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Delineation of aquifer boundary by two vertical superconducting gravimeters in a karst hydrosystem, France

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Abstract

Mass distribution on Earth is continuously changing due to various physical processes beneath the Earth's surface or on the surface. Some of the primary sources for these mass displacements are tidal forces, atmospheric and oceanic loading, and seasonal changes in continental water distribution. The development of relative cryogenic gravimeters, the Superconducting Gravimeters (SGs), has made it possible to characterize and monitor such mass variations at orders of magnitudes as small as a few nm/s^2 ($1 \text{ nm/s}^2 \sim 10^{-10} \text{ g}$ where g is the mean gravity at the Earth's surface). Our study focuses on the hydrodynamics of the 900 m thick unsaturated zone of the low-noise underground research laboratory (Laboratoire Souterrain à Bas Bruit, LSBB) located in Rustrel (France) using a unique configuration of two SGs vertically arranged 520 m depth apart. The installation of an SG (iGrav31) at the site surface several years after installing the first (iOSG24) inside a tunnel has provided several new insights into the understanding of the hydrological processes occurring in the LSBB. By comparing differential and residual gravity time-series together with global hydrological loading models, we find that most water-storage changes occur in the unsaturated zone between both SGs. The misfit between the observed gravity time-series and the gravity effect corresponding to local hydrological contribution calculated from global hydrological models can be explained by large lateral fluxes and rapid runoff occurring in the LSBB site. Finally, we implement a rectangular prism method to compute forward gravity responses to water storage changes for a homogeneous water-layer following the site topography using a 5-m digital elevation model. In particular, we analyse the sensitivity of the differential record from both SGs to the extent and depth of the water storage changes by computing the corresponding 2D admittances. This gravity difference is sensitive to an extension up to about 2500 m laterally before tending towards an asymptotic value corresponding to the Bouguer plate approximation. We show that the zone of water-storage changes that best fits observed differential gravity signal is located at depths larger than 500 m (below iOSG24). This fitting is improving when the integration radius increases with depth. This is the first time that hydrological processes are investigated when the baseline configuration of two SGs is vertical.

Keywords: hydro-gravimetry, superconducting gravimeters, hydrogeology, gravity time-series, karst aquifer

1. Introduction

Keeping in mind global warming and growing water demand, the importance of characterizing and monitoring the hydrological cycle has rapidly increased. The scientific community is continuously developing more robust monitoring techniques in order to assess the depletion of water resources. Karst systems represent an essential resource of fresh water but are

51 heterogeneous matrices with fractures and open voids revealing specific hydrodynamic
52 behaviours. Classical hydrogeological methods (e.g., piezometers) provide direct
53 measurements of aquifer properties, but they are invasive and characterize a local region of a
54 larger heterogeneous karstic aquifer system. Other methods, like flow rate and chemical
55 measurements at the spring, provide an overall hydrologic information at the basin scale. Non-
56 invasive geophysical methods appear relevant for investigating the spatial variability of karst
57 systems covering a few square kilometers (Chalikakis et al., 2011). Among the geophysical
58 methods, gravimetry provides a direct quantification of the water-mass fluctuations at the
59 catchment scale (Kroner et al., 2006; Hasan et al., 2008; Jacob et al., 2010; Hector et al., 2013;
60 Hemmings et al., 2016; Imanishi et al., 2006; Jacob et al., 2009; Pool & Eychaner, 1995; Van
61 Camp et al., 2006; Wilson et al., 2011). Besides, hydrological models require calibration that
62 can be achieved by assimilating gravity measurements (Naujoks, et al., 2010; Creutzfeldt et al.
63 2012; Chaffaut et al. 2020). Moreover, gravity gradiometry or multiple gravimeters across the
64 local area can further improve the assessment of temporal water mass changes (Naujoks et al.,
65 2008; Kennedy et al. 2014), by mitigating the common non-uniqueness issues in single-
66 gravimeter surveys.

67 The low noise underground laboratory, the so-called LSBB (“Laboratoire Souterrain à Bas
68 Bruit”) of Rustrel (France) is located in a karstic massif at the southern edge of the Fontaine de
69 Vaucluse impluvium. The LSBB constitutes a typical experimental site for underground karstic
70 water-fractured hydrosystems (Carrière et al. 2016). The LSBB site is well-equipped with
71 various instrumentations, as for our concern, with two Superconducting Gravimeters (SGs): an
72 observatory SG, the iOSG24, recording data since September 2015 and an iGrav instrument
73 (Warburton et al. 2010), the iGrav31 continuously recording since February 2019. SGs are
74 relative instruments that record temporal gravity variations at the Earth's surface and have a
75 wide range of geodynamics applications (Hinderer et al., 2007). SGs have precision and long-
76 term stability appropriate for monitoring water-storage changes (e.g., Kroner and Weise, 2011;
77 Hector et al., 2014; Fores et al., 2017; Güntner et al., 2017; Chaffaut et al., 2020). The set-up
78 of both SGs at the LSBB site is such that iGrav31 is 520 m nearly vertically above the iOSG24,
79 with horizontal offset of about 90 m. This specific configuration of SGs is unique in the world
80 and constrains the local water masses influencing SG measurements (Mouyen et al., 2019).
81 This is the first time that hydrological processes are investigated when the baseline
82 configuration of two SGs is vertical. Previously, several significant findings were made based
83 on signals from horizontal baselines (Naujoks et al., 2008, Kennedy et al., 2014). Here, the
84 vertical configuration of two gravimeters has the advantage to increase the sensitivity to water
85 mass changes in the area in between, by a factor two when considering a simple Bouguer plate
86 approximation. Indeed, the gravity change associated with an equivalent liquid water thickness
87 h is $\Delta g = 2\pi G\rho h$ where ρ is the volumetric mass of water and G is universal gravitational
88 constant. The gravimeter at the surface measures $\Delta g(\text{iGrav31}) = 2\pi G\rho h$ while the one below
89 the water layer measures $\Delta g(\text{iOSG24}) = -2\pi G\rho h$ since the water masses are located above the
90 sensor. The differential gravity measured by these two SGs $\Delta g(\text{iGrav31}) - \Delta g(\text{iOSG24})$ is then
91 $4\pi G\rho h$, twice the Bouguer plate effect. Another advantage of the vertical configuration of two
92 SGs is that a single gravimeter is insensitive to water mass changes at the same height of the
93 instrument while the vertical configuration of two SGs solves this issue (see e.g., Fig. S2 in
94 Carrière et al., 2022).

95
96 In this paper, we first describe the hydrogeological setting of the LSBB site and the
97 unique configuration of two vertically arranged SGs. We then explain the SG data processing
98 used to retrieve gravity time-residuals compared with two global hydrological models. A
99 hydro-gravimetric modelling based on integration of rectangular prisms is used to compute
100 hydrological admittances (i.e., gravity response to a 10-cm water layer thickness). Then, we

101 use the obtained admittances to calculate the gravity effects induced by a homogeneous water
102 layer spread over the topography with water head changes inferred from the global hydrological
103 model. Finally, the sensitivity of the differential gravity signal to the extent and depth of the
104 main water storage change is discussed.

105 **2. Experimental Site: Geological and Hydrological Setting**

106 Fontaine de Vaucluse Hydrosystem

107 The Fontaine de Vaucluse (FdV) is a karst hydrosystem located in the southeast of France.
108 Groundwater discharges from the system at the FdV spring with an average flow-rate of 23
109 m³/s (Cognard-Plancq et al., 2006). The impluvium spreads over an area of 1,115 km², and the
110 average thickness of its unsaturated zone (UZ) is around 800 m (Puig, 1987). This karst system
111 consists of massive and continuous lower Cretaceous limestone and its base extends to a depth
112 of 1500 m (Masse, 1969).

113 The reef limestone part of this carbonate platform has thickness of around 450 m in the study
114 area. The Urgonian facies present in the reef limestone has high total matrix porosity up to 25-
115 30 percent among limestone and is present in half of the FdV (Masse, 1969; Masse and Masse,
116 2011). The stable low flow rate, the thick UZ, and some small permanent outlets in the
117 limestone plateau indicate that the low flow discharge occurs mainly in the UZ (Garry, 2007).
118 The base of this karst system lies on Valanginian and -Upper Hauterivian impermeable marls.
119 The presence of dolines and dry valleys cutting through the plateau indicates strong
120 karstification of the FdV hydrosystem due to limestones dissolution (Blavoux et al., 1992).

121 Low Noise Underground Laboratory

122 The LSBB is a 3.8 km long, nearly horizontal underground tunnel in the north of Rustrel
123 village, France (Fig. 1). It is a former component of the French nuclear missile system, which
124 has been turned into a cross disciplinary underground research laboratory. The LSBB is a
125 unique low noise underground set-up because of its initial conception, aiming to overcome
126 nuclear blast, and its location far from anthropic noise (Waysand et al. 2002). This tunnel passes
127 through the massif and randomly intersects karstic features. It also goes through some channels
128 in the UZ. The maximum thickness of rock cover over the tunnel reaches 519 m. The saturated
129 zone (SZ) in this catchment lies approximately 400 m beneath the LSBB tunnel. The fracturing
130 and karstification are very diverse throughout the LSBB gallery. Garry (2007) carried out
131 several hydro-chemical and hydrodynamic investigations in this site and developed different
132 hydrodynamic models of these flow points. Carrière et al. (2016) demonstrated the effects of
133 the water in the rock using an integrated hydro-geophysical approach.

134 **3. SG Observations and Global Hydrological Models**

135 **3.1. Gravity data processing**

136 The iOSG24 has been recording data since September 2015 (Rosat et al., 2018), whereas the
137 iGrav31 was set up more recently and started recording data since May 2019. The recorded SG
138 data correspond to the feedback voltage that balances the position of the levitating sphere at a
139 1 Hz sampling frequency. The pre-processing of SG data includes the change of voltage to
140 acceleration through a calibration factor using FG5 parallel measurements, signal processing,
141 and drift correction. Many studies related to SG calibration using parallel absolute gravity
142 measurements can be cited (e.g., Francis, 1997; Imanishi et al., 2002). SG calibrations are
143 known to be very stable over time (e.g., Goodkind, 1991; Calvo et al., 2014). Two absolute
144 gravity campaigns were carried out during the considered time epoch, in October 2019 and in
145 September 2020 for both instruments. The obtained calibrations factors are -451 ± 3 nm/s²/V
146 for iOSG24 (Rosat et al., 2018) and -851 ± 7 nm/s²/V for the iGrav31, that is an accuracy better
147
148
149

150 than 1 %. Please note that the uncertainty of the scale factor for iGrav31 is larger because of
151 its installation at the surface, where environmental noise is larger than inside the tunnel.

152 The instrumental drift was removed using a linear trend since we considered data recorded
153 several months after the installation, when the drift has become linear. Discussions about SG
154 drift were published in Van Camp and Francis (1997) and more recently in Hinderer et al.
155 (2022). Contributions from signals that are not due to local hydrology are finally removed.
156 These signals are listed below:

- 157 • Solid and oceanic tides: they were both removed using a local tidal model resulting
158 from a least-squares adjustment to SG data so that it contains both solid and oceanic
159 tides. This tidal analysis was performed with ETERNA3.4 (Wenzel, 1996) software
160 based on Hartmann and Wenzel (1995) potential catalogue (as in Rosat et al. 2018) and
161 for which semi-annual and annual tides were set to the nominal 1.16 gravimetric factors
162 (else part of the seasonal hydrological signal would be artificially reduced).
- 163 • Atmospheric pressure masses: these effects were calculated using atmospheric pressure
164 data from the second Modern-Era Retrospective analysis for Research and Applications
165 (MERRA2) assuming an inverted barometer response of the oceans. This atmospheric
166 loading (including non-tidal oceanic loading) is calculated using a Green's function
167 formalism as in Boy et al. (2002); the atmospheric loading consists of two
168 contributions: a local contribution corresponding to an integration within 0.10° around
169 the gravimeter and converted to a gravity signal using a barometric value of -2.21047
170 $\text{nm/s}^2/\text{hPa}$, a non-local contribution resulting from the Green function loading
171 integrating deformation beyond the 0.10° radius.
- 172 • Polar motion: the gravity effect for pole tides is calculated using Earth's rotation
173 parameters taken from IERS (<ftp://hpiers.obspm.fr/iers/eop/eopc04/>) using elastic
174 gravimetric factor as in Hinderer et al. (2007).
- 175 • Non-local hydrological loading: the "non-local" contribution of hydrology is computed
176 using the Green's function (Farrell, 1972) as in Boy and Hinderer (2006). The non-local
177 part corresponds to a region outside 20-30 km radius from the station, and its amplitudes
178 are about 15 percent of that of the local signal. Values are calculated using either
179 MERRA2 (Reichle et al., 2017a) or ERA5 (Hersbach et al., 2020) hydrological models
180 (soil moisture, and snow components).

181

182 The hence obtained gravity residuals are plotted on Fig. 2. In the following, we assume that air
183 mass changes were perfectly reduced from the gravity time-series, so that the remaining gravity
184 signals can be interpreted directly in terms of water storage changes.

185

186 **3.2. Comparison of gravity data with global hydrological models**

187 In this section, we compare gravity residuals of the iOSG24 ($\Delta g(\text{iOSG24})$) and iGrav31
188 ($\Delta g(\text{iGrav31})$) (Fig. 3a) between May 2019 and June 2020 with corresponding local
189 contributions from MERRA2 and ERA5 hydrological models. The local hydrological effect
190 was calculated using a Bouguer plate approximation with a nominal admittance value of
191 $\pm 4.2677 \text{ nm}\cdot\text{s}^{-2} / \text{cm}$ (the sign depends on the location of the station with respect to the ground:
192 it is minus when the station is underground). We recall that gravity residuals should only
193 contain local hydrological effects.

194

195 The MERRA2 reanalysis is a data product of the National Aeronautics and Space
196 Administration (NASA) Global Modelling and Assimilation Office (Gelaro et al., 2017). The
197 MERRA2 contains hourly-sampled data from 1980 to the present with a horizontal resolution
198 of approximately 50 km. In MERRA2, the land surface water budget was estimated using

199 observation-based precipitation (Reichle et al. 2017b). This analysis is calculated on a latitude-
200 longitude grid at the exact spatial resolution as the atmospheric model.

201

202 ERA5, a fifth-generation reanalysis data product, is the most recent release by the European
203 Centre for Medium-Range Weather Forecasts (ECMWF). Several significant improvements
204 were incorporated compared to its previous ERA-Interim (Hersbach et al., 2020). This is a 1-
205 hourly sampled data and was computed using a more advanced 4D VAR assimilation
206 algorithm. The horizontal resolution of ERA5 is approximately 30 km. The ERA5 contains
207 data from 1979 to the present.

208

209 The temperature and precipitation biases of ERA5 are found to be more accurate relative to
210 MERRA2 for hydrological modelling. Local hydrological contribution to gravity loading at the
211 LSBB site is slightly smaller for MERRA2 than for ERA5 model (Fig. 3).

212

213 In Fig. 3(a), we plot gravity residuals recorded by both SGs with corresponding local effects
214 from both hydrological models ERA5 and MERRA2. We see that the ERA5 and MERRA2
215 models predict well the main features of gravity changes at the surface (iGrav31) while for
216 iOSG24, after taking into account the sign inversion, the predicted variability is overestimated.
217 Particularly between November 2019 and January 2020, the predicted gravity fluctuations
218 reach ~ 75 nm/s² while the observed ones are less than 20 nm/s² underground and larger than
219 100 nm/s² at the surface.

220 In Fig. 3(b), we compare differential gravity time-series ($\Delta g(iGrav31) - \Delta g(iOSG24)$) with
221 twice the corresponding values from the ERA5 and the MERRA2. Note that we doubled ERA5
222 and MERRA2 local hydrological contribution because, as we explained in the introduction, the
223 vertical configuration of two gravimeters increases the gravitational effect of the water located
224 between the gravimeters by a factor two, when considering a simple Bouguer plate
225 approximation and assuming a homogeneous mass distribution.

226 When the difference $\Delta g(iGrav31) - \Delta g(iOSG24)$ is negative (Fig. 3(b)), we can say that most of
227 the water storage change is located below the iOSG24. While when $\Delta g(iGrav31) - \Delta g(iOSG24)$
228 is positive, most of the water mass change is temporarily stored in the UZ between the two
229 SGs. The overall agreement between the hydrological models and the differential gravity time-
230 series is quantified by a RMS of 44.3 nm/s² while for each time series taken separately the
231 RMS is of 72.4 nm/s² for iOSG24 and 28.5 nm/s² for iGrav31. This suggests that the
232 hydrological models better represent the groundwater storage variation from the surface down
233 to the UZ when the differential $\Delta g(iGrav31) - \Delta g(iOSG24)$ gravity signal is considered. Indeed,
234 under Bouguer plate approximation, the contribution of the water below both gravimeters
235 cancels out when we compute the differential gravity. Since the Bouguer plate is insensitive to
236 depth, the water-mass contribution is the same on both gravimeters.

237

238 However, misfits between local hydrological models and differential gravity time-series are
239 still visible, highlighting the effects of lateral flows and the role of topography, which mitigate
240 the Bouguer plate approximation and the adequateness of the global hydrological models in
241 such a mountainous karstic system.

242 Subsequently, we zoomed in the residual gravity time-series from Dec 1, 2019 to Jan 10, 2020
243 (Fig. 4) and plotted the observations-based precipitation as provided within MERRA2 products
244 (Reichle et al. 2017b). We marked five major timings (numbered 1 to 5 in Fig.4) where the
245 residual gravity trends of both SGs change abruptly. An important criterion for our analysis is
246 whether the gravity variations measured at each gravimeter at a given time, have the same sign
247 or are opposite. If they are opposite, it means that water storage variation occurred at a depth
248 between the two gravimeters. If they have the same sign, then the water storage variation

249 occurred below both gravimeters. For timing 1 corresponding to a large rain event, $\Delta g(iGrav31)$
 250 increases significantly whereas $\Delta g(iOSG24)$ decreases slightly, because of the SGs
 251 configuration. However, for timing 2, which does not correspond to any precipitation event,
 252 the gravity trends are reversed indicating that all the surface water was seeped into the zone
 253 below both the SGs. This transition from rainfall to infiltration has taken approximately 6 days,
 254 which is very quick in terms of hydrological events. This shows that the UZ is highly porous
 255 and permeable. Timing 3 illustrates another rain event smaller than event 1, with an amount of
 256 gravity increase seen by iGrav31 smaller than the amount of gravity decrease seen by iOSG24.
 257 Similarly, timing 4 marks another precipitation occurrence whereas timing 5 is for infiltration.

258 We also see that the slope of $\Delta g(iGrav31)$ during infiltration periods is much larger than
 259 observed water storage changes ($\Delta g(iOSG24)$), indicating faster water mass transfers as seen
 260 by gravimetry.

261

262 **4. Forward Gravity Modelling**

263 **4.1. Methodology**

264 Here we present a rectangular prism method developed to compute time-lapse gravity signals
 265 from the hydrological models. In this forward problem, the hydrological model cells are
 266 assumed to be incremental prismatic mass storages, and corresponding gravity effects are
 267 computed by integrating over all cells within a given a priori radius. However, we consider a
 268 homogenous water layer change spread over the topography, so our modelling is not 3D since
 269 we do not consider the possibility to have different water head changes in different cells. We
 270 used the RGE ALTI digital elevation model (DEM) provided by the National Institute of
 271 Geographical and Forestry Information in France (IGN-F) on a regular grid with 1 m resolution
 272 but decimated to 5 m in order to save computation time. This method comprises three formulas:
 273 the prism formula, the MacMillan formula, and the point-mass formula. These three equations
 274 were derived from Newton's law of gravity with different approximations depending on the
 275 distance to the mass change (see Leiriao et al., 2009). They are used to calculate gravity
 276 changes due to an unconfined aquifer with varying hydraulic heads. From previous studies, we
 277 know that a single prism causes gravity changes only if its vertical length varies. The DEM
 278 (Fig. 5) is discretized into rectangular prisms of sizes 5 m x 5 m x 5 m to incorporate the site
 279 topography. We consider the lateral offset between both SGs in the modelling by computing
 280 the forward gravity effects at the exact locations of each SG.

281

282 The gravity variation is then calculated using all three formulas depending upon the radial
 283 distance to the location of the SG (Forsberg, 1984). The entire computation process involves
 284 four steps comprising applying the prism formulas and summing up the obtained results. The
 285 Forsberg formula is used when the SG is close to the mass change:

286

$$287 \quad \Delta g = \gamma \rho s_y \left\| \left\| x \log(y + r) + y \log(x + r) - z \tan^{-1} \frac{xy}{zr} \right\| \right\|,$$

288 where the vertical lines indicate that the integration of the mass element is performed in the x, y
 289 and z directions between the two points of coordinates that define the prism. x, y and z are the
 290 components of the instrument-prism vector (see Leiriao et al., 2009 for further details).

291

292 For intermediate radial distances, the spatial detail in the discretization decreases. This leads to
 293 the application of the MacMillan formula (MacMillan, 1958):

294

$$295 \quad \Delta g = -\gamma \rho s_y \Delta x \Delta y \Delta z \left(-\frac{z}{r^3} - \frac{5}{24} \frac{(\alpha x^2 + \beta y^2 + \omega z^2)z}{r^7} + \frac{1}{12} \frac{\omega z}{r^5} \right)$$

296 For farther radial distances, it is assumed that the whole mass is in the centre of the cell.
 297 Consequently, the 3D point-mass formula is used in the remaining model grid:
 298

$$299 \quad dg = -\gamma\rho\Delta s dx dy \frac{(h - z_0)}{(x^2 + y^2 + (h - z_0)^2)^{3/2}}$$

300
 301 Finally, the total gravity change is a cumulative sum of the contributions from applications of
 302 all three formulas (Forsberg, MacMillan, and point-mass).
 303

304 The formulation of the rectangular prism method is sketched in Fig. 6 where $d=r$ i.e., the
 305 distance between the prism and the instrument, and Δh are hydraulic heads representing the
 306 vertical changes in water layer thickness. Fig. 6 is basically a schematic design indicating
 307 manifestations of the Newton's forward gravity equation.
 308

309 The total gravity response for a hydro-system is the cumulative effect of all the prisms.
 310 Therefore, it is crucial to decide which of the three formulas to use when a prism is located
 311 randomly away from the SGs location. The normalized switching factor has been defined using
 312 the size of the prism and its relative distance to include all three formulas. The switching factor
 313 f^2 is defined below:

$$314 \quad f^2 = \frac{r^2}{dr^2}$$

315
 316 Where $r^2 = x^2 + y^2 + z^2$ is the distance between the prism and the instrument, $dr^2 = dx^2 +$
 317 $dy^2 + dz^2$ is the square of length of the diagonal line of a prism, and other symbols are
 318 described in Leiriao et al. (2009). Here in this work, the prism formula is used for f^2 below 25,
 319 the MacMillan formula is used for f^2 between 25 and 36, and the point-mass formula for f^2
 320 above 36 as done in Leiriao et al. (2009) and Chaffaut et al. (2020).
 321

322 4.2. Hydrological Admittances

323 The hydrological admittance usually refers to an increment in the observed gravitational
 324 acceleration due to an infinitely extended horizontal layer of water (infinite Bouguer plate),
 325 with the SG at any distance above the layer. It is approximately equal to $4.2677 \text{ nms}^{-2}\text{cm}^{-1}$
 326 for the density of water. For practical reasons, if the distance between the sensor and the layer
 327 is less than or equal to one-tenth of the horizontal size, the layer is considered infinite. The
 328 admittance value enables quantifying the contribution from local hydrology and is obtained
 329 using a simple Bouguer approximation in its infinite limit. Using the prism method described
 330 in section 4.1, we computed admittances for 10-cm thick layers of water that follow the
 331 topographic model. These admittances are 2D transfer functions that depend on the radius of
 332 integration and depth of the water layer. In the infinite limit for the radius of integration, they
 333 correspond to the classical admittance. In the following, we denote this 2D transfer function as
 334 “admittance”.

335 The obtained hydrological admittances (Fig. 7) show that the sensitivity of iOSG24 to water-
 336 storage changes tends to a Bouguer plate with the asymptotic value of $-3.4 \text{ nm s}^{-2}\text{cm}^{-1}$ for
 337 radii larger than 4 km, while for iGrav31, the admittance tends to an asymptotic 3.86
 338 $\text{nm s}^{-2}\text{cm}^{-1}$ for radii larger than 6 km. Please note that for depths of 450 m and larger, the
 339 admittance computed at the iOSG24 location becomes positive since, upon incorporating the
 340 topography height at this depth, the water layer is located below the SG.
 341

342 4.3. Relative contribution of Unsaturated and Saturated Zones

343 In the previous section, we performed the integration for all depths independently to the fact
344 the water layer was located in the UZ or SZ. However, SG observations integrate gravity
345 signals from changes in water distribution in both SZ and UZ, in particular with different
346 sensitivities for iGrav31 and iOSG24 due to their respective distances to the SZ. In this section,
347 we want to check the importance of the SZ w.r.t. the UZ to gravity changes. For that, we divided
348 the forward gravity modelling in two steps: one for the integration in the UZ and one for the
349 integration in the SZ. The temporal gravity variation in the UZ is computed by dividing the soil
350 into several layers between 0 and 900 m depth, and then contributions from each layer are
351 summed up together. A similar integration is performed for the SZ by summing up the
352 contributions of water layers located between 900 and 1500 m deep.

353
354 The time-variability of the hydraulic head is simulated by using the local gravimetric effect
355 computed for ERA5 model divided by the nominal admittance of $4.2677 \text{ nm s}^2 \text{ cm}^{-1}$ to convert
356 it into water-head changes. The obtained water layer height changes are then multiplied by the
357 respective hydrological 2D admittances as computed in the section 4.1 for various integration
358 radii. Please remember that we call admittance not the limit for an infinite Bouguer plate (large
359 radii) but the Newtonian gravity effect integrated along various radii and depths. The
360 admittance is then a 2D function of radius and depth and can be directly multiplied with the
361 time-varying water-level changes to simulate the time-dependent gravity effect. An integration
362 radius of 150 m predicts gravity changes that best fit $\Delta g(\text{iGrav31})$ (Fig. 8). For iOSG24, finding
363 a best-fit solution is more complicated since the influence of the SZ becomes larger and the UZ
364 is located partly below and partly above the instrument. A specific inverse problem could be
365 developed but is beyond the scope of this paper, which is to focus on the use of the gravity
366 difference $\Delta g(\text{iGrav31}) - \Delta g(\text{iOSG24})$, indeed, to mitigate the uncertainty linked to the use of a
367 unique gravimeter.

368
369 The water storage within the SZ in karst conduits plays a significant role in regulating flow in
370 hydro-systems (Mangin, 1975). We see that the water mass fluctuations in the SZ are, as
371 expected, much less than the effects of the UZ on the iGrav31, 96% of the gravity contribution
372 coming from the UZ (Fig. 8). As seen in most hydro-systems, main contributions to the gravity
373 signals are from the time-varying soil moisture content in the UZ, between both SGs. This
374 result matches with Mangin's schema (Mangin, 1975), which claimed the presence of an
375 epikarst zone in this site. This zone lies at the top of the UZ, especially above deep infiltration
376 zone. Therefore, the simulated gravity signals in the UZ are corresponding to the epikarst and
377 the infiltration zone. The hydrological models are also essentially informing on the epikarst in
378 this area, given the agreement obtained with the differential gravity signal $\Delta g(\text{iGrav31}) -$
379 $\Delta g(\text{iOSG24})$ (section 3.2).

380 381 **4.4. Sensitivity of the differential gravity**

382 Our aim here is to delineate major water storage in the unsaturated zone using differential
383 gravity data. The underlined idea is to find the zone where the water-mass changes best explain
384 observed gravity fluctuations. Therefore, we computed the differential gravity effect
385 $\Delta g(\text{iGrav31}) - \Delta g(\text{iOSG24})$ for various depths and radii of integration (Fig. 9(a)) that we
386 compare with the observed $\Delta g(\text{iGrav31}) - \Delta g(\text{iOSG24})$. The RMS of the difference between
387 predicted and observed $\Delta g(\text{iGrav31}) - \Delta g(\text{iOSG24})$ shows the depths and integration radii for
388 which the RMS is minimum (dark blue zone in Fig. 9(b)). Some examples for specific radii of
389 integration and depths of the water layer are illustrated in Fig. 10. We can get a few major
390 findings upon thoroughly analysing Fig. 9 and Fig. 10. Firstly, the largest RMS values above
391 iOSG24 suggests that residual data of both SGs have significant differences. This is mostly
392 because the water present in this zone (i.e., from surface to 500 m depth) lies below the iGrav31

393 and above the iOSG24. The differential gravity signal in this upper part of the UZ quickly
394 extends to a Bouguer plate approximation when the radius of integration exceeds 400 m (Fig.
395 9(a)) that is closer than for an individual gravimeter as inferred from the computed admittances
396 (Fig. 7). At depths between 0 and 400 m (above iOSG24), the admittance ranges from 2 to 4
397 $\text{nm}^2\text{cm}^{-1}$ indicating that the surface iGrav31 fluctuations dominate in this zone. The largest
398 RMS difference of observed differential gravity with simulated one is located in the zone
399 located at depths between 300 and 500 m and radii between 150 and 500 m (Fig. 9(b)) where
400 the admittance is closer to zero (Fig. 9(a)) indicating a lack of sensitivity of the differential
401 gravity to water-storage changes in this zone. We indeed see that an integration radius of 200
402 m at a depth of 300 m is not sufficient to explain most of the observed $\Delta g(\text{iGrav31})-$
403 $\Delta g(\text{iOSG24})$ (Fig. 10(a)). A radius between 1000 and 1500 m would better explain the
404 amplitude of the gravity fluctuations. When considering a water-layer below iOSG24, at a
405 depth of 600 m (Fig. 10(b)), the integration radius needs to be larger than 1500 m to fit the
406 amplitude of the observed differential gravity fluctuations. When the integration radius is too
407 small (less than 600 m), whatever the depth of the water layer, the amplitude of the simulated
408 $\Delta g(\text{iGrav31})-\Delta g(\text{iOSG24})$ is too small to explain the observed one (Fig. 10(a) and Fig. 10(c)).
409 When the depth of the water layer is close to the depth of iOSG24 (between 400 and 500 m),
410 the predicted differential gravity signal is close to zero (Fig. 10(d)) when the integration radius
411 is large enough (here 1000 m). This is because we are closer to iOSG24 and there is a
412 cancellation of the gravity effect due to the water layer, which is partly above and partly below
413 iOSG24.

414

415 **5. Discussion**

416 Comparison of gravity observations with local contributions from global hydrological models

417 We observe that both local hydrological contributions of ERA5 and MERRA2 show a few
418 trends of overestimating and underestimating gravity effects (Fig. 3(a)) compared to the
419 residual gravity time-series recorded by the iGrav31. This could be due to the following
420 reasons:

- 421 i) The excessive lateral fluxes of groundwater may be inhibiting the vertical
422 infiltration of rainwater into the ground.
- 423 ii) The SGs are located at the south border of the impluvium, and hence, the water
424 mass distributions outside it may partly cancel the gravity effect.
- 425 iii) The time lags between observed rainfalls at the site, residual gravity and the
426 hydrological loading model may be the cause for this misfit.
- 427 iv) The water storage variations given by the hydrological models are not strictly
428 localised at the surface (where the iGrav31 is) but integrate a greater thickness of
429 ground

430

431 However, when we compare the ERA5 and MERRA2 signals with differential gravity time-
432 series $\Delta g(\text{iGrav31})-\Delta g(\text{iOSG24})$, we could see that those hydrological models are better
433 mimicking the observed differential gravity fluctuations (Fig. 3b). Due to the above reason, for
434 further investigations, we mostly used differential signals ($\Delta g(\text{iGrav31})-\Delta g(\text{iOSG24})$).

435

436 Comparison of surface gravity changes with gravity changes inside the tunnel after rainfalls

437 By comparing the two gravity time series at the surface and inside the tunnel, we can estimate
438 the transfer time of the water from the surface to the SZ after major rain events. Indeed, when
439 we have a sign concordance between the two gravimetric series, after a rain event, that implies
440 that the water is below both the instruments. For instance, during the period between Dec 3,
441 2019, and Dec 22, 2019 (Fig. 4), we can infer that the peak of flow inside the tunnel occurs as
442 approximately 6 days after the major rain as already observed in past studies (e.g. Garry 2007).

443 Please note that since the amplitude of the gravity effect is different at the surface and at the
444 depth of iOSG24, the differential gravity (Fig. 3(b)) cannot show such a transfer simply as a
445 negative anomaly. A few days of transfer indicates the presence of a high porosity infiltration
446 zone corresponding to the epikarst overlying the infiltration zone. This zone in the LSBB tunnel
447 was already identified as a fast circulation flow point with a dripping within the LSBB tunnel
448 occurring a few days after a rain event (Garry et al., 2008) but other flow points were identified
449 representative either of a karstic discharge or of a fractured circulation. The spatial sensitivity
450 of both SGs raises the problem of the representativity of such gravity data at the system scale.
451 Besides, this time is not an infiltration or transfer time since it corresponds to a mass transfer
452 without taking into account the identity of these water masses. In other words, geochemistry
453 measurements are required to identify the origin of this water mass reaching iOSG24 and
454 resulting in a dripping inside the LSBB tunnel (e.g. Blondel et al., 2012).

455

456 Sensitivity of the gravity signals to water storage changes

457 In section 4.2, we have computed the 2D hydrological admittance using a prism method for a
458 uniform water layer of 10 cm as a function of the radius of integration and depth of the water
459 layer (Fig. 9). This modelling shows that the area of largest sensitivity for the differential
460 gravity lies from 450 m to 700 m deep and in the epikarst zone between 0 and 400 m deep
461 approximately (Fig. 9(a)). This result is very crucial in the estimation of the total water storage
462 in the LSBB site. When we further compare the observed and simulated differential gravity
463 time-series for different depths and radii of integration, we can make the following remarks

- 464 i) The simulated signals best fit the observed gravity data for a water-layer depth
465 below 500 m that is below iOSG24 (Fig. 9(b) in dark blue and Fig. 10).
- 466 ii) The RMS of the difference between predicted and observed $\Delta g(\text{iGrav31}) -$
467 $\Delta g(\text{iOSG24})$ is largest in the UZ between 400 and 500 m depth and for radii smaller
468 than 500 m (Fig. 10(b)) where the admittance is close to zero (Fig. 10(a)). In this
469 zone, induced gravity effects on iGrav31 and on iOSG24 cancel out.
- 470 iii) We note that as we go deeper below iOSG24, the predicted differential gravity
471 signal fits better the observed one when the integration radius increases (Fig. 9(b)).

472

473 **6. Conclusions**

474 In this study, we put complete emphasis on investigating the hydrological processes occurring
475 at the LSBB site. These processes were inferred from the observed differential gravity time-
476 series from a vertical dipole of two superconducting gravimeters. The observed mass transfer
477 time of 6 days gives an estimate of the vertical hydraulic conductivity of the infiltration zone.
478 The residual and differential gravity time series of the SG vertical dipole clearly show the
479 groundwater redistribution at seasonal time-scales at the LSBB site. The simulated gravity
480 responses from a global hydrological model indicate that most groundwater is stored in the
481 unsaturated zone between both SGs. This paper also presented the relative contribution from
482 the saturated and unsaturated zones separately. Firstly, most of the contributions in gravity
483 changes are due to water in the unsaturated zone. Secondly, since the saturated zone lies farther
484 away from the SGs, the gravity contributions are very low. We also computed true and
485 asymptotic admittance values using a rectangular prism method matching the topography. We
486 started with simple comparisons of recorded gravity time-series with the gravity effects of
487 global hydrological models and then investigated the following observations further. Finally,
488 we mapped the entire aquifer zone by minimizing the misfit between the observed and
489 simulated differential gravity signals from both saturated and unsaturated zones. We could
490 conclude that a part of the aquifer lies above the iOSG24, whereas the remaining part lies below
491 iOSG24 in the unsaturated zone. Even though we successfully delineated the groundwater
492 boundary in this catchment, further work is needed to improve the simplified hydro-gravimetric

493 model proposed in this study. Mainly, the mask effect due to the shelter at the iGrav31 location
494 and the mask effect due to the tunnel around iOSG24 should be considered in the gravity
495 modelling by removing prism cells around the instruments (e.g., Chaffaut et al. 2020). Mask
496 effects could have important consequences on the parameters of the unsaturated zone, which is
497 shallower.

498 The main hypothesis made in this study is that all other contributions, particularly air mass
499 changes, were completely and accurately reduced from gravity observations. However, it is
500 well-known that the dynamics of atmospheric mass changes is also complex with turbulent
501 effects, particularly during heavy rain events, that are not considered in the loading models
502 used in this study (see, for example, Neumeyer et al., 2004; Gitlein et al., 2013). Some small
503 atmospheric contribution in the gravity residuals may remain.

504
505 We have considered in this paper the problem of determining water layer depth and extent in
506 the unsaturated zone from the vertical gradient of time-varying gravity changes using pre-
507 imposed water head changes with a uniform layer fitting the topography. Using the
508 hydrogeological knowledge that has been accumulated among the years at the LSBB karstic
509 site by the numerous previous studies, we should in the future extend the prism modelling into
510 3D by considering heterogeneous water content field (Chaffaut et al. 2022). Finally, we have
511 shown how the vertical gradient of gravity could infer properties on water storage content
512 within a mountainous karstic environment and the potential improvements that could be done
513 from this initial work.

514

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522

523 **References**

524 Blavoux, B., Mudry, J., Puig, J.M. (1992). The karst system of the Fontaine de Vaucluse
525 (Southeastern France), *Environ. Geol. Water Sci.*, 19(3), 215-225.

526 Blondel T., Emblanch, C., Batiot-Guilhe, C., Dudal, Y., Boyer, D. (2012). Punctual and
527 continuous estimation of transit time from dissolved organic matter fluorescence properties in
528 karst aquifers, application to groundwaters of ‘Fontaine de Vaucluse’ experimental basin (SE
529 France), *Environ Earth Sci*, 65, 2299-2309, doi:10.1007/s12665-012-1562-x

530 Boy, J.-P., Hinderer, J. (2006). Study of the seasonal gravity signal in superconducting
531 gravimeter data, *J. of Geodyn.*, 41(1-3), 227-233, <http://doi.org/10.1016/j.jog.2005.08.035>

532 Boy, J.-P., Gegout, P., Hinderer, J. (2002). Reduction of surface gravity data from global
533 atmospheric pressure loading. *Geophys. J. Int.* 149, 534-545.

534 Calvo, M., Hinderer, J., Rosat, S., Legros, H., Boy, J.-P., Ducarme, B., Zürn, W. (2014). Time
535 stability of spring and superconducting gravimeters through the analysis of very long gravity
536 record. *J. Geodyn.* 80, 20-33, <http://dx.doi.org/10.1016/j.jog.2014.04.009>

537 Carrière, S. D., Chalikakis, K., Danquigny, C., Davi, H., Mazzilli, N., Ollivier, C. and C.
538 Emblanch (2016). The role of porous matrix in water flow regulation within a karst unsaturated
539 zone: an integrated hydrogeophysical approach, *Hydrogeol. J.*, 24, 1905-1918.

540 Chaffaut Q., Hinderer J., Masson F., Viville D., Bernard J-D., Cotel S., Pierret M-C., Lesparre
541 N., Jeannot B. (2020). Continuous Monitoring with a Superconducting Gravimeter As a Proxy
542 for Water Storage Changes in a Mountain Catchment, *Int. Association of Geodesy*
543 *Symposium*, http://doi.org/10.1007/1345_2020_105

544 Chaffaut Q, Lesparre N, Masson F, Hinderer J, Viville D, Bernard J-D, Ferhat G, Cotel S.
545 (2022). Hybrid Gravimetry to Map Water Storage Dynamics in a Mountain Catchment, *Front.*
546 *Water*, 3, 715298, <https://doi.org/10.3389/frwa.2021.715298>

547 Chalikakis, K., Plagnes, V., Guerin, R., Valois, R. & Bosch, F.P. (2011). Contribution of
548 geophysical methods to karst-system exploration: an overview, *Hydrogeol. J.*, 19(6), 1169-
549 1180.

550
551 Cognard-Plancq AL, Gevaudan C, Emblanch C (2006). Historical monthly rainfall-runoff
552 database on Fontaine de Vaucluse karst system: review and lessons. In: Duràn JJ, Andreo B,
553 Carrasco F (eds) *Karst, cambio climático y aguas submediterraneas [Karst, climate change and*
554 *submediterranean waters]*. Publicaciones del Instituto Geológico y Minero de España, Madrid,
555 pp 465-475

556
557 Creutzfeldt B., Ferre T., Troch P., Merz B., Wziontek H., Güntner A. (2012). Total water
558 storage dynamics in response to climate variability and extremes: Inference from long-term
559 terrestrial gravity measurement, *J. of Geophys. Res.*, 117,
560 <https://doi.org/10.1029/2011JD016472>

561
562 Farrell, W.E. (1972). Deformation of the Earth by surface loads. *Rev. Geophys. Space Phys.*
563 10, 761-797.

564
565 Fores B., Champollion, C., Le Moigne, N., Bayer, R. and J. Chéry (2017). Assessing the
566 precision of the iGrav superconducting gravimeter for hydrological models and karstic
567 hydrological process identification, *Geophys. J. Int.*, 208, 269-280,
568 <https://doi.org/10.1093/gji/ggw396>

569
570 Forsberg, R. (1984). *A study of Terrain Reductions Density Anomalies and Geophysical*
Inversion Methods in Gravity Field Modelling. Ohio State University.

571
572 Francis, O. (1997). Calibration of the C021 superconducting gravimeter in Membach
573 (Belgium) using 47 days of absolute gravity measurements. In: *International Association of*
Geodesy Symposia, vol. 117, Springer-Verlag, pp. 212-219.

574
575 Garry, B. (2007). *Etude des processus d'écoulement de la zone non saturée pour la*
576 *modélisation des aquifères karstiques. Expérimentation hydrodynamique et hydrochimique sur*
577 *les sites du Laboratoire Souterrain à Bas Bruit (LSBB) de Rustrel et de Fontaine de Vaucluse.*
578 *PhD Thesis, Université d'Avignon et des Pays du Vaucluse, Avignon, France*

579
580 Garry, B., Blondel, T., Emblanch, C., Sudre, C., Bilgot, S., Cavaillou, A., Boyer, D. & Auguste,
581 M. (2008). Contribution of artificial galleries to knowledge of karstic system behaviour in
582 addition to natural cavern data, *International Journal of Speleology*, 37(1): 75-82.

582 Gelaro, R., McCarty, W., Suarez, M.J., Todling, R., Molod, A.M., Takacs, L.L., Randles, C.,
583 Darnenov, A., Bosilovich, M.G., Reichle, R.H., Wargan, K., Coy, L., Cullather, R.I., Akella,
584 S.R., Bachard, V., Conaty, A.L., da Silva, A., Gu, W., Koster, R.D., Lucchesi, R.A., Merkova,
585 D., Partyka, G.S., Pawson, S., Putman, W.M., Rienecker, M.M., Schubert, S.D., Sienkiewicz,
586 M.E., Zhao, B. (2017). The modern-era retrospective analysis for research and applications,
587 version-2 (MERRA-2). *J. Clim.* 30, 5419-5454, <http://dx.doi.org/10.1175/JCLI-D-16-0758.1>

588 Gitlein, O., Timmen, L., Müller, J. (2013). Modeling of Atmospheric Gravity Effects for High-
589 Precision Observations. *International Journal of Geosciences* 04, 663-671.,
590 <https://doi.org/10.4236/ijg.2013.44061>

591 Goodkind, J.M. (1991). The Superconducting Gravimeters. principles of operation, current
592 performance, and future prospects. In: Proc. of the Workshop Won-Tidal Gravity Changes.
593 Intercomparison Between Absolute and Superconducting Gravimeters'. Cahiers du Centre
594 Europeen de Geodynamique et de Seismologie, vol. 3, Luxemburg, pp. 81-90.

595 Güntner, A., Reich, M., Mikolaj, M., Creutzfeld, B., Schroeder, S. and H. Wziontek (2017).
596 Landscape-scale water balance monitoring with an iGrav superconducting gravimeter in a field
597 enclosure, *Hydrol. Earth Syst. Sci.*, 21, 3167-3182, <https://doi.org/10.5194/hess-21-3167-2017>
598

599 Hartmann, T. and Wenzel, H.-G. (1995). The HW95 tidal potential catalog. *Geophys. Res.*
600 *Let.*, 22(24), 3553-3556.
601

602 Hasan, S., Troch, P.A., Bogaart, P.W. & Kroner, C. (2008). Evaluating catchment-scale
603 hydrological modelling by means of terrestrial gravity observations, *Water Resour. Res.*, 44(8),
604 <http://doi.org/10.1029/2007WR006321>
605

606 Hector, B., Séguis, L., Hinderer, J., Descloitres, M., Vouillamoz, J.M., Wubda, M. & Le
607 Moigne, N. (2013). Gravity effect of water storage changes in a weathered hard-rock aquifer
608 in West Africa: results from joint absolute gravity, hydrological monitoring and geophysical
609 prospection, *Geophys. J. Int.*, 194(2), 737-750.
610

611 Hector, B., Hinderer, J., Séguis, L., Boy, J.-P., Calvo, M., Descloitres, M., Rosat, S., Galle, S.,
612 Riccardi, U. (2014). Hydro-gravimetry in West-Africa: first results from the Djougou (Benin)
613 superconducting gravimeter, *J. Geodyn.*, 80, 34-49.

614 Hemmings B., Gottsmann J., Whitaker F., Coco A. (2016). Investigating hydrological
615 contributions to volcano monitoring signals: A time-lapse gravity example, *Geophys. J. Int.*,
616 207, <https://doi.org/10.1093/gji/ggw266>

617 Hersbach, H., Bell, B., Berrisford, P., Hirahara, S., Horányi, A., Muñoz-Sabater, J., Nicolas, J.,
618 Peubey, C., Radu, R., Schepers, D., Simmons, A., Soci, C., Abdalla, S., Abellan, X., Balsamo,
619 G., Bechtold, P., Biavati, G., Bidlot, J., Bonavita, M., De Chiara, G., Dahlgren, P., Dee, D.,
620 Diamantakis, M., Dragani, R., Flemming, J., Forbes, R., Fuentes, M., Geer, A., Haimberger,
621 L., Healy, S., Hogan, R. J., Hólm, E., Janisková, M., Keeley, S., Laloyaux, P., Lopez, P., Lupu,
622 C., Radnoti, G., de Rosnay, P., Rozum, I., Vamborg, F., Villaume, S., Thépaut, J.-N. (2020).
623 The ERA5 global reanalysis. *Q. J. R. Meteorol Soc.*, 146, 1999-2049, doi:10.1002/qj.3803.

624 Hinderer, J., Crossley, D., Warburton, R.J. (2007). Herring, T., Schubert, G. (Eds.),
625 Superconducting Gravimetry in Treatise on Geophysics. (Geodesy), vol. 3. Elsevier,
626 Amsterdam, pp. 65-122.

627
628 Hinderer, J., Warburton, R. J., Rosat, S., Riccardi, U., Boy, J.-P., Forster, F., Jousset, P.,
629 Güntner, A., Erbas, K., Littel, F. and J.-D. Bernard (2022). Intercomparing Superconducting
630 Gravimeter Records in a Dense Meter-Scale Network at the J9 Gravimetric Observatory of
631 Strasbourg, France, *Pure and Applied Geophys.*, <https://doi.org/10.1007/s00024-022-03000-4>

632 Imanishi, Y., Higashi, T., Fukuda, Y. (2002). Calibration of the superconducting gravimeter
633 T011 by parallel observation with the absolute gravimeter FG5#210 - a Bayesian approach.
634 *Geophys. J. Int.* 151, 867-878.

635 Imanishi Y., Kokubo K., Tatehata H. (2006). Effect of underground water on gravity
636 observation Matsushiro, Japan, *J. of Geodyn.*, vol-41, pg. 221-226,
637 <https://doi.org/10.1016/j.jog.2005.08.031>

638 Jacob T., Chery J., Bayer R., Moigne N. L., Boy J-P., Vernant P., Boudin F. (2009). Time-
639 lapse surface to depth gravity measurements on a karst system reveal the dominant role of the
640 epikarst as a water storage entity, *Geophys. J. Int.*, 177, 347-360,
641 <https://doi.org/10.1111/j.1365-246X.2009.04118.x>.

642 Jacob, T., Bayer, R., Chery, J. & Le Moigne, N. (2010). Time-lapse microgravity surveys
643 reveal water storage heterogeneity of a karst aquifer, *J. Geophys. Res.*, 115(B6),
644 <http://doi.org/10.1029/2009JB006616>

645
646 Kazama T., Okubo S. (2009). Hydrological modelling of groundwater disturbances to observed
647 gravity: Theory and application to Asama Volcano, Central Japan, *J. Geophys. Res.*, 114,
648 B08402, <http://doi.org/10.1029/2009JB006391>

649
650 Kennedy J., T. P. A. Ferre, A. Güntner, M. Abe, and B. Creutzfeldt (2014). Direct measurement
651 of subsurface mass change using the variable baseline gravity gradient method, *Geophys. Res.*
652 *Let.*, 41, 2827-2834.

653
654 Kroner, C., T. Jahr, M. Naujoks, A. Weise (2006). Hydrological signals in gravity - foe or
655 friend? *Dynamic Planet*, IAG Symposia Series 130, Springer, ISBN 978-3-540-49349-5, 504-
656 510.

657
658 Kroner, C., A. Weise (2011). Sensitivity of superconducting gravimeters in central Europe on
659 variations in regional river and drainage basins. *J. Geod.* 85(10), 651-659, doi:
660 10.1007/s00190-011-0471-1.

661
662 Leiriao S., X. He, L. Christiansen, O.B. Anderson, and P. Bauer-Gottwein (2009). Calculation
663 of the temporal gravity variation from spatially variable water storage change in soils and
664 aquifers, *J. Hydrol.*, 365, 302-309.

665
666 MacMillan, W.D. (1958). *The Theory of Potential. Theoretical Mechanics*, vol. 2. Dover, New
667 York

668
669 Mangin, A. (1975). *Contribution à l'étude hydrodynamique des aquifères karstiques*, Ph.D
670 thesis. Université de Dijon, 124 pp.

671

672 Masse, J.-P. (1969). Contribution à l'étude de l'Urgonien (Barrémien - Bédoulien) des Monts
673 de vacluse et du Luberon. [Contribution to the study of the Urgonian (Barremian-Bedoulian)
674 of the Vaucluse and the Luberon mountains]. Bureau de Recherches Géologiques et Minières,
675 Orléans, France, 59 pp
676

677 Masse, J.-P. (1976). Les calcaires urgoniens de Provence, Valanginien - Aptien inférieur, tome
678 1: stratigraphie - paléontologie; tome 2: Les paléoenvironnements et leur évolution [Urgonian
679 Limestones of Provence, Valanginian – Lower Aptian, vol 1: stratigraphy – paleontology, vol
680 2: paleoenvironments and their evolution]. PhD Thesis, Univ. D'Aix-Marseille, Marseille,
681 France, 445 pp
682

683 Masse J.-P., Fenerci-Masse M (2011). Drowning discontinuities and stratigraphic correlation
684 in platform carbonates: the Late Barremian-Early Aptian record of southeast France. *Crétacé*
685 *Res.*, 32(6):659-684.
686

687 Mouyen, M., Longuevergne, L., Chalikakis, K., Mazzilli, N., Ollivier, C., Rosat, S., Hinderer,
688 J., Champollion, C. (2019). Monitoring groundwater redistribution in a karst aquifer using a
689 superconducting gravimeter, E3S Web of Conf., 88, 03001,
690 <https://doi.org/10.1051/e3sconf/20198803001>

691 Naujoks, M., A. Weise, C. Kroner, T. Jahr (2008). Detection of small hydrological variations
692 in gravity by repeated observations with relative gravimeters. *J. Geod.* 82(9), 543-553, doi:
693 10.1007/s00190-007-0202-9.

694 Naujoks, M., C. Kroner, A. Weise, T. Jahr, P. Krause, S. Eisner (2010). Evaluating local
695 hydrological modelling by temporal gravity observations and a gravimetric three-dimensional
696 model. *Geophys. J. Int.* 182(1), 233-249, doi: 10.1111/j.1365-246X.2010.04615.x.

697 Neumeyer, J., Hagedoorn, J., Leitloff, J., Schmidt, T. (2004). Gravity reduction with three-
698 dimensional atmospheric pressure data for precise ground gravity measurements. *J. of Geodyn.*,
699 38, 437-450. <https://doi.org/10.1016/j.jog.2004.07.006>

700 Pool D. R., Eyechaner J.H. (1995). Measurements of Aquifer-Storage Change and Specific
701 Yield Using Gravity Surveys, *Groundwater*, 33, 425-432, [https://doi.org/10.1111/j.1745-](https://doi.org/10.1111/j.1745-6584.1995.tb00299.x)
702 [6584.1995.tb00299.x](https://doi.org/10.1111/j.1745-6584.1995.tb00299.x)

703 Puig JM (1987) Le système karstique de la Fontaine de Vaucluse [The karst system of the
704 Fontaine de Vaucluse]. PhD Thesis, Univ. D'Avignon et des Pays de Vaucluse, France, 207 pp
705

706 Reichle, R.H., Draper, C.S., Liu, Q., Giroto, M., Mahanama, S.P.P., Koster, R.D., De Lannoy,
707 G.J.M., (2017a). Assessment of MERRA-2 land surface hydrology estimates, *J. Clim.*, 30,
708 2937-2960, <http://dx.doi.org/10.1175/JCLI-D-16-0720.1>
709

710 Reichle, R. H., Liu, Q., Koster, R. D., Draper, C. S., Mahanama, S. P. P., & Partyka, G. S.
711 (2017b). Land Surface Precipitation in MERRA-2, *J. of Climate*, 30(5), 1643-1664
712

713 Rosat, S., Hinderer, J., Boy, J.-P., Littel, F., Bernard, J.-D., Boyer, D., Mémin, A., Rogister, Y.
714 and S. Gaffet (2018). A two-year analysis of the iOSG24 superconducting gravimeter at the
715 low noise underground laboratory (LSBB URL) of Rustrel, France: environmental noise
716 estimate, *J. of Geodyn.*, 119, 1-8, <https://doi.org/10.1016/j.jog.2018.05.009>

717 Van Camp, M., and Francis, O. (2007). Is the instrumental drift of superconducting gravimeters
718 a linear or exponential function of time?, *J Geodesy*, 81(5):337-344.
719 <http://doi.org/10.1007/s00190-006-0110-4>

720 Van Camp M., Vanclooster M., Crommen, O., Petermans, T., Verbeeck, K., Meurers, B., van
721 Dam, T., Dassargues, A. (2006). Hydrogeological investigations at the Membach station,
722 Belgium, and application to correct long periodic gravity variations, *J. of Geophys. Res.*, 111,
723 <https://doi.org/10.1029/2006JB004405>.

724 Warburton, R.J., Pillai, H. and Reineman, R.C. (2010). Initial results with the new GWR
725 iGravTM superconducting gravity meter, in Extended Abstract Presented at 2nd Asia Workshop
726 on Superconducting Gravimetry, Taipei, Taiwan.

727

728 Waysand, G., Gaffet, S., Virieux, J., Chwala, A., Auguste, M., Boyer, D., Cavaillou, A.,
729 Guglielmi, Y., Rodrigues, D., Waysand, G., Gaffet, S., Virieux, J., Chwala, A., Auguste, M.,
730 Boyer, D., Cavaillou, A., Guglielmi, Y., & Rodrigues, D. (2002). The Laboratoire Souterrain
731 Bas Bruit (lsbb) In Rustrel-pays D’apt (france): A Unique Opportunity For Low-noise
732 Underground Science, *EGSGA*, 3869,
733 <https://ui.adsabs.harvard.edu/abs/2002EGSGA..27.3869W/abstract>

734

735 Wenzel, H.-G. (1996). The nanogal software: earth tide data processing package ETERNA
736 3.30. *Bull. Inf. Marées Terr.* 124, 9425-9439.

737 Wilson C.R., Scanlon B., Sharp J., Longuevergne L., Wu H. (2011). Field Test of the
738 Superconducting Gravimeter as a Hydrologic Sensor, *Groundwater*, 50, 442-449,
739 <https://doi.org/10.1111/j.1745-6584.2011.00864.x>

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741 **Statements and Declarations**

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748 **Competing Interests**

749 The authors have no relevant financial or non-financial interests to disclose.

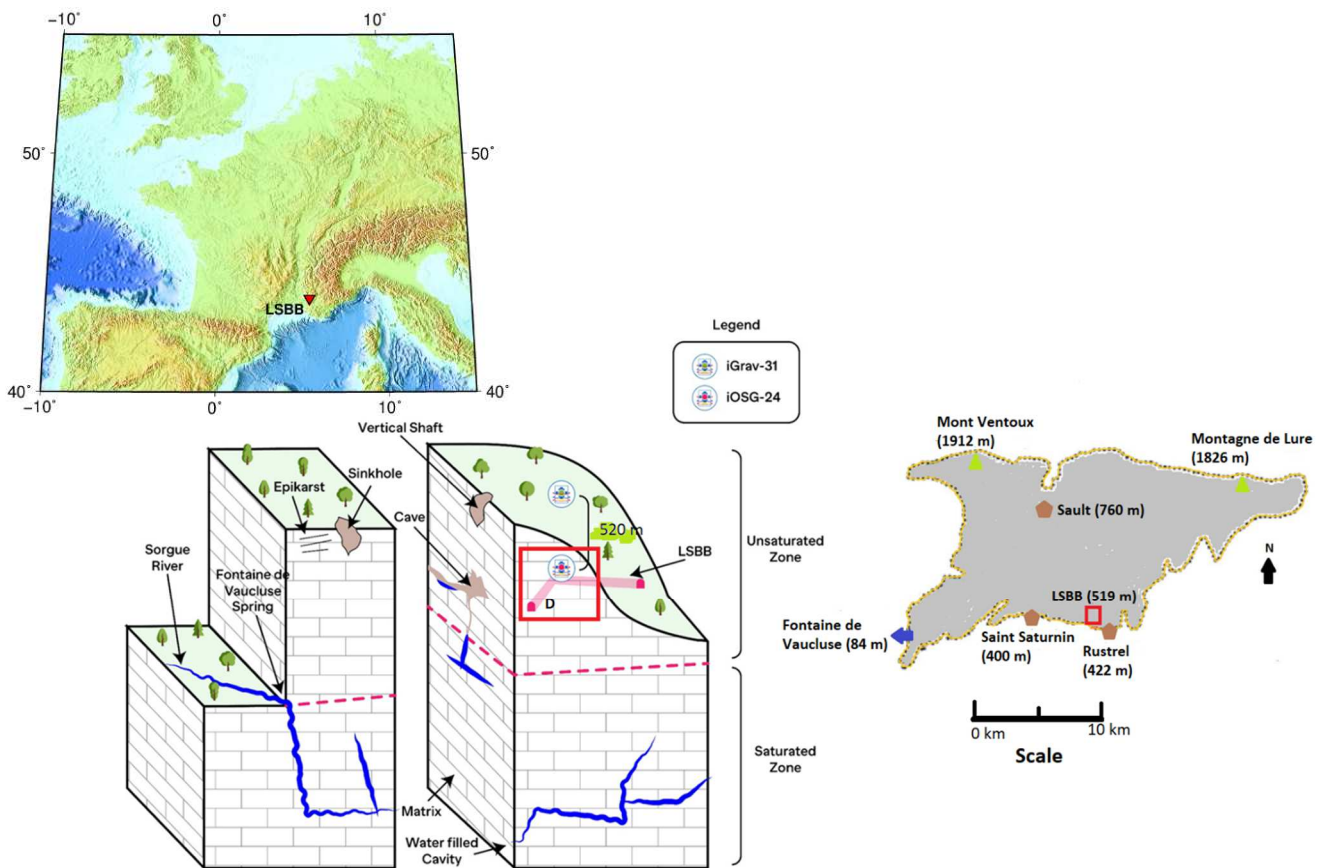
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751 **Author Contributions**

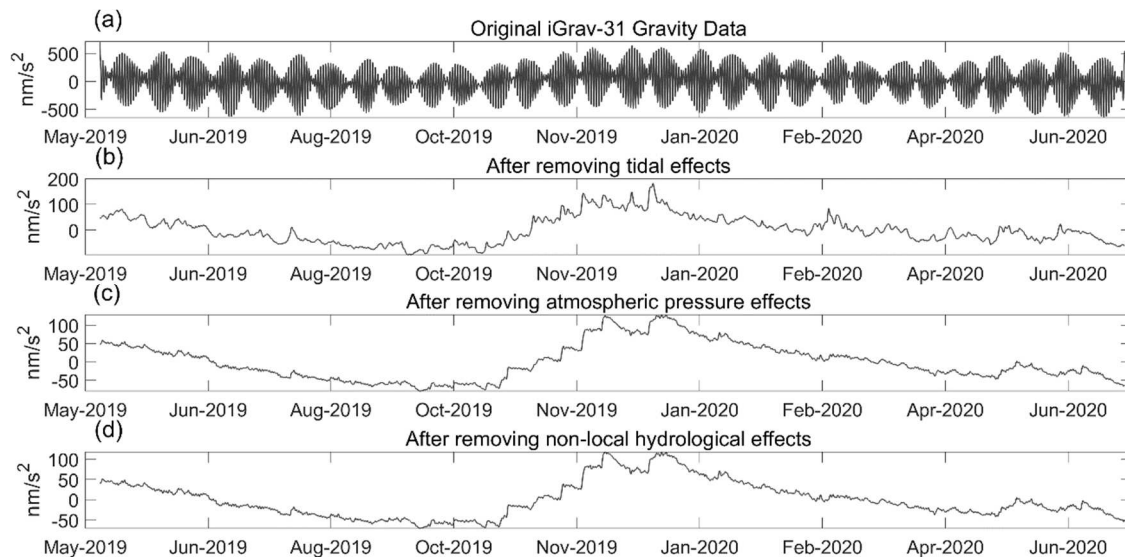
752 JH and SR contributed to the study conception and design. Material preparation, data collection
753 and analysis were performed by SK and SR. The first draft of the manuscript was written by
754 SK and SR. Major revisions were handled by SR. All authors commented on and corrected
755 previous versions of the manuscript. All authors read and approved the final manuscript.

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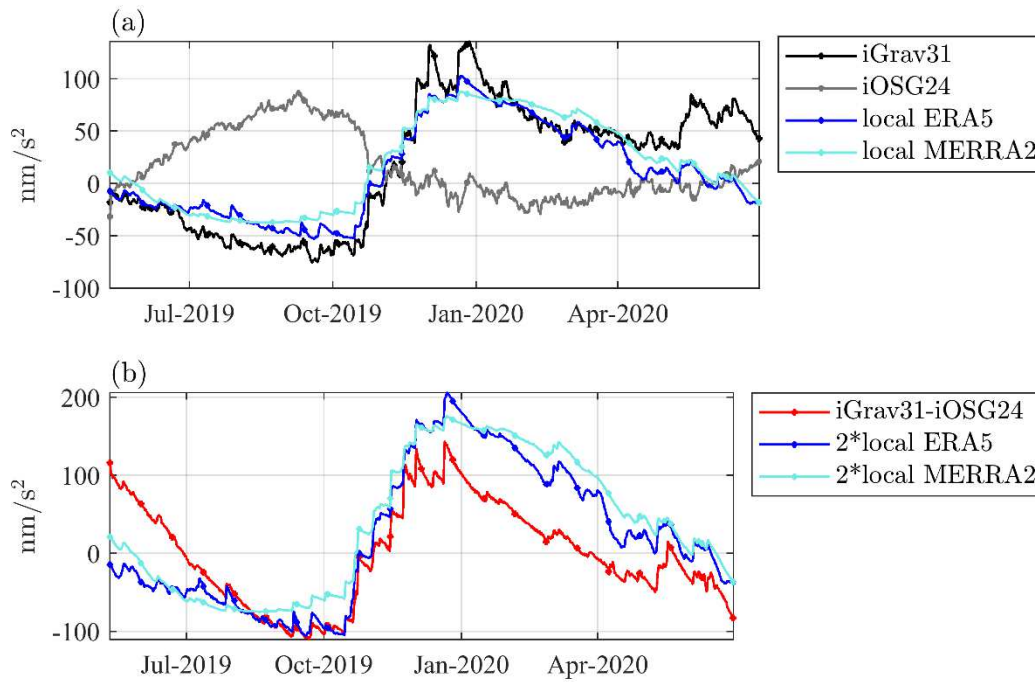
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 763 **Figure 1** A simplistic model of the Fontaine de Vaucluse hydrosystem with the LSBB site
 764 (modified from Carrière et al., 2016) in the south-east of France. The location of the two
 765 Superconducting Gravimeters (iGrav31 and iOSG24) is indicated: iOSG24 is located within
 766 the LSBB tunnel while iGrav31 is at the surface. The separation between the unsaturated and
 767 saturated zones is at around 800 m depth.
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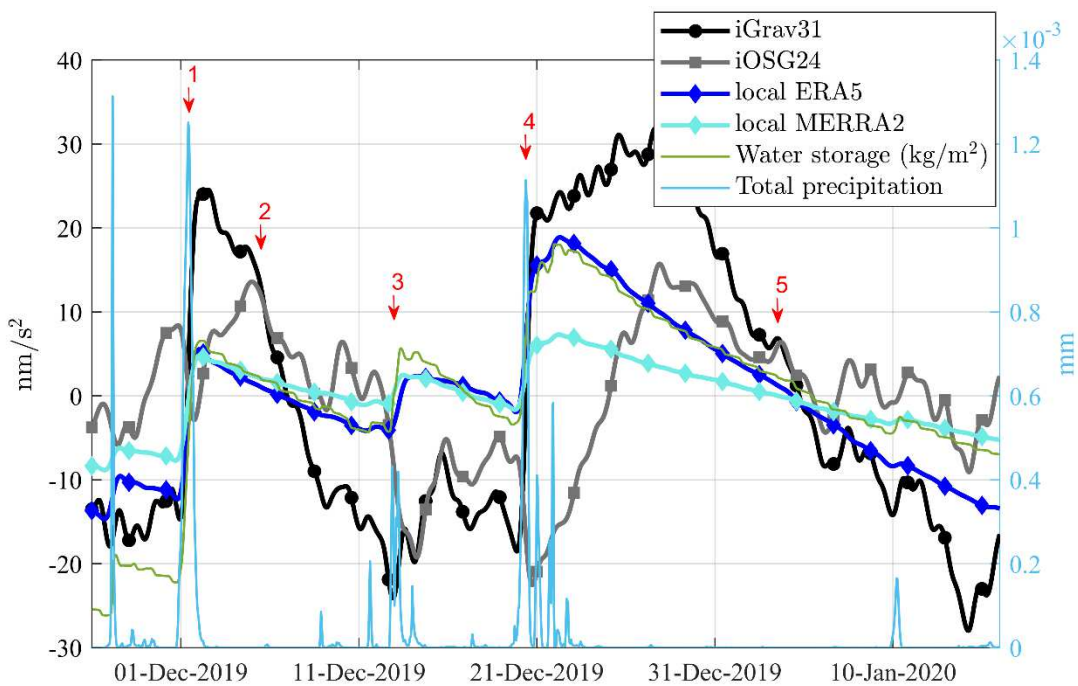


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 770 **Figure 2** Time-varying gravity recorded by iGrav31. (a) Original data showing mainly tidal
 771 variations. Gravity residuals after subtraction of (b) tides and polar motion effects. (c) Tides,
 772 polar motion and air mass effects (d) Tides, polar motion, atmospheric and non-local
 773 hydrological loading effects.
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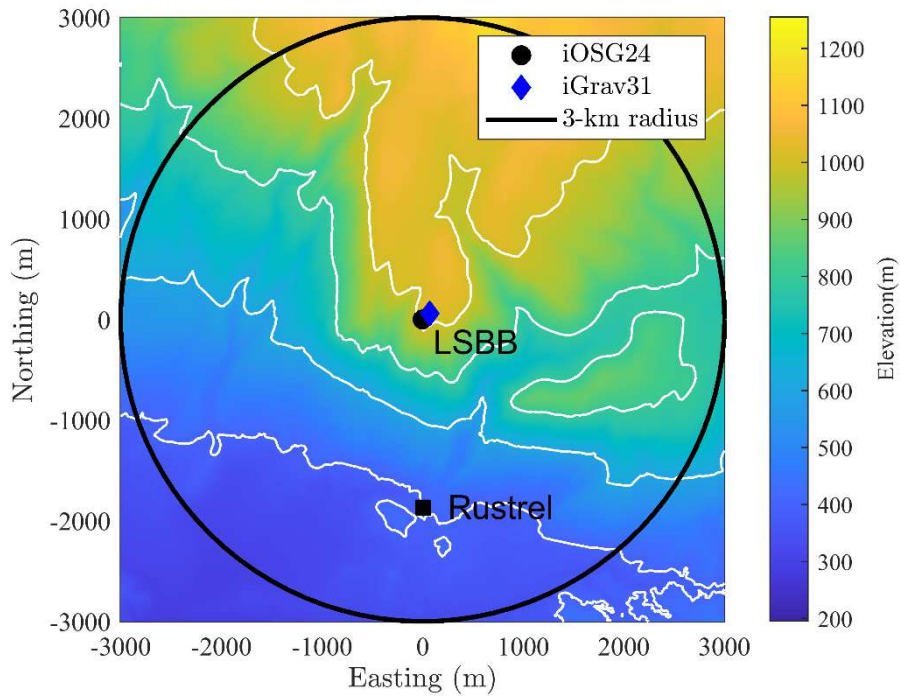
Figure 3 (a) Residual gravity time-series recorded by iGrav31 (black) and by iOSG24 (gray) compared to the gravity effect induced by the local contribution calculated from ERA5 (blue) and MERRA2 (cyan) hydrological models; (b) Differential gravity signal (red) compared to twice the gravity effect computed for iGrav31 corresponding to the local hydrological contributions from ERA5 (blue) and MERRA2 (cyan) global models.



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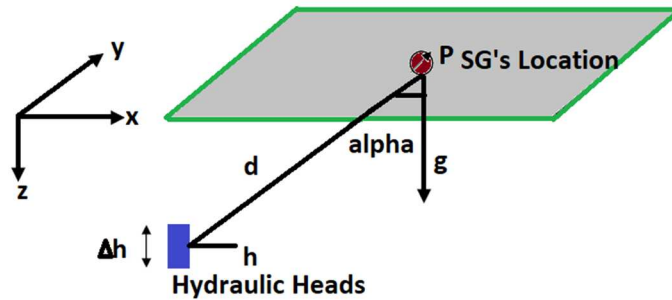
Figure 4 Zoom-in view of the residual gravity time-series recorded by iGrav31 (black) and by iOSG24 (gray) between December 2019 to January 2020 compared with the local gravity effect induced by hydrology computed from ERA5 (blue) and MERRA2 (cyan) global models. Observed total precipitation (mm) and water storage (kg/m²) used in MERRA2 modeling are also plotted in cyan and green.

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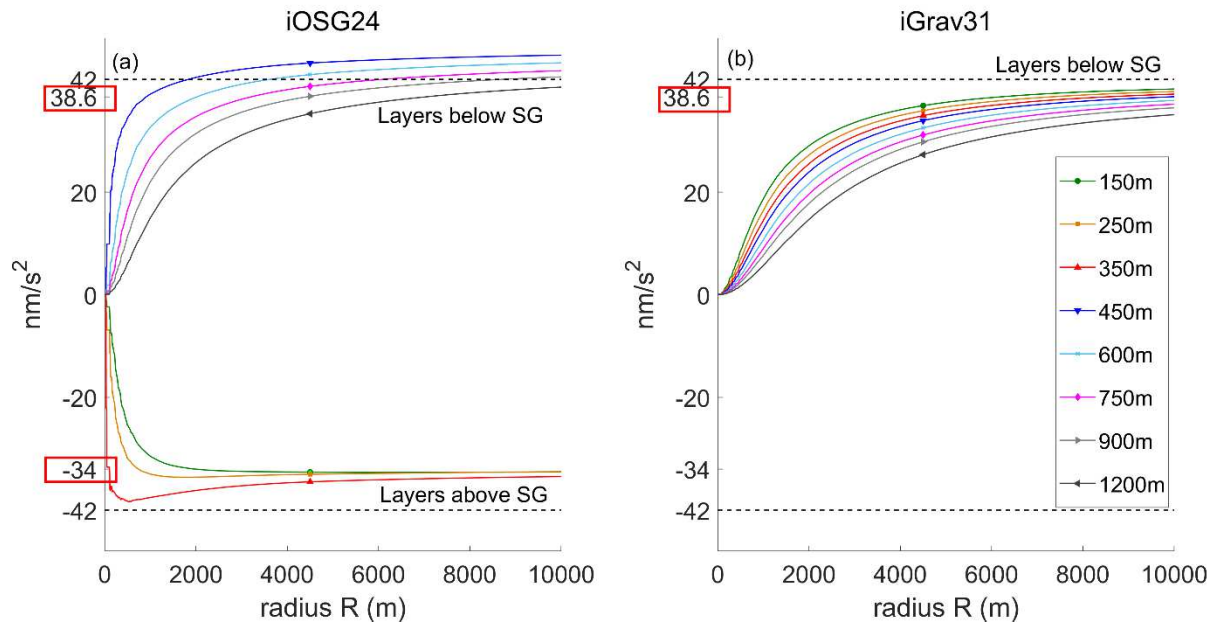
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Figure 5 Digital elevation model of the LSBB site centered on iOSG24 (black dot). The black circle indicates the radius of integration of a supposed water layer around iOSG24.



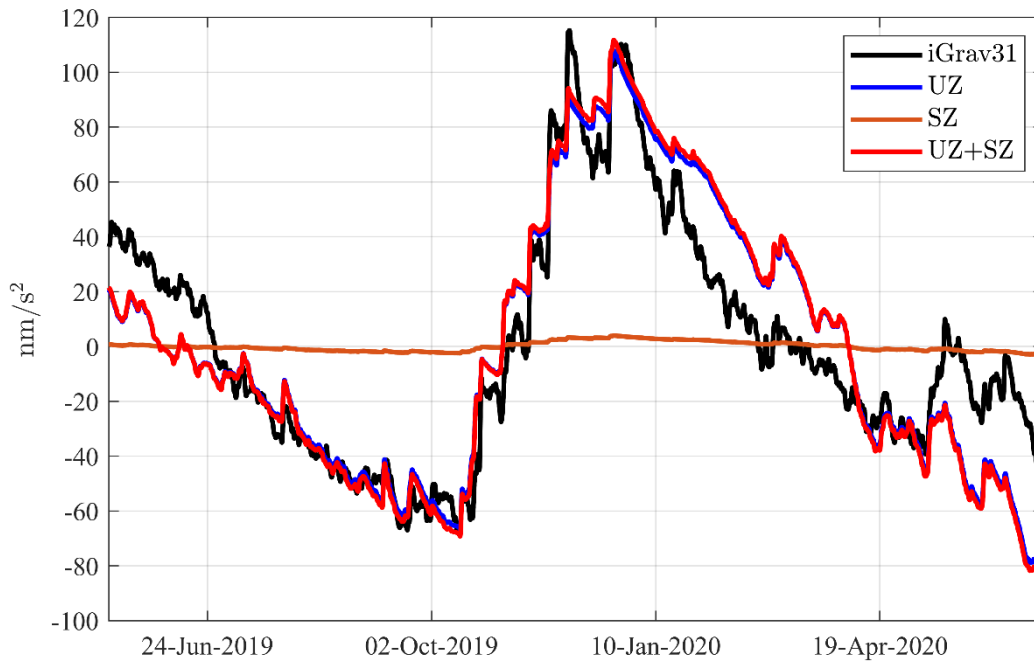
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Figure 6 Schematic view of the parameters used in the forward gravity modelling of the gravity effect due to water storage change in an aquifer.



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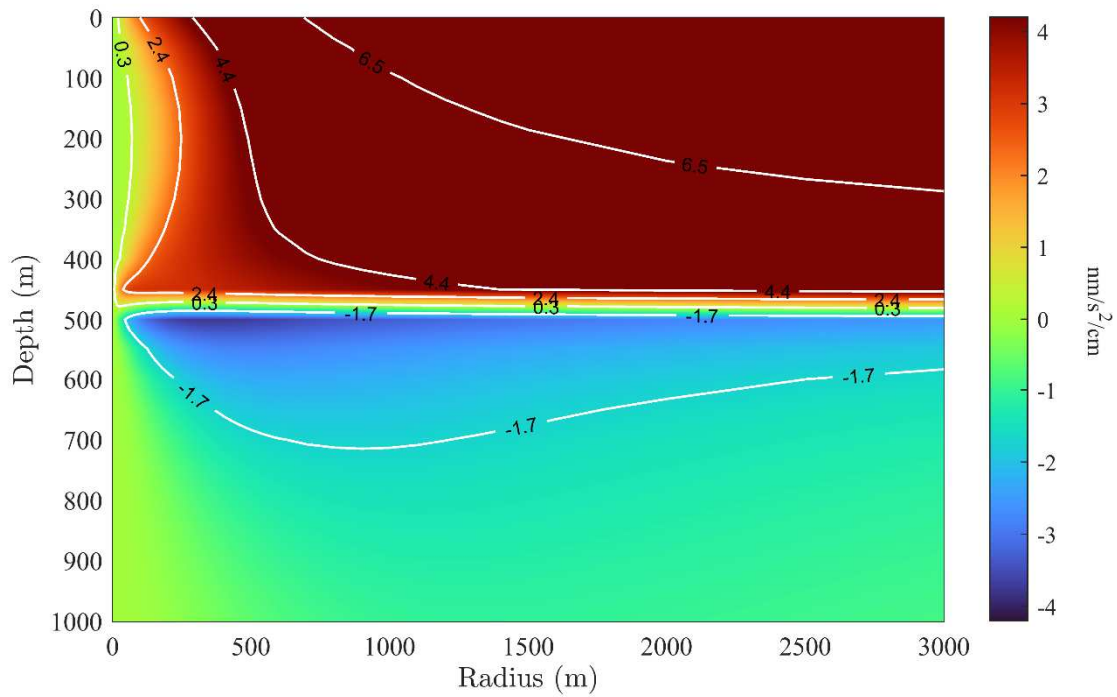
Figure 7 Gravity effects of the 10-cm uniform water layer w.r.t radii, for various depths (one color per depth). Red boxes mark the asymptotic gravity value for layers above/below the SGs.



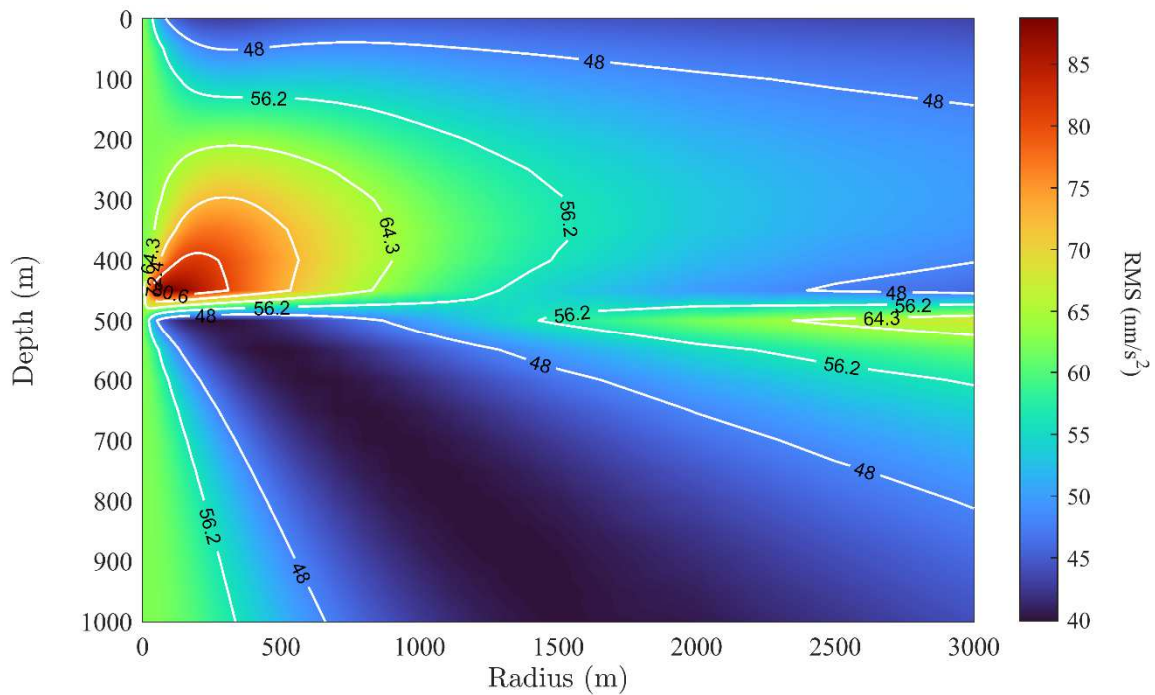
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Figure 8 Residual gravity time-series recorded by iGrav31 and the simulated gravity effects due to saturated (SZ), unsaturated (UZ), or sum of the two zones (UZ+SZ).

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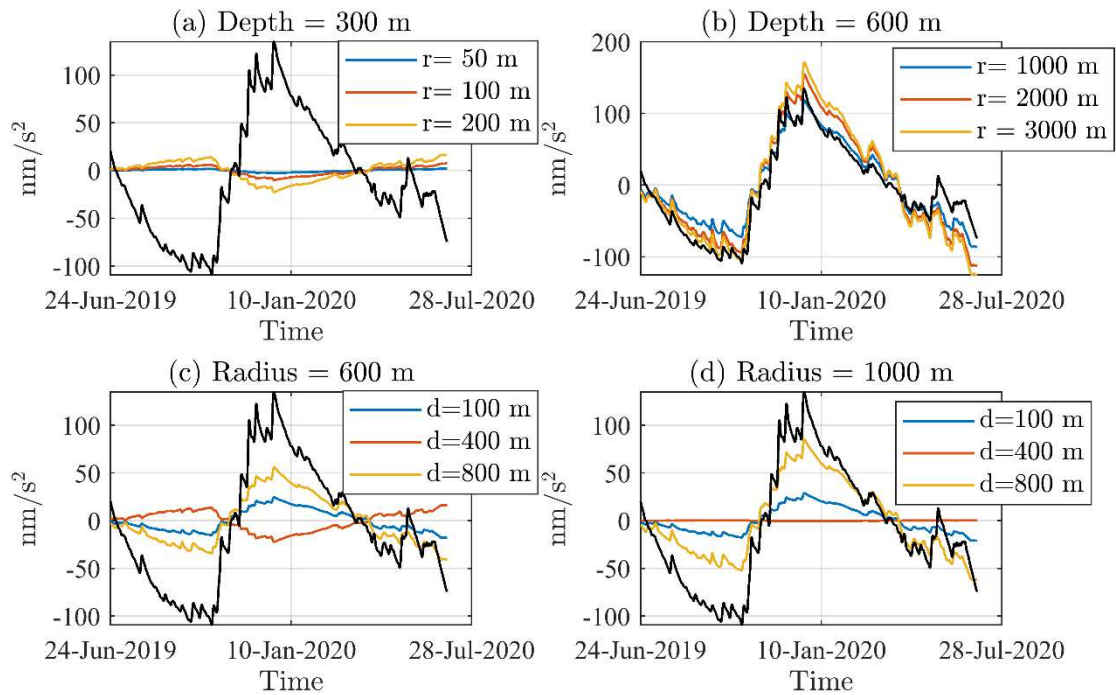


817 (b)
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819 **Figure 9** Predicted $\Delta g(iGrav31)-\Delta g(iOSG24)$ computed for various depths of the water layer
820 and various radii of integration. (a) The 2D admittance in $nm/s^2/cm$. The contour lines are
821 isolines for the admittance values in $nm/s^2/cm$. (b) root mean square (RMS) of the difference
822 between observed and predicted $\Delta g(iGrav31)-\Delta g(iOSG24)$ The contour lines are isolines for
823 the RMS values in nm/s^2 .
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Figure 10 Observed (black) and predicted (colored) $\Delta g(\text{iGrav31}) - \Delta g(\text{iOSG24})$ gravity signal for various depths of the water layer and integration radii. (a) For a depth of 300 m and integration radii of 50, 100 and 200 m; (b) For a depth of 600 m and integration radii of 1000, 2000 and 3000 m; (c) For an integration radius of 600 m and a water of layer at depths 100, 400 and 800 m; (d) For an integration radius of 1000 m and a water layer at depths 100, 400 and 800 m.