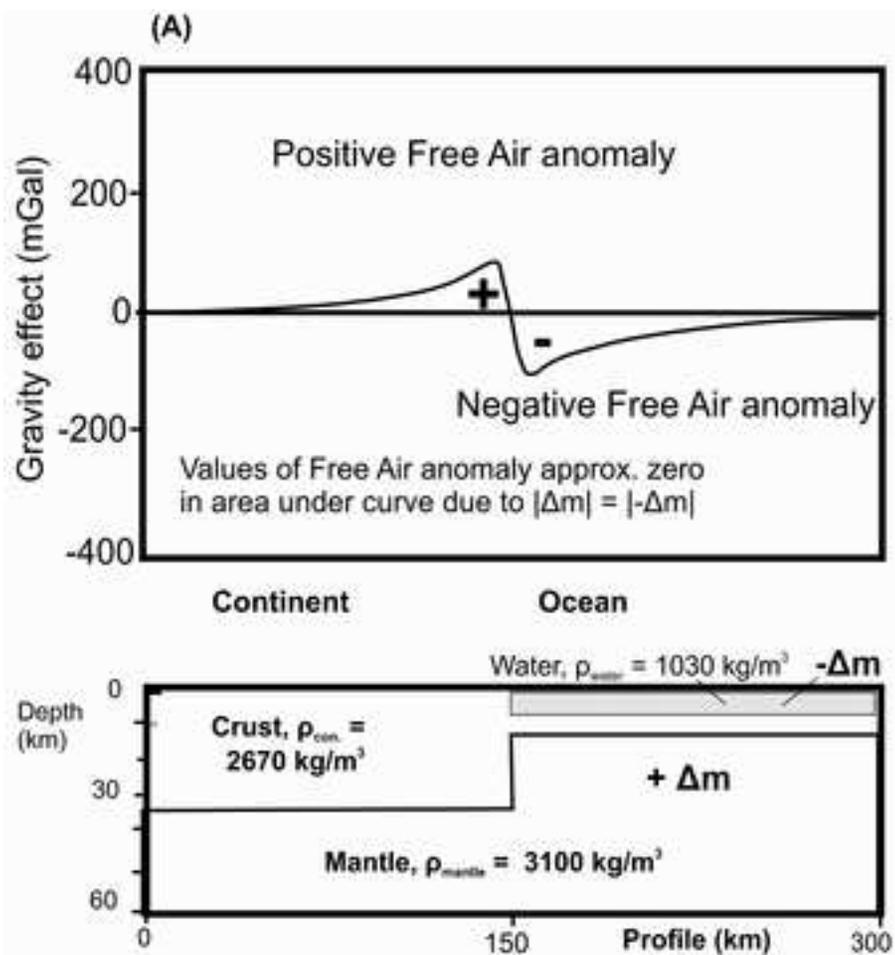
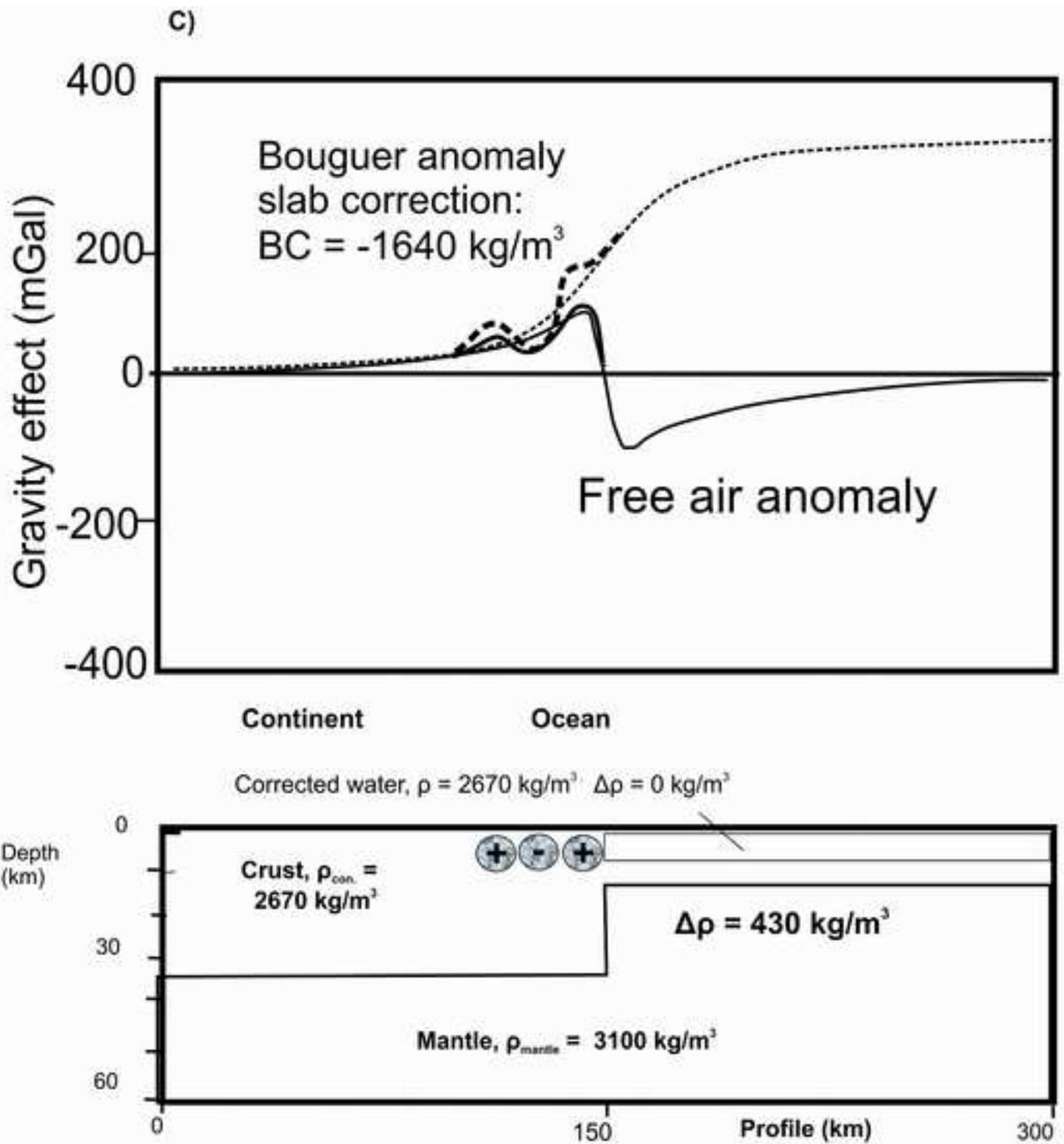
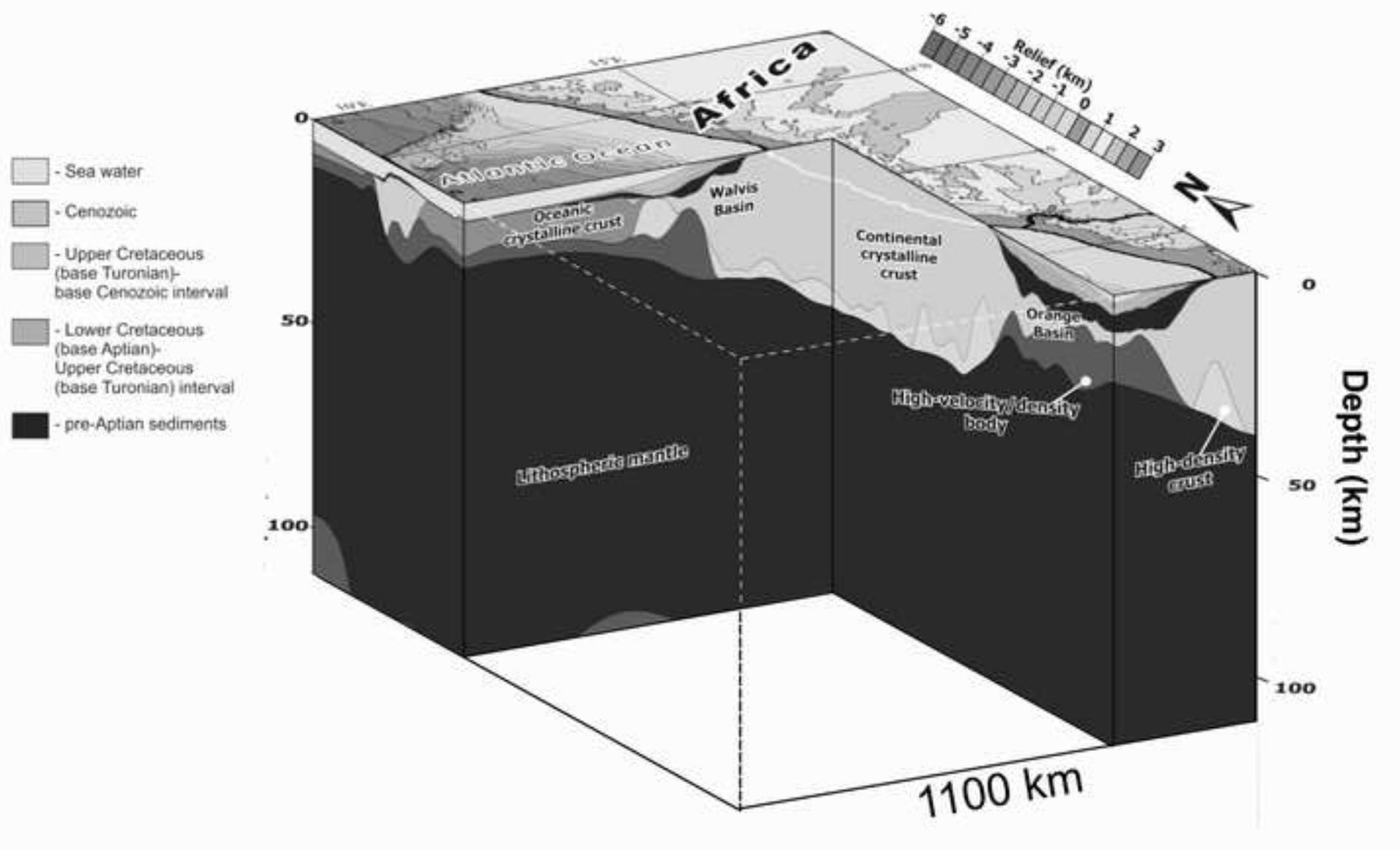


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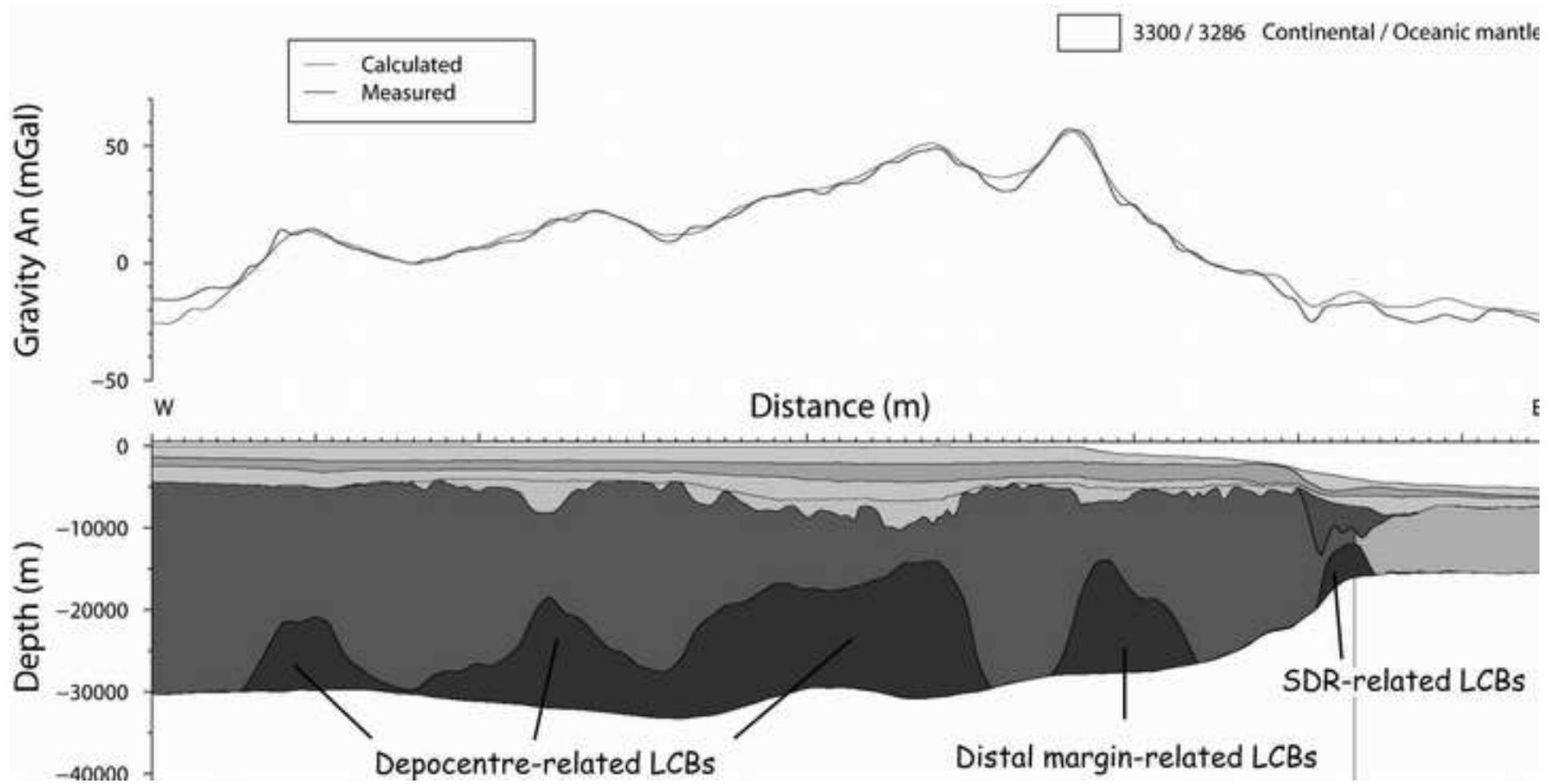
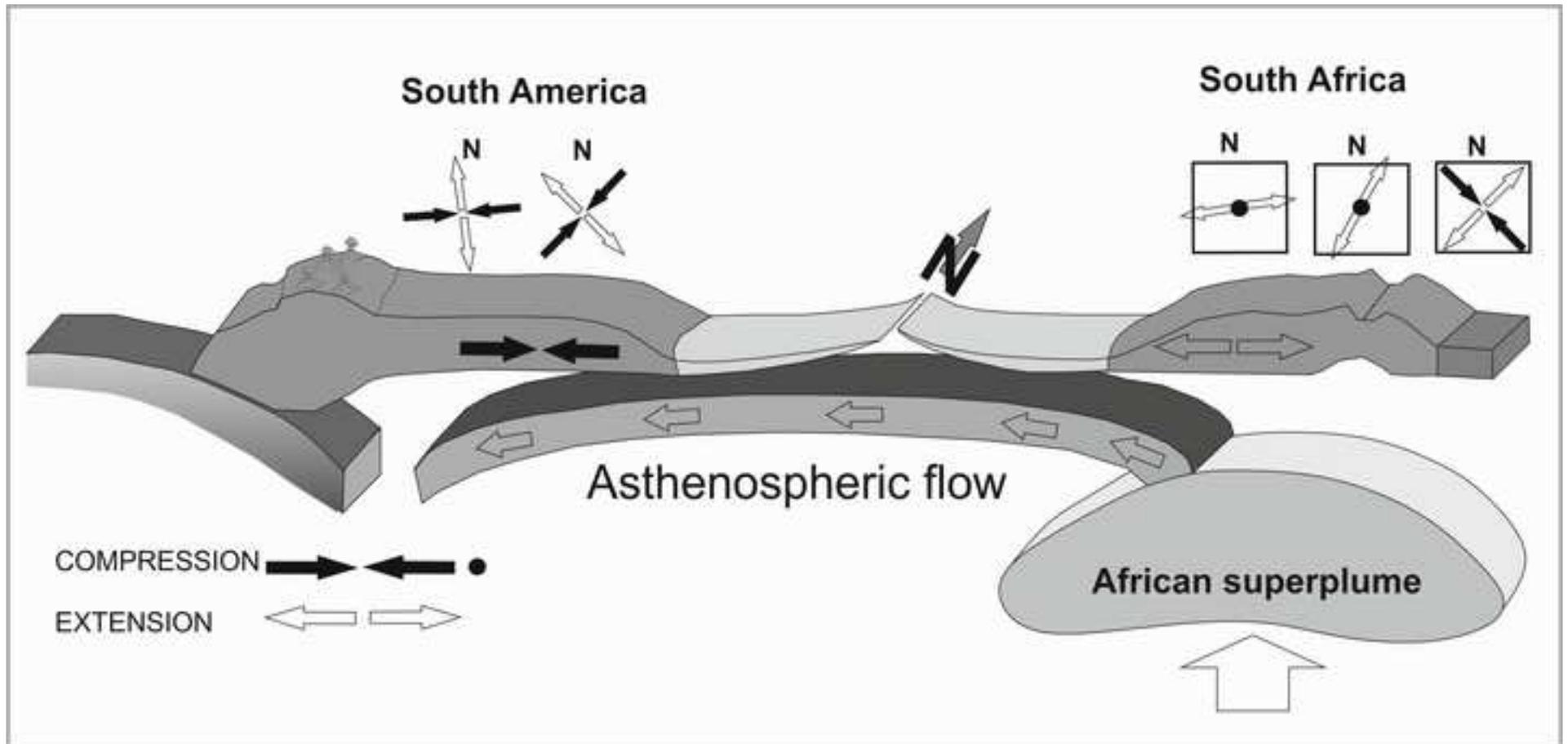


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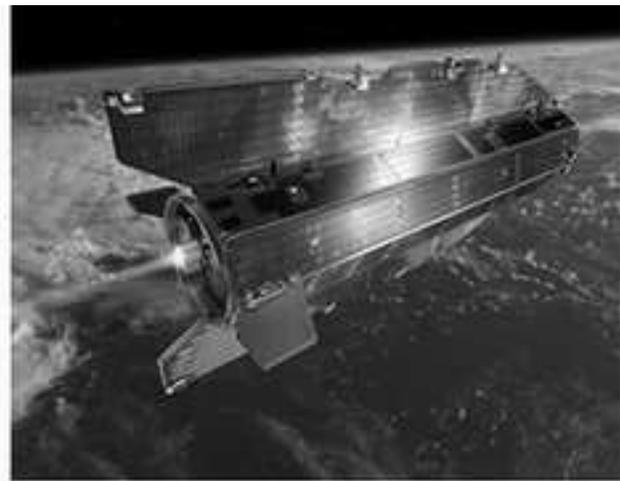
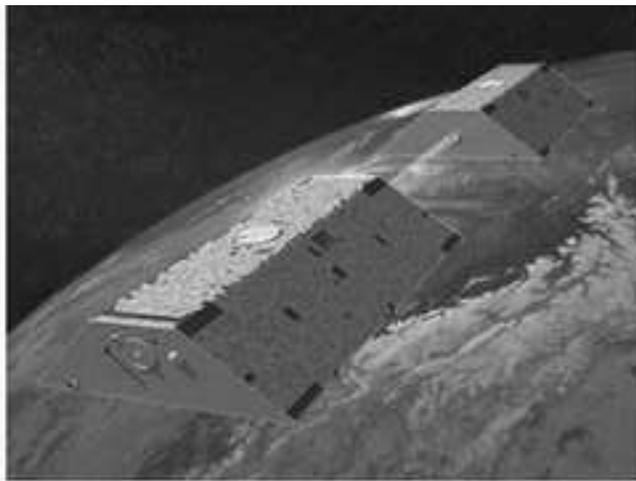


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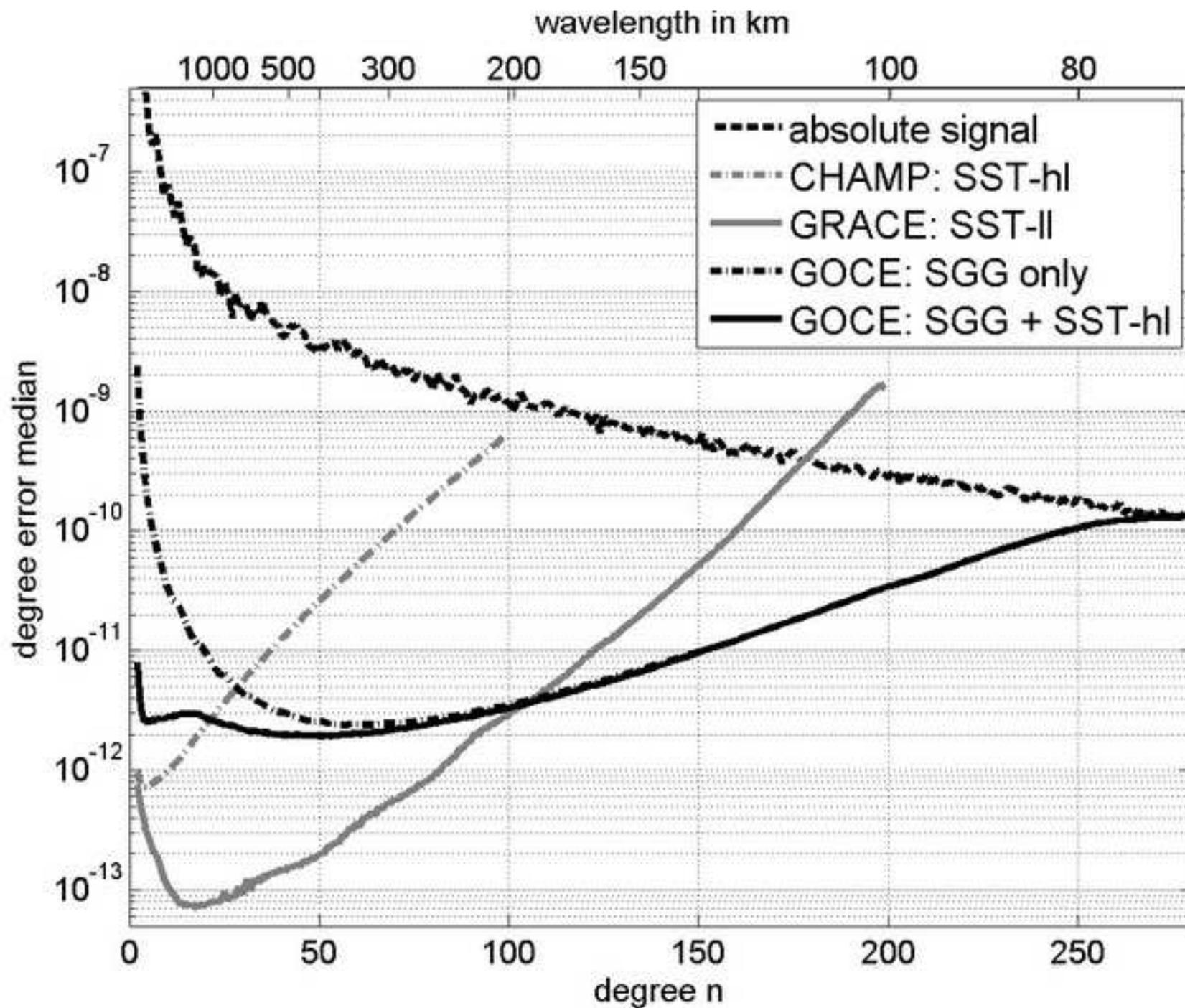


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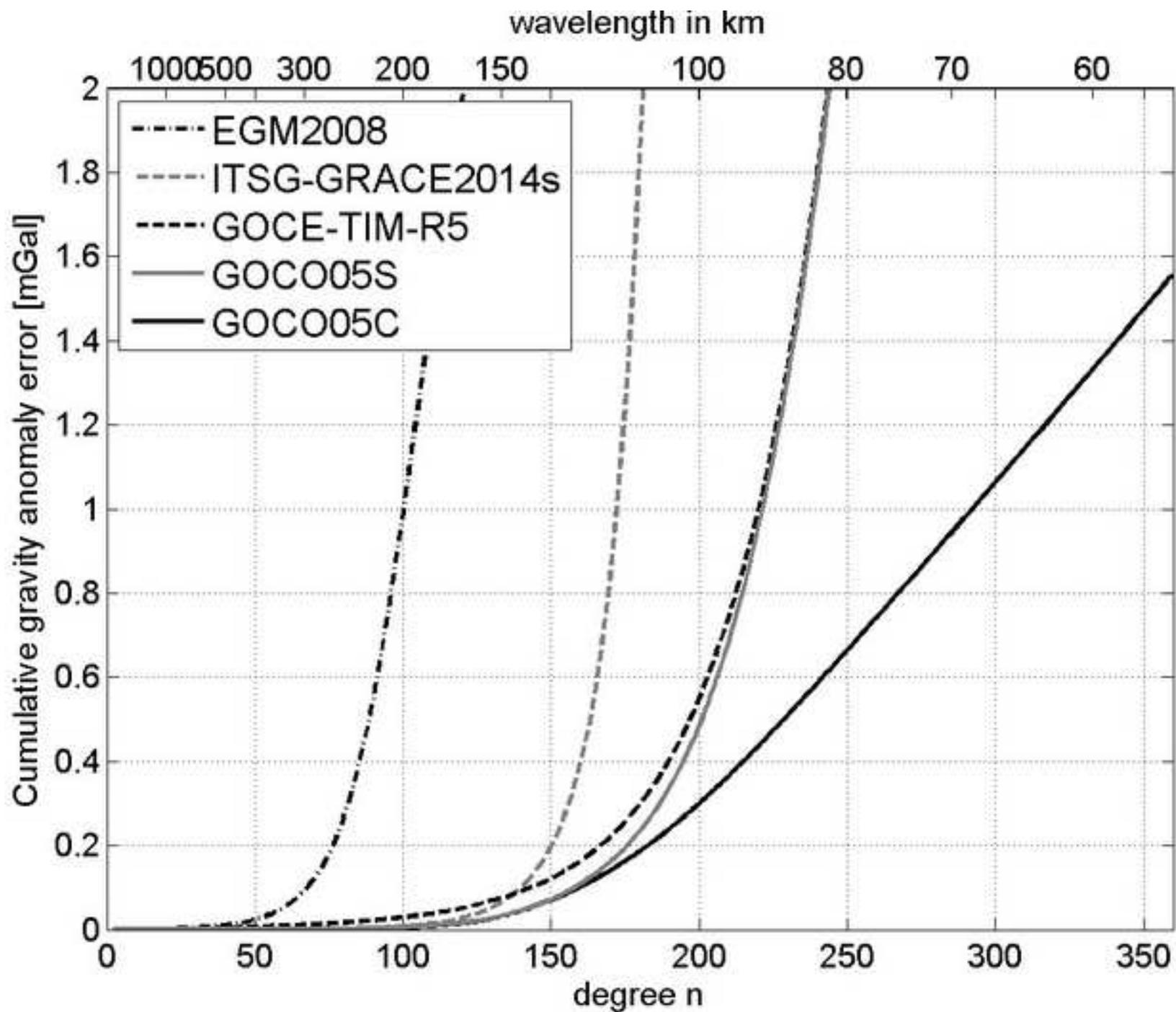
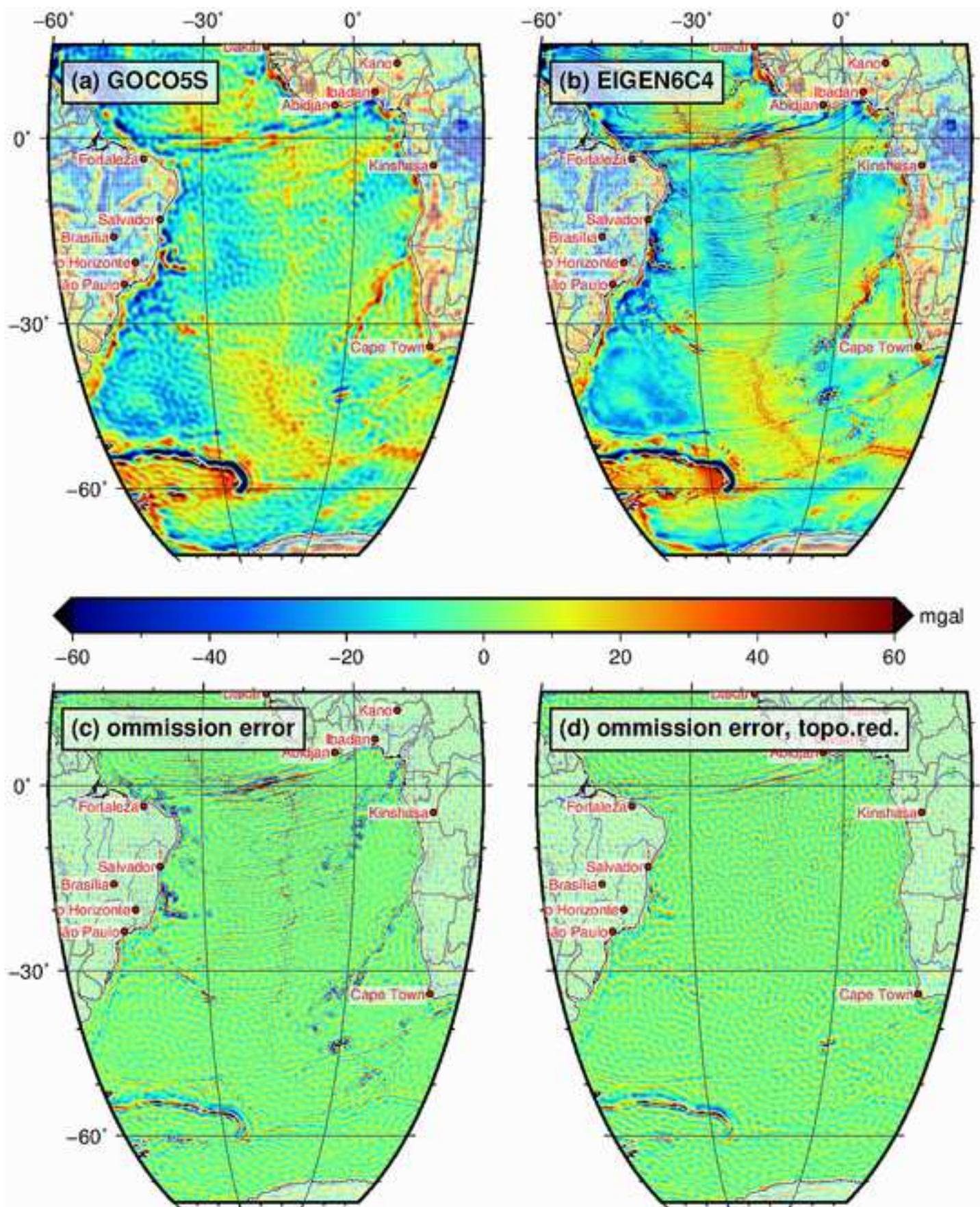
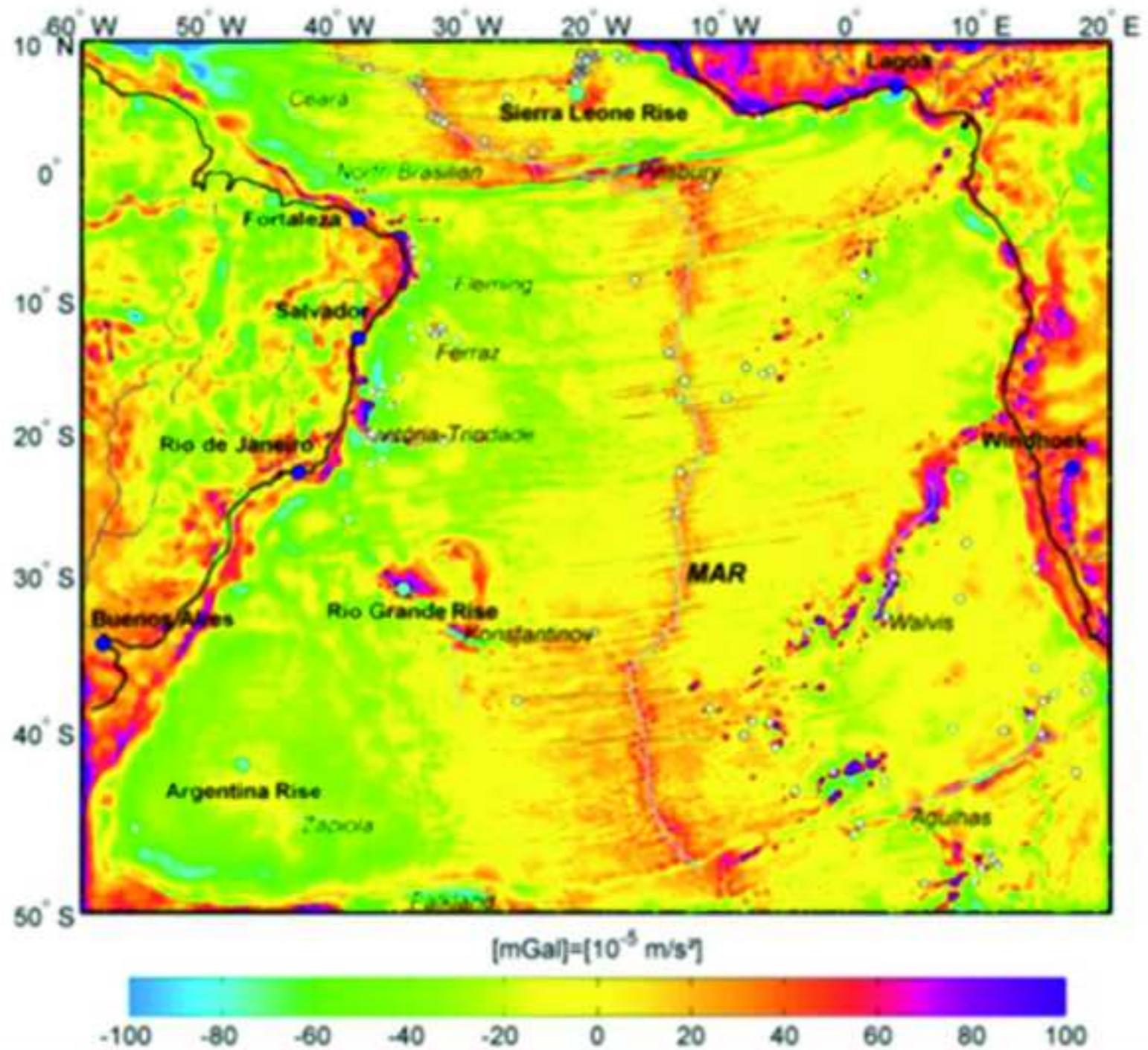


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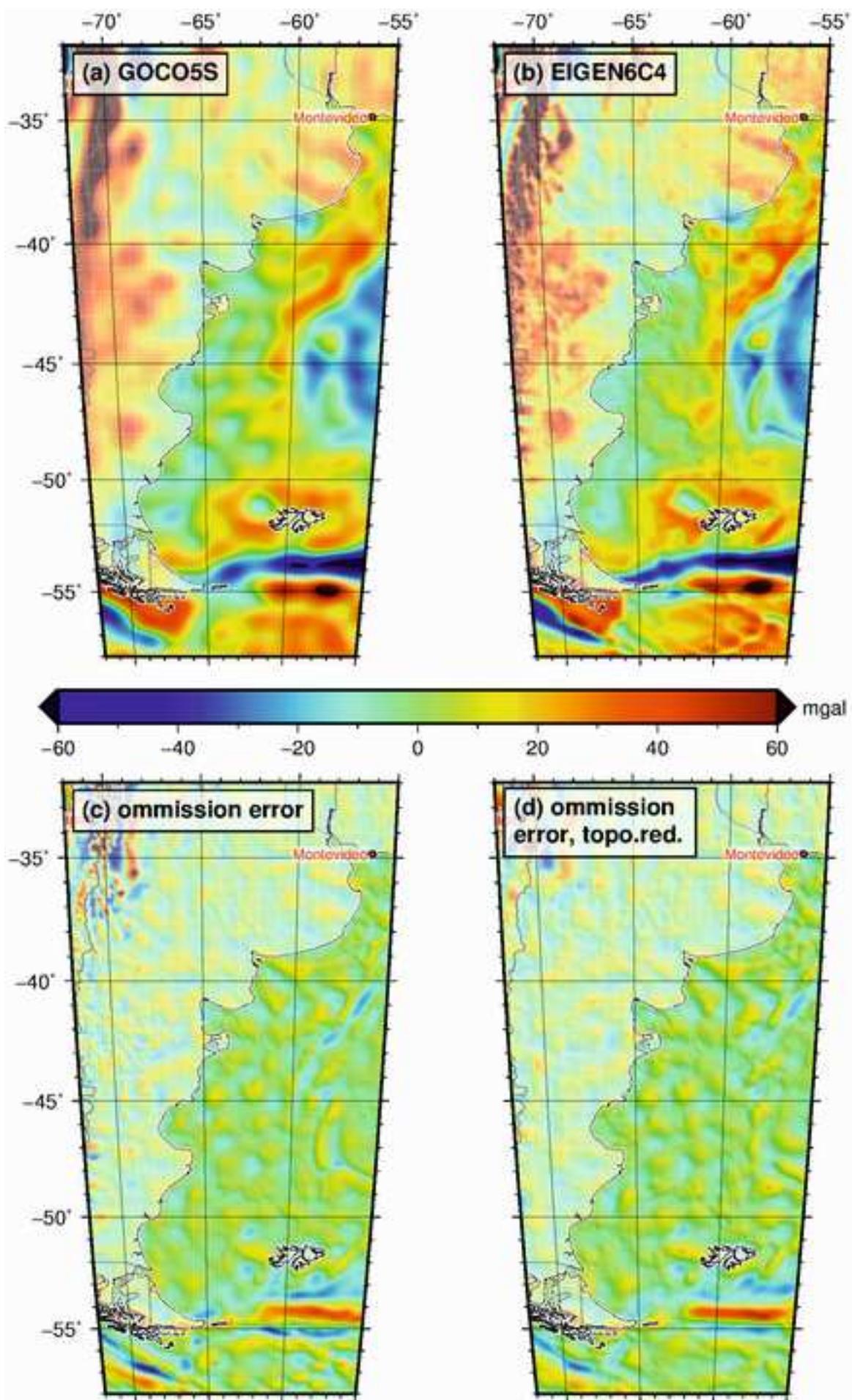


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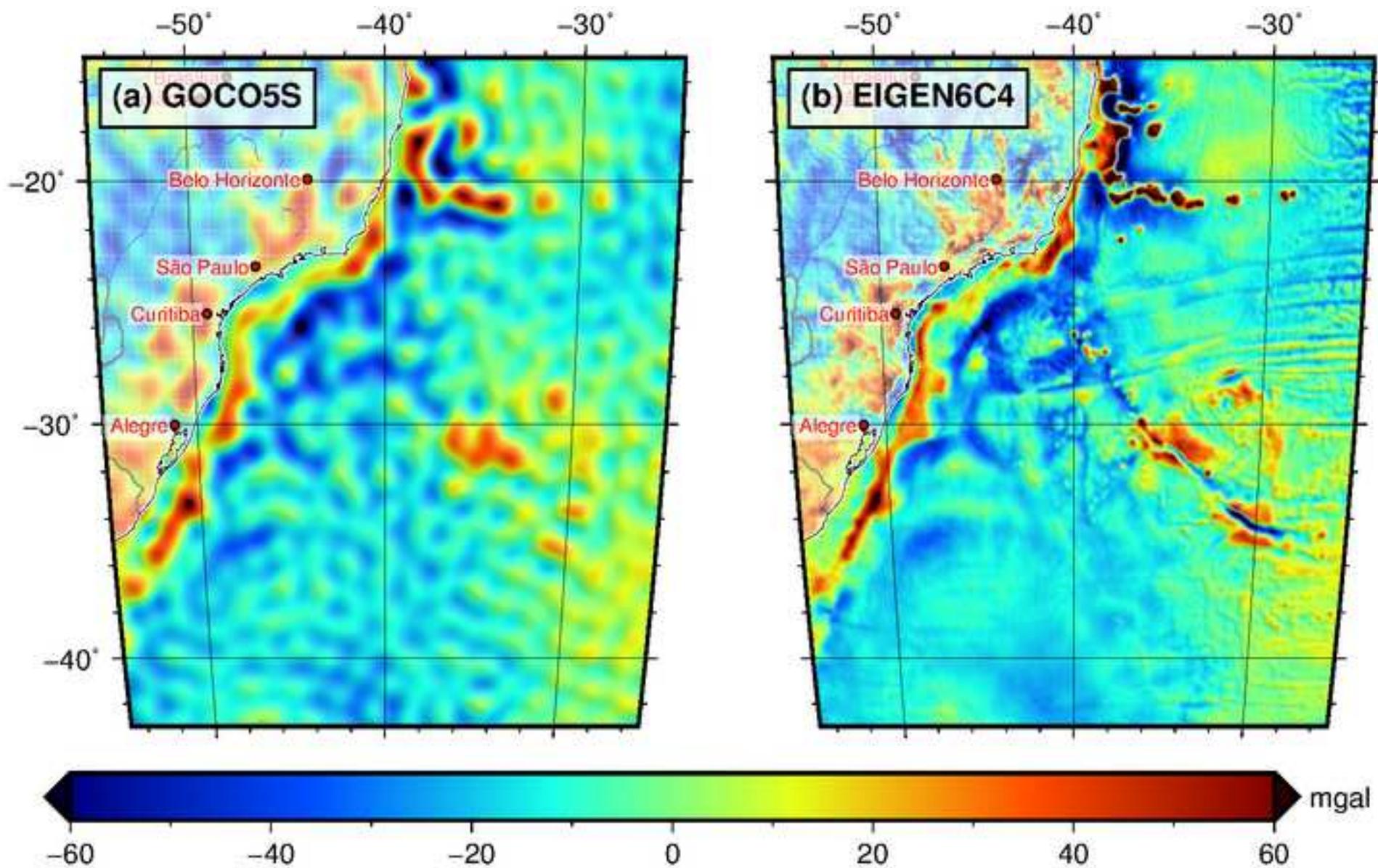


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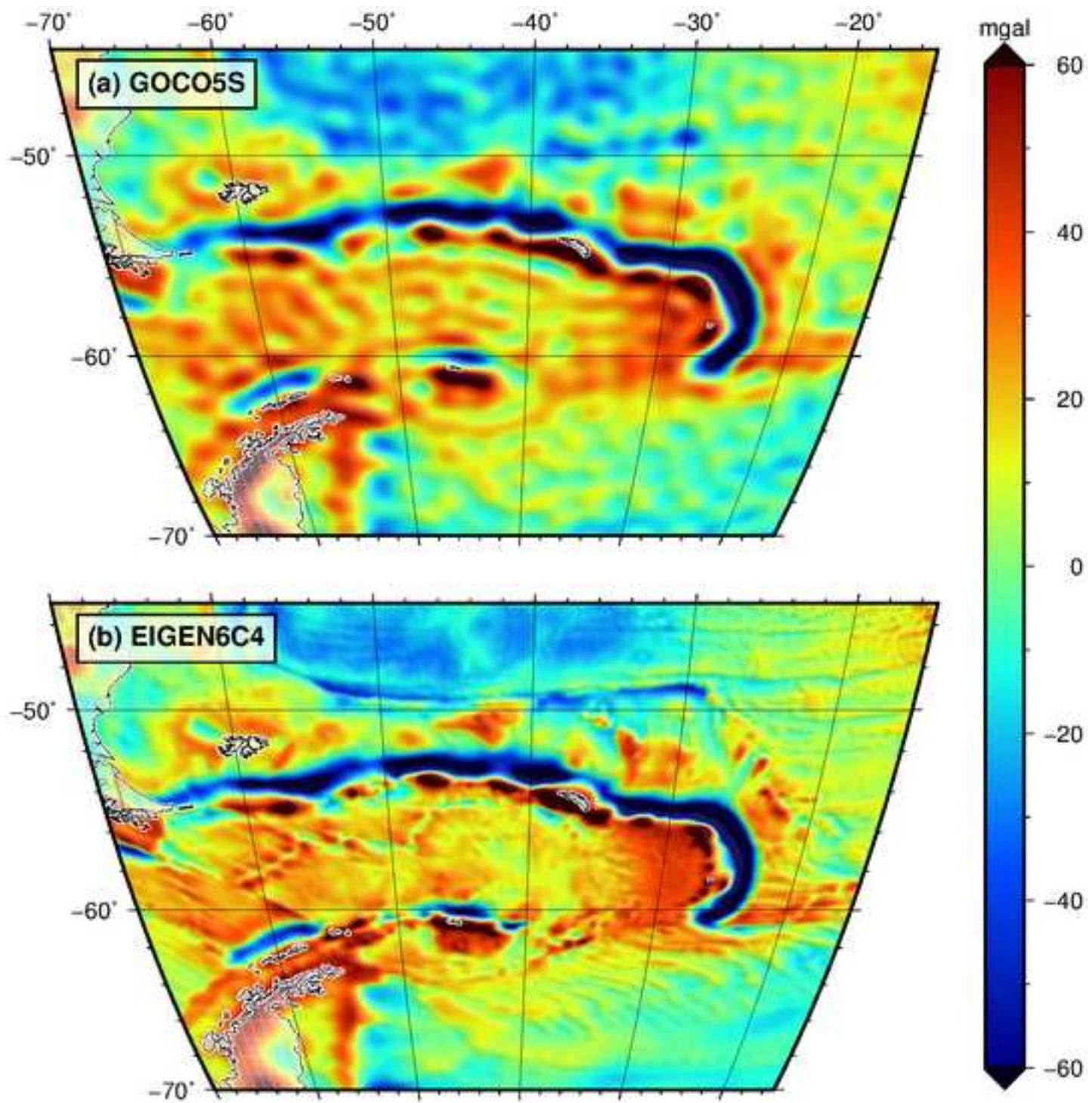


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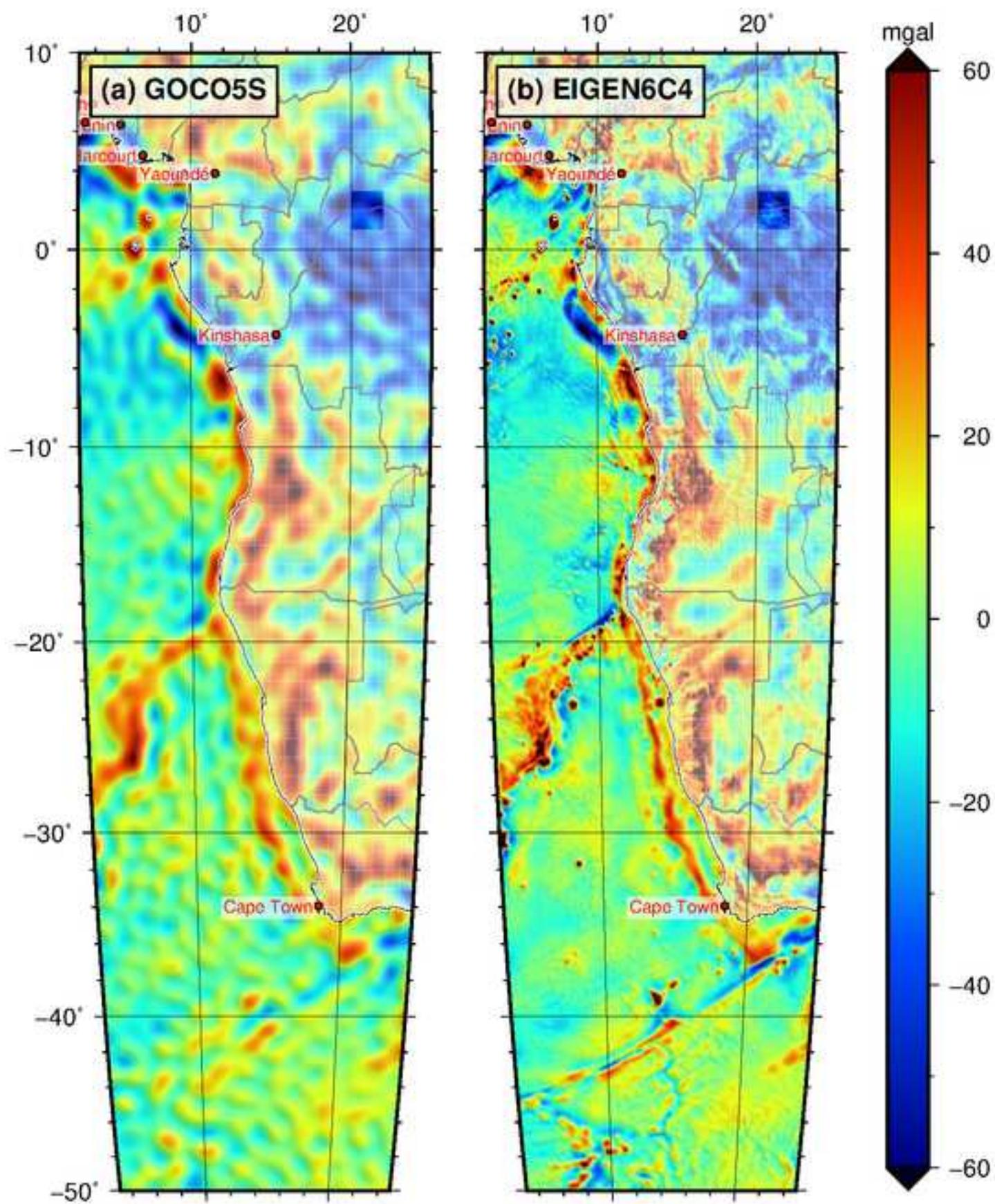


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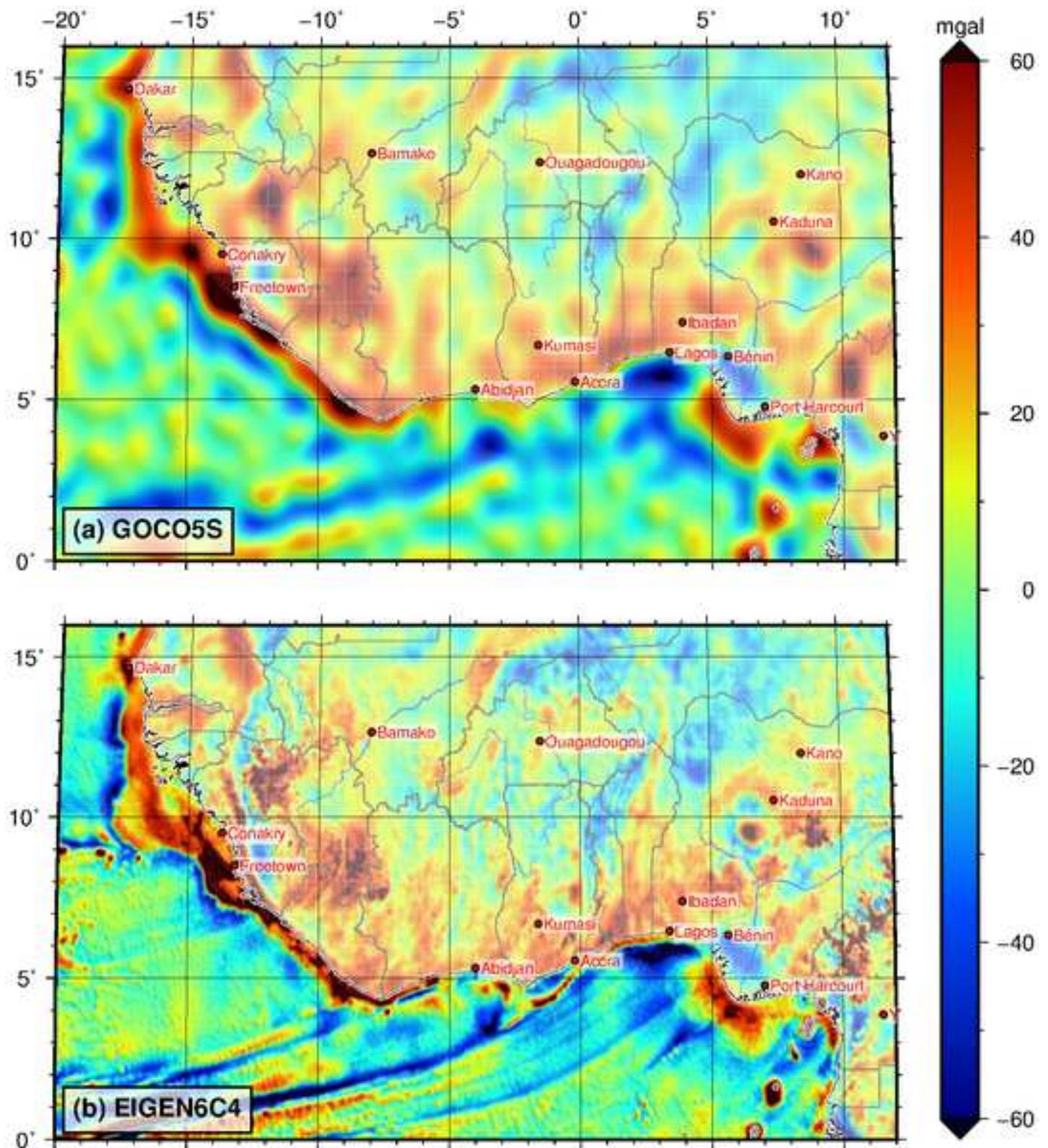
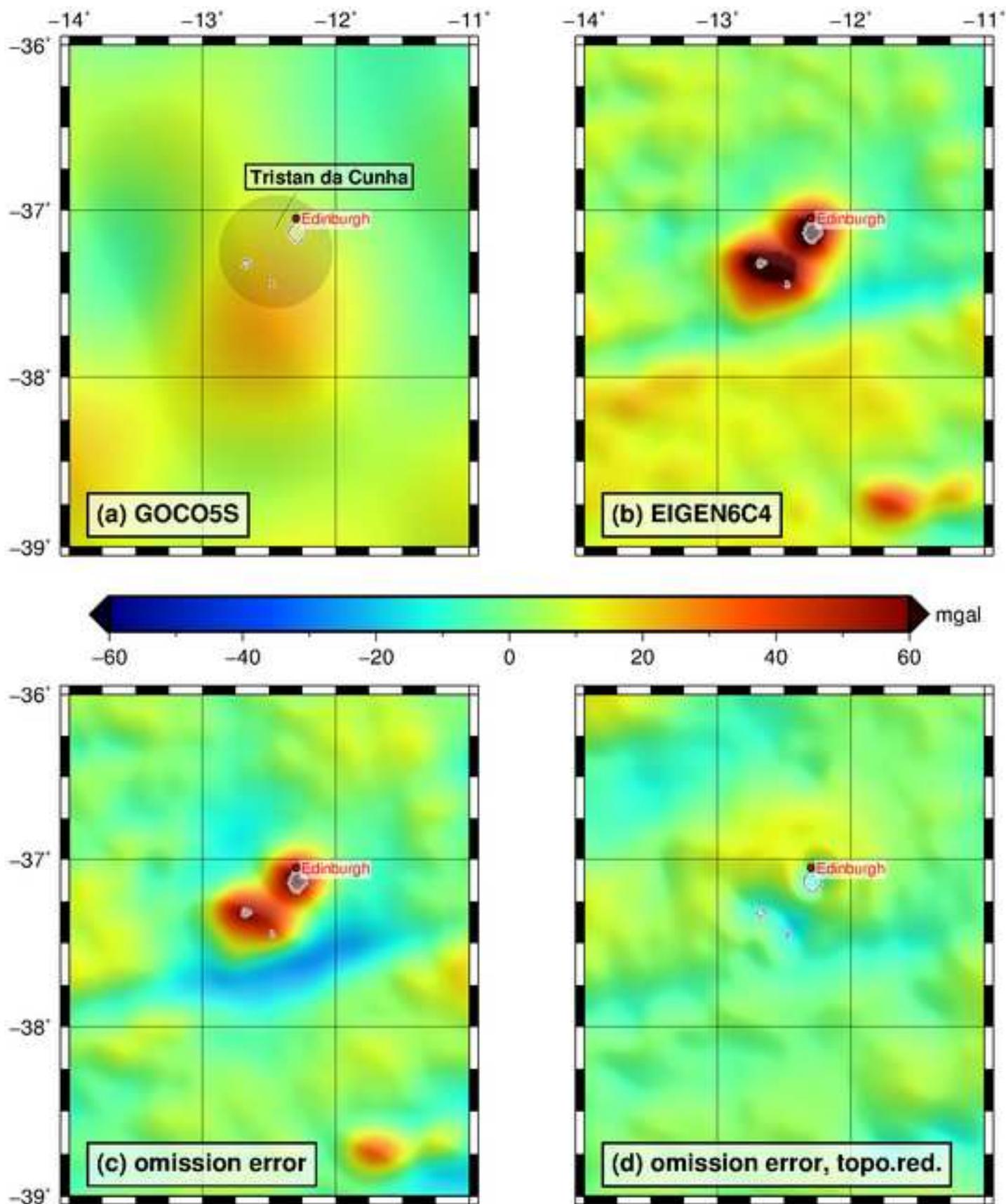
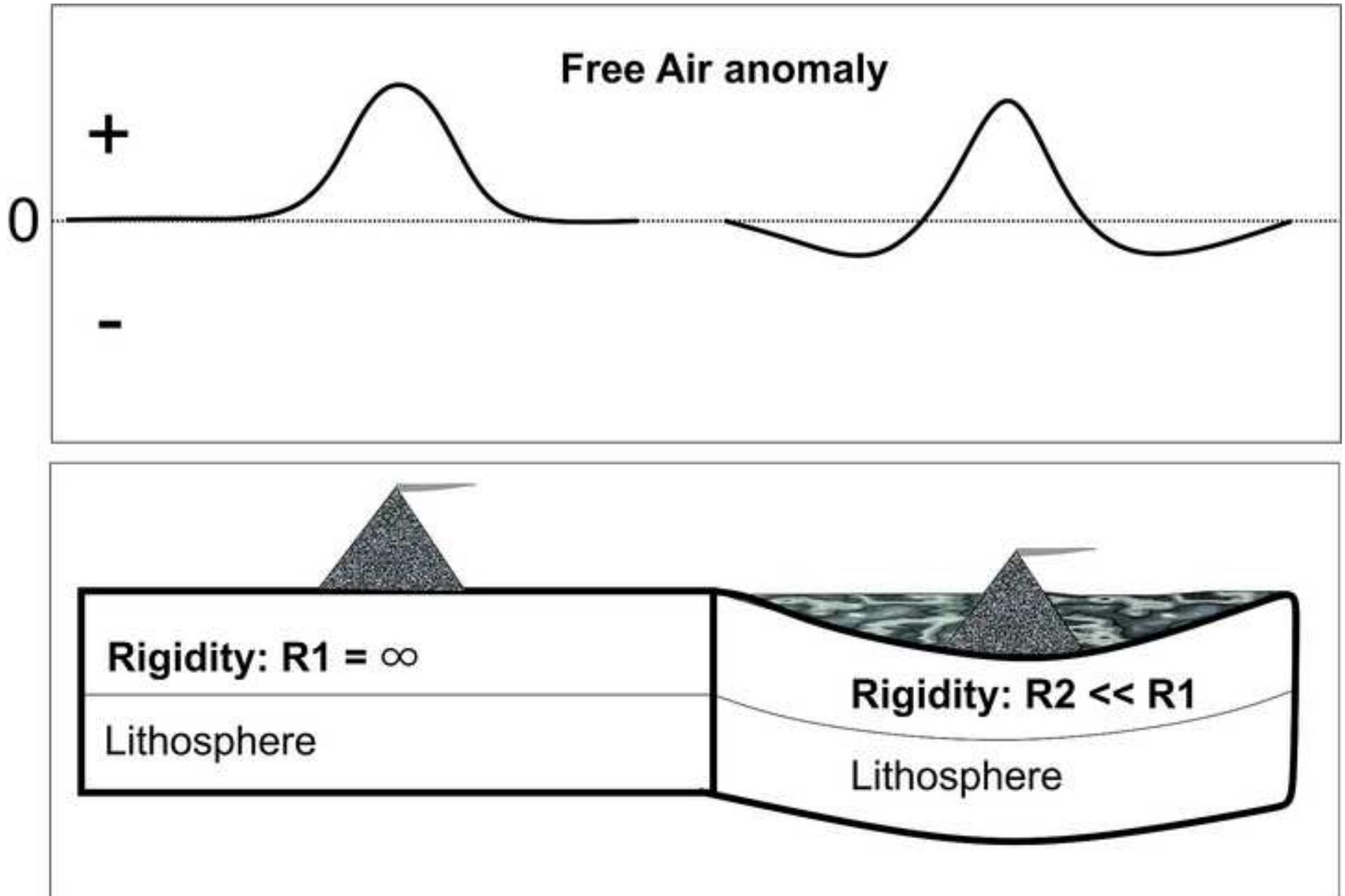


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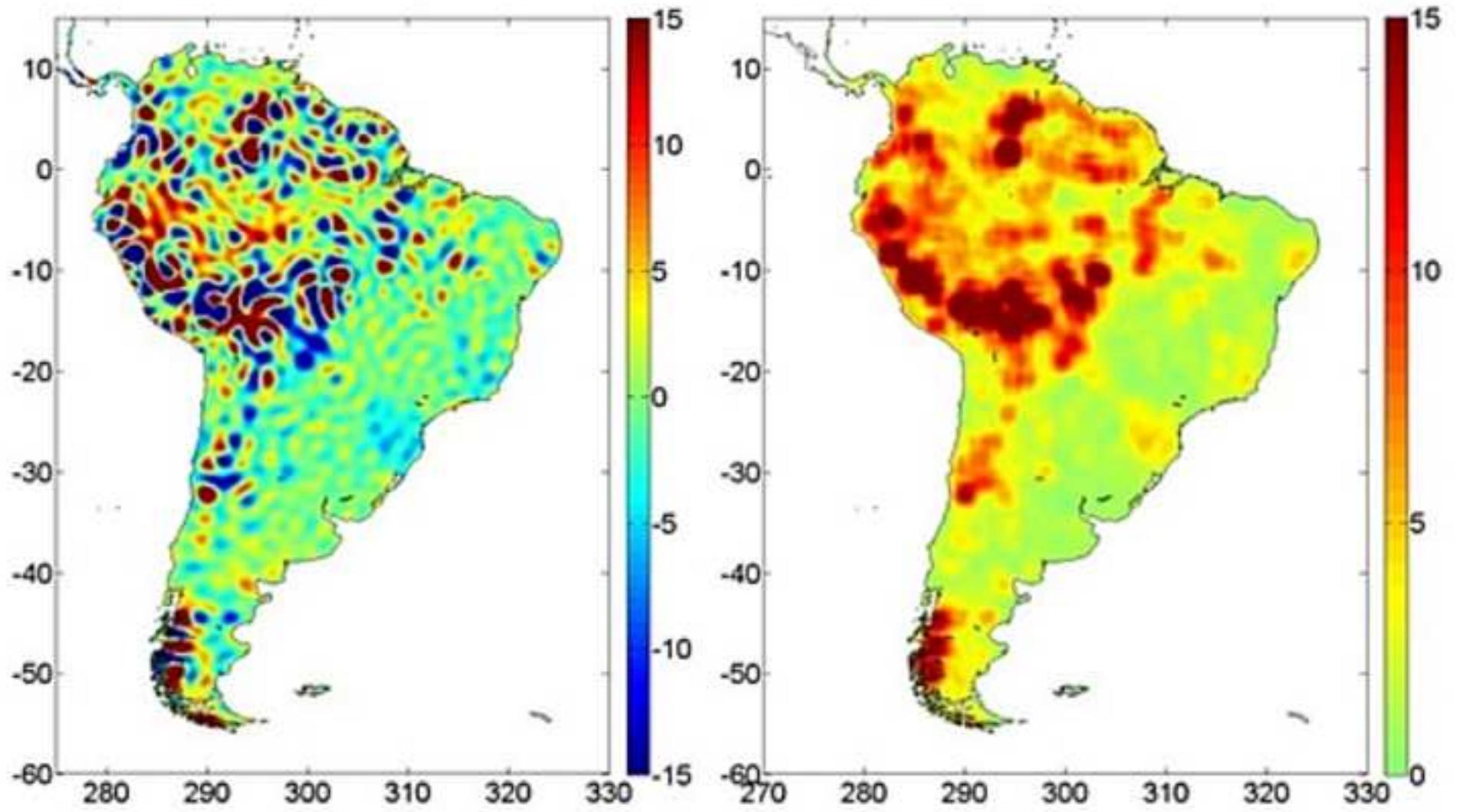
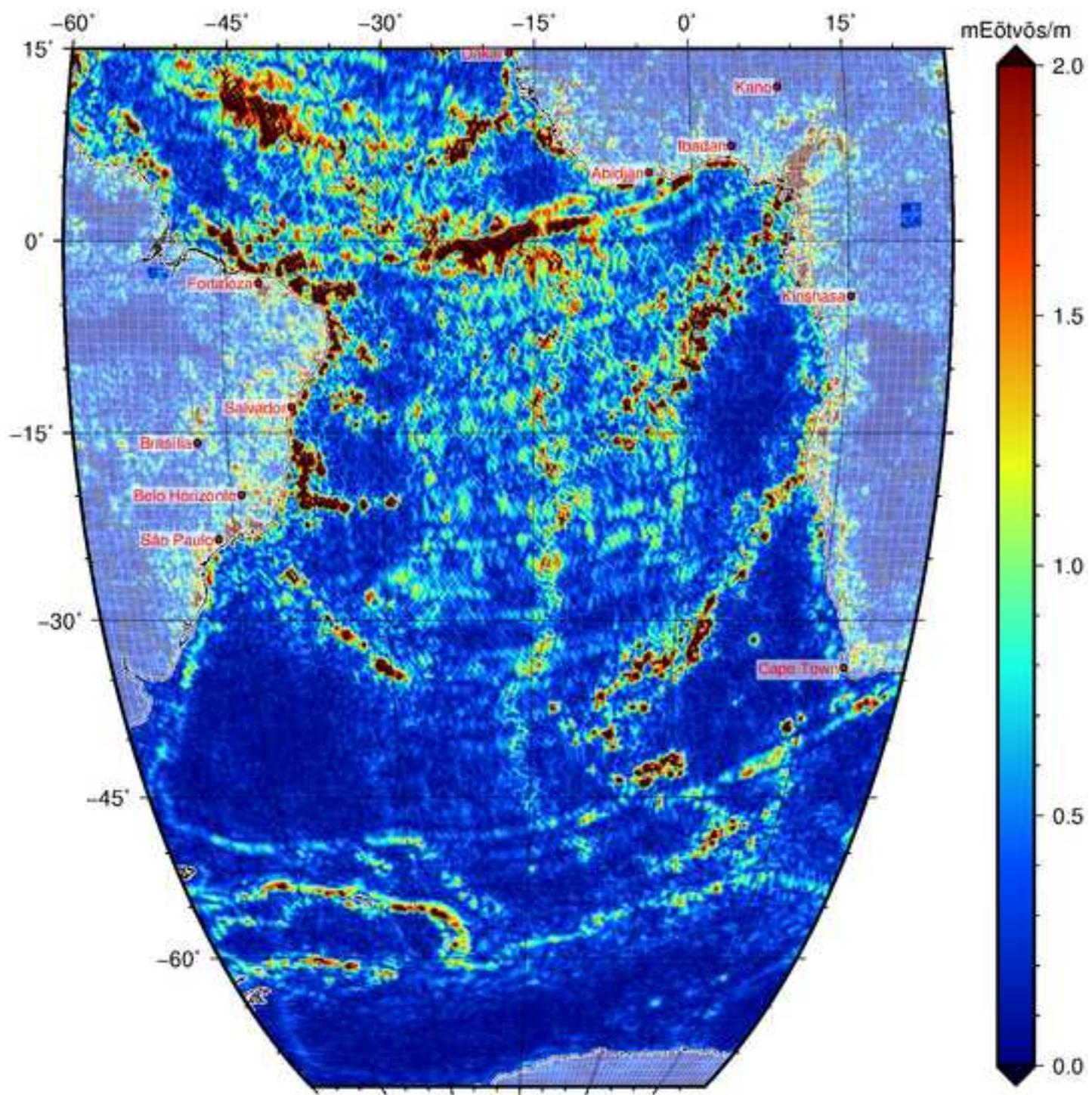


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**Response to Reviewers/Editors**

We corrected Line 324: „weak“ instaed of „week“.

We replaced in line 186 “Bocher et al.” and cited “Colli et al., 2016”

H.-J. Götze and R. Pail

1  
2 1 **Insights from recent gravity satellite missions in the density structure of**  
3 2 **continental margins – with focus on the passive margins of the South**  
4 3 **Atlantic**

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6 4  
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8 5 **Abstract**

9  
10 6 We focus on new gravity and gravity gradient data sets from modern satellite  
11 7 missions GOCE, GRACE and CHAMP, and their geophysical interpretation at  
12 8 passive continental margins of the South Atlantic. Both sides, South Africa  
13 9 and South America, have been targets of hydrocarbon exploration and  
14 10 academic research of the German Priority Program SAMPLE (South Atlantic  
15 11 Margin Processes and Links with onshore Evolution). The achievable spatial  
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19 15 evaluation of the omission error, which is not always in the focus of  
20 16 interpreters. It results from high-frequency signals of very rough topographic  
21 17 and bathymetric structures, which cannot be resolved by satellite gravimetry  
22 18 due to the exponential signal attenuation with altitude. The omission error is  
23 19 estimated from the difference of the combined gravity model EIGEN-6C4 and  
24 20 the satellite-only model GOCO05S. It can be significantly reduced by  
25 21 topographic reductions. Simple 2D density models and their related  
26 22 mathematical formulas provide insights in the magnitude of the gravity effect  
27 23 of masses that form a passive continental margin. They are contrasted with  
28 24 results from satellite-only and combined gravity models. Example geophysical  
29 25 interpretations are given for the western and eastern margin of the South  
30 26 Atlantic Ocean, where standard deviations vary from 25 – 16 mGal and 21 –  
31 27 11 mGal, respectively. It could be demonstrated, that modern satellite gravity

1  
2 28 data provide significant added value in the geophysical gravity data  
3  
4 29 processing domain and in the validation of heterogeneous terrestrial data  
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6 30 bases. Combined models derived from high-resolution terrestrial gravity and  
7  
8 31 homogeneous satellite data will lead to more detailed and better constrained  
9  
10 32 lithospheric density models, and hence will improve our knowledge about  
11  
12 33 structure, evolution and state of stress in the lithosphere.  
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34

### 35 **1. Motivation and the concept of Plate Tectonics**

36

37 Geosciences are striving for an interdisciplinary perception to combine  
38 their basic findings in a world embracing synthesis to understand global  
39 processes in the Earth interior and at its surface. Most of these processes are  
40 generally geothermally driven, and it is easy to accept that their origin lies  
41 below the lithosphere, in the Earth's mantle (among others Stadler et al.,  
42 2010). Today, the theory of plate tectonics enables us to draw a coherent  
43 picture of the Earth's lithosphere. Interactions between the plates at their plate  
44 boundaries are responsible for most of the earthquakes that occur here  
45 (among many other publications and websites:

46 <http://earthquake.usgs.gov/earthquakes/?source=sitenav>,

47 <http://www.isc.ac.uk/about/> or <http://geofon.gfz-potsdam.de/>).

48 This paper will review the status of satellite gravity missions and terrestrial  
49 data, as well as global gravity models, fields and gradients derived from them.  
50 It will focus on their accuracy, resolution and the omission error – which is out  
51 of focus of many earth scientists. It is structured as follows: Section 1 defines  
52 continent-ocean transitions and recalls some basics in the context of Plate  
53 Tectonics concept and passive margins in particular. For those readers who  
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Field Code Changed

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1  
2 54 are not familiar with the interpretation of gravity anomalies at a continental  
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4 55 margin in Section 2 the basic concepts are illustrated; this section can be  
5  
6 56 skipped by experts. Further on the focus is set on the question how (satellite)  
7  
8 57 gravity interpretations can help to explore these passive margins (Section  
9  
10 58 2.3). In the course of this paper we will notice later to what extent the new  
11  
12 59 fields and gradients from recent satellite gravity missions can support  
13  
14 60 research at the passive margins of the South Atlantic (Section 2.4).  
15  
16

17  
18 61 In Section 3 we will demonstrate how this new information augmented  
19  
20 62 our view on the density structures of the lithosphere particularly at passive  
21  
22 63 margins of the South Atlantic. At last we describe the benefits for combined  
23  
24 64 interpretations in Section 4, and merge information from both terrestrial and  
25  
26 65 satellite gravity fields.  
27  
28

29 66

### 30 31 32 67 *1.1 Short introduction to history of plate margins tectonics* 33 34 68

35  
36 69 Considering the history of the Plate Tectonic concepts in the early  
37  
38 70 1960s, Wegener's view on the „the continental drift“ (Wegener, 1920) began  
39  
40 71 to be accepted after it was refined and confirmed by geophysical observations  
41  
42 72 namely by early seismological studies on deep earthquakes (Wadati, 1929;  
43  
44 73 Benioff, 1954), later by Isacks et al. (1968), Oliver and Isacks (1967) and  
45  
46 74 paleo-magnetic shipborne observations (Hess, 1962; Vine and Matthews 1963  
47  
48 75 among others). Together with the techniques of radiometric dating (age  
49  
50 76 determination) published first by Boltwood (1907) geophysicists were able to  
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1  
2 77 date the magnetic mid oceanic reversals by precise physical measurements.

3  
4 78 They helped to get the modern concept of “plate tectonics” fully accepted.

5  
6 79 It provides the framework for the interpretation of structures, the history  
7  
8 80 and composition of continental margins. Plate movements and the differences  
9  
10 81 in density of oceanic and continental crust types led to the structural pattern of  
11  
12 82 continental margins and result in a tectonic classification of coastlines as  
13  
14 83 **active** or **passive** margins (among others Wefer et al., 2003). *Active margins*  
15  
16 84 are typical units of the “Circum-Pacific Ring of Fire” in the Pacific where plates  
17  
18 85 are converging and coincide with plate boundaries in a subduction zone.

19  
20 86 These margins are called *active* (e.g. Oncken et al., 2006; Lallemand, 2014)  
21  
22 87 due to the big variety of tectonic, magmatic and metamorphic processes that  
23  
24 88 occur here. If continental margins mark only the boundary to the oceanic  
25  
26 89 portion of the same tectonic plate, they are called *passive margins*. Passive  
27  
28 90 margins are typical of the Atlantic Ocean between Scandinavia and Greenland  
29  
30 91 (Scheck-Wenderoth and Maystrenko 2008), Iberic peninsula and the East  
31  
32 92 American coast or between Africa and South America (Blaich et al., 2011 and  
33  
34 93 many papers within).

35  
36  
37 94

#### 38 39 95 *1.2 Passive Margins*

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41 96

42  
43 97 Although we assume that most of the readers are familiar with the  
44  
45 98 concept of plate tectonics we will recall briefly some basics. In particular  
46  
47 99 *passive continental margins* are characterized by a transition of continental  
48  
49 100 into oceanic crust within the same plate. It emerges from the splitting up of  
50  
51 101 continents and the following divergent plate drift that forms new oceanic  
52  
53 102 lithosphere by seafloor spreading at the divergent plate boundary. At the  
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2 103 edges of the Atlantic geophysical investigations identify a laterally 10-15 km  
3  
4 104 thick crustal transition between the thick crust of the continents and the thin  
5  
6 105 oceanic crust. It is interpreted as relicts of thinned, mafic magmas which  
7  
8 106 intruded into continental crust. In addition, passive continental margins are  
9  
10 107 often stretched by intensive fault tectonics. They have a tectonically thinned  
11  
12 108 continental crust, which is characterized by listric faults and tilted fault blocks.  
13  
14 109 Tectonic crustal expansion starts at the time of installation of the continental  
15  
16 110 drift rift system and refines the passive continental margin further (Blaich et  
17  
18 111 al., 2011). These margins are marked by smooth relief due to tectonic  
19  
20 112 inactivity and major sediment accumulation. This phenomenon is due to  
21  
22 113 thermal cooling and sediment loading that led to conditions of subsidence and  
23  
24 114 sediment accumulation, because the margins move away from the spreading  
25  
26 115 center. Irregular subsidence and different sediment load often cause the  
27  
28 116 accumulation of salt diapirs in the sediments of passive continental margins.  
29  
30 117 The tectonic-sedimentary conditions are also favorable for the formation of  
31  
32 118 hydrocarbon deposits and large salt deposits (Mohriak, 2014).  
33  
34  
35  
36

37 120 Modern passive margins (Fig. 1) border the oceans formed by the  
38  
39 121 spreading caused by the break-up of the Gondwana supercontinent (e.g.  
40  
41 122 Bond et al., 1984). From Figure 1 one can see that the margins of the South  
42  
43 123 Atlantic Ocean (Cappelletti et al., 2011; Blaich et al., 2011; Mohriak, 2014),  
44  
45 124 the western Indian Ocean, the Arctic and Norwegian seas (Scheck-Wenderoth  
46  
47 125 and Maystrenko, 2008; Ebbing et al, 2006; Skogseid et al., 2000), the magma  
48  
49 126 poor rifted margins of the North and Central Atlantic Ocean (Reston, 2009;  
50  
51 127 Mohriak, 2014) and the margins of Antarctica are part of this system (Kalberg,  
52  
53 128 2016). A rising convection cell or a plume in the rifting area caused initial  
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1  
2 129 rifting and a regional uplift as well as volcanic activities above or below  
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4 130 oceanic sea level after an initial period of crustal thinning and erosion. Basile  
5  
6 131 (2015) describe another type of margin: “transform continental margins” by  
7  
8 132 simple kinematic models of transform faulting which cause among two other  
9  
10 133 types “passive transform margins”. The satellite gravity picture of the  
11  
12 134 Equatorial Atlantic Ocean will be shown in Section 4.2, Fig. 20.

13  
14 135

15  
16 136 Figure 1: full page width

17  
18 137

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20  
21 138 **Fig. 1.** Continental margins on Earth. Blue lines mark passive continental  
22  
23 139 margins mainly surrounding the Atlantic Ocean, the Antarctic Seas, and Indian  
24  
25 140 Ocean; red lines indicate active margins (subduction zones). Continental  
26  
27 141 margins were taken from from Frisch und Meschede (2005).The underlying  
28  
29 142 gravity field is the map of “gravity disturbance” calculated on base of the  
30  
31 143 EIGEN-6C4 model (Förste et al., 2014). Gravity was calculated on a grid of  
32  
33 144 0.5° x 0.5°.

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39 146 *1.3 Volcanic passive margins*

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45 148 These margins present distinctive genetic and structural features, e.g.,  
46  
47 149 high-rate extension of the lithosphere is associated with catastrophic mantle  
48  
49 150 melting responsible for the accretion of a thick igneous crust (Geoffroy, 2005).  
50  
51 151 Typical rifted “magma-dominated” margins are characterized by large volumes  
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53 152 of flood basalts which flow across the continental hinterlands during  
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1  
2 153 continental breakup (among others refer to: Hopper et al., 2004; Eldholm et  
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4 154 al., 2000; Gernigon et al., 2006; White and Smith, 2009; O'Connor et al.,  
5  
6 155 2012). Underlying the extrusive lavas at the continent-ocean transition zone,  
7  
8 156 these margins exhibit high seismic velocities in the lower crust of some The  
9  
10 157 deeper crust is characterized by two areas of high seismic velocity (7.2 to 7.4  
11  
12 158 km/s; Franke et al., 2010)., which are associated with voluminous igneous  
13  
14 159 rocks intruded into the lower crust.

16  
17 160 In recent time the question arose in the context of fixing international  
18  
19 161 boundaries offshore of the continents due to economic interests because e.g.  
20  
21 162 hydrocarbon exploration moved further offshore e.g. to explore deep water  
22  
23 163 resources. Toward the definition of this "continent-ocean-boundary, COB" at  
24  
25 164 passive margins one of the crucial questions is how to define these  
26  
27 165 boundaries. Torsvik et al. (2009) described how COB for the South Atlantic  
28  
29 166 margins at both sides can be defined: by the interpretation of seismic,  
30  
31 167 gravimetric, magnetic, bathymetric and geological information. Any  
32  
33 168 identification of the COB is also important for the definition of plate boundaries  
34  
35 169 at the time of break-up which allows the reconstruction of geometry and earlier  
36  
37 170 position of former continents - e.g. in the reconstruction of Pangaea. It is not  
38  
39 171 the aim of this paper to recalculate COBs. However we trust that the  
40  
41 172 processing of gravity field and its second derivatives for recalculation of COB  
42  
43 173 (Torsvik et al., 2009) will benefit from the increase of high resolution satellite  
44  
45 174 derived gravity fields. Later we will return to this point in the conclusion  
46  
47 175 (Section 5) suggesting calculating the 2<sup>nd</sup> derivative of satellite gravity  
48  
49 176 (derivative of the gravity gradient).

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1  
2 **178 2. Geophysical characteristics of margins**  
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7  
8 180 To investigate both active and passive continental margins almost all  
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10 181 geophysical methods can help to discover their lithospheric structures:  
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12 182 Seismic, potential, electro-magnetic and electrical fields. Recent studies (e.g.  
13  
14 183 Torsvik et al. 2009) show that processes in the lithosphere are linked to the  
15  
16 184 dynamic mantle and dynamic processes have an important influence on the  
17  
18 185 evolution of lithospheric plates, which is manifested in the formation of  
19  
20 186 dynamic topography (e.g. Flament et al., 2013; [Bocher et al., 2016](#) [Colli et al.,](#)  
21  
22 187 [2016](#)). A key example for this interaction is the opening history of the Atlantic,  
23  
24 188 where asthenospheric material reaches the surface at the mid-ocean  
25  
26 189 spreading center. The uplift of hot material of the asthenosphere leads to  
27  
28 190 seafloor spreading which manifests itself in the spreading anomalies in the  
29  
30 191 magnetic field. The spreading axes are expressed either by topographic  
31  
32 192 (heights above sea level, e.g. Iceland) or bathymetric heights (below sea level  
33  
34 193 e.g. Atlantic ocean ridge). The worldwide offshore stripe pattern of magnetic  
35  
36 194 anomalies which are located parallel to mid-ocean ridges allow a temporal  
37  
38 195 classification of developed oceanic crust. The EMAG2 total magnetic field  
39  
40 196 model has a reasonable resolution of two arc-minutes (Maus et al. 2009). The  
41  
42 197 model was compiled from ship-, air-borne and satellite data (CHAMP). For the  
43  
44 198 purpose of interpolation of magnetic field anomalies the oceanic crust ages of  
45  
46 199 Müller et al. (2008) were employed (Maus et al. 2009). Hydro-carboniferous  
47  
48 200 exploration targeted most of the passive margins of the world with both  
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50 201 refraction and/or reflection seismic.  
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2 202 Seismic onshore-offshore investigations of a passive continental margin  
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4 203 aim to investigate the transition zone between the oceans and continents, e.g.  
5  
6 204 SE Atlantic and the African continent (Franke et al., 2006; 2010; Bauer et al.,  
7  
8 205 2000; Hirsch et al., 2009; Schnabel et al., 2008). These studies reveal the  
9  
10 206 history and mechanisms of the break-up and its relation to the driving  
11  
12 207 magmatic processes (plume – crust interactions) in the underlying mantle and  
13  
14 208 lithosphere. The South American side of the southern Atlantic was target of  
15  
16 209 multiple seismic investigations since many years (Mohriak, 2014; Blaiçh et al.,  
17  
18 210 2011; Franke et al. 2006; Heit et al. 2007; Hinz et al., 1999 and many more) to  
19  
20 211 investigate the shelf areas offshore, the continent-ocean transition and the  
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22 212 seaward dipping reflectors (SDR) which are characterized by high seismic  
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24 213 velocities and related high densities.  
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30 215 *2.1 The role of gravity field interpretations and prerequisites*  
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33 216 Remark:

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35 217 The units used for gravity field values and its gradients are given in  
36  
37 218 mGal (Milligals) and E (Eötvös) – as it is still common use in  
38  
39 219 Geophysics and Geodesy. Converting these units into SI-units, there  
40  
41 220 are:  $1 \text{ mGal} = 10^{-5} \text{ m/s}^2$  and  $E = 10^{-9} \text{ s}^{-2}$ .

42  
43 221 Rock densities are always given in  $\text{kg/m}^3$ .

44  
45 222 First, the situation will be analyzed which was typical before the era of  
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47 223 satellite missions which started in the year 2000 with the launch of the  
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49 224 CHAMP satellite (refer to Chapter 4). Modelling and interpretation of the  
50  
51 225 Earth's gravity field and its derivatives often had to deal with a merge of data  
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53 226 sets with rather different resolution, different age and quality, coverage and  
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2 227 wavelength content. This has been documented in a long list of papers  
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4 228 (among many others: Schaller et al., 2015; Bouman et al., 2014; Hosse et al.,  
5  
6 229 2014; Gutknecht et al., 2014; Köther et al., 2012; Tašárová, 2007). These  
7  
8 230 papers have shown that lithospheric models of study regions suffer mainly  
9  
10 231 from two facts:  
11  
12 232 (1) Nearly all information for constrained gravity field modelling of the  
13  
14 233 lithospheric is based on irregularly distributed profiles and  
15  
16 234 (2) large data gaps and partly low data quality due to very limited  
17  
18 235 access and infrastructure.  
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22  
23 237 The global gravity model EGM2008 (Pavlis et al. 2012) inherited – and still  
24  
25 238 inherits - these problems related with available terrestrial databases. Pavlis et  
26  
27 239 al. (2008) described the compilation of the EGM2008 gravity which consists of  
28  
29 240 different sources: up to a spatial resolution of approximately 140 km GRACE  
30  
31 241 (Gravity Recovery and Climate Experiment) data have been used, data from  
32  
33 242 140 km to approx. 10 km spatial resolution are derived from terrestrial and  
34  
35 243 satellite altimetry and so called “fill-in” data (Section 4.2). Today we know that  
36  
37 244 the relative weighting of GRACE with respect to the other datasets was too  
38  
39 245 low, leading to a dominance of terrestrial data errors in spectral regions where  
40  
41 246 better GRACE data would already have been available.  
42

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44 247

45 248 Figure 2: Full page width

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49  
50 250 **Fig. 2.** This figure portrays the terrestrial data inconsistency which is rather  
51  
52 251 typical for gravity surveys in remote areas (here across the South American  
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54 252 continent at a swath between 36°S – 42°S from the passive margin in the East  
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2 253 to the active margin in the West). As an example the different sources of  
3  
4 254 gravity data of the Southern Central American continent are shown together  
5  
6 255 with the visualization of big data gaps (Tašárová, 2007). The green and red  
7  
8 256 points in Chile (CH) and Argentina (AR) are stations in the ARANEDA I and II  
9  
10 257 (University of Chile) datasets; black points in southern Chile and the Arauco  
11  
12 258 Peninsula are ENAP data (Chilean oil industry); the gray dense network in  
13  
14 259 Argentina (AR) show the YPF data (Argentine oil company); yellow points:  
15  
16 260 stations of the MIGRA 2000 dataset, and the MIGRA 2002 data (both  
17  
18 261 measured by the gravity group at the University Kiel) are shown in blue. The  
19  
20 262 brown lines offshore denote the ship-borne gravity data profiles from the  
21  
22 263 German research vessel "Sonne".

24  
25 264 Fig. 2 shows the situation some 20 years ago at the central South  
26  
27 265 American subcontinent. For a continental gravity field study the field had to be  
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29 266 compiled from very different data sources often without any meta data  
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31 267 information - e.g. for gravity data which were measured on behalf of oil  
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33 268 companies (grey area in Fig. 2). In other areas, e.g., in the eastern part of  
34  
35 269 Argentina (yellow dots in Fig. 2) data are missing due to difficult or impossible  
36  
37 270 access.

39 271 At the end of this Section we will pose the question how large  
40  
41 272 lithospheric structures and how big density differences to their surrounding  
42  
43 273 have to be in order to cause a detectable signal at orbit height of a satellite.  
44  
45 274 This consideration describes the situation at active and passive margins in an  
46  
47 275 analogues manner. In Fig. 3 the effects of gravity and gradients are shown for  
48  
49 276 a simple model. For density contrasts between the values  $\Delta\rho = 10^1 - 10^3$   
50  
51 277  $\text{kg/m}^3$  Gutknecht et al. (2011) calculated gravity and gradients at the GOCE  
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53 278 satellite orbit height of 255 km resulting from a sphere with minimum  
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2 279 diameters between  $d = 20 - 200$  km ( $b = 10 - 100$  km, refer to Fig. 3) to be  
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4 280 tangent to the Earth surface). Within the range of the assumed parameters the  
5  
6 281 minimum diameter required to produce signal differences of  $1 \times 10^{-5} \text{ ms}^{-2}$  and  
7  
8 282  $12 \times 10^{-12} \text{ s}^{-2}$  at orbit height. This is rather close to the expected accuracies of  
9  
10 283 the gravity and vertical gravity gradient of the GOCE mission (recent values  
11  
12 284 are: 0.45 mE for GOCE-only, 0.2 mE for GOCE+GRACE models, cf. also Fig.  
13  
14 285 13). Fig. 3 conveys that a structure with a diameter of some 45 km and a  
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16 286 density contrast of  $240 \text{ kg m}^{-3}$  could be detected in satellite gravity at orbit  
17  
18 287 height. If the diameter of the model sphere increases to 90 km, its density  
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20 288 contrast should not be less than  $33 \text{ kg m}^{-3}$ .

22  
23 289 The simple model described above fits rather well the dimension of the  
24  
25 290 Jurassic arc batholiths at the Northern Chilean continental margin with  
26  
27 291 diameters of roughly 60 – 120 km (Sobiesiak et al., 2007). This supports the  
28  
29 292 idea that batholithic structures e.g. intrusions at continental margins, can be  
30  
31 293 detected using data of the modern satellite missions– both gravity and  
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33 294 gradients.

35 295

37 296 Figure 3: one and a half page width

39 297

41 298 **Fig. 3.** Gravity and gradient signal caused by a minimum diameter of a sphere  
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43 299 with given density contrast in the orbit height of 255 km (Gutknecht et al.,  
44  
45 300 2011). The thick solid and dotted lines represent gravity signals of 1 and 10  
46  
47 301 mGal at orbit height. The thin dashed and dash-dotted lines represent  
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49 302 gradients of 12 and 1000 mE, respectively. The grey shaded area shows  
50  
51 303 results which are based on a possible combination of geometry and density  
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53 304 parameters of the causing mass anomaly.

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5 306 *2.2 Gravity anomalies and isostasy*

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10 308 A first look at Fig. 4 provides already helpful information on the gravity  
11  
12 309 gradients of the Earth at the continental margins. The figure bases on the  
13  
14 310 evaluation of gravity field maps from the recent gravity missions. A complete  
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16 311 Bouguer anomaly (corrected by the effect of elevation, spherical slab,  
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18 312 topography both on- and offshore) was used to calculate the first derivative of  
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21 313 the field – the dip curvature of gravity.

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25  
26 315 Figure 4: Full page width

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32 317 **Fig 4.** Global horizontal gravity gradients. Deep blueish colors mark regions  
33  
34 318 on Earth where the dip curvature (horizontal gradient) of the worldwide  
35  
36 319 Bouguer gravity field (EIGEN-6C4) is small or even zero. The more reddish  
37  
38 320 the colors are the steeper the gradients. The strongest gradients are observed  
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40 321 at the active continental margin of Central South America. On the contrary at  
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42 322 most of the passive continental margins (Fig. 1) the gradients are rather small.

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48 324 Dark blueish colors indicate rather weak dip curvature which means that  
49  
50 325 horizontal gravity gradients are small, on the contrary reddish tones point to  
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52 326 strong dip curvature and therefore strong gradients. The active continental  
53  
54 327 margins in the area of the Circum-Pacific are generally marked by reddish

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2 328 colors – strong gradients – which are related with extended density variations

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4 329 in crust and mantle in the transition from oceanic to continental margins.

5  
6 330 Mostly low gradients – light blueish colors - are typical for passive continental

7  
8 331 margins. In order to understand this relationship two questions arise:

9  
10 332 (1) How was the gravity *anomaly* calculated and how big is its magnitude of

11  
12 333 the anomaly caused by the mass distribution (both topography/bathymetry

13  
14 334 and densities) at continental margins and

15  
16 335 (2) How was the *field* observed and what is known about accuracy and

17  
18 336 homogeneity of gravity field observations?

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20 337 With regard to the first question one has to consider that gravity

21  
22 338 observations at the Earth's surface and the Earth near space vary from the

23  
24 339 theoretical "normal" field value. Their magnitude is caused not only by the

25  
26 340 latitude effect but by elevation of observational points, the density of cap

27  
28 341 beneath the station in the underground and the topographic masses in the

29  
30 342 neighborhood. After correction of earth tidal effects and air pressure

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32 343 variations the remaining time invariable parts of the measured signal are:

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34 344 Normal gravity  $\gamma_0$

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36 345 Effect of topographic masses  $\delta g_{TOP}$

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38 346 Bouguer slab  $\delta g_{BPL}$

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40 347 Elevation effect (free air term)  $\delta g_{NIV}$

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42 348 Effect of the crustal root  $\delta g_R$

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44 349 Known mass inhomogeneities  $\delta g_{GEOL}$

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2 351 Common representations of gravity measurements in maps and profiles  
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4 352 depend on what has been calculated as stated in the table above and added  
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6 353 as corrections to the measured gravity field values. From all measured gravity  
7  
8 354 values the normal gravity (in the height and position of the observable) is  
9  
10 355 subtracted, and therefore the term “gravity anomaly” is defined. This is simply  
11  
12 356 the difference between the observable values to “normal” gravity. In  
13  
14 357 Geophysics we distinguish mainly between three anomalies Free air- (FA),  
15  
16 358 Bouguer- (BA) and isostatic anomaly (ISA) which define gravity in the height  
17  
18 359 of the observation (e.g. Li and Götze, 2001; Hackney and Featherstone, 2003;  
19  
20 360 Naudy et al., 1965; LaFehr, 1991 and many others).

21  
22  
23 361 With  $\delta g_R$  we describe the effect of a mountain root and define Free Air,  
24  
25 362 Bouguer and isostatic anomalies such as:

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28 363 
$$FA = \Delta g'_0 = g_{obs} + \delta g_{NIV} [+ \delta g_{TOP}] - \gamma_0$$

29  
30 364 
$$BA = \Delta g''_0 = g_{obs} + \delta g_{NIV} + \delta g_{TOP} + \delta g_{BPL} - \gamma_0$$

31  
32 365 
$$ISA = \Delta g_{ISA} = g_{obs} + \delta g_{NIV} + \delta g_{TOP} + \delta g_{BPL} + \delta g_R - \gamma_0$$

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38 367 Calculations of the individual correction terms depend on the objective of the  
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40 368 survey and are variously complicated to handle.  $g_{obs}$  denotes the measured  
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42 369 gravity field value. The calculation of the topographical reduction ( $\delta g_{TOP}$ )  
43  
44 370 requires precise knowledge of the terrain and surface near densities and  
45  
46 371 today it is most likely calculated by the aid of digital elevation models (DEM),  
47  
48 372 among others refer to (Holzrichter, 2013 and Szwillus and Götze, 2016); it  
49  
50 373 requires the use of a computer and digital elevation data. The effect of  
51  
52 374 topographic masses is normally calculated in a surrounding circular area of 50  
53  
54 375 km up to 167 km (e.g. La Fehr, 1991; Hinze et al., 2006). However, Mikuška et

1  
2 376 al. (2006), Szwillus and Götze (2016), Szwillus et al., (2016) point to “long  
3  
4 377 distant relief effects” and propose the calculation of *all* topographic masses on  
5  
6 378 Earth. The gravity effect of a Bouguer slab with a thickness which is defined  
7  
8 379 by the difference between the physical station height and the reference level  
9  
10 380 (normally the geoid) should always be calculated by a spherical cap (e.g.  
11  
12 381 Baeschlin, 1948). For smaller areas (with a reduction radius  $R < 20$  km) and  
13  
14 382 moderate terrain roughness the Bouguer slab can easily be calculated by:

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17 383 
$$\delta g_{BPL} = -2 \pi G \rho (H_S - H_B)$$

18  
19  
20 384 with:

21  
22 385  $G$  = Gravitational constant ( $6.672 \cdot 10^{-11} \text{ m}^3 \text{ s}^{-2} \text{ kg}^{-1}$ ),

23  
24 386  $\rho$  = crustal density ( $2\,670 \text{ kg/m}^3$ ); offshore: rock-equivalent  
25  
26 387 density

27  
28 388  $H_S$  = station height; offshore: ocean depth  $H_D$ ,

29  
30 389  $H_B$  = reference height (usually it is referred to the geoid).  
31

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34  
35 391 In the last step the free air effect  $\delta g_{NIV}$  is calculated by use of the  
36  
37 392 “normal gradient” ( $0.3085 \text{ mGal/m}$ ):

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39  
40 393 
$$\delta g_{NIV} = 0.3085 \cdot (H_S - H_B) \text{ mGal/m}$$

41  
42 394 It has to emphasized that the above used constant gravity gradient for large  
43  
44 395 scale investigations has to be replaced by equivalent values of derivation of  
45  
46 396 closed mathematical expression of the normal gravity which is always latitude  
47  
48 397 and height dependent. It was also recommended to calculate an “atmospheric  
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50 398 correction” (Wenzel, 1985) in order to eliminate long wavelengths errors from  
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52 399 the observed gravity field.  
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2 400 Note: the terms of “flat” Bouguer slab in  $\delta g_{\text{BPL}}$  and the “constant vertical  
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4 401 gradient” in  $\delta g_{\text{NIV}}$  are in the above formulas are used here for simplification  
5  
6 402 only. Modern satellite gravity field processing at large spatial scales requires a  
7  
8 403 spherical cap calculation and the consideration of latitude and height  
9  
10 404 dependent calculation of the vertical gradient.

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12  
13 405 Special emphasis has to be put on the situation in oceanic regions. If  
14  
15 406 we assume that station heights are equal to geoidal heights ( $h = 0$  m) Bouguer  
16  
17 407 and Free Air anomaly is equal due to:

$$20 \quad 408 \quad \text{BA} = \text{FA} + \delta g_{\text{BPL}} = \text{FA} + [-2 \pi G \rho (H_s - H_B)]$$

21  
22 409 with:  $H_s - H_B = 0$  m;

23  
24 410 it follows that  $\text{BA} = \text{FA}$ , in case the FA is already corrected by  $\delta g_{\text{TOP}}$  which  
25  
26 411 contains gravity effect of masses at the ocean floor.

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28 412 In oceanic areas the slab density in the Bouguer slab correction term  
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30 413 must be modified due to the known water depth “D” and the difference in  
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32 414 water- and rock density (Fig. 5). If we assume a crustal rock density of 2 670  
33  
34 415  $\text{kg/m}^3$  and a water density of 1030  $\text{kg/m}^3$  the resulting density for calculations  
35  
36 416 of offshore Bouguer anomalies is  $\rho^* = -1\,640 \text{ kg/m}^3$ .

37  
38  
39 417 This results in:

$$40 \quad 418 \quad \text{BA} = \text{FA} + \delta g_{\text{BPL}} = \text{FA} + 2 \pi G \rho^* (H_s - H_D)$$

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45 420 Figure 5: Full page width

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49 422 **Fig. 5.** Illustration for describing the calculation of Bouguer anomalies on  
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51 423 continents (A), at the ocean (B) and from satellite (C). (A): On land the  
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53 424 reduction density  $\rho$  is commonly taken as  $2\,670 \text{ kg m}^{-3}$ . The effect of  
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2 425 topography is already removed. The thickness of Bouguer slab equals the  
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4 426 station height ( $H_s$ ). (B): In contrast the reduction density at sea is  $-1\,640\text{ kg}$   
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6 427  $\text{m}^{-3}$ . It is the difference between the sea water density of  $1\,030\text{ kg m}^{-3}$  and the  
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8 428 rock density of  $2\,670\text{ kg m}^{-3}$ ; thickness of the slab now is equal to the different  
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10 429 water depths (D). (C): Calculating a Bouguer anomaly in case of satellite  
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12 430 gravity a “mass correction” is calculated:  $\delta g_{\text{Mass}} = \delta g_{\text{Top}} + \delta g_{\text{BPL}}$ .

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### 15 16 432 *2.3 Gravity at passive continental margins*

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22 434 In Fig. 6 a very simple Airy-Isostasy model of a passive continental margin is  
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24 435 shown. The continental crust is much thicker than the oceanic crust and above  
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26 436 the oceanic crust there is the water cover of a few 1000 meters. The specific  
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28 437 geometry of the „crust – mantle – water cover“ constellation plays an  
29  
30 438 important role on the trend of the gravity field here. Because of the fact that in  
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32 439 the example of Fig. 6 there is no topography, the continental margin remains  
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34 440 in an isostatic equilibrium, and one can assume that no Free Air anomaly  
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36 441 exists.

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42 443 Figure 6: One and half page width

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48 445 **Fig. 6.** Airy isostatic model at a continent – ocean transition. Notice the thick  
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50 446 continental crust and the thin oceanic crust at a passive continental margin.  
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52 447 Crust and mantle densities are simplified. Reasonable contrasts which cause  
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54 448 large gravity anomalies are related to the water – continent density contrast  
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2 449 and crust – mantle density at the continent of approx.  $430 \text{ kg m}^{-3}$ . Refer to text  
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4 450 for more information.  
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10 452 However, in Fig. 7 (A) strong gravity anomaly results from the same  
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12 453 density model (Fig. 7C) which was shown in Fig. 6. The Fig. 7(A) shows the  
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14 454 modelled anomaly only for water effect of gravity: related to a model of a “half-  
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16 455 indefinite” plate the resulting anomaly is negative and is caused by a strong  
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18 456 gradient. Fig. 7 (B) on the other side contains model results which have been  
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20 457 done only for the oceanic mantle: now the anomaly is positive and it has a  
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22 458 gentle increase because its position is far deeper. Finally Fig. 7(C)  
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24 459 demonstrates how the total Free Air anomaly results from the superposition of  
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26 460 both effects: the Free Air anomaly is zero in the continental area and over the  
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28 461 ocean as well. However, exactly above the margin the gravity field is  
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30 462 characterized by a maximum and a minimum that follows. This distribution is  
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32 463 a so called „edge“ or boundary effect of the Free Air anomaly and is effected  
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34 464 by the difference of the steep gradients in the model.  
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40 466 Figure 7: one and a half page width  
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46 468 **Fig. 7.** The principal effects on the gravity field at continental margins have  
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48 469 equal gravity magnitudes but different gradients. In (A) the water effect causes  
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50 470 a steep gradient and in (B) the density surplus of the oceanic mantle is a  
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52 471 deeper seated effect which causes only a gentle gradient. In (C) it is explained  
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54 472 that a Free Air anomaly at a continental margin is caused by both a negative  
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2 473 and positive “edge effect” due to the superposition of contributions that have  
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4 474 equal magnitudes but different gradients.  
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10 476 The trends of a Free air and Bouguer anomalies are shown in Fig. 8.

11 477 Here modelling again gets use of a „half indefinite“ thin plate for the offshore

12 478 area (water). It is “zero” over the continent zero and over the oceanic area

13  
14 479 “positive” ( $\rho_w = 1030 \text{ kg/m}^3$ ). The half of the whole Free Air anomaly maximum

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17 480 is accomplished exactly over the edge of the continent.  
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23 482 Figure 8a-b: Full page width  
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26 483 Figure 8c: One and a half page width  
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32 485 **Fig. 8.** Free Air anomaly and Bouguer anomaly at continental margins which

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34 486 is also in an isostatic balance. (A) The absolute value of the excess mass

35  
36 487  $|\Delta m|$  is equal to the absolute value of the deficient mass  $|\Delta m|$ . Therefore

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38 488 the integral of gravity change with respect to the x-coordinate is zero:  $\int \Delta g \, dx =$

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40 489 0. (B) The Bouguer correction at the ocean (see Fig. 3) applied to the Free Air

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42 490 anomaly in (A) yields the general form of the Bouguer anomaly at passive

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44 491 continental margins. (C) The “geological” mass inhomogeneities at the

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46 492 continental margins (seaward dipping reflectors, magmatic remnants, salt

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48 493 structures etc.) cause rather local gravity anomalies which superimpose the

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50 494 regional gravity wavelengths – which are effected by the “simple” structures in

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52 495 (A) and (B).  
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2 496 The gravity fields in Figs. 7 and 8 are caused by the over simplified  
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4 497 density structure at the “modelled margin” in Fig. 6. In the real world these  
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6 498 margins show a rather complicated picture of gravity distribution due to mass  
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8 499 inhomogeneities in the Earth’s crust and lithosphere (and even in the mantle)  
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10 500 which are the results of the long-lasting history of the breakup of the  
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12 501 Gondwana supercontinent. This becomes quite clear if looking at the  
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14 502 processed gravity fields which are shown in the series of figures (Fig. 14  
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16 503 through 21) in Section 4. Their interpretation in terms of regional tectonic and  
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18 504 distribution of rock densities will help to provide a rather detailed insight into  
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20 505 the causing structure (geometry) and density distribution of the passive  
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22 506 margins in the South Atlantic region.

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25 507 Most aspects of the calculated anomalies, both Free Air and Bouguer,  
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27 508 which were discussed before, are typical for nearly all of the continental  
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29 509 passive and active margins on Earth. In the next Section the focus will be set  
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31 510 on the situation in the Southern Atlantic between Africa and South America –  
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33 511 the research area of the German Priority Program 1375 “SAMPLE” of the  
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35 512 German Science Foundation – DFG (<https://www.sample-spp.de/>). The  
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37 513 acronym stands for “**S**outh **A**tlantic **M**argin **P**rocesses and **L**inks with onshore  
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39 514 **E**volution”. In this interdisciplinary project the primary research areas are the  
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41 515 mantle dynamics and magmatic processes, the lithospheric structure,  
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43 516 deformation processes and rifted margin formation, the post-rift topographic  
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45 517 evolution and many more. In the following Section we will concentrate on this  
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47 518 part of the world because a big variety of data and information is available to  
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49 519 responds to one of the key questions – how modern satellite missions can  
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51 520 contribute to the interpretations and to the understanding of the transition from  
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53 521 continental to oceanic lithosphere.  
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5 523 *2.4 Focus region: South Atlantic passive margins*  
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10 525 To study deeper structures and the overall evolution of conjugate  
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12 526 passive continental margins of the South African and South American  
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14 527 continents 3D structural models have been designed and evaluated by  
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16 528 SAMPLE scientists and their international partners: They constructed detailed  
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18 529 density models at both sides of the Southern Atlantic Ocean and a rather  
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21 530 preliminary density model for the oceanic part in course of a master thesis  
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23 531 (Klinge, 2016). These models are constrained by information and data from  
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25 532 boreholes, refraction and reflection seismic, seismological tomography and  
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27 533 potential field data – mainly gravity field data. Geophysical fields and  
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29 534 observations map geometry and distribution of physical properties of the  
30  
31 535 transitional structures of both crust and lithospheric mantle. Model results (Fig.  
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33 536 9) show (Maystrenko et al., 2013; Autin et al., 2016) that basin centers at the  
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35 537 western (Argentinean) side are oriented west-east and therefore oblique to the  
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37 538 mid ocean rift axis while at the other (African) side basin centers extend  
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39 539 parallel to the ocean rift in north-south direction.  
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45 541 Figure 9: One and a half page width  
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2 543 **Fig. 9.** For illustration this figure portrays a 3D density model of the SW  
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4 544 African continental margin (left, modified after Maystrenko et al., 2013), and  
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6 545 the density structure at the Argentinean side (right, Autin et al., 2016).  
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11 547       Apart from these structural differences both sides of the Southern  
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13 548 Atlantic reveal similar distributions in temperature and density. Small  
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15 549 thicknesses and density modifications in the lithospheric mantle point to small  
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17 550 lateral variations of heat transfer into the overlying crust. However, more  
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19 551 relevant for the crustal heat field are lateral thickness changes of the  
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21 552 crystalline crust which produce the bigger part of radiogenic heat. This  
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23 553 contrasts observations and modelling results at passive continental margins in  
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25 554 the area of the Northern Atlantic (Scheck-Wenderoth and Maystrenko, 2008).  
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27 555 They found that the oceanic part of lithospheric mantle is much thinner and  
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29 556 characterized by smaller densities which cause higher temperatures in the  
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31 557 upper crust of the ocean.  
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35 558       To contrasting large scale paleostress fields on the correlating margins  
36  
37 559 of the South Atlantic Salomon et al. (2014) point to in their studies of the  
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39 560 South Atlantic. They asked themselves “how passive” continental margins  
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41 561 across the globe currently are. Following the results of several other studies  
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43 562 these margins experience a variety of stress states and undergo significant  
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45 563 vertical movements, as they were deduced from studies of paleo-stresses at  
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47 564 both sides of the Southern Atlantic. Here, the bounding continents consist of  
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49 565 very different recent geological histories: Africa experiencing continental rifting  
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51 566 whereas South America is influenced by subduction on the Pacific side. It is  
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53 567 not clear to what extent the Atlantic continental margins are subject to the  
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2 568 same stresses and vertical motions as the main continents. Their results show  
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4 569 that the tectonic evolution of the continental margins of the South Atlantic is  
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6 570 not only passive and that both margins vary significantly in structural style and  
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8 571 stress fields, indicating that variable plate boundary forces play a major role in  
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10 572 margin evolution. In Fig. 10 we show the situation at the S-American and S-  
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12 573 African margin with reference of the paleo-stress field, as it was published by  
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14 574 Salomon et al. (2014). Their findings demand careful modelling of both  
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16 575 continental margins and a geophysical database which is able to resolve even  
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18 576 very small modifications of physical parameters and their structures; refer also  
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20 577 to Fig 9 (a) and (b) and the 3D density modelling of lithospheric by  
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22 578 Maystrenko et al. (2013) and Autin et al. (2016).

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34 582 **Fig. 10.** The sketch (Salomon et al., 2014) portrays an E-W cross-Section  
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36 583 between South Africa and South America which summarize the situation of  
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38 584 their obtained paleo stresses. It shows that the African margin is controlled by  
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40 585 extension while compression characterizes the situation at the South  
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42 586 American side. Salomon et al. (2014) explained the extensional state in the  
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44 587 east by the existing “African superplume” and the compression in the west by  
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46 588 the Andean subduction zone.

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52 590 Novel satellite gravity missions aim at a breakthrough in recovering the  
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54 591 Earth’s gravity and magnetic fields, their gradients as well as their temporal

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2 592 variation. Static anomalies in potential fields (refer to Figs. 14 through 21) are  
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4 593 caused by irregular mass distribution on and within the Earth, temporal  
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6 594 variations of the gravity field are associated with mass transport processes in  
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8 595 the Earth system, such as dynamic processes on the Earth's surface, in  
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10 596 lithosphere and upper mantle.

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15 598 **3. Modern satellite gravity missions**16  
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19 600 The launch of the first generation of satellite gravity missions (Fig. 11) has  
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21 601 revolutionized our knowledge of the global Earth's gravity field and its temporal  
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23 602 changes. The German CHAMP (Challenging Minisatellite payload; mission period  
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25 603 2000-2010; Reigber et al., 2002; <http://op.gfz-potsdam.de/champ/>) mission, the  
26  
27 604 US/German GRACE (Gravity Recovery and Climate Experiment; mission period  
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29 605 2002-ongoing; Tapley et al., 2004; <http://www.csr.utexas.edu/grace/>) mission, and the  
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31 606 European GOCE (Gravity field and steady-state Ocean Circulation Explorer; mission  
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33 607 period 2009-2013; Drinkwater et al., 2003;  
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35 608 [http://www.esa.int/Our\\_Activities/Observing\\_the\\_Earth/The\\_Living\\_Planet\\_Programm](http://www.esa.int/Our_Activities/Observing_the_Earth/The_Living_Planet_Programme/Earth_Explorers/GOCE/ESA_s_gravity_mission_GOCE)  
36  
37 609 [e/Earth\\_Explorers/GOCE/ESA\\_s\\_gravity\\_mission\\_GOCE](http://www.esa.int/Our_Activities/Observing_the_Earth/The_Living_Planet_Programme/Earth_Explorers/GOCE/ESA_s_gravity_mission_GOCE)) operated by the European  
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39 610 Space Agency (ESA), improved significantly the coverage and availability of high  
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41 611 resolution and precisely measured data. These gravity missions are the only  
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43 612 measurement technique that can directly observe mass changes on a global scale,  
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45 613 and thus they provide a unique observation system for monitoring mass transport in  
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47 614 the Earth system. For modern magnetic field observation, apart from the CHAMP  
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49 615 mission (2000-2010), with ESA's three SWARM satellites that have been successfully  
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51 616 launched in November 2013 also gradients observations have become available  
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53 617 (<http://esamultimedia.esa.int/multimedia/publications/BR-302/>).

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6 621 **Fig. 11.** Satellite gravity missions CHAMP (left), GRACE (center) und GOCE (right).

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8 622 (Sources: CHAMP: GFZ Potsdam, GRACE: NASA, GOCE: ESA Medialab)

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11  
12 624 In these missions, three measurement concepts are implemented:

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14 625 1. Observation of orbit perturbations of low-flying satellites due to the varying  
15 626 gravitational attraction, by Global Positioning System (GPS), with an accuracy  
16 627 of 2-3 cm. Non-gravitational forces acting on the satellite, such as drag of the  
17 628 residual atmosphere or solar radiation pressure, are measured by an  
18 629 accelerometer and corrected for in the frame of the gravity field modelling.  
19 630 This satellite tracking technique between a low Earth orbiter (LEO) and high-  
20 631 flying GPS satellites is called satellite-to-satellite tracking in high-low mode  
21 632 (SST-hl), and is implemented in all three missions CHAMP, GRACE and  
22 633 GOCE. It is the primary measurement technique of CHAMP.

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26 635 2. Observation of orbit differences (ranges) and their temporal change (range  
27 636 rates) between two LEO satellites. This satellite-to-satellite tracking in low-low  
28 637 mode (SST-ll) concept is realized by the GRACE mission. It consists of two  
29 638 identical satellites following each other on the same orbit with an average  
30 639 distance of 200 km. The inter-satellite ranging is performed by means of a K-  
31 640 band microwave system with micrometer accuracy, and shall be done by laser  
32 641 interferometry in future gravity missions in order to further increase the  
33 642 ranging accuracy.

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37 644 3. Observation of acceleration differences on very short baselines (satellite  
38 645 gravity gradiometry, SGG), representing second order derivatives of the  
39 646 gravitational potential  $V$  in all three spatial directions. This concept was

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2 647 applied by the GOCE mission. Its core measurement, the gravity gradiometer,  
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4 648 was composed of 6 accelerometers fixed on 3 orthogonal axes symmetrically  
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6 649 around the center of mass of the satellite, measuring acceleration differences  
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8 650 on very short baselines of only half a meter in all three spatial dimensions.

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10 651 The achievable performance of satellite gravity missions depends mainly on the  
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12 652 observation technique and the orbit altitude. Fig. 12 shows the performance of  
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14 653 different mission concepts in terms of the degree error median, which describes the  
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16 654 average signal or noise amplitude at a certain degree  $n$  of the spherical harmonic  
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18 655 series expansion of the gravitational potential  $V$  in spherical coordinates (with radius  
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20 656  $r$ , co-latitude  $\vartheta$ , longitude  $\lambda$ ):

$$V(r, \vartheta, \lambda) = \frac{GM}{R} \sum_{n=0}^{N_{\max}} \left( \frac{R}{r} \right)^{n+1} \sum_{m=0}^n \bar{P}_{nm}(\cos \vartheta) [\bar{C}_{nm} \cos(m\lambda) + \bar{S}_{nm} \sin(m\lambda)]$$

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28 660 where  $G$  is the gravitational constant,  $M$  the mass of the Earth,  $R$  the mean Earth  
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30 661 radius,  $\bar{P}_{nm}$  the fully normalized Legendre polynomials of degree  $n$  and order  $m$ , and  
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32 662  $\{\bar{C}_{nm}, \bar{S}_{nm}\}$  the corresponding (Stokes) coefficients (e.g Torge, 2001). Therefore, the  
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34 663 degree error median describes the achievable gravity field accuracy at a certain  
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36 664 spatial (half) wavelength  $\lambda$ . The wavelength  $\lambda$  is linked to the harmonic degree  $n$  by

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$$\lambda = 20\,000 \text{ km}/n$$

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42 667  
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44 668 As an example, a harmonic degree of  $n = 200$ , which was the minimum target  
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46 669 resolution for the GOCE mission, corresponds to a spatial wavelength of  $\lambda =$   
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48 670  $20\,000 \text{ km}/n = 100 \text{ km}$ .

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50 671 As a reference, the stippled black curve in Fig. 12 shows the gravity field  
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52 672 signal itself. Correspondingly, the cross-over point of a mission performance curve

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2 673 with the black stippled curve indicates at which harmonic degree the signal-to-noise  
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4 674 ratio is '1'.

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11 678 **Fig. 12.** Absolute gravity signal and error estimates of different observation concepts  
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13 679 as a function of the harmonic degree  $n$  (bottom axis) and spatial wavelength  $\lambda$  (top  
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15 680 axis).

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19 682 From the orbit information (SST-hl) only the long-wavelength features of the gravity  
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21 683 field can be extracted. Although this observation type is not a direct gravity field  
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23 684 functional, it can be interpreted as disturbing acceleration acting on the orbit, and  
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25 685 thus the first order spatial derivative of the gravitational potential  $\partial V/\partial x_i$ . As a  
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27 686 representative of this measurement concept, the grey dot-and-dashed line curve  
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29 687 shows the performance of the CHAMP-only model AIUB-CHAMP 03S (Prange,  
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31 688 2011), which is based on 8 years of CHAMP kinematic orbit data.

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33 689 The grey curve shows the performance of the recent GRACE-only model  
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35 690 ITSG-Grace2014 (Mayer-Gürr et al., 2014), which is based on almost 11 years of K-  
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37 691 band inter-satellite ranging data following the SST-II concept (and supported by SST-  
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39 692 hl in the very low degrees). Compared to CHAMP, the superior measurement  
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41 693 principle of SST-II results in a significantly better accuracy in the low to medium  
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43 694 degree range as well as a higher spatial resolution. This can be explained by the fact  
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45 695 that the SST-II concept can be interpreted as a measurement of acceleration  
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47 696 differences on long baselines of about 200 km. The excellent performance of GRACE  
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49 697 in this spectral range makes this mission sensitive to the tiny temporal variations of  
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51 698 the Earth's gravity field, which are 4 - 5 magnitudes smaller than the static signal.

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53 699 The black solid curve shows the performance of GOCE, represented by the  
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55 700 GOCE-only model GOCE-TIM-R5 (Brockmann et al., 2014). It is mainly based on the  
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2 701 measurement technique of SGG and again SST-hl in the low degrees, because SGG  
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4 702 alone (green curve) is weak in this spectral range due to the specific noise  
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6 703 characteristics of the gravity gradiometer instrument. Measuring acceleration  
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8 704 differences on very short baselines of about half a meter, which approximate second  
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10 705 order derivatives of the gravitational potential  $\partial^2 V/(\partial x_i \partial x_j)$ , enables a further increase  
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12 706 of sensitivity for high-frequency signals. GOCE starts to become superior over  
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14 707 GRACE approximately at degree  $n = 115$ .

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17 709 *3.1 Global Gravity Field Models*18  
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21 711 Gravity field models including GOCE data from the complete mission period  
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23 712 are meanwhile available. While the model GOCE-TIM-R5, which is based on the  
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25 713 time-wise approach (Pail et al., 2011), is based purely on GOCE data, GOCE-DIR-  
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27 714 R5, which is based on the direct method (Bruinsma et al., 2014), contains also  
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29 715 GRACE and satellite laser ranging (SLR) data. Further satellite-only models are, e.g.,  
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31 716 EIGEN-6S2 (Rudenko et al., 2014), or the S-models of the GOCO series (Pail et al.,  
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33 717 2010). The maximum degree of expansion of these models is driven by the resolution  
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35 718 of GOCE, and varies from  $n = 280$  to  $300$ , corresponding to about  $70$  km spatial  
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37 719 wavelength. This makes clear that all medium scaled geological structures at  
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39 720 continental margins and elsewhere in the world which cause a significant gravity  
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41 721 effect can be detected (resolved) in the GOCE gravity field.

42 722 Combination models (notice the "C" in the field identifier) including also  
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44 723 terrestrial, air- and shipborne as well as altimetric gravity are, e.g., the already  
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46 724 mentioned pre-GOCE model EGM2008 (Pavlis et al. 2012), EIGEN-6C4 (Förste et  
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48 725 al., 2014), and GOCO05C (Fecher et al., 2013, 2016). These models extend the  
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50 726 spatial resolution beyond degree  $2000$  (which corresponds to  $10$  km wavelength).  
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52 727 However, it should be noticed that there are many regions with sparse and/or low-  
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54 728 quality terrestrial data, where it has to be questioned if such a high resolution is

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2 729 justified. This holds for many areas worldwide, e.g., the Central Andes in South  
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4 730 America and also for the passive continental margins of the South Atlantic.

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8 732 *3.2 Products for use in Earth sciences interpretation*

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11 734 Specifically regarding GOCE-related data, modelers and other users have the  
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13 735 choice among basically three representations of gravity field products:

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17 737 *3.2.1 Spherical harmonic coefficients*

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21 739 The most commonly used representation of the global gravitational potential  $V$   
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23 740 is its series expansion into spherical harmonics (Section 3). There corresponding fully  
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25 741 normalized spherical harmonic (Stokes) coefficients  $\{\bar{c}_{nm}, \bar{s}_{nm}\}$  represent the model  
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27 742 parameters, and are usually the target quantity when deriving the model from the  
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29 743 original gravity field data. The advantage of using this representation is that it can be  
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31 744 considered as a weighted average of the original measurement data, so that the  
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33 745 original noise level is significantly reduced due to this averaging. Based on the set of  
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35 746 spherical harmonic coefficients any arbitrary gravity anomaly can be derived at the  
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37 747 Earth's surface or at any height in outer space. All the global gravity models  
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39 748 discussed above are given in this parametrization.

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42 750 *3.2.2 Original gravity gradients along the satellite's orbit*

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45 752 In principle, also the gravity gradient time series for all six tensor components  
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47 753 measured along GOCE's satellite orbit can be used for geophysical modelling (refer  
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49 754 e.g. to Fig. 3). They represent the most original measurements. However, it has to be  
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51 755 considered that they are measured in a rotating reference frame, the so-called  
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53 756 "gradiometer reference frame" (GRF), which means that tensor rotations of the base

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2 757 functions have to be applied to exploit them to the best possible extent. Additionally,  
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4 758 they are affected by the colored noise characteristics of the GOCE gradiometer (Pail  
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6 759 et al., 2011), so that a single point-wise gravity gradient observation is affected by  
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8 760 large instrument noise, and therefore by itself has a low signal-to-noise ratio. All  
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10 761 these drawbacks make it difficult to use this data type directly for geophysical  
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12 762 modelling.

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### 14 764 3.2.3 Gravity gradient grids

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18 766 A reasonable compromise between the use of spherical harmonics and  
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20 767 original gravity gradients (see above) results in the use of *gravity gradient grids*,  
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22 768 which are usually defined in a well-oriented radial (North-East-up) frame at a constant  
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24 769 altitude. They are computed from the original gravity gradients defined in the GRF by  
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26 770 means of regional gravity processing methods. In fact, they are the spatial equivalent  
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28 771 of the spherical harmonic representation, but much easier to use and interpret. Pure  
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30 772 GOCE gravity gradient grids result from the space-wise method (Gatti and  
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32 773 Reguzzoni, 2015). In the frame of the ESA project GEOExplore global grids of all six  
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34 774 components of the gravity gradient tensor, based on a combination of GOCE and  
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36 775 GRACE data, and defined in a radial Earth-fixed reference frame at two altitudes of  
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38 776 225 km and 255 km, have been derived (Bouman et al., 2015). Since these grid  
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40 777 values are products of “averaging” original gradient data, the error level should be  
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42 778 similar as that of gradients synthesized from global spherical harmonic models.

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44 779 There is an ongoing discussion whether the gradient data contain more (high-  
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46 780 frequency) signal than global gravity models that have been derived from them. The  
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48 781 answer to this question lies in the constraints applied to these models. Constraints  
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50 782 applied to global gravity models are usually designed to optimize the signal-to-noise  
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52 783 ratio on a global scale. This means that in regions of very rough topography and  
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54 784 therefore high-frequency gravity signals there is the tendency to constrain the system

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2 785 too strongly. Regional gravity solutions techniques, which are usually applied to  
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4 786 generate gridded gravity gradient products, allow for regionally optimized constraints,  
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6 787 but on the cost of global homogeneity. In Pail et al. (2015b) it could be shown, that  
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8 788 compared to global models the gravity gradient grids are affected by a higher noise  
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10 789 level.

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12 790 The achievable accuracy and sensitivity of current gravity field models or  
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14 791 corresponding gravity (gradient) grids can be expressed by cumulative quantities,  
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16 792 which describe the estimated cumulative error at a certain harmonic degree (or the  
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18 793 corresponding spatial wavelength). Fig. 13 shows cumulative gravity anomaly errors  
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20 794 (a), as well as cumulative vertical gravity gradient errors at GOCE satellite altitude of  
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22 795 250 km (b), and ground level (c), for the GRACE models ITSG-GRACE2014s, the  
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24 796 pure GOCE model GOCE-TIM-R5, the combined satellite-only model GOCO05S and  
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26 797 the combined models EGM2008 (pre-GOCE) and GOCO05C (including GOCE data).

27 798 From Fig. 13 we can learn which geological structure at passive continental  
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29 799 margins (or elsewhere) can be resolved by the different gravity model types.  
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31 800 Assuming that the geological structure/mass anomaly generates a *gravity anomaly*  
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33 801 with a certain spatial wavelength on the Earth's surface, Fig. 13a then provides the  
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35 802 accuracy in mGal with which this anomaly can be captured. (The connection from  
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37 803 the size of a disturbing body to the resulting gravity signal is made in Fig. 3.). As an  
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39 804 example, a gravity signal with 100 km spatial wavelength at the Earth's surface could  
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41 805 be measured by satellites with an accuracy of about 0.5 mGal (black dashed and  
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43 806 solid grey curves). It can clearly be seen that the accuracy for shorter wavelength  
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45 807 signals dramatically decreases, and is already larger than 2 mGal for gravity signals  
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47 808 with approx. 80 km spatial wavelength. Beyond this resolution, satellites cannot  
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49 809 significantly contribute anymore, and high-accuracy terrestrial information, as it was  
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51 810 included, e.g., in

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7 816 **Fig. 13.** Cumulative gravity anomaly errors in (mGal) (a); vertical gravity gradient

8 817 errors in (mE) at 250 km (b), and ground level (c). This figure shows the generally

9 818 dramatic increase of the gravity gradient errors at ground level as a result of

10 819 downward continuation.

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17 821 GOCO05C (solid black curve), is necessary to resolve smaller-scale

18 822 geological structures. This becomes immediately clear if looking at the series of Figs.

19 823 14 through 21: most of the anomaly sizes at the margins of the South Atlantic are

20 824 smaller than 80 km. On the other hand we state that the regional gravity field caused

21 825 at the ocean-continent transition can satisfyingly be resolved by satellite only models

22 826 (S models). Fig. 13a also shows the major step forward due to satellite missions

23 827 compared to pre-GOCE models such as EGM2008 (black dot-and-dashed line)

24 828 especially in the long to medium wavelengths for gross interpretations at a

25 829 continental scale.

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28 831 Fig. 13b shows a similar representation when using *gravity gradients* at

29 832 satellite altitude as basis information for geophysical modelling of geological

30 833 structures. Pure GOCE-only models such as GOCE-TIM-R5 (dashed black curve)

31 834 provide gravity gradients at satellite altitude with standard deviations of 0.45 mE for

32 835 gravity signals with a spatial resolution of 100 km. These values can be decreased

33 836 further to 0.25 mE by combination with GRACE information, as it was done, e.g., in

34 837 the GOCO05S model (solid grey curve). Evidently, GRACE alone (dashed grey

35 838 curve) results in very high error amplitudes in the higher degrees, demonstrating the

36 839 dominant impact of GOCE at shorter wavelengths. Modern combined gravity models

37 840 such as GOCO05C (solid black curve) further increase the performance in the short-

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2 841 wavelength range by complementing the satellite data by ground data over the  
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4 842 continents and satellite altimetry over the oceans. Also here the improvement  
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6 843 compared to pre-GOCE combined models such as EGM2008 (black dot-and-dashed  
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8 844 line curve) is significant. Recently gradients of the satellite gravity field came into the  
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10 845 focus for modelling purposes which can support interdisciplinary interpretations  
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12 846 (Ebbing et al., 2013; Schaller et al., 2015; Götze, 2015).

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14 847 In Fig. 13b a very interesting feature is the flat curve of the combined gravity  
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16 848 model GOCO05C beyond degree 250. This results from the fact that beyond this  
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18 849 degree the signal amplitude of gravity gradients is already below the mE level, i.e.,  
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20 850 due to signal attenuation with altitude there is no significant gradient signal left in orbit  
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22 851 altitude beyond this degree, because most parts have been “filtered out” due to  
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24 852 upward continuation. Inversely, this also means that GOCE has captured 97% of the  
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26 853 amplitude of the gradient signal that exists in orbit altitude.

27 854 The picture changes completely (Fig. 13c) when continuing the gradient  
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29 855 information down to ground level. Here the GOCE model (black dashed curve) and  
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31 856 the GOCE+GRACE combination (solid grey curve) perform practically identical, again  
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33 857 showing the dominance of GOCE compared to GRACE at shorter scales. However,  
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35 858 also here for gravity signals with spatial scales below 80 - 100 km a combination with  
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37 859 terrestrial/airborne gravity information is necessary to achieve acceptable accuracies  
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39 860 (solid black curve), so that the gravity field information can be used for local  
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41 861 geophysical modelling of short-scale density structures.

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### 44 863 *3.3 Not always in focus: the omission error*

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48 865 In order to perform a complete evaluation of the impact of modern satellite  
49  
50 866 missions for deriving density structure of continental margins, one of the most  
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52 867 important aspects is the evaluation of the *omission error*. It results from high-  
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54 868 frequency signals, which cannot be resolved by satellite gravimetry due to the  
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2 869 exponential signal attenuation with altitude. These missing signals of satellite-only  
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4 870 models are an important issue for the determination of near-surface density  
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6 871 variations, but also shallow lithospheric structures.

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8 872 Fig. 14 shows gravity anomaly fields for the South Atlantic region. Fig. 14 a is  
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10 873 based on the GOCO05S model resolved up to its maximum resolution of degree  $n =$   
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12 874 280 (~ 70 km), while Fig. 14 b displays the free-air gravity anomalies based on  
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14 875 EIGEN-6C4 with its maximum resolution of degree  $n = 2160$  (10 km). Comparing  
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16 876 these two figures, the current limits of satellite-only models regarding their spatial  
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18 877 resolution becomes evident, and can only be coped with by combination with  
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20 878 complementary data sources from terrestrial/airborne/shipborne gravimetry, and  
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22 879 satellite altimetry over the oceans, as it was done in EIGEN-6C4. An estimate of the  
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24 880 omission error (Fig. 14 c) for satellite gravity models is given by the difference of  
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26 881 EIGEN-6C4 and GOCO05S, being equivalent to the difference of the Figs. 14 a and  
27  
28 882 b. Evidently, very rough topographic and bathymetric structures, generating high-  
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30 883 frequency gravity field anomalies and steep slopes, cannot be resolved by the  
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32 884 satellite data. However, usually these topographic features are not the main focus of  
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34 885 geophysical modelling and interpretation, but rather sub-surface lithospheric  
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36 886 structures. Therefore, a topographic reduction was applied, using the  
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38 887 RWI\_TOPO\_2015 topographic potential model (Grombein et al., 2015) and thus  
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40 888 taking away the effect of topographic masses up to zero level:  $\delta g_{TOP} + \delta g_{BPL}$ . The so  
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42 889 called “mass reduction effect” was already introduced in Fig. 6. The result is a  
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44 890 significantly reduced omission error (Fig. 14 d).

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46 891 This difference field in Figs. 14 c and d can be considered as errors made  
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48 892 when computing Bouguer anomalies from pure satellite models, which are then  
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50 893 further used for lithospheric modelling. Table 1 gives an overview of the main  
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52 894 statistical parameters of the gravity anomaly fields shown in Fig. 14.

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56 896 Figure 14: Full page width  
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898 **Fig. 14.** Free-air gravity anomalies (mGal) of the South Atlantic region based on  
 899 satellite-only model GOCO05S (a) resolved up to degree 280 , combined gravity  
 900 model EIGEN-6C4 (b) resolved up to degree 2160, omission error of a satellite-only  
 901 model (c) and omission error after reduction of topographic signals (d).

902

903 **Table 1:** Main statistical parameters of gravity fields of the South Atlantic region.

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Gravity field	Figure	min (mGal)	max (mGal)	std.dev. (mGal)
GOCO05S (d/o 280)	14 a	-199.8	116.1	18.8
EIGEN-6C4 (d/o 2160)	14 b	-227.7	453.8	21.3
GOCO05S omission error	14 c	-166.7	415.1	10.8
GOCO05S omission error, topo-reduced	14 d	-143.1	112.8	7.1

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#### 907 **4. Benefits for combined interpretations**

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909 However, in relation with the two key questions asked in Section 2.3  
 910 (processing, quality and secondly availability for interpretations at continental  
 911 margins) we have to respond to them in the light of interpretations of solid Earth  
 912 structures. For example a precise geoid can be used to identify global and deep  
 913 anomalies related to mantle lithosphere and deeper structures. Gravity anomalies,  
 914 being first order radial derivatives of the gravitational potential, are sensitive to gravity  
 915 effects of the entire lithosphere, and in particular to the crustal and upper crustal  
 916 structures and density variations e.g. at active and passive continental margins. As it  
 917 has been shown above (Fig. 14), each combination of satellite gravity data with

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2 918 terrestrial gravity data can be used for all interdisciplinary interpretations techniques,  
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4 919 e.g., “back stripping” in basin modelling at the African continental margin (Dressel et  
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6 920 al. 2015) which also includes thermal subsidence in the reconstruction of the passive  
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8 921 margins through time or 3D modelling of Moho undulations. The new database was  
9  
10 922 also used to reconstruct the Gondwana continent (Braitenberg, 2015). Fig. 15 refers  
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12 923 to the isostatic residual anomaly in the Southern Atlantic. It was calculated by Klinge  
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14 924 (2016) on base of the corresponding formula for “ISA” in Section 2.3 and the EIGEN-  
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16 925 6C4 model also portrayed in Fig. 14 (b).

17 926 Both anomaly maps are rather similar and caused by the main tectonic  
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19 927 features of the South Atlantic: the “highs” which are caused by the Mid Atlantic ridge,  
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21 928 the extended “lows” of the four basins in front of South America (Argentinean and  
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23 929 Brazilian basins) and South Africa (Cape and Angola basins). The hotspot trail (e.g.  
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25 930 Torsvik et al., 2009 among others) is visible in the structure of the SW-NE trending  
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27 931 Walvis Ridge offshore South Africa and the corresponding trace of the Rio Grande  
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29 932 Rise at the western side. To the North of the Romanche Fracture Zone between  
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31 933 Fortaleza in the west and Lagos in the East the Sierra Leone Rise is located. Even  
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33 934 the regions of salt deposits offshore Brazil and West Africa (blueish colors indicating  
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35 935 low gravity values) and the magmatic margins at both sides of the margins (reddish  
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37 936 colors and high gravity values) can be distinguished in the satellite derived fields. The  
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39 937 very short wavelengths in the gravity field correspond to masses that are located in  
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41 938 the crust and lithosphere – they were already mentioned in the sketch of Fig. 8 and  
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43 939 mark places of different density contrast at the margins. Other examples were given  
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45 940 in Bouman et al. (2014), Gutknecht et al. (2014), and Hosse et al. (2014).

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47 942 Figure 15: one and a half page width

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51 944 **Fig. 15.** The isostatic residual field was calculated by Klinge (2016) in the framework  
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53 945 of his MSc thesis. Reference depth  $T_0 = 30\text{km}$  and  $T_e = 20\text{ km}$  (elastic thickness  
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2 946 which was kept constant over the entire area). The figure shows the residual gravity  
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4 947 field in the Southern Atlantic Ocean of the combined EIGEN-6C4 model (Förste et al.,  
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6 948 2014). It correlates well with bathymetric/topographic structures e.g. the Mid Atlantic  
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8 949 Rift (MAR) and portrays also the effect of geological bodies: the positive anomalies in  
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10 950 the area of Windhoek and Buenos Aires. Along the Mid-ocean rift axes positive  
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12 951 anomalies of up to 40 mGal exist. MAR = Mid Atlantic rift.

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15 953 *4.1 The continental margins of the South Atlantic*

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19 955 New light can be shed on the gravity structures of South Atlantic oceanic  
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21 956 margins at regional (Figs. 16, 17, 19 and 20) and more local scales (Figs. 18 and 21).  
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23 957 By the help of these new compiled maps we will show that modern satellite gravity  
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25 958 fields described in Section 4 can support (1) interpretations of the lithospheric  
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27 959 structures in the South Atlantic and its passive margins and (2) provide much more  
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29 960 details in the gravity field than it was showed along the oversimplified profiles of Figs.  
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31 961 7 and 8.

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33 962 With reference to the Fig. 14, the following sequence of Figs. (16 – 21) contains  
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35 963 always the same information for comparative reasons: the two gravity fields based on  
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37 964 the “satellite only” model GOCO05S (a) and the “combined model” EIGEN-6C4 (b),  
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39 965 and additionally Figs. 18 and 19 include the omission errors *without* (c) and *with* (d)  
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41 966 calculated topographic reductions.

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43 967 Table 2 provides a summary of the standard deviations of the gravity fields  
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45 968 shown in these figures. **d/o** refer to the spherical harmonic analysis: to **d**egree and  
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47 969 **o**rders of the expansion.

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51 971 **Table 2:** Standard deviations (mGal) of the gravity fields shown in Figures 16 to 21.

Region	Figure	GOCO05S	EIGEN-6C4	GOCO05S	GOCO05S

		(d/o 280)	(d/o 2160)	omission error	omission error, topo- reduced
Argentinean coast	16	22.4	24.6	9.8	7.1
Brazilian coast	17	15.9	18.8	10.8	6.3
Falkland Bank	18	24.4	25.3	7.3	5.9
African coast	19	19.3	21.5	10.6	7.3
Equatorial African margin	20	19.3	23.9	14.3	9.8
Tristan da Cunha isle	21	6.9	12.3	10.3	4.8

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973

974 Continental gravity edge effects indicate a fast change from positive to  
975 negative anomalies as it is normal for the transition from oceanic to continental crust.  
976 In Fig. 16 the positive anomalies indicate in the offshore area the seaward dipping  
977 reflectors (SDR) which are of magmatic origin (e.g. Blaich, 2011; Franke et al. 2006;  
978 Section 2). The negative anomalies (greenish and blueish colors) offshore are  
979 caused by the sedimentary infill of the margin basins e.g. Colorado and Salado (e.g.  
980 Autin et al., 2013; 2016). Onshore positive anomalies follow W-E trending  
981 topographic features (Salado and Colorado Basin) and in the western continental part  
982 of the maps the topography of the Southern Central Andes.

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984 Figure 16: Full page width

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986 **Fig. 16.** Detailed picture of the free-air gravity field along the Argentinean coast  
987 compiled by the GOCO05S (a) and the EIGEN-6C4 (b) models. The lower two

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2 988 figures indicate the omission errors without (c) and with topographic correction (d).  
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4 989 For more information refer to manuscript. The continental areas are marked by  
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6 990 transparent overlays.

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10 992 In general the series of the following figures will portray similar gravity  
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12 993 anomalies (both magnitude and trend of anomalies). It is no wonder that all EIGEN-  
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14 994 6C4 compilations consist of more structural details than the GOCO05S models which  
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16 995 base on data in the orbit height of some 250 km where small local gravity anomalies  
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18 996 are not detectable.

19 997 Figs. 17 (a) and (b) shows that the “central Atlantic segment” is dominated by  
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21 998 high density rocks which cause positive anomalies. The positive gravity offshore  
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23 999 between 40° - 30° longitude is caused by the “Rio Grande High” which marks the  
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25 1000 most western edge of the hot spot trace which starts at the position of the Tristan da  
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27 1001 Cunha hotspot area. The negative gravity anomalies close to the Brazilian coast are  
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29 1002 caused by negative densities of salt accumulation here (Mohriak, 2014). The SDRs  
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31 1003 with their high rock densities (Section 2.2) of the southern segment are not  
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33 1004 documented here with high resolution; they are too small to be resolved *in detail* as  
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35 1005 we show already in Fig. 3. However, at a larger scale the belt of positive gravity  
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37 1006 marks the area of SRDs quite well.

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40 1008 Figure 17: Full page width

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44 1010 **Fig. 17.** The gravity fields (a) GOCO05S and (b) EIGEN6C4 along the Brazilian coast  
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46 1011 and offshore regions of the “central segment” of the southern Atlantic. Figures  
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48 1012 content is equal to Figs. 14 and 16. However the omission errors are not portrayed  
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50 1013 here.

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2 1015 The resolution of gravity anomaly in the off-shore area of the *Falkland Bank*  
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4 1016 and the Scotia Plate with the Eastern Sandwich trench allows the separation of  
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6 1017 subduction related trench lithosphere, the eastern border of the Sandwich Plate, and  
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8 1018 the southern rim with the Antarctic Plate (low gravity corresponding to blueish colors  
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10 1019 in Fig. 18) from high density rocks of the Scotia Plate and Sandwich Isles (yellow and  
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12 1020 reddish colors in Fig. 18 (a and b). Exactly here in a region with rather complex  
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14 1021 interplay of different plates the resolution of gravity fields before the era of the  
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16 1022 modern satellite missions was extremely low and often hindered a tectonic  
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18 1023 interpretation of lithosphere at medium scale. The Scotia Plate in the center (reddish  
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20 1024 colors in Fig. 18) is clearly separated from the other plates of the region (South  
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22 1025 American plate to the North, Antarctic plate to the South, Scotia plate to center.

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25 1027 Figure 18: One and a half page width26  
27 102828  
29 1029 **Fig. 18.** The gravity field of the Falkland bank and the Scotia plate with the eastern  
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31 1030 South Sandwich trench after the processing of new satellite gravity (GOCO5S: (a)  
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33 1031 and EIGEN6C4: (b). Figures content is equal to Fig. 14.34  
35 103236  
37 1033 Due to the symmetry of evolution of the South American and South African  
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39 1034 margins also the gravity field of the western African margin shows the same general  
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41 1035 features as it was exemplified for the South American margin: in the southern  
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43 1036 segment the magma dominated structures cause small positive gravity anomalies  
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45 1037 and North of the Walvis Ridge the area of salt layers is characterized by negative  
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47 1038 anomalies (blueish colors) in Fig. 19. The SW – NE trending Walvis Ridge separates  
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49 1039 the domains of magmatic material from salt layers. Positive gravity anomalies of the  
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51 1040 ridge clearly indicate the Tristan da Cunha hotspot trace – as it was already  
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53 1041 explained for the western part of the Southern Atlantic. More to the South at the  
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55 1042 South African tip of the Cape a second ridge (Agulhas Ridge) can be identified.  
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2 1043 Onshore at the African continent close to the equator the extended gravity low (-50  
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4 1044 mGal) of the Congo Basin with its thick sediments dominates the gravity picture.

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8 1046 Figure 19: full page width

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11 1048 **Fig. 19.** The gravity field of the African margin in the central and southern segment  
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13 1049 (Fig.15) after the GOCE gravity (a) and the EIGEN-6C4 data (b) processing. The  
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15 1050 series of maps correspond to the displayed formats of figures before; transparent  
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17 1051 overlay mark continental area.

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19 1052 One of the most spectacular fracture zones in the Equatorial and Northern  
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21 1053 South Atlantic connecting Africa and South America is illustrated in Fig. 20 (a and b).  
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23 1054 Fairhead and Wilson (2005) explain the formation of the fracture system with  
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25 1055 processes which were related with the opening of the Central and South Atlantic.  
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27 1056 They state that a differential motion between plate segments was absorbed in the  
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29 1057 Caribbean and West and Central African rift systems. The fracture system developed  
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31 1058 due to the temporal different opening phases of the northern and southern Atlantic.  
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33 1059 Then the two independent spreading centers joined a major shear zone developed  
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35 1060 between West Africa and the northern margin of Brazil. The maps of satellite gravity  
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37 1061 image impressively this major shear zone. The gravity map of the EIGEN6C4 model  
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39 1062 provides a clear and sharp picture of the fractures zones.

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42 1064 Figure 20: One a a half page width

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45 1066 **Fig. 20.** The two gravity fields (GOCO5S (a) and EIGEN (b)) of the Equatorial Atlantic  
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47 1067 Ocean map major transform structures offshore the African margin. Sequence of  
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49 1068 maps corresponds with displayed formats of figures before; transparent overlay mark  
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51 1069 continental area.

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2 1071 The limits of resolution of modern satellite only gravity fields (S models) can  
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4 1072 nicely be demonstrated by the gravity field of the Tristan da Cunha Isle, whose  
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6 1073 gravity field signal is at the edge of the spatial resolution of current satellite gravity  
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8 1074 missions. Fig. 21 shows that although GOCE is able to detect the gravity field signal  
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10 1075 of this island, it is significantly damped. It should be emphasized, that a constraint  
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12 1076 has been applied to the GOCO05S model in the frame of the gravity modelling  
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14 1077 procedure in order to improve the signal-to-noise ratio at higher degrees, i.e. noise is  
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16 1078 filtered out at the cost of damping also the signal. As already discussed in  
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18 1079 Section 3.2, the strength of constraining the solution was optimized on a global scale.  
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20 1080 Therefore, it is not tailored to small regions with strong gravity field signal, where a  
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22 1081 weaker constraint would be preferable due to a larger signal-to-noise ratio in this  
23  
24 1082 region compared to the global average. If the satellite gravity solution were optimized  
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26 1083 for this specific region, it can be expected that in such a regionally tailored solution  
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28 1084 slightly more signal could be retained. The series of Figs. 21 (a) – 21 (d) shows that  
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30 1085 the satellite gravity fields of both GOCE and EIGEN-6c4 are mainly caused by the  
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32 1086 topography of the island. Perfectly seen is the “ring” of negative anomalies in Fig. 21  
33  
34 1087 (b) which can be explained by the flexure of oceanic lithosphere due to isostatic  
35  
36 1088 response of the loaded isle masses. After calculating a topographic correction (Fig.  
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38 1089 21 (d)) an anomaly of some 20 mGal appears. One may speculate if this negative  
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40 1090 anomaly is caused by a mass deficit which is related to the hot spot or to crustal  
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42 1091 thickening..

43 1092 The statistics in Table 2 shows that for such rather small-scale structures the  
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45 1093 amplitude of the omission error can be larger than the signal captured by GOCE.  
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47 1094 However, Fig. 21d shows that significant parts of this high-frequency gravity signal  
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49 1095 result from topography.

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51 1097 Figure 21: one a half page width

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2 1099 **Fig. 21.** The Free Air gravity field of the Tristan de Cunha area after the GOCE (a)  
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4 1100 gravity and EIGEN-6C4 (b) data processing. Sequence of the four maps corresponds  
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6 1101 with displayed formats of figures before; omission error without (c) and (d) with  
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8 1102 topographic correction. The limits of resolution of satellite observations can nicely be  
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10 1103 demonstrated by these gravity fields.

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13 1105 Figure 21a: Small column size

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17 1107 **Fig. 21a.** In addition to what we interpret in Fig. 21 this sketch can explain the typical  
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19 1108 negative ring around the positive anomaly in the last figure: the central mass causes  
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21 1109 the positive anomaly while the sediments around the central mass cause a  
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23 1110 symmetrical gravity low. The extent of deformed crust below the mass crust depends  
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25 1111 on the rigidity of the surrounding crust: the left situation (rigidity R1) demonstrates a  
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27 1112 case with extreme high rigidity, on the right a lower crustal rigidity R2 was assumed.

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30 1114 *4.2 Validation of terrestrial gravity by GOCE data*

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34 1116 The GOCE mission provided not only new geoid and gravity fields, but also  
35  
36 1117 gravity gradient data. Representing the second derivatives of the gravitational  
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38 1118 potential, they are more sensitive to the density structures of the upper crust than  
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40 1119 gravity data normally are. Additionally, gravity gradients provide a better resolution of  
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42 1120 flanks of geological structures, faults, lineaments or even large intrusions at  
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44 1121 continental margins. Gradient data from satellite missions have the potential to  
45  
46 1122 identify the extent of different structures with varying densities even in the lower crust  
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48 1123 (e.g. Ebbing et al., 2013). Panet et al. (2014) even identify correlations of certain  
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50 1124 components of the gravity gradient tensor with lower mantle structures.

51 1125 For gravity interpretations at larger wavelengths the new satellite gravity  
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53 1126 database will help to identify a density zonation and segmentation in horizontal and

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2 1127 vertical directions in the lithosphere. As shown in Section 3.2, GOCE satellite-only  
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4 1128 gradient data provide a spatial (horizontal) resolution in the range of less than 100  
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6 1129 km. However, for many structures - in particular for offshore studies of Applied  
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8 1130 Geophysics - this spatial resolution is not yet sufficient, because smaller crustal  
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10 1131 structures cause anomalies with smaller spatial wavelengths. Therefore, terrestrial  
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12 1132 and airborne gravity measurements have not become obsolete even in the modern  
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14 1133 satellite era, but on the contrary they complement satellite observations on the short-  
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16 1134 wavelength scale where satellite data lack sensitivity.

17 1135 In addition to their very high accuracy in the long to medium wavelength  
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19 1136 range, modern satellite gravity data definitely provide significant added value in the  
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21 1137 geophysical gravity fields processing domain, especially for:

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23 1138 a) Validation of heterogeneous terrestrial gravity data bases and identification  
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25 1139 of outliers;

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27 1140 b) Fill-in of regions with sparse terrestrial data coverage or even data gaps.

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29 1141 As an example of the first task (a), Fig. 22 a shows the difference between a  
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31 1142 terrestrial gravity data base of South America and GOCO05S, resolved up to  
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33 1143 degree 200. To bring them to the same spatial resolution, the terrestrial data have  
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35 1144 been expanded as part of a global  $0.25^\circ \times 0.25^\circ$  terrestrial gravity anomaly grid into a  
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37 1145 spherical harmonic series to degree 720, and then have been cut at degree 200. Fig.  
38  
39 1146 22 a clearly indicates systematic differences, which can be attributed to errors of the  
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41 1147 terrestrial data, because of the globally homogeneous accuracy of less than 1 mGal  
42  
43 1148 for the satellite model. Based on this result, the terrestrial database can be further  
44  
45 1149 screened for outliers and suspicious observations (either of the gravity value itself or  
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47 1150 the attached height information, Hosse et al., 2014 ). This information can then be  
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49 1151 used to derive empirical error estimates of the terrestrial dataset, which can be further  
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51 1152 used for a spatially depending weighting scheme in the frame of a combined solution  
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53 1153 with satellite data (Fecher et al., 2013, 2016). The implicit assumption is that the data  
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55 1154 quality of a terrestrial observation is already reflected in its long-wavelength  
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2 1155 component. By means of this procedure, satellite data get a higher weight in regions  
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4 1156 where a lower accuracy of terrestrial data is suspected.

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8 1158 Figure 22: Full page width

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11 1160 **Fig 22.** (a) Gravity anomaly differences (mGal) between a South American terrestrial  
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13 1161 database (kindly provided by the US National Geospatial-Intelligence Agency) and  
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15 1162 GOCO05S, consistently resolved up to degree 200; b) empirical error estimates  
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17 1163 (mGal) derived from the difference field (after: Fecher et al. (2015), modified).

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21 1165 It should be emphasized, that this validation procedure can be applied in any  
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23 1166 region on Earth. Thereby, a globally uniform satellite gravity model provides the  
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25 1167 chance to estimate a-posteriori the accuracy and reliability even of historical gravity  
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27 1168 data bases (terrestrial, ship- and air-borne), for which only incomplete or even no  
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29 1169 meta-information about the measurement process and conditions is available.

30 1170 Also Bomfim et al. (2013) describe how gradients of the GOCE mission can  
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32 1171 help to estimate systematic errors in terrestrial gravity data in the cratonic basins  
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34 1172 (e.g. Amazon and Parnaiba Basins) in Brazil. Here they calculate an average value of  
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36 1173 terrestrial gravity anomaly and compare its long- and medium-wavelength content of  
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38 1174 the terrestrial gravity with the GOCE gravity field. The analysis shows that where  
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40 1175 terrestrial data are sparse and therefore require an improvement in data coverage,  
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42 1176 satellite data can be substituted in order to represent the gravity field correctly. The  
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44 1177 method they proposed can be used directly to control other gravity databases and  
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46 1178 constitutes as a tool for the quality assessment of terrestrial gravity observations,  
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48 1179 both on- and offshore.

49 1180 The second task (b) also addressed the heterogeneity of terrestrial gravity  
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51 1181 data. There are many regions worldwide where terrestrial data are of very bad quality  
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2 1182 or not available/accessible at all (refer to Fig. 2). In these regions, data from satellite  
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4 1183 gravity missions are the only available data source.

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6 1184         These examples demonstrate, that although satellite missions provide (only)  
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8 1185 long to medium wavelength gravity field data, they are able to provide new gravity  
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10 1186 field information especially in regions where up to now the gravity field has been  
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12 1187 practically unknown. This regional model can then been used as constraint for an  
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14 1188 improved lithospheric density model and the derivation of the state of stress of the  
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16 1189 subduction zone (Gutknecht et al., 2014), clearly demonstrating the added value of  
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18 1190 GOCE especially in these data-critical onshore regions.

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## 20 21 1192 **5. Conclusions and outlook**

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25 1194         The resolution of “satellite only data” up to now does not fall below a resolution of  
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27 1195 80 - 100 km. This is still the borderline for studies presented in the above mentioned  
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29 1196 Sections. In summary we have to say that rather small complex structures related  
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31 1197 e.g. to the “seaward dipping reflectors at passive continental margins (SDR)” with  
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33 1198 small size and density contrast cannot be resolved as separated anomalies in the  
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35 1199 orbit heights of recent satellite missions (e.g. Schaller et al. 2015). For this purpose  
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37 1200 terrestrial gravity data have to be combined with satellite data in gravity models e.g.  
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39 1201 GOCO5C. The interpretations of Section 4.1. showed that gravity and its gradients  
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41 1202 from the modern satellite missions support interpretations at a medium scale – at  
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43 1203 passive continental margins and elsewhere.

44 1204         Modelers of lithospheric structures at continental margins hope that medium scale  
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46 1205 gravity data from the recent and future satellite missions (GRACE and/or GOCE;  
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48 1206 GRACE follow on) can support combined interpretation together with seismological  
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50 1207 and gravity studies. For rather local models (wavelengths of gravity anomalies are  
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52 1208 smaller than 20 km) both resolution and quality of satellite only gravity data have to  
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54 1209 be seen still reluctant until today. However, there is no doubt that combinations with  
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2 1210 terrestrial gravity data bases and satellite gravity with a spatial resolution of 10 – 20  
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4 1211 km can provide detailed insight in the structural behavior of continental margins.

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6 1212 For modelling at continental scales Fig. 2 demonstrated that terrestrial databases  
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8 1213 often are of inhomogeneous distribution (e.g. in South America), just if gravity data  
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10 1214 are sampled over long time-consuming field campaigns with big human efforts: there  
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12 1215 remain big gaps in the data base. They are mainly caused by limited access to the  
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14 1216 terrain in remote areas of the world – high mountains, deserts, swamps and jungle.  
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16 1217 Even more field procedures and technical instrumentation varied over time and  
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18 1218 together with missing other metadata a homogeneous data base can be established  
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20 1219 only with big effort and high costs. Here the new data bases already helped in a  
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22 1220 spectacular way: Hosse et al. (2014) and Gutknecht et al. (2014) replaced the  
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24 1221 incomplete terrestrial gravity data base by homogeneously measured satellite gravity  
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26 1222 and gravity gradient data for lithospheric modelling. New data were applied to the  
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28 1223 calculation of GPE (gravity potential energy), stress distributions and combined  
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30 1224 interpretation of complex geologic structures. Satellite gravity information was also  
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32 1225 used for validation and cleaning of inhomogeneous gravity databases taking the  
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34 1226 benefit of very homogeneous error characteristics and accuracy of global satellite  
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36 1227 gravity data (Hosse et. al, 2014; Bomfim et al., 2013). The high spatial resolution of  
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38 1228 terrestrial gravity *combined* with the homogeneous lower-orbit satellite data leads to  
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40 1229 more detailed and better-constrained lithospheric density models, and hence  
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42 1230 improves our knowledge about structure, evolution and state of stress in the  
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44 1231 lithosphere basing on the consistency in the long-to-medium wavelengths, down to  
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46 1232 10 – 50 km.

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48 1233  
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50 1234 At the beginning (Section 1.1) we mentioned the calculation/recalculation of the  
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52 1235 COB from an integrated interpretation of gravity, magnetic, seismic, electrical  
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54 1236 methods and geology (Torsvik et al., 2009). We did not deal with the calculation of  
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56 1237 COB, however, we think that the combined satellite fields can successfully replace  
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2 1238 the terrestrial gravity data which have to be used in former times. Because we  
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4 1239 analyzed Free Air gravity from the GOCE mission, in Fig. 23 topographic features on-  
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6 1240 and offshore are enhanced. These enhancements indicate clearly the slopes of the  
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8 1241 continental shelf regions of the Southern Atlantic.

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11 1243 Figure 23: Full page width

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15 1245 **Fig. 23.** The third derivations of the gravitational potential and the resulting total  
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17 1246 gradient of the vertical gravity gradient were calculated from the GOCO5C model  
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19 1247 (expansion to degree 720). It provides already good insight into local gravity field  
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21 1248 structures particularly at the margins by the derivations of the vertical gradient ( $V_{zz}$ ).

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24  
25 1250 In the near future complementary information from seismic and magnetics  
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27 1251 could be included in a joint inversion for lithospheric modelling also at passive  
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29 1252 continental margins. ESA's magnetic field mission *Swarm* was successfully launched  
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31 1253 in November 2013 and provides valuable information of the long to medium  
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33 1254 wavelength Earth's magnetic field and its temporal variations with an accuracy on the  
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35 1255 nT- (nano tesla) level (<http://esamultimedia.esa.int/multimedia/publications/BR-302/>).  
36  
37 1256 The value of the mission for the determination of the crustal remanent magnetic field  
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39 1257 will increase in the future, because the three satellites will continuously lower their  
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41 1258 orbit altitudes during mission lifetime, thus also increasing their sensitivity for detail  
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43 1259 magnetic field structures. However, a joint interpretation of remanent magnetic and  
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45 1260 gravity field is only possible in the case of common sources, i.e. similar contrasts in  
46  
47 1261 density and magnetization. In this case Poisson's equation can be applied, which  
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49 1262 links the magnetic and gravity potential fields. *Swarm* is already now a very valuable  
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51 1263 tool to determine the electric conductivity of the Earth's mantle and thus provides very  
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53 1264 important information on the thermochemical and compositional structure of the  
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55 1265 Earth.

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2 1266 Several concepts for future satellite mission constellations to explore the  
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4 1267 Earth's potential fields are under development and investigation. A strong need by  
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6 1268 the user communities was expressed in terms of a joint IUGG resolution adopted at  
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8 1269 the IUGG General Assembly 2015 (IUGG, 2015). The science requirements and user  
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10 1270 needs for a future gravity field mission constellation were consolidated (Pail et al.,  
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12 1271 2015a) also under active participation of the geophysical user community. In addition  
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14 1272 to an improved temporal resolution for the detection of co- and post-seismic  
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16 1273 deformation, an increased spatial resolution together with an improved accuracy will  
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18 1274 shift the capabilities to use satellite-based gravity observations for geophysical  
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20 1275 interpretation in passive continental margins, and elsewhere, to even more small-  
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22 1276 scale structures.

23 1277

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25 1279

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1 **Insights from recent gravity satellite missions in the density structure of**  
2 **continental margins – with focus on the passive margins of the South**  
3 **Atlantic**

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7 **5 Abstract**

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10 6 We focus on new gravity and gravity gradient data sets from modern satellite  
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12 7 missions GOCE, GRACE and CHAMP, and their geophysical interpretation at  
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14 8 passive continental margins of the South Atlantic. Both sides, South Africa  
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16 9 and South America, have been targets of hydrocarbon exploration and  
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18 10 academic research of the German Priority Program SAMPLE (South Atlantic  
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20 11 Margin Processes and Links with onshore Evolution). The achievable spatial  
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22 12 resolution, driven by GOCE, is 70 - 80 km. Therefore, most of the geological  
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24 13 structures, which cause a significant gravity effect (by both size and density  
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26 14 contrast), can be resolved. However, one of the most important aspects is the  
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28 15 evaluation of the omission error, which is not always in the focus of  
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30 16 interpreters. It results from high-frequency signals of very rough topographic  
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32 17 and bathymetric structures, which cannot be resolved by satellite gravimetry  
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34 18 due to the exponential signal attenuation with altitude. The omission error is  
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36 19 estimated from the difference of the combined gravity model EIGEN-6C4 and  
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38 20 the satellite-only model GOCO05S. It can be significantly reduced by  
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40 21 topographic reductions. Simple 2D density models and their related  
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42 22 mathematical formulas provide insights in the magnitude of the gravity effect  
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44 23 of masses that form a passive continental margin. They are contrasted with  
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46 24 results from satellite-only and combined gravity models. Example geophysical  
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48 25 interpretations are given for the western and eastern margin of the South  
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50 26 Atlantic Ocean, where standard deviations vary from 25 – 16 mGal and 21 –  
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52 27 11 mGal, respectively. It could be demonstrated, that modern satellite gravity  
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28 data provide significant added value in the geophysical gravity data  
29 processing domain and in the validation of heterogeneous terrestrial data  
30 bases. Combined models derived from high-resolution terrestrial gravity and  
31 homogeneous satellite data will lead to more detailed and better constrained  
32 lithospheric density models, and hence will improve our knowledge about  
33 structure, evolution and state of stress in the lithosphere.

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### 35 **1. Motivation and the concept of Plate Tectonics**

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37       Geosciences are striving for an interdisciplinary perception to combine  
38 their basic findings in a world embracing synthesis to understand global  
39 processes in the Earth interior and at its surface. Most of these processes are  
40 generally geothermally driven, and it is easy to accept that their origin lies  
41 below the lithosphere, in the Earth's mantle (among others Stadler et al.,  
42 2010). Today, the theory of plate tectonics enables us to draw a coherent  
43 picture of the Earth's lithosphere. Interactions between the plates at their plate  
44 boundaries are responsible for most of the earthquakes that occur here  
45 (among many other publications and websites:

46 <http://earthquake.usgs.gov/earthquakes/?source=sitenav>,

47 <http://www.isc.ac.uk/about/> or <http://geofon.gfz-potsdam.de/>).

48 This paper will review the status of satellite gravity missions and terrestrial  
49 data, as well as global gravity models, fields and gradients derived from them.  
50 It will focus on their accuracy, resolution and the omission error – which is out  
51 of focus of many earth scientists. It is structured as follows: Section 1 defines  
52 continent-ocean transitions and recalls some basics in the context of Plate  
53 Tectonics concept and passive margins in particular. For those readers who

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are not familiar with the interpretation of gravity anomalies at a continental margin in Section 2 the basic concepts are illustrated; this section can be skipped by experts. Further on the focus is set on the question how (satellite) gravity interpretations can help to explore these passive margins (Section 2.3). In the course of this paper we will notice later to what extent the new fields and gradients from recent satellite gravity missions can support research at the passive margins of the South Atlantic (Section 2.4).

In Section 3 we will demonstrate how this new information augmented our view on the density structures of the lithosphere particularly at passive margins of the South Atlantic. At last we describe the benefits for combined interpretations in Section 4, and merge information from both terrestrial and satellite gravity fields.

### *1.1 Short introduction to history of plate margins tectonics*

Considering the history of the Plate Tectonic concepts in the early 1960s, Wegener's view on the „the continental drift“ (Wegener, 1920) began to be accepted after it was refined and confirmed by geophysical observations namely by early seismological studies on deep earthquakes (Wadati, 1929; Benioff, 1954), later by Isacks et al. (1968), Oliver and Isacks (1967) and paleo-magnetic shipborne observations (Hess, 1962; Vine and Matthews 1963 among others). Together with the techniques of radiometric dating (age determination) published first by Boltwood (1907) geophysicists were able to

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77 date the magnetic mid oceanic reversals by precise physical measurements.

78 They helped to get the modern concept of “plate tectonics” fully accepted.

79 It provides the framework for the interpretation of structures, the history  
80 and composition of continental margins. Plate movements and the differences  
81 in density of oceanic and continental crust types led to the structural pattern of  
82 continental margins and result in a tectonic classification of coastlines as  
83 **active** or **passive** margins (among others Wefer et al., 2003). *Active margins*  
84 are typical units of the “Circum-Pacific Ring of Fire” in the Pacific where plates  
85 are converging and coincide with plate boundaries in a subduction zone.

86 These margins are called *active* (e.g. Oncken et al., 2006; Lallemand, 2014)  
87 due to the big variety of tectonic, magmatic and metamorphic processes that  
88 occur here. If continental margins mark only the boundary to the oceanic  
89 portion of the same tectonic plate, they are called *passive margins*. *Passive*  
90 margins are typical of the Atlantic Ocean between Scandinavia and Greenland  
91 (Scheck-Wenderoth and Maystrenko 2008), Iberic peninsula and the East  
92 American coast or between Africa and South America (Blaich et al., 2011 and  
93 many papers within).

94

## 95 *1.2 Passive Margins*

96

97 Although we assume that most of the readers are familiar with the  
98 concept of plate tectonics we will recall briefly some basics. In particular  
99 *passive continental margins* are characterized by a transition of continental  
100 into oceanic crust within the same plate. It emerges from the splitting up of  
101 continents and the following divergent plate drift that forms new oceanic  
102 lithosphere by seafloor spreading at the divergent plate boundary. At the

103 edges of the Atlantic geophysical investigations identify a laterally 10-15 km  
104 thick crustal transition between the thick crust of the continents and the thin  
105 oceanic crust. It is interpreted as relicts of thinned, mafic magmas which  
106 intruded into continental crust. In addition, passive continental margins are  
107 often stretched by intensive fault tectonics. They have a tectonically thinned  
108 continental crust, which is characterized by listric faults and tilted fault blocks.  
109 Tectonic crustal expansion starts at the time of installation of the continental  
110 drift rift system and refines the passive continental margin further (Blaich et  
111 al., 2011). These margins are marked by smooth relief due to tectonic  
112 inactivity and major sediment accumulation. This phenomenon is due to  
113 thermal cooling and sediment loading that led to conditions of subsidence and  
114 sediment accumulation, because the margins move away from the spreading  
115 center. Irregular subsidence and different sediment load often cause the  
116 accumulation of salt diapirs in the sediments of passive continental margins.  
117 The tectonic-sedimentary conditions are also favorable for the formation of  
118 hydrocarbon deposits and large salt deposits (Mohriak, 2014).

119

120 Modern passive margins (Fig. 1) border the oceans formed by the  
121 spreading caused by the break-up of the Gondwana supercontinent (e.g.  
122 Bond et al., 1984). From Figure 1 one can see that the margins of the South  
123 Atlantic Ocean (Cappelletti et al., 2011; Blaich et al., 2011; Mohriak, 2014),  
124 the western Indian Ocean, the Arctic and Norwegian seas (Scheck-Wenderoth  
125 and Maystrenko, 2008; Ebbing et al, 2006; Skogseid et al., 2000), the magma  
126 poor rifted margins of the North and Central Atlantic Ocean (Reston, 2009;  
127 Mohriak, 2014) and the margins of Antarctica are part of this system (Kalberg,  
128 2016). A rising convection cell or a plume in the rifting area caused initial

129 rifting and a regional uplift as well as volcanic activities above or below  
130 oceanic sea level after an initial period of crustal thinning and erosion. Basile  
131 (2015) describe another type of margin: “transform continental margins” by  
132 simple kinematic models of transform faulting which cause among two other  
133 types “passive transform margins”. The satellite gravity picture of the  
134 Equatorial Atlantic Ocean will be shown in Section 4.2, Fig. 20.

135

136 Figure 1: double column

137

138 **Fig. 1.** Continental margins on Earth. Blue lines mark passive continental  
139 margins mainly surrounding the Atlantic Ocean, the Antarctic Seas, and Indian  
140 Ocean; red lines indicate active margins (subduction zones). Continental  
141 margins were taken from from Frisch und Meschede (2005).The underlying  
142 gravity field is the map of “gravity disturbance” calculated on base of the  
143 EIGEN-6C4 model (Förste et al., 2014). Gravity was calculated on a grid of  
144  $0.5^\circ \times 0.5^\circ$ .

145

146 *1.3 Volcanic passive margins*

147

148 These margins present distinctive genetic and structural features, e.g.,  
149 high-rate extension of the lithosphere is associated with catastrophic mantle  
150 melting responsible for the accretion of a thick igneous crust (Geoffroy, 2005).  
151 Typical rifted “magma-dominated” margins are characterized by large volumes  
152 of flood basalts which flow across the continental hinterlands during

153 continental breakup (among others refer to: Hopper et al., 2004; Eldholm et  
154 al., 2000; Gernigon et al., 2006; White and Smith, 2009; O'Connor et al.,  
155 2012). Underlying the extrusive lavas at the continent-ocean transition zone,  
156 these margins exhibit high seismic velocities in the lower crust of some The  
157 deeper crust is characterized by two areas of high seismic velocity (7.2 to 7.4  
158 km/s; Franke et al., 2010)., which are associated with voluminous igneous  
159 rocks intruded into the lower crust.

160 In recent time the question arose in the context of fixing international  
161 boundaries offshore of the continents due to economic interests because e.g.  
162 hydrocarbon exploration moved further offshore e.g. to explore deep water  
163 resources. Toward the definition of this "continent-ocean-boundary, COB" at  
164 passive margins one of the crucial questions is how to define these  
165 boundaries. Torsvik et al. (2009) described how COB for the South Atlantic  
166 margins at both sides can be defined: by the interpretation of seismic,  
167 gravimetric, magnetic, bathymetric and geological information. Any  
168 identification of the COB is also important for the definition of plate boundaries  
169 at the time of break-up which allows the reconstruction of geometry and earlier  
170 position of former continents - e.g. in the reconstruction of Pangaea. It is not  
171 the aim of this paper to recalculate COBs. However we trust that the  
172 processing of gravity field and its second derivatives for recalculation of COB  
173 (Torsvik et al., 2009) will benefit from the increase of high resolution satellite  
174 derived gravity fields. Later we will return to this point in the conclusion  
175 (Section 5) suggesting calculating the 2<sup>nd</sup> derivative of satellite gravity  
176 (derivative of the gravity gradient).

177

## 178 2. Geophysical characteristics of margins

179

180 To investigate both active and passive continental margins almost all  
181 geophysical methods can help to discover their lithospheric structures:  
182 Seismic, potential, electro-magnetic and electrical fields. Recent studies (e.g.  
183 Torsvik et al. 2009) show that processes in the lithosphere are linked to the  
184 dynamic mantle and dynamic processes have an important influence on the  
185 evolution of lithospheric plates, which is manifested in the formation of  
186 dynamic topography (e.g. Flament et al., 2013; Colli et al., 2016). A key  
187 example for this interaction is the opening history of the Atlantic, where  
188 asthenospheric material reaches the surface at the mid-ocean spreading  
189 center. The uplift of hot material of the asthenosphere leads to seafloor  
190 spreading which manifests itself in the spreading anomalies in the magnetic  
191 field. The spreading axes are expressed either by topographic (heights above  
192 sea level, e.g. Iceland) or bathymetric heights (below sea level e.g. Atlantic  
193 ocean ridge). The worldwide offshore stripe pattern of magnetic anomalies  
194 which are located parallel to mid-ocean ridges allow a temporal classification  
195 of developed oceanic crust. The EMAG2 total magnetic field model has a  
196 reasonable resolution of two arc-minutes (Maus et al. 2009). The model was  
197 compiled from ship-, air-borne and satellite data (CHAMP). For the purpose of  
198 interpolation of magnetic field anomalies the oceanic crust ages of Müller et al.  
199 (2008) were employed (Maus et al. 2009). Hydro-carboniferous exploration  
200 targeted most of the passive margins of the world with both refraction and/or  
201 reflection seismic.

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202            Seismic onshore-offshore investigations of a passive continental margin  
203   aim to investigate the transition zone between the oceans and continents, e.g.  
204   SE Atlantic and the African continent (Franke et al., 2006; 2010; Bauer et al.,  
205   2000; Hirsch et al., 2009; Schnabel et al., 2008). These studies reveal the  
206   history and mechanisms of the break-up and its relation to the driving  
207   magmatic processes (plume – crust interactions) in the underlying mantle and  
208   lithosphere. The South American side of the southern Atlantic was target of  
209   multiple seismic investigations since many years (Mohriak, 2014; Blaich et al.,  
210   2011; Franke et al. 2006; Heit et al. 2007; Hinz et al., 1999 and many more) to  
211   investigate the shelf areas offshore, the continent-ocean transition and the  
212   seaward dipping reflectors (SDR) which are characterized by high seismic  
213   velocities and related high densities.

214

### 215    *2.1 The role of gravity field interpretations and prerequisites*

216    Remark:

217            The units used for gravity field values and its gradients are given in  
218            mGal (Milligals) and E (Eötvös) – as it is still common use in  
219            Geophysics and Geodesy. Converting these units into SI-units, there  
220            are:  $1 \text{ mGal} = 10^{-5} \text{ m/s}^2$  and  $E = 10^{-9} \text{ s}^{-2}$ .

221            Rock densities are always given in  $\text{kg/m}^3$ .

222            First, the situation will be analyzed which was typical before the era of  
223            satellite missions which started in the year 2000 with the launch of the  
224            CHAMP satellite (refer to Chapter 4). Modelling and interpretation of the  
225            Earth's gravity field and its derivatives often had to deal with a merge of data  
226            sets with rather different resolution, different age and quality, coverage and

227 wavelength content. This has been documented in a long list of papers  
228 (among many others: Schaller et al., 2015; Bouman et al., 2014; Hosse et al.,  
229 2014; Gutknecht et al., 2014; Köther et al., 2012; Tašárová, 2007). These  
230 papers have shown that lithospheric models of study regions suffer mainly  
231 from two facts:

- 232 (1) Nearly all information for constrained gravity field modelling of the  
233 lithospheric is based on irregularly distributed profiles and
- 234 (2) large data gaps and partly low data quality due to very limited  
235 access and infrastructure.

236

237 The global gravity model EGM2008 (Pavlis et al. 2012) inherited – and still  
238 inherits - these problems related with available terrestrial databases. Pavlis et  
239 al. (2008) described the compilation of the EGM2008 gravity which consists of  
240 different sources: up to a spatial resolution of approximately 140 km GRACE  
241 (Gravity Recovery and Climate Experiment) data have been used, data from  
242 140 km to approx. 10 km spatial resolution are derived from terrestrial and  
243 satellite altimetry and so called “fill-in” data (Section 4.2). Today we know that  
244 the relative weighting of GRACE with respect to the other datasets was too  
245 low, leading to a dominance of terrestrial data errors in spectral regions where  
246 better GRACE data would already have been available.

247

248 Figure 2: double column

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250 **Fig. 2.** This figure portrays the terrestrial data inconsistency which is rather  
251 typical for gravity surveys in remote areas (here across the South American  
252 continent at a swath between 36°S – 42°S from the passive margin in the East

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253 to the active margin in the West). As an example the different sources of  
254 gravity data of the Southern Central American continent are shown together  
255 with the visualization of big data gaps (Tašárová, 2007). The green and red  
256 points in Chile (CH) and Argentina (AR) are stations in the ARANEDA I and II  
257 (University of Chile) datasets; black points in southern Chile and the Arauco  
258 Peninsula are ENAP data (Chilean oil industry); the gray dense network in  
259 Argentina (AR) show the YPF data (Argentine oil company); yellow points:  
260 stations of the MIGRA 2000 dataset, and the MIGRA 2002 data (both  
261 measured by the gravity group at the University Kiel) are shown in blue. The  
262 brown lines offshore denote the ship-borne gravity data profiles from the  
263 German research vessel "Sonne".

264 Fig. 2 shows the situation some 20 years ago at the central South  
265 American subcontinent. For a continental gravity field study the field had to be  
266 compiled from very different data sources often without any meta data  
267 information - e.g. for gravity data which were measured on behalf of oil  
268 companies (grey area in Fig. 2). In other areas, e.g., in the eastern part of  
269 Argentina (yellow dots in Fig. 2) data are missing due to difficult or impossible  
270 access.

271 At the end of this Section we will pose the question how large  
272 lithospheric structures and how big density differences to their surrounding  
273 have to be in order to cause a detectable signal at orbit height of a satellite.  
274 This consideration describes the situation at active and passive margins in an  
275 analogues manner. In Fig. 3 the effects of gravity and gradients are shown for  
276 a simple model. For density contrasts between the values  $\Delta\rho = 10^1 - 10^3$   
277  $\text{kg/m}^3$  Gutknecht et al. (2011) calculated gravity and gradients at the GOCE  
278 satellite orbit height of 255 km resulting from a sphere with minimum

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279 diameters between  $d = 20 - 200$  km ( $b = 10 - 100$  km, refer to Fig. 3) to be  
280 tangent to the Earth surface). Within the range of the assumed parameters the  
281 minimum diameter required to produce signal differences of  $1 \times 10^{-5} \text{ ms}^{-2}$  and  
282  $12 \times 10^{-12} \text{ s}^{-2}$  at orbit height. This is rather close to the expected accuracies of  
283 the gravity and vertical gravity gradient of the GOCE mission (recent values  
284 are: 0.45 mE for GOCE-only, 0.2 mE for GOCE+GRACE models, cf. also Fig.  
285 13). Fig. 3 conveys that a structure with a diameter of some 45 km and a  
286 density contrast of  $240 \text{ kg m}^{-3}$  could be detected in satellite gravity at orbit  
287 height. If the diameter of the model sphere increases to 90 km, its density  
288 contrast should not be less than  $33 \text{ kg m}^{-3}$ .

289         The simple model described above fits rather well the dimension of the  
290 Jurassic arc batholiths at the Northern Chilean continental margin with  
291 diameters of roughly 60 – 120 km (Sobiesiak et al., 2007). This supports the  
292 idea that batholithic structures e.g. intrusions at continental margins, can be  
293 detected using data of the modern satellite missions– both gravity and  
294 gradients.

295

296 Figure 3: one and a half page width

297

298 **Fig. 3.** Gravity and gradient signal caused by a minimum diameter of a sphere  
299 with given density contrast in the orbit height of 255 km (Gutknecht et al.,  
300 2011). The thick solid and dotted lines represent gravity signals of 1 and 10  
301 mGal at orbit height. The thin dashed and dash-dotted lines represent  
302 gradients of 12 and 1000 mE, respectively. The grey shaded area shows  
303 results which are based on a possible combination of geometry and density  
304 parameters of the causing mass anomaly.

305

306 *2.2 Gravity anomalies and isostasy*

307

308 A first look at Fig. 4 provides already helpful information on the gravity  
309 gradients of the Earth at the continental margins. The figure bases on the  
310 evaluation of gravity field maps from the recent gravity missions. A complete  
311 Bouguer anomaly (corrected by the effect of elevation, spherical slab,  
312 topography both on- and offshore) was used to calculate the first derivative of  
313 the field – the dip curvature of gravity.

314

315 Figure 4: Full page width

316

317 **Fig 4.** Global horizontal gravity gradients. Deep blueish colors mark regions  
318 on Earth where the dip curvature (horizontal gradient) of the worldwide  
319 Bouguer gravity field (EIGEN-6C4) is small or even zero. The more reddish  
320 the colors are the steeper the gradients. The strongest gradients are observed  
321 at the active continental margin of Central South America. On the contrary at  
322 most of the passive continental margins (Fig. 1) the gradients are rather small.

323

324 Dark blueish colors indicate rather weak dip curvature which means that  
325 horizontal gravity gradients are small, on the contrary reddish tones point to  
326 strong dip curvature and therefore strong gradients. The active continental  
327 margins in the area of the Circum-Pacific are generally marked by reddish

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328 colors – strong gradients – which are related with extended density variations

329 in crust and mantle in the transition from oceanic to continental margins.

330 Mostly low gradients – light blueish colors - are typical for passive continental

331 margins. In order to understand this relationship two questions arise:

332 (1) How was the gravity *anomaly* calculated and how big is its magnitude of

333 the anomaly caused by the mass distribution (both topography/bathymetry

334 and densities) at continental margins and

335 (2) How was the *field* observed and what is known about accuracy and

336 homogeneity of gravity field observations?

337 With regard to the first question one has to consider that gravity

338 observations at the Earth's surface and the Earth near space vary from the

339 theoretical "normal" field value. Their magnitude is caused not only by the

340 latitude effect but by elevation of observational points, the density of cap

341 beneath the station in the underground and the topographic masses in the

342 neighborhood. After correction of earth tidal effects and air pressure

343 variations the remaining time invariable parts of the measured signal are:

344	Normal gravity	$\gamma_0$
345	Effect of topographic masses	$\delta g_{TOP}$
346	Bouguer slab	$\delta g_{BPL}$
347	Elevation effect (free air term)	$\delta g_{NIV}$
348	Effect of the crustal root	$\delta g_R$
349	Known mass inhomogeneities	$\delta g_{GEOL}$

350

351 Common representations of gravity measurements in maps and profiles  
 352 depend on what has been calculated as stated in the table above and added  
 353 as corrections to the measured gravity field values. From all measured gravity  
 354 values the normal gravity (in the height and position of the observable) is  
 355 subtracted, and therefore the term “gravity anomaly” is defined. This is simply  
 356 the difference between the observable values to “normal” gravity. In  
 357 Geophysics we distinguish mainly between three anomalies Free air- (FA),  
 358 Bouguer- (BA) and isostatic anomaly (ISA) which define gravity in the height  
 359 of the observation (e.g. Li and Götze, 2001; Hackney and Featherstone, 2003;  
 360 Naudy et al., 1965; LaFehr, 1991 and many others).

361 With  $\delta g_R$  we describe the effect of a mountain root and define Free Air,  
 362 Bouguer and isostatic anomalies such as:

$$363 \quad FA = \Delta g'_0 = g_{obs} + \delta g_{NIV} [+ \delta g_{TOP}] - \gamma_0$$

$$364 \quad BA = \Delta g''_0 = g_{obs} + \delta g_{NIV} + \delta g_{TOP} + \delta g_{BPL} - \gamma_0$$

$$365 \quad ISA = \Delta g_{ISA} = g_{obs} + \delta g_{NIV} + \delta g_{TOP} + \delta g_{BPL} + \delta g_R - \gamma_0$$

366

367 Calculations of the individual correction terms depend on the objective of the  
 368 survey and are variously complicated to handle.  $g_{obs}$  denotes the measured  
 369 gravity field value. The calculation of the topographical reduction ( $\delta g_{TOP}$ )  
 370 requires precise knowledge of the terrain and surface near densities and  
 371 today it is most likely calculated by the aid of digital elevation models (DEM),  
 372 among others refer to (Holzrichter, 2013 and Szwillus and Götze, 2016); it  
 373 requires the use of a computer and digital elevation data. The effect of  
 374 topographic masses is normally calculated in a surrounding circular area of 50  
 375 km up to 167 km (e.g. La Fehr, 1991; Hinze et al., 2006). However, Mikuška et

376 al. (2006), Szwillus and Götze (2016), Szwillus et al., (2016) point to “long  
 377 distant relief effects” and propose the calculation of *all* topographic masses on  
 378 Earth. The gravity effect of a Bouguer slab with a thickness which is defined  
 379 by the difference between the physical station height and the reference level  
 380 (normally the geoid) should always be calculated by a spherical cap (e.g.  
 381 Baeschlin, 1948). For smaller areas (with a reduction radius  $R < 20$  km) and  
 382 moderate terrain roughness the Bouguer slab can easily be calculated by:

$$383 \quad \delta g_{\text{BPL}} = -2 \pi G \rho (H_S - H_B)$$

384 with:

385  $G$  = Gravitational constant ( $6.672 \cdot 10^{-11} \text{ m}^3 \text{ s}^{-2} \text{ kg}^{-1}$ ),

386  $\rho$  = crustal density ( $2\,670 \text{ kg/m}^3$ ); offshore: rock-equivalent  
 387 density

388  $H_S$  = station height; offshore: ocean depth  $H_D$ ,

389  $H_B$  = reference height (usually it is referred to the geoid).

390

391 In the last step the free air effect  $\delta g_{\text{NIV}}$  is calculated by use of the  
 392 “normal gradient” ( $0.3085 \text{ mGal/m}$ ):

$$393 \quad \delta g_{\text{NIV}} = 0.3085 \cdot (H_S - H_B) \text{ mGal/m}$$

394 It has to emphasized that the above used constant gravity gradient for large  
 395 scale investigations has to be replaced by equivalent values of derivation of  
 396 closed mathematical expression of the normal gravity which is always latitude  
 397 and height dependent. It was also recommended to calculate an “atmospheric  
 398 correction” (Wenzel, 1985) in order to eliminate long wavelengths errors from  
 399 the observed gravity field.

400 Note: the terms of “flat” Bouguer slab in  $\delta g_{\text{BPL}}$  and the “constant vertical  
 401 gradient” in  $\delta g_{\text{NIV}}$  are in the above formulas are used here for simplification  
 402 only. Modern satellite gravity field processing at large spatial scales requires a  
 403 spherical cap calculation and the consideration of latitude and height  
 404 dependent calculation of the vertical gradient.

405 Special emphasis has to be put on the situation in oceanic regions. If  
 406 we assume that station heights are equal to geoidal heights ( $h = 0$  m) Bouguer  
 407 and Free Air anomaly is equal due to:

$$408 \quad \text{BA} = \text{FA} + \delta g_{\text{BPL}} = \text{FA} + [-2 \pi G \rho (H_S - H_B)]$$

$$409 \quad \text{with: } H_S - H_B = 0 \text{ m;}$$

410 it follows that  $\text{BA} = \text{FA}$ , in case the FA is already corrected by  $\delta g_{\text{TOP}}$  which  
 411 contains gravity effect of masses at the ocean floor.

412 In oceanic areas the slab density in the Bouguer slab correction term  
 413 must be modified due to the known water depth “D” and the difference in  
 414 water- and rock density (Fig. 5). If we assume a crustal rock density of 2 670  
 415  $\text{kg/m}^3$  and a water density of 1030  $\text{kg/m}^3$  the resulting density for calculations  
 416 of offshore Bouguer anomalies is  $\rho^* = -1\,640 \text{ kg/m}^3$ .

417 This results in:

$$418 \quad \text{BA} = \text{FA} + \delta g_{\text{BPL}} = \text{FA} + 2 \pi G \rho^* (H_S - H_D)$$

419

420 Figure 5: Full page width

421

422 **Fig. 5.** Illustration for describing the calculation of Bouguer anomalies on  
 423 continents (A), at the ocean (B) and from satellite (C). (A): On land the  
 424 reduction density  $\rho$  is commonly taken as 2 670  $\text{kg m}^{-3}$ . The effect of

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425 topography is already removed. The thickness of Bouguer slab equals the  
426 station height ( $H_s$ ). (B): In contrast the reduction density at sea is  $-1\,640\text{ kg m}^{-3}$   
427  $\text{m}^{-3}$ . It is the difference between the sea water density of  $1\,030\text{ kg m}^{-3}$  and the  
428 rock density of  $2\,670\text{ kg m}^{-3}$ ; thickness of the slab now is equal to the different  
429 water depths (D). (C): Calculating a Bouguer anomaly in case of satellite  
430 gravity a “mass correction” is calculated:  $\delta g_{\text{Mass}} = \delta g_{\text{Top}} + \delta g_{\text{BPL}}$ .

431

### 432 *2.3 Gravity at passive continental margins*

433

434 In Fig. 6 a very simple Airy-Isostasy model of a passive continental margin is  
435 shown. The continental crust is much thicker than the oceanic crust and above  
436 the oceanic crust there is the water cover of a few 1000 meters. The specific  
437 geometry of the „crust – mantle – water cover“ constellation plays an  
438 important role on the trend of the gravity field here. Because of the fact that in  
439 the example of Fig. 6 there is no topography, the continental margin remains  
440 in an isostatic equilibrium, and one can assume that no Free Air anomaly  
441 exists.

442

443 Figure 6: One and half page width

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445 **Fig. 6.** Airy isostatic model at a continent – ocean transition. Notice the thick  
446 continental crust and the thin oceanic crust at a passive continental margin.  
447 Crust and mantle densities are simplified. Reasonable contrasts which cause  
448 large gravity anomalies are related to the water – continent density contrast

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449 and crust – mantle density at the continent of approx.  $430 \text{ kg m}^{-3}$ . Refer to text  
450 for more information.

451

452           However, in Fig. 7 (A) strong gravity anomaly results from the same  
453 density model (Fig. 7C) which was shown in Fig. 6. The Fig. 7(A) shows the  
454 modelled anomaly only for water effect of gravity: related to a model of a “half-  
455 indefinite” plate the resulting anomaly is negative and is caused by a strong  
456 gradient. Fig. 7 (B) on the other side contains model results which have been  
457 done only for the oceanic mantle: now the anomaly is positive and it has a  
458 gentle increase because its position is far deeper. Finally Fig. 7(C)  
459 demonstrates how the total Free Air anomaly results from the superposition of  
460 both effects: the Free Air anomaly is zero in the continental area and over the  
461 ocean as well. However, exactly above the margin the gravity field is  
462 characterized by a maximum and a minimum that follows. This distribution is  
463 a so called „edge“ or boundary effect of the Free Air anomaly and is effected  
464 by the difference of the steep gradients in the model.

465

466 Figure 7: one and a half page width

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468 **Fig. 7.** The principal effects on the gravity field at continental margins have  
469 equal gravity magnitudes but different gradients. In (A) the water effect causes  
470 a steep gradient and in (B) the density surplus of the oceanic mantle is a  
471 deeper seated effect which causes only a gentle gradient. In (C) it is explained  
472 that a Free Air anomaly at a continental margin is caused by both a negative

473 and positive “edge effect” due to the superposition of contributions that have  
 474 equal magnitudes but different gradients.

475

476 The trends of a Free air and Bouguer anomalies are shown in Fig. 8.

477 Here modelling again gets use of a „half indefinite“ thin plate for the offshore

478 area (water). It is “zero” over the continent zero and over the oceanic area

479 “positive” ( $\rho_w = 1030 \text{ kg/m}^3$ ). The half of the whole Free Air anomaly maximum

480 is accomplished exactly over the edge of the continent.

481

482 Figure 8a-b: Full page width

483 Figure 8c: One and a half page width

484

485 **Fig. 8.** Free Air anomaly and Bouguer anomaly at continental margins which

486 is also in an isostatic balance. (A) The absolute value of the excess mass

487  $|\Delta m|$  is equal to the absolute value of the deficient mass  $|\Delta m|$ . Therefore

488 the integral of gravity change with respect to the x-coordinate is zero:  $\int \Delta g \, dx =$

489 0. (B) The Bouguer correction at the ocean (see Fig. 3) applied to the Free Air

490 anomaly in (A) yields the general form of the Bouguer anomaly at passive

491 continental margins. (C) The “geological” mass inhomogeneities at the

492 continental margins (seaward dipping reflectors, magmatic remnants, salt

493 structures etc.) cause rather local gravity anomalies which superimpose the

494 regional gravity wavelengths – which are effected by the “simple” structures in

495 (A) and (B).

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496           The gravity fields in Figs. 7 and 8 are caused by the over simplified  
497 density structure at the “modelled margin” in Fig. 6. In the real world these  
498 margins show a rather complicated picture of gravity distribution due to mass  
499 inhomogeneities in the Earth’s crust and lithosphere (and even in the mantle)  
500 which are the results of the long-lasting history of the breakup of the  
501 Gondwana supercontinent. This becomes quite clear if looking at the  
502 processed gravity fields which are shown in the series of figures (Fig. 14  
503 through 21) in Section 4. Their interpretation in terms of regional tectonic and  
504 distribution of rock densities will help to provide a rather detailed insight into  
505 the causing structure (geometry) and density distribution of the passive  
506 margins in the South Atlantic region.

507           Most aspects of the calculated anomalies, both Free Air and Bouguer,  
508 which were discussed before, are typical for nearly all of the continental  
509 passive and active margins on Earth. In the next Section the focus will be set  
510 on the situation in the Southern Atlantic between Africa and South America –  
511 the research area of the German Priority Program 1375 “SAMPLE” of the  
512 German Science Foundation – DFG (<https://www.sample-spp.de/>). The  
513 acronym stands for “**S**outh **A**tlantic **M**argin **P**rocesses and **L**inks with onshore  
514 **E**volution”. In this interdisciplinary project the primary research areas are the  
515 mantle dynamics and magmatic processes, the lithospheric structure,  
516 deformation processes and rifted margin formation, the post-rift topographic  
517 evolution and many more. In the following Section we will concentrate on this  
518 part of the world because a big variety of data and information is available to  
519 responds to one of the key questions – how modern satellite missions can  
520 contribute to the interpretations and to the understanding of the transition from  
521 continental to oceanic lithosphere.

522

523 *2.4 Focus region: South Atlantic passive margins*

524

525 To study deeper structures and the overall evolution of conjugate  
526 passive continental margins of the South African and South American  
527 continents 3D structural models have been designed and evaluated by  
528 SAMPLE scientists and their international partners: They constructed detailed  
529 density models at both sides of the Southern Atlantic Ocean and a rather  
530 preliminary density model for the oceanic part in course of a master thesis  
531 (Klinge, 2016). These models are constrained by information and data from  
532 boreholes, refraction and reflection seismic, seismological tomography and  
533 potential field data – mainly gravity field data. Geophysical fields and  
534 observations map geometry and distribution of physical properties of the  
535 transitional structures of both crust and lithospheric mantle. Model results (Fig.  
536 9) show (Maystrenko et al., 2013; Autin et al., 2016) that basin centers at the  
537 western (Argentinean) side are oriented west-east and therefore oblique to the  
538 mid ocean rift axis while at the other (African) side basin centers extend  
539 parallel to the ocean rift in north-south direction.

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541 Figure 9o: One and a half page width

542 Figure 9u: One and a half page width

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544 **Fig. 9.** For illustration this figure portrays a 3D density model of the SW  
545 African continental margin (above, modified after Maystrenko et al., 2013),  
546 and the density structure at the Argentinean side (below, Autin et al., 2016).

547

548           Apart from these structural differences both sides of the Southern  
549 Atlantic reveal similar distributions in temperature and density. Small  
550 thicknesses and density modifications in the lithospheric mantle point to small  
551 lateral variations of heat transfer into the overlaying crust. However, more  
552 relevant for the crustal heat field are lateral thickness changes of the  
553 crystalline crust which produce the bigger part of radiogenic heat. This  
554 contrasts observations and modelling results at passive continental margins in  
555 the area of the Northern Atlantic (Scheck-Wenderoth and Maystrenko, 2008).  
556 They found that the oceanic part of lithospheric mantle is much thinner and  
557 characterized by smaller densities which cause higher temperatures in the  
558 upper crust of the ocean.

559           To contrasting large scale paleostress fields on the correlating margins  
560 of the South Atlantic Salomon et al. (2014) point to in their studies of the  
561 South Atlantic. They asked themselves “how passive” continental margins  
562 across the globe currently are. Following the results of several other studies  
563 these margins experience a variety of stress states and undergo significant  
564 vertical movements, as they were deduced from studies of paleo-stresses at  
565 both sides of the Southern Atlantic. Here, the bounding continents consist of  
566 very different recent geological histories: Africa experiencing continental rifting  
567 whereas South America is influenced by subduction on the Pacific side. It is  
568 not clear to what extent the Atlantic continental margins are subject to the

1 569 same stresses and vertical motions as the main continents. Their results show  
2 570 that the tectonic evolution of the continental margins of the South Atlantic is  
3  
4 571 not only passive and that both margins vary significantly in structural style and  
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7 572 stress fields, indicating that variable plate boundary forces play a major role in  
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9 573 margin evolution. In Fig. 10 we show the situation at the S-American and S-  
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11 574 African margin with reference of the paleo-stress field, as it was published by  
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13 575 Salomon et al. (2014). Their findings demand careful modelling of both  
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15 576 continental margins and a geophysical database which is able to resolve even  
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17 577 very small modifications of physical parameters and their structures; refer also  
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19 578 to Fig 9 (a) and (b) and the 3D density modelling of lithospheric by  
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21 579 Maystrenko et al. (2013) and Autin et al. (2016).  
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37 **Fig. 10.** The sketch (Salomon et al., 2014) portrays an E-W cross-Section  
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39 584 between South Africa and South America which summarize the situation of  
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41 585 their obtained paleo stresses. It shows that the African margin is controlled by  
42  
43 586 extension while compression characterizes the situation at the South  
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45 587 American side. Salomon et al. (2014) explained the extensional state in the  
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47 588 east by the existing “African superplume” and the compression in the west by  
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49 589 the Andean subduction zone.  
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58 591 Novel satellite gravity missions aim at a breakthrough in recovering the  
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60 592 Earth’s gravity and magnetic fields, their gradients as well as their temporal  
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593 variation. Static anomalies in potential fields (refer to Figs. 14 through 21) are  
594 caused by irregular mass distribution on and within the Earth, temporal  
595 variations of the gravity field are associated with mass transport processes in  
596 the Earth system, such as dynamic processes on the Earth's surface, in  
597 lithosphere and upper mantle.

598

### 599 **3. Modern satellite gravity missions**

600

601           The launch of the first generation of satellite gravity missions (Fig. 11) has  
602 revolutionized our knowledge of the global Earth's gravity field and its temporal  
603 changes. The German CHAMP (Challenging Minisatellite payload; mission period  
604 2000-2010; Reigber et al., 2002; <http://op.gfz-potsdam.de/champ/>) mission, the  
605 US/German GRACE (Gravity Recovery and Climate Experiment; mission period  
606 2002-ongoing; Tapley et al., 2004; <http://www.csr.utexas.edu/grace/>) mission, and the  
607 European GOCE (Gravity field and steady-state Ocean Circulation Explorer; mission  
608 period 2009-2013; Drinkwater et al., 2003;  
609 [http://www.esa.int/Our\\_Activities/Observing\\_the\\_Earth/The\\_Living\\_Planet\\_Programm](http://www.esa.int/Our_Activities/Observing_the_Earth/The_Living_Planet_Programme/Earth_Explorers/GOCE/ESA_s_gravity_mission_GOCE)  
610 [e/Earth\\_Explorers/GOCE/ESA\\_s\\_gravity\\_mission\\_GOCE](http://www.esa.int/Our_Activities/Observing_the_Earth/The_Living_Planet_Programme/Earth_Explorers/GOCE/ESA_s_gravity_mission_GOCE)) operated by the European  
611 Space Agency (ESA), improved significantly the coverage and availability of high  
612 resolution and precisely measured data. These gravity missions are the only  
613 measurement technique that can directly observe mass changes on a global scale,  
614 and thus they provide a unique observation system for monitoring mass transport in  
615 the Earth system. For modern magnetic field observation, apart from the CHAMP  
616 mission (2000-2010), with ESA's three SWARM satellites that have been successfully  
617 launched in November 2013 also gradients observations have become available  
618 (<http://esamultimedia.esa.int/multimedia/publications/BR-302/>).

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620 Figure 11: one and a half page width

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622 **Fig. 11.** Satellite gravity missions CHAMP (left), GRACE (center) und GOCE (right).

623 (Sources: CHAMP: GFZ Potsdam, GRACE: NASA, GOCE: ESA Medialab)

624

625 In these missions, three measurement concepts are implemented:

626 1. Observation of orbit perturbations of low-flying satellites due to the varying  
627 gravitational attraction, by Global Positioning System (GPS), with an accuracy  
628 of 2-3 cm. Non-gravitational forces acting on the satellite, such as drag of the  
629 residual atmosphere or solar radiation pressure, are measured by an  
630 accelerometer and corrected for in the frame of the gravity field modelling.

631 This satellite tracking technique between a low Earth orbiter (LEO) and high-  
632 flying GPS satellites is called satellite-to-satellite tracking in high-low mode  
633 (SST-hl), and is implemented in all three missions CHAMP, GRACE and  
634 GOCE. It is the primary measurement technique of CHAMP.

635

636 2. Observation of orbit differences (ranges) and their temporal change (range  
637 rates) between two LEO satellites. This satellite-to-satellite tracking in low-low  
638 mode (SST-ll) concept is realized by the GRACE mission. It consists of two  
639 identical satellites following each other on the same orbit with an average  
640 distance of 200 km. The inter-satellite ranging is performed by means of a K-  
641 band microwave system with micrometer accuracy, and shall be done by laser  
642 interferometry in future gravity missions in order to further increase the  
643 ranging accuracy.

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645 3. Observation of acceleration differences on very short baselines (satellite  
646 gravity gradiometry, SGG), representing second order derivatives of the  
647 gravitational potential  $V$  in all three spatial directions. This concept was

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648 applied by the GOCE mission. Its core measurement, the gravity gradiometer,  
 649 was composed of 6 accelerometers fixed on 3 orthogonal axes symmetrically  
 650 around the center of mass of the satellite, measuring acceleration differences  
 651 on very short baselines of only half a meter in all three spatial dimensions.

652 The achievable performance of satellite gravity missions depends mainly on the  
 653 observation technique and the orbit altitude. Fig. 12 shows the performance of  
 654 different mission concepts in terms of the degree error median, which describes the  
 655 average signal or noise amplitude at a certain degree  $n$  of the spherical harmonic  
 656 series expansion of the gravitational potential  $V$  in spherical coordinates (with radius  
 657  $r$ , co-latitude  $\vartheta$ , longitude  $\lambda$ ):

$$V(r, \vartheta, \lambda) = \frac{GM}{R} \sum_{n=0}^{N_{\max}} \left(\frac{R}{r}\right)^{n+1} \sum_{m=0}^n \bar{P}_{nm}(\cos \vartheta) [\bar{C}_{nm} \cos(m\lambda) + \bar{S}_{nm} \sin(m\lambda)]$$

660 where  $G$  is the gravitational constant,  $M$  the mass of the Earth,  $R$  the mean Earth  
 661 radius,  $\bar{P}_{nm}$  the fully normalized Legendre polynomials of degree  $n$  and order  $m$ , and  
 662  $\{\bar{C}_{nm}, \bar{S}_{nm}\}$  the corresponding (Stokes) coefficients (e.g Torge, 2001). Therefore, the  
 663 degree error median describes the achievable gravity field accuracy at a certain  
 664 spatial (half) wavelength  $\lambda$ . The wavelength  $\lambda$  is linked to the harmonic degree  $n$  by  
 665

$$666 \quad \lambda = 20\,000 \text{ km}/n$$

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 668  
 669 As an example, a harmonic degree of  $n = 200$ , which was the minimum target  
 670 resolution for the GOCE mission, corresponds to a spatial wavelength of  $\lambda =$   
 671  $20\,000 \text{ km}/n = 100 \text{ km}$ .

672 As a reference, the stippled black curve in Fig. 12 shows the gravity field  
 673 signal itself. Correspondingly, the cross-over point of a mission performance curve

674 with the black stippled curve indicates at which harmonic degree the signal-to-noise  
675 ratio is '1'.

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677 Figure 12: One and a half page width

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679 **Fig. 12.** Absolute gravity signal and error estimates of different observation concepts  
680 as a function of the harmonic degree  $n$  (bottom axis) and spatial wavelength  $\lambda$  (top  
681 axis).

682

683 From the orbit information (SST-hl) only the long-wavelength features of the gravity  
684 field can be extracted. Although this observation type is not a direct gravity field  
685 functional, it can be interpreted as disturbing acceleration acting on the orbit, and  
686 thus the first order spatial derivative of the gravitational potential  $\partial V/\partial x_i$ . As a  
687 representative of this measurement concept, the grey dot-and-dashed line curve  
688 shows the performance of the CHAMP-only model AIUB-CHAMP 03S (Prange,  
689 2011), which is based on 8 years of CHAMP kinematic orbit data.

690 The grey curve shows the performance of the recent GRACE-only model  
691 ITSG-Grace2014 (Mayer-Gürr et al., 2014), which is based on almost 11 years of K-  
692 band inter-satellite ranging data following the SST-II concept (and supported by SST-  
693 hl in the very low degrees). Compared to CHAMP, the superior measurement  
694 principle of SST-II results in a significantly better accuracy in the low to medium  
695 degree range as well as a higher spatial resolution. This can be explained by the fact  
696 that the SST-II concept can be interpreted as a measurement of acceleration  
697 differences on long baselines of about 200 km. The excellent performance of GRACE  
698 in this spectral range makes this mission sensitive to the tiny temporal variations of  
699 the Earth's gravity field, which are 4 - 5 magnitudes smaller than the static signal.

700 The black solid curve shows the performance of GOCE, represented by the  
701 GOCE-only model GOCE-TIM-R5 (Brockmann et al., 2014). It is mainly based on the

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702 measurement technique of SGG and again SST-hl in the low degrees, because SGG  
703 alone (green curve) is weak in this spectral range due to the specific noise  
704 characteristics of the gravity gradiometer instrument. Measuring acceleration  
705 differences on very short baselines of about half a meter, which approximate second  
706 order derivatives of the gravitational potential  $\partial^2 V/(\partial x_i \partial x_j)$ , enables a further increase  
707 of sensitivity for high-frequency signals. GOCE starts to become superior over  
708 GRACE approximately at degree  $n = 115$ .

709

### 710 *3.1 Global Gravity Field Models*

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712 Gravity field models including GOCE data from the complete mission period  
713 are meanwhile available. While the model GOCE-TIM-R5, which is based on the  
714 time-wise approach (Pail et al., 2011), is based purely on GOCE data, GOCE-DIR-  
715 R5, which is based on the direct method (Bruinsma et al., 2014), contains also  
716 GRACE and satellite laser ranging (SLR) data. Further satellite-only models are, e.g.,  
717 EIGEN-6S2 (Rudenko et al., 2014), or the S-models of the GOCO series (Pail et al.,  
718 2010). The maximum degree of expansion of these models is driven by the resolution  
719 of GOCE, and varies from  $n = 280$  to  $300$ , corresponding to about  $70$  km spatial  
720 wavelength. This makes clear that all medium scaled geological structures at  
721 continental margins and elsewhere in the world which cause a significant gravity  
722 effect can be detected (resolved) in the GOCE gravity field.

723 Combination models (notice the “C” in the field identifier) including also  
724 terrestrial, air- and shipborne as well as altimetric gravity are, e.g., the already  
725 mentioned pre-GOCE model EGM2008 (Pavlis et al. 2012), EIGEN-6C4 (Förste et  
726 al., 2014), and GOCO05C (Fecher et al., 2013, 2016). These models extend the  
727 spatial resolution beyond degree  $2000$  (which corresponds to  $10$  km wavelength).  
728 However, it should be noticed that there are many regions with sparse and/or low-  
729 quality terrestrial data, where it has to be questioned if such a high resolution is

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2 730 justified. This holds for many areas worldwide, e.g., the Central Andes in South  
3 731 America and also for the passive continental margins of the South Atlantic.

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6 733 *3.2 Products for use in Earth sciences interpretation*

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11 735 Specifically regarding GOCE-related data, modelers and other users have the  
12 736 choice among basically three representations of gravity field products:

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16 738 *3.2.1 Spherical harmonic coefficients*

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22 740 The most commonly used representation of the global gravitational potential  $V$   
23 741 is its series expansion into spherical harmonics (Section 3). There corresponding fully  
24 742 normalized spherical harmonic (Stokes) coefficients  $\{\bar{C}_{nm}, \bar{S}_{nm}\}$  represent the model  
25 743 parameters, and are usually the target quantity when deriving the model from the  
26 744 original gravity field data. The advantage of using this representation is that it can be  
27 745 considered as a weighted average of the original measurement data, so that the  
28 746 original noise level is significantly reduced due to this averaging. Based on the set of  
29 747 spherical harmonic coefficients any arbitrary gravity anomaly can be derived at the  
30 748 Earth's surface or at any height in outer space. All the global gravity models  
31 749 discussed above are given in this parametrization.

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35 751 *3.2.2 Original gravity gradients along the satellite's orbit*

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40 753 In principle, also the gravity gradient time series for all six tensor components  
41 754 measured along GOCE's satellite orbit can be used for geophysical modelling (refer  
42 755 e.g. to Fig. 3). They represent the most original measurements. However, it has to be  
43 756 considered that they are measured in a rotating reference frame, the so-called  
44 757 "gradiometer reference frame" (GRF), which means that tensor rotations of the base

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758 functions have to be applied to exploit them to the best possible extent. Additionally,  
759 they are affected by the colored noise characteristics of the GOCE gradiometer (Pail  
760 et al., 2011), so that a single point-wise gravity gradient observation is affected by  
761 large instrument noise, and therefore by itself has a low signal-to-noise ratio. All  
762 these drawbacks make it difficult to use this data type directly for geophysical  
763 modelling.

764

### 765 *3.2.3 Gravity gradient grids*

766

767 A reasonable compromise between the use of spherical harmonics and  
768 original gravity gradients (see above) results in the use of *gravity gradient grids*,  
769 which are usually defined in a well-oriented radial (North-East-up) frame at a constant  
770 altitude. They are computed from the original gravity gradients defined in the GRF by  
771 means of regional gravity processing methods. In fact, they are the spatial equivalent  
772 of the spherical harmonic representation, but much easier to use and interpret. Pure  
773 GOCE gravity gradient grids result from the space-wise method (Gatti and  
774 Reguzzoni, 2015). In the frame of the ESA project GEOExplore global grids of all six  
775 components of the gravity gradient tensor, based on a combination of GOCE and  
776 GRACE data, and defined in a radial Earth-fixed reference frame at two altitudes of  
777 225 km and 255 km, have been derived (Bouman et al., 2015). Since these grid  
778 values are products of “averaging” original gradient data, the error level should be  
779 similar as that of gradients synthesized from global spherical harmonic models.

780 There is an ongoing discussion whether the gradient data contain more (high-  
781 frequency) signal than global gravity models that have been derived from them. The  
782 answer to this question lies in the constraints applied to these models. Constraints  
783 applied to global gravity models are usually designed to optimize the signal-to-noise  
784 ratio on a global scale. This means that in regions of very rough topography and  
785 therefore high-frequency gravity signals there is the tendency to constrain the system

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786 too strongly. Regional gravity solutions techniques, which are usually applied to  
787 generate gridded gravity gradient products, allow for regionally optimized constraints,  
788 but on the cost of global homogeneity. In Pail et al. (2015b) it could be shown, that  
789 compared to global models the gravity gradient grids are affected by a higher noise  
790 level.

791         The achievable accuracy and sensitivity of current gravity field models or  
792 corresponding gravity (gradient) grids can be expressed by cumulative quantities,  
793 which describe the estimated cumulative error at a certain harmonic degree (or the  
794 corresponding spatial wavelength). Fig. 13 shows cumulative gravity anomaly errors  
795 (a), as well as cumulative vertical gravity gradient errors at GOCE satellite altitude of  
796 250 km (b), and ground level (c), for the GRACE models ITSG-GRACE2014s, the  
797 pure GOCE model GOCE-TIM-R5, the combined satellite-only model GOCO05S and  
798 the combined models EGM2008 (pre-GOCE) and GOCO05C (including GOCE data).

799         From Fig. 13 we can learn which geological structure at passive continental  
800 margins (or elsewhere) can be resolved by the different gravity model types.  
801 Assuming that the geological structure/mass anomaly generates a *gravity anomaly*  
802 with a certain spatial wavelength on the Earth's surface, Fig. 13a then provides the  
803 accuracy in mGal with which this anomaly can be captured. (The connection from  
804 the size of a disturbing body to the resulting gravity signal is made in Fig. 3.). As an  
805 example, a gravity signal with 100 km spatial wavelength at the Earth's surface could  
806 be measured by satellites with an accuracy of about 0.5 mGal (black dashed and  
807 solid grey curves). It can clearly be seen that the accuracy for shorter wavelength  
808 signals dramatically decreases, and is already larger than 2 mGal for gravity signals  
809 with approx. 80 km spatial wavelength. Beyond this resolution, satellites cannot  
810 significantly contribute anymore, and high-accuracy terrestrial information, as it was  
811 included, e.g., in

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815 Figure 13: Full page width

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817 **Fig. 13.** Cumulative gravity anomaly errors in (mGal) (a); vertical gravity gradient

818 errors in (mE) at 250 km (b), and ground level (c). This figure shows the generally

819 dramatic increase of the gravity gradient errors at ground level as a result of

820 downward continuation.

821

822 GOCO05C (solid black curve), is necessary to resolve smaller-scale

823 geological structures. This becomes immediately clear if looking at the series of Figs.

824 14 through 21: most of the anomaly sizes at the margins of the South Atlantic are

825 smaller than 80 km. On the other hand we state that the regional gravity field caused

826 at the ocean-continent transition can satisfyingly be resolved by satellite only models

827 (S models). Fig. 13a also shows the major step forward due to satellite missions

828 compared to pre-GOCE models such as EGM2008 (black dot-and-dashed line)

829 especially in the long to medium wavelengths for gross interpretations at a

830 continental scale.

831

832 Fig. 13b shows a similar representation when using *gravity gradients* at

833 satellite altitude as basis information for geophysical modelling of geological

834 structures. Pure GOCE-only models such as GOCE-TIM-R5 (dashed black curve)

835 provide gravity gradients at satellite altitude with standard deviations of 0.45 mE for

836 gravity signals with a spatial resolution of 100 km. These values can be decreased

837 further to 0.25 mE by combination with GRACE information, as it was done, e.g., in

838 the GOCO05S model (solid grey curve). Evidently, GRACE alone (dashed grey

839 curve) results in very high error amplitudes in the higher degrees, demonstrating the

840 dominant impact of GOCE at shorter wavelengths. Modern combined gravity models

841 such as GOCO05C (solid black curve) further increase the performance in the short-

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842 wavelength range by complementing the satellite data by ground data over the  
843 continents and satellite altimetry over the oceans. Also here the improvement  
844 compared to pre-GOCE combined models such as EGM2008 (black dot-and-dashed  
845 line curve) is significant. Recently gradients of the satellite gravity field came into the  
846 focus for modelling purposes which can support interdisciplinary interpretations  
847 (Ebbing et al., 2013; Schaller et al., 2015; Götze, 2015).

848 In Fig. 13b a very interesting feature is the flat curve of the combined gravity  
849 model GOCO05C beyond degree 250. This results from the fact that beyond this  
850 degree the signal amplitude of gravity gradients is already below the mE level, i.e.,  
851 due to signal attenuation with altitude there is no significant gradient signal left in orbit  
852 altitude beyond this degree, because most parts have been “filtered out” due to  
853 upward continuation. Inversely, this also means that GOCE has captured 97% of the  
854 amplitude of the gradient signal that exists in orbit altitude.

855 The picture changes completely (Fig. 13c) when continuing the gradient  
856 information down to ground level. Here the GOCE model (black dashed curve) and  
857 the GOCE+GRACE combination (solid grey curve) perform practically identical, again  
858 showing the dominance of GOCE compared to GRACE at shorter scales. However,  
859 also here for gravity signals with spatial scales below 80 - 100 km a combination with  
860 terrestrial/airborne gravity information is necessary to achieve acceptable accuracies  
861 (solid black curve), so that the gravity field information can be used for local  
862 geophysical modelling of short-scale density structures.

863

### 864 *3.3 Not always in focus: the omission error*

865

866 In order to perform a complete evaluation of the impact of modern satellite  
867 missions for deriving density structure of continental margins, one of the most  
868 important aspects is the evaluation of the *omission error*. It results from high-  
869 frequency signals, which cannot be resolved by satellite gravimetry due to the

1 870 exponential signal attenuation with altitude. These missing signals of satellite-only  
2 871 models are an important issue for the determination of near-surface density  
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4 872 variations, but also shallow lithospheric structures.  
5

6 873 Fig. 14 shows gravity anomaly fields for the South Atlantic region. Fig. 14 a is  
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8 874 based on the GOCO05S model resolved up to its maximum resolution of degree  $n =$   
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10 875 280 (~ 70 km), while Fig. 14 b displays the free-air gravity anomalies based on  
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12 876 EIGEN-6C4 with its maximum resolution of degree  $n = 2160$  (10 km). Comparing  
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14 877 these two figures, the current limits of satellite-only models regarding their spatial  
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16 878 resolution becomes evident, and can only be coped with by combination with  
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18 879 complementary data sources from terrestrial/airborne/shipborne gravimetry, and  
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20 880 satellite altimetry over the oceans, as it was done in EIGEN-6C4. An estimate of the  
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22 881 omission error (Fig. 14 c) for satellite gravity models is given by the difference of  
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24 882 EIGEN-6C4 and GOCO05S, being equivalent to the difference of the Figs. 14 a and  
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26 883 b. Evidently, very rough topographic and bathymetric structures, generating high-  
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28 884 frequency gravity field anomalies and steep slopes, cannot be resolved by the  
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30 885 satellite data. However, usually these topographic features are not the main focus of  
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32 886 geophysical modelling and interpretation, but rather sub-surface lithospheric  
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34 887 structures. Therefore, a topographic reduction was applied, using the  
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36 888 RWI\_TOPO\_2015 topographic potential model (Grombein et al., 2015) and thus  
37  
38 889 taking away the effect of topographic masses up to zero level:  $\delta g_{\text{TOP}} + \delta g_{\text{BPL}}$ . The so  
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40 890 called “mass reduction effect” was already introduced in Fig. 6. The result is a  
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42 891 significantly reduced omission error (Fig. 14 d).  
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48 892 This difference field in Figs. 14 c and d can be considered as errors made  
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50 893 when computing Bouguer anomalies from pure satellite models, which are then  
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52 894 further used for lithospheric modelling. Table 1 gives an overview of the main  
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54 895 statistical parameters of the gravity anomaly fields shown in Fig. 14.  
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899 **Fig. 14.** Free-air gravity anomalies (mGal) of the South Atlantic region based on  
 900 satellite-only model GOCO05S (a) resolved up to degree 280 , combined gravity  
 901 model EIGEN-6C4 (b) resolved up to degree 2160, omission error of a satellite-only  
 902 model (c) and omission error after reduction of topographic signals (d).

903

904 **Table 1:** Main statistical parameters of gravity fields of the South Atlantic region.

905

Gravity field	Figure	min (mGal)	max (mGal)	std.dev. (mGal)
GOCO05S (d/o 280)	14 a	-199.8	116.1	18.8
EIGEN-6C4 (d/o 2160)	14 b	-227.7	453.8	21.3
GOCO05S omission error	14 c	-166.7	415.1	10.8
GOCO05S omission error, topo-reduced	14 d	-143.1	112.8	7.1

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907

#### 908 **4. Benefits for combined interpretations**

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910 However, in relation with the two key questions asked in Section 2.3  
 911 (processing, quality and secondly availability for interpretations at continental  
 912 margins) we have to respond to them in the light of interpretations of solid Earth  
 913 structures. For example a precise geoid can be used to identify global and deep  
 914 anomalies related to mantle lithosphere and deeper structures. Gravity anomalies,  
 915 being first order radial derivatives of the gravitational potential, are sensitive to gravity  
 916 effects of the entire lithosphere, and in particular to the crustal and upper crustal  
 917 structures and density variations e.g. at active and passive continental margins. As it  
 918 has been shown above (Fig. 14), each combination of satellite gravity data with

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919 terrestrial gravity data can be used for all interdisciplinary interpretations techniques,  
920 e.g., “back stripping” in basin modelling at the African continental margin (Dressel et  
921 al. 2015) which also includes thermal subsidence in the reconstruction of the passive  
922 margins through time or 3D modelling of Moho undulations. The new database was  
923 also used to reconstruct the Gondwana continent (Braitenberg, 2015). Fig. 15 refers  
924 to the isostatic residual anomaly in the Southern Atlantic. It was calculated by Klinge  
925 (2016) on base of the corresponding formula for “ISA” in Section 2.3 and the EIGEN-  
926 6C4 model also portrayed in Fig. 14 (b).

927 Both anomaly maps are rather similar and caused by the main tectonic  
928 features of the South Atlantic: the “highs” which are caused by the Mid Atlantic ridge,  
929 the extended “lows” of the four basins in front of South America (Argentinean and  
930 Brazilian basins) and South Africa (Cape and Angola basins). The hotspot trail (e.g.  
931 Torsvik et al., 2009 among others) is visible in the structure of the SW-NE trending  
932 Walvis Ridge offshore South Africa and the corresponding trace of the Rio Grande  
933 Rise at the western side. To the North of the Romanche Fracture Zone between  
934 Fortaleza in the west and Lagos in the East the Sierra Leone Rise is located. Even  
935 the regions of salt deposits offshore Brazil and West Africa (blueish colors indicating  
936 low gravity values) and the magmatic margins at both sides of the margins (reddish  
937 colors and high gravity values) can be distinguished in the satellite derived fields. The  
938 very short wavelengths in the gravity field correspond to masses that are located in  
939 the crust and lithosphere – they were already mentioned in the sketch of Fig. 8 and  
940 mark places of different density contrast at the margins. Other examples were given  
941 in Bouman et al. (2014), Gutknecht et al. (2014), and Hosse et al. (2014).

942  
943 Figure 15: one and a half page width

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945 **Fig. 15.** The isostatic residual field was calculated by Klinge (2016) in the framework  
946 of his MSc thesis. Reference depth  $T_0 = 30\text{km}$  and  $T_e = 20\text{ km}$  (elastic thickness

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947 which was kept constant over the entire area). The figure shows the residual gravity  
948 field in the Southern Atlantic Ocean of the combined EIGEN-6C4 model (Förste et al.,  
949 2014). It correlates well with bathymetric/topographic structures e.g. the Mid Atlantic  
950 Rift (MAR) and portrays also the effect of geological bodies: the positive anomalies in  
951 the area of Windhoek and Buenos Aires. Along the Mid-ocean rift axes positive  
952 anomalies of up to 40 mGal exist. MAR = Mid Atlantic rift.

953

#### 954 *4.1 The continental margins of the South Atlantic*

955

956 New light can be shed on the gravity structures of South Atlantic oceanic  
957 margins at regional (Figs. 16, 17, 19 and 20) and more local scales (Figs. 18 and 21).  
958 By the help of these new compiled maps we will show that modern satellite gravity  
959 fields described in Section 4 can support (1) interpretations of the lithospheric  
960 structures in the South Atlantic and its passive margins and (2) provide much more  
961 details in the gravity field than it was showed along the oversimplified profiles of Figs.  
962 7 and 8.

963 With reference to the Fig. 14, the following sequence of Figs. (16 – 21) contains  
964 always the same information for comparative reasons: the two gravity fields based on  
965 the “satellite only” model GOCO05S (a) and the “combined model” EIGEN-6C4 (b),  
966 and additionally Figs. 18 and 19 include the omission errors *without* (c) and *with* (d)  
967 calculated topographic reductions.

968 Table 2 provides a summary of the standard deviations of the gravity fields  
969 shown in these figures. **d/o** refer to the spherical harmonic analysis: to **d**egree and  
970 **o**order of the expansion.

971

972 **Table 2:** Standard deviations (mGal) of the gravity fields shown in Figures 16 to 21.

Region	Figure	GOCO05S	EIGEN-6C4	GOCO05S	GOCO05S
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		(d/o 280)	(d/o 2160)	omission error	omission error, topo- reduced
Argentinean coast	16	22.4	24.6	9.8	7.1
Brazilian coast	17	15.9	18.8	10.8	6.3
Falkland Bank	18	24.4	25.3	7.3	5.9
African coast	19	19.3	21.5	10.6	7.3
Equatorial African margin	20	19.3	23.9	14.3	9.8
Tristan da Cunha isle	21	6.9	12.3	10.3	4.8

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985 Figure 16: Full page width

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Continental gravity edge effects indicate a fast change from positive to negative anomalies as it is normal for the transition from oceanic to continental crust. In Fig. 16 the positive anomalies indicate in the offshore area the seaward dipping reflectors (SDR) which are of magmatic origin (e.g. Blaich, 2011; Franke et al. 2006; Section 2). The negative anomalies (greenish and blueish colors) offshore are caused by the sedimentary infill of the margin basins e.g. Colorado and Salado (e.g. Autin et al., 2013; 2016). Onshore positive anomalies follow W-E trending topographic features (Salado and Colorado Basin) and in the western continental part of the maps the topography of the Southern Central Andes.

**Fig. 16.** Detailed picture of the free-air gravity field along the Argentinean coast compiled by the GOCO05S (a) and the EIGEN-6C4 (b) models. The lower two

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989 figures indicate the omission errors without (c) and with topographic correction (d).

990 For more information refer to manuscript. The continental areas are marked by

991 transparent overlays.

992

993 In general the series of the following figures will portray similar gravity

994 anomalies (both magnitude and trend of anomalies). It is no wonder that all EIGEN-

995 6C4 compilations consist of more structural details than the GOCO05S models which

996 base on data in the orbit height of some 250 km where small local gravity anomalies

997 are not detectable.

998 Figs. 17 (a) and (b) shows that the “central Atlantic segment” is dominated by

999 high density rocks which cause positive anomalies. The positive gravity offshore

1000 between 40° - 30° longitude is caused by the “Rio Grande High” which marks the

1001 most western edge of the hot spot trace which starts at the position of the Tristan da

1002 Cunha hotspot area. The negative gravity anomalies close to the Brazilian coast are

1003 caused by negative densities of salt accumulation here (Mohriak, 2014). The SDRs

1004 with their high rock densities (Section 2.2) of the southern segment are not

1005 documented here with high resolution; they are too small to be resolved *in detail* as

1006 we show already in Fig. 3. However, at a larger scale the belt of positive gravity

1007 marks the area of SRDs quite well.

1008

1009 Figure 17: Full page width

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1011 **Fig. 17.** The gravity fields (a) GOCO5S and (b) EIGEN6C4 along the Brazilian coast

1012 and offshore regions of the “central segment” of the southern Atlantic. Figures

1013 content is equal to Figs. 14 and 16. However the omission errors are not portrayed

1014 here.

1015

1016 The resolution of gravity anomaly in the off-shore area of the *Falkland Bank*  
1017 and the Scotia Plate with the Eastern Sandwich trench allows the separation of  
1018 subduction related trench lithosphere, the eastern border of the Sandwich Plate, and  
1019 the southern rim with the Antarctic Plate (low gravity corresponding to blueish colors  
1020 in Fig. 18) from high density rocks of the Scotia Plate and Sandwich Isles (yellow and  
1021 reddish colors in Fig. 18 (a and b). Exactly here in a region with rather complex  
1022 interplay of different plates the resolution of gravity fields before the era of the  
1023 modern satellite missions was extremely low and often hindered a tectonic  
1024 interpretation of lithosphere at medium scale. The Scotia Plate in the center (reddish  
1025 colors in Fig. 18) is clearly separated from the other plates of the region (South  
1026 American plate to the North, Antarctic plate to the South, Scotia plate to center.

1027

1028 Figure 18: One and a half page width

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1030 **Fig. 18.** The gravity field of the Falkland bank and the Scotia plate with the eastern  
1031 South Sandwich trench after the processing of new satellite gravity (GOCO5S: (a)  
1032 and EIGEN6C4: (b). Figures content is equal to Fig. 14.

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1034 Due to the symmetry of evolution of the South American and South African  
1035 margins also the gravity field of the western African margin shows the same general  
1036 features as it was exemplified for the South American margin: in the southern  
1037 segment the magma dominated structures cause small positive gravity anomalies  
1038 and North of the Walvis Ridge the area of salt layers is characterized by negative  
1039 anomalies (blueish colors) in Fig. 19. The SW – NE trending Walvis Ridge separates  
1040 the domains of magmatic material from salt layers. Positive gravity anomalies of the  
1041 ridge clearly indicate the Tristan da Cunha hotspot trace – as it was already  
1042 explained for the western part of the Southern Atlantic. More to the South at the  
1043 South African tip of the Cape a second ridge (Agulhas Ridge) can be identified.

1044 Onshore at the African continent close to the equator the extended gravity low (-50  
1045 mGal) of the Congo Basin with its thick sediments dominates the gravity picture.

1046

1047 Figure 19: full page width

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1049 **Fig. 19.** The gravity field of the African margin in the central and southern segment  
1050 (Fig.15) after the GOCE gravity (a) and the EIGEN-6C4 data (b) processing. The  
1051 series of maps correspond to the displayed formats of figures before; transparent  
1052 overlay mark continental area.

1053 One of the most spectacular fracture zones in the Equatorial and Northern  
1054 South Atlantic connecting Africa and South America is illustrated in Fig. 20 (a and b).  
1055 Fairhead and Wilson (2005) explain the formation of the fracture system with  
1056 processes which were related with the opening of the Central and South Atlantic.  
1057 They state that a differential motion between plate segments was absorbed in the  
1058 Caribbean and West and Central African rift systems. The fracture system developed  
1059 due to the temporal different opening phases of the northern and southern Atlantic.  
1060 Then the two independent spreading centers joined a major shear zone developed  
1061 between West Africa and the northern margin of Brazil. The maps of satellite gravity  
1062 image impressively this major shear zone. The gravity map of the EIGEN6C4 model  
1063 provides a clear and sharp picture of the fractures zones.

1064

1065 Figure 20: One a a half page width

1066

1067 **Fig. 20.** The two gravity fields (GOCO5S (a) and EIGEN (b)) of the Equatorial Atlantic  
1068 Ocean map major transform structures offshore the African margin. Sequence of  
1069 maps corresponds with displayed formats of figures before; transparent overlay mark  
1070 continental area.

1071

1072           The limits of resolution of modern satellite only gravity fields (S models) can  
1073 nicely be demonstrated by the gravity field of the Tristan da Cunha Isle, whose  
1074 gravity field signal is at the edge of the spatial resolution of current satellite gravity  
1075 missions. Fig. 21 shows that although GOCE is able to detect the gravity field signal  
1076 of this island, it is significantly damped. It should be emphasized, that a constraint  
1077 has been applied to the GOCO05S model in the frame of the gravity modelling  
1078 procedure in order to improve the signal-to-noise ratio at higher degrees, i.e. noise is  
1079 filtered out at the cost of damping also the signal. As already discussed in  
1080 Section 3.2, the strength of constraining the solution was optimized on a global scale.  
1081 Therefore, it is not tailored to small regions with strong gravity field signal, where a  
1082 weaker constraint would be preferable due to a larger signal-to-noise ratio in this  
1083 region compared to the global average. If the satellite gravity solution were optimized  
1084 for this specific region, it can be expected that in such a regionally tailored solution  
1085 slightly more signal could be retained. The series of Figs. 21 (a) – 21 (d) shows that  
1086 the satellite gravity fields of both GOCE and EIGEN-6c4 are mainly caused by the  
1087 topography of the island. Perfectly seen is the “ring” of negative anomalies in Fig. 21  
1088 (b) which can be explained by the flexure of oceanic lithosphere due to isostatic  
1089 response of the loaded isle masses. After calculating a topographic correction (Fig.  
1090 21 (d)) an anomaly of some 20 mGal appears. One may speculate if this negative  
1091 anomaly is caused by a mass deficit which is related to the hot spot or to crustal  
1092 thickening..

1093           The statistics in Table 2 shows that for such rather small-scale structures the  
1094 amplitude of the omission error can be larger than the signal captured by GOCE.  
1095 However, Fig. 21d shows that significant parts of this high-frequency gravity signal  
1096 result from topography.

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1098 Figure 21: one a half page width

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1100 **Fig. 21.** The Free Air gravity field of the Tristan de Cunha area after the GOCE (a)  
1101 gravity and EIGEN-6C4 (b) data processing. Sequence of the four maps corresponds  
1102 with displayed formats of figures before; omission error without (c) and (d) with  
1103 topographic correction. The limits of resolution of satellite observations can nicely be  
1104 demonstrated by these gravity fields.

1105

1106 Figure 21a: Small column size

1107

1108 **Fig. 21a.** In addition to what we interpret in Fig. 21 this sketch can explain the typical  
1109 negative ring around the positive anomaly in the last figure: the central mass causes  
1110 the positive anomaly while the sediments around the central mass cause a  
1111 symmetrical gravity low. The extent of deformed crust below the mass crust depends  
1112 on the rigidity of the surrounding crust: the left situation (rigidity R1) demonstrates a  
1113 case with extreme high rigidity, on the right a lower crustal rigidity R2 was assumed.

1114

1115 *4.2 Validation of terrestrial gravity by GOCE data*

1116

1117 The GOCE mission provided not only new geoid and gravity fields, but also  
1118 gravity gradient data. Representing the second derivatives of the gravitational  
1119 potential, they are more sensitive to the density structures of the upper crust than  
1120 gravity data normally are. Additionally, gravity gradients provide a better resolution of  
1121 flanks of geological structures, faults, lineaments or even large intrusions at  
1122 continental margins. Gradient data from satellite missions have the potential to  
1123 identify the extent of different structures with varying densities even in the lower crust  
1124 (e.g. Ebbing et al., 2013). Panet et al. (2014) even identify correlations of certain  
1125 components of the gravity gradient tensor with lower mantle structures.

1126 For gravity interpretations at larger wavelengths the new satellite gravity  
1127 database will help to identify a density zonation and segmentation in horizontal and

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1128 vertical directions in the lithosphere. As shown in Section 3.2, GOCE satellite-only  
1129 gradient data provide a spatial (horizontal) resolution in the range of less than 100  
1130 km. However, for many structures - in particular for offshore studies of Applied  
1131 Geophysics - this spatial resolution is not yet sufficient, because smaller crustal  
1132 structures cause anomalies with smaller spatial wavelengths. Therefore, terrestrial  
1133 and airborne gravity measurements have not become obsolete even in the modern  
1134 satellite era, but on the contrary they complement satellite observations on the short-  
1135 wavelength scale where satellite data lack sensitivity.

1136 In addition to their very high accuracy in the long to medium wavelength  
1137 range, modern satellite gravity data definitely provide significant added value in the  
1138 geophysical gravity fields processing domain, especially for:

1139 a) Validation of heterogeneous terrestrial gravity data bases and identification  
1140 of outliers;

1141 b) Fill-in of regions with sparse terrestrial data coverage or even data gaps.

1142 As an example of the first task (a), Fig. 22 a shows the difference between a  
1143 terrestrial gravity data base of South America and GOCO05S, resolved up to  
1144 degree 200. To bring them to the same spatial resolution, the terrestrial data have  
1145 been expanded as part of a global  $0.25^\circ \times 0.25^\circ$  terrestrial gravity anomaly grid into a  
1146 spherical harmonic series to degree 720, and then have been cut at degree 200. Fig.  
1147 22 a clearly indicates systematic differences, which can be attributed to errors of the  
1148 terrestrial data, because of the globally homogeneous accuracy of less than 1 mGal  
1149 for the satellite model. Based on this result, the terrestrial database can be further  
1150 screened for outliers and suspicious observations (either of the gravity value itself or  
1151 the attached height information, Hosse et al., 2014 ). This information can then be  
1152 used to derive empirical error estimates of the terrestrial dataset, which can be further  
1153 used for a spatially depending weighting scheme in the frame of a combined solution  
1154 with satellite data (Fecher et al., 2013, 2016). The implicit assumption is that the data  
1155 quality of a terrestrial observation is already reflected in its long-wavelength

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1156 component. By means of this procedure, satellite data get a higher weight in regions  
1157 where a lower accuracy of terrestrial data is suspected.

1158

1159 Figure 22: Full page width

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1161 **Fig 22.** (a) Gravity anomaly differences (mGal) between a South American terrestrial  
1162 database (kindly provided by the US National Geospatial-Intelligence Agency) and  
1163 GOCO05S, consistently resolved up to degree 200; b) empirical error estimates  
1164 (mGal) derived from the difference field (after: Fecher et al. (2015), modified).

1165

1166 It should be emphasized, that this validation procedure can be applied in any  
1167 region on Earth. Thereby, a globally uniform satellite gravity model provides the  
1168 chance to estimate a-posteriori the accuracy and reliability even of historical gravity  
1169 data bases (terrestrial, ship- and air-borne), for which only incomplete or even no  
1170 meta-information about the measurement process and conditions is available.

1171 Also Bomfim et al. (2013) describe how gradients of the GOCE mission can  
1172 help to estimate systematic errors in terrestrial gravity data in the cratonic basins  
1173 (e.g. Amazon and Parnaiba Basins) in Brazil. Here they calculate an average value of  
1174 terrestrial gravity anomaly and compare its long- and medium-wavelength content of  
1175 the terrestrial gravity with the GOCE gravity field. The analysis shows that where  
1176 terrestrial data are sparse and therefore require an improvement in data coverage,  
1177 satellite data can be substituted in order to represent the gravity field correctly. The  
1178 method they proposed can be used directly to control other gravity databases and  
1179 constitutes as a tool for the quality assessment of terrestrial gravity observations,  
1180 both on- and offshore.

1181 The second task (b) also addressed the heterogeneity of terrestrial gravity  
1182 data. There are many regions worldwide where terrestrial data are of very bad quality

1  
2 1183 or not available/accessible at all (refer to Fig. 2). In these regions, data from satellite  
3 1184 gravity missions are the only available data source.

4 1185         These examples demonstrate, that although satellite missions provide (only)  
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6 1186 long to medium wavelength gravity field data, they are able to provide new gravity  
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8 1187 field information especially in regions where up to now the gravity field has been  
9  
10 1188 practically unknown. This regional model can then been used as constraint for an  
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12 1189 improved lithospheric density model and the derivation of the state of stress of the  
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14 1190 subduction zone (Gutknecht et al., 2014), clearly demonstrating the added value of  
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16 1191 GOCE especially in these data-critical onshore regions.  
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## 21 1193 **5. Conclusions and outlook**

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26 1195         The resolution of “satellite only data” up to now does not fall below a resolution of  
27  
28 1196 80 - 100 km. This is still the borderline for studies presented in the above mentioned  
29  
30 1197 Sections. In summary we have to say that rather small complex structures related  
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32 1198 e.g. to the “seaward dipping reflectors at passive continental margins (SDR)” with  
33  
34 1199 small size and density contrast cannot be resolved as separated anomalies in the  
35  
36 1200 orbit heights of recent satellite missions (e.g. Schaller et al. 2015). For this purpose  
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38 1201 terrestrial gravity data have to be combined with satellite data in gravity models e.g.  
39  
40 1202 GOCO5C. The interpretations of Section 4.1. showed that gravity and its gradients  
41  
42 1203 from the modern satellite missions support interpretations at a medium scale – at  
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44 1204 passive continental margins and elsewhere.

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47 1205         Modelers of lithospheric structures at continental margins hope that medium scale  
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49 1206 gravity data from the recent and future satellite missions (GRACE and/or GOCE;  
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51 1207 GRACE follow on) can support combined interpretation together with seismological  
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53 1208 and gravity studies. For rather local models (wavelengths of gravity anomalies are  
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55 1209 smaller than 20 km) both resolution and quality of satellite only gravity data have to  
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58 1210 be seen still reluctant until today. However, there is no doubt that combinations with  
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1211 terrestrial gravity data bases and satellite gravity with a spatial resolution of 10 – 20  
1212 km can provide detailed insight in the structural behavior of continental margins.

1213 For modelling at continental scales Fig. 2 demonstrated that terrestrial databases  
1214 often are of inhomogeneous distribution (e.g. in South America), just if gravity data  
1215 are sampled over long time-consuming field campaigns with big human efforts: there  
1216 remain big gaps in the data base. They are mainly caused by limited access to the  
1217 terrain in remote areas of the world – high mountains, deserts, swamps and jungle.  
1218 Even more field procedures and technical instrumentation varied over time and  
1219 together with missing other metadata a homogeneous data base can be established  
1220 only with big effort and high costs. Here the new data bases already helped in a  
1221 spectacular way: Hosse et al. (2014) and Gutknecht et al. (2014) replaced the  
1222 incomplete terrestrial gravity data base by homogeneously measured satellite gravity  
1223 and gravity gradient data for lithospheric modelling. New data were applied to the  
1224 calculation of GPE (gravity potential energy), stress distributions and combined  
1225 interpretation of complex geologic structures. Satellite gravity information was also  
1226 used for validation and cleaning of inhomogeneous gravity databases taking the  
1227 benefit of very homogeneous error characteristics and accuracy of global satellite  
1228 gravity data (Hosse et. al, 2014; Bomfim et al., 2013). The high spatial resolution of  
1229 terrestrial gravity *combined* with the homogeneous lower-orbit satellite data leads to  
1230 more detailed and better-constrained lithospheric density models, and hence  
1231 improves our knowledge about structure, evolution and state of stress in the  
1232 lithosphere basing on the consistency in the long-to-medium wavelengths, down to  
1233 10 – 50 km.

1234  
1235 At the beginning (Section 1.1) we mentioned the calculation/recalculation of the  
1236 COB from an integrated interpretation of gravity, magnetic, seismic, electrical  
1237 methods and geology (Torsvik et al., 2009). We did not deal with the calculation of  
1238 COB, however, we think that the combined satellite fields can successfully replace

1239 the terrestrial gravity data which have to be used in former times. Because we  
1240 analyzed Free Air gravity from the GOCE mission, in Fig. 23 topographic features on-  
1241 and offshore are enhanced. These enhancements indicate clearly the slopes of the  
1242 continental shelf regions of the Southern Atlantic.

1243

1244 Figure 23: Full page width

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1246 **Fig. 23.** The third derivations of the gravitational potential and the resulting total  
1247 gradient of the vertical gravity gradient were calculated from the GOCO5C model  
1248 (expansion to degree 720). It provides already good insight into local gravity field  
1249 structures particularly at the margins by the derivations of the vertical gradient ( $V_{zz}$ ).

1250

1251 In the near future complementary information from seismic and magnetics  
1252 could be included in a joint inversion for lithospheric modelling also at passive  
1253 continental margins. ESA's magnetic field mission *Swarm* was successfully launched  
1254 in November 2013 and provides valuable information of the long to medium  
1255 wavelength Earth's magnetic field and its temporal variations with an accuracy on the  
1256 nT- (nano tesla) level (<http://esamultimedia.esa.int/multimedia/publications/BR-302/>).  
1257 The value of the mission for the determination of the crustal remanent magnetic field  
1258 will increase in the future, because the three satellites will continuously lower their  
1259 orbit altitudes during mission lifetime, thus also increasing their sensitivity for detail  
1260 magnetic field structures. However, a joint interpretation of remanent magnetic and  
1261 gravity field is only possible in the case of common sources, i.e. similar contrasts in  
1262 density and magnetization. In this case Poisson's equation can be applied, which  
1263 links the magnetic and gravity potential fields. *Swarm* is already now a very valuable  
1264 tool to determine the electric conductivity of the Earth's mantle and thus provides very  
1265 important information on the thermochemical and compositional structure of the  
1266 Earth.

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1267           Several concepts for future satellite mission constellations to explore the  
1268 Earth's potential fields are under development and investigation. A strong need by  
1269 the user communities was expressed in terms of a joint IUGG resolution adopted at  
1270 the IUGG General Assembly 2015 (IUGG, 2015). The science requirements and user  
1271 needs for a future gravity field mission constellation were consolidated (Pail et al.,  
1272 2015a) also under active participation of the geophysical user community. In addition  
1273 to an improved temporal resolution for the detection of co- and post-seismic  
1274 deformation, an increased spatial resolution together with an improved accuracy will  
1275 shift the capabilities to use satellite-based gravity observations for geophysical  
1276 interpretation in passive continental margins, and elsewhere, to even more small-  
1277 scale structures.

1278

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1293

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Ladies and Gentlemen,

This is an invited manuscript by me and Roland Pail for the "Passive margin" issue.

Please, start with the review process.

Regards,

H.-J. Götze