Measuring integral soil moisture variations using a geoelectrical resistivity meter

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1 Introduction

The attractional effect of time-varying mass distributions of environmental origin is a limiting factor in high resolution terrestrial gravimetry. Beside atmospheric air mass fluctuations, which is the largest environmental signal, mass fluctuations of water in the underground play an important role (e.g. Harnisch & Harnisch 2005, and references cited herein). After removing the Earth tides, the pole tide, and air pressure effects, the residual time series of superconducting gravimeters often show signals showing some correlation with hydrological data. One example from the Wettzell gravimeter shows a clear improvement of the residual gravity when groundwater corrections are applied (figure 1). At other stations a groundwater correction doesn't work well or groundwater data are not available. In some cases the modeled gravity effect of precipitation using the charge/discharge model of Crossley et al. (1998) helps reducing the hydrological signal. But both sensors do not contain the whole information how much water is stored in the underground.



Figure 1: Hydrological correction of residual gravity at Wettzell using groundwater data

Regarding the basic equation of the hydrological cycle (figure 2)

Precipitation = Evapotranspiration + Surface Runoff + Subsurface Runoff

it is very difficult to determine the present distribution of water in the different subsystems, because many parameters controlling the water flow are time dependent. The amount of water going back to the atmosphere by evapotranspiration is strongly affected by temperature, wind and vegetation cover. The surface runoff depends on the slope, soil type, rain intensity, vegetation and soil moisture, whereas the latter three are also functions of time. Subsurface runoff is extremely difficult to estimate.



Figure 2: Sketch illustrating the hydrological cycle

In terms of measurements, there is easy access to precipitation and groundwater level. But how much water is stored in the unsaturated zone? One appoach to solve this is the use of hydrological models, where the unknown quantities in the hydrological cycle are estimated on the base of meteorological, geographical and geological data (e.g. J2000 hydrological model, Krause et al. 2006). The other approach is to measure soil moisture.

2 Measuring soil moisture

There are many soil moisture sensors commercially available working in different ways and measuring different quantities. An overview of different sensor types is given in table 1. However, all sensors measure a physical property in the immediate vicinity of the sensor, generally probing less than 1 liter. How representative are these data for the whole matter affecting the gravimeter, and can the data extrapolated vertically or laterally? One could tackle this problem by an 3-D array of soil moisture sensors, but besides the costs, the installation is time-consuming and not practicable at depths greater than 2-3 m. As an alternative, electrical resistivity survey techniques acting at the surface and penetrating the underground yield an integral information of the electrical conductivity and easily reaches depths of several meters without digging or drilling.

Sensor type	Measured quantity
Gypsum block	Electrical conductivity
Tensiometer	Water tension
Time domain reflectometry (TDR)	Propagation delay of electromagnetic pulses (ϵ)
Frequency domain reflectometry (FDR)	Frequency shift of electromagnetic waves (ϵ)
Heat dissipation sensor	Thermal conductivity
Neutron probe	Energy loss of neutrons (H-atoms)

Table 1: Sensor types for measuring soil moisture

3 Electrical resistivity survey

The basic principle of electrical resitivity survey is sketched in figure 3. Two current electrodes C1 and C2 are supplied with a constant electrical current of DC or AC type. The potential drop between the two electrodes P1 and P2 being located in a defined distance between (or even outside) the current electrodes is a function of the resistivity of the soil. There is a sensitivity maximum at a certain depth, depending on the electrode configuration. The bigger the spacing of the electrodes is, the deeper is the sensitivity maximum. In this way information of different depth in gained. A 2-D or even 3-D mapping of the specific resistivity of the underground can be done by varying systematically the spacing and the location of the electrodes. The unit is $[\Omega m]$, which is the 1-D specific resistivity $[\Omega/m]$ times the cross sectional area of the probed body $[m^2]$.



Figure 3: Basic principle of electric resistivity survey

The electrical resistivity is mainly controlled by the material. Some rough values are 1-30 Ω m for clay, 10-50 Ω m for silt, 50-2000 Ω m for Sand and 10³-10⁶ Ω m for crystalline rocks. The material is taken as constant in time. Time-variable parameters affecting the electrical resistivity of the soil are water content and temperature. Variations of the salinity also affect the resistivity, should however play a minor role in silicate rocks.

Standard electrical resistivity meters have two current and two potential electrodes. Using this configuration, information at 1 location and 1 depth is obtained. In order to get spatial information, re-location of the electrodes is required. This procedure is not practicable for countinuous monitoring.

Since several years multielectrode devices are available, where a big number of electrodes are permanently installed. The configuration can be set up software-controlled that each electrode can operate as current or potential electrode. More than 100 electrodes can be operated simultaneously. All possible configurations are scanned automatically yielding data for different locations and depths. The obtained data set has to be inverted to get the true spatial information of the specific resistivity. An example of a section inversion using the RES2DINV software is given in figure 4. The measurement has been performed at the Fundamental Station Wettzell using 50 electrodes with 1 m spacing (figure 5).



Figure 4: Inverted section of geoelectric test line Wettzell; location given in figure 5



Figure 5: Map of the Fundamental Station Wettzell showing the locations of the geoelectrical test line, the groundwater and rain gauge, the soil moisture sensor and the gravimeter site.

4 Continuous monitoring

Electrical resistivity survey is normally used to map the kind and distribution of the underground material. The software controlling the scanning process is interactive and stops after a full cycle is complete. In order to create time series, the software has been modified in a way that a new cycle starts automatically after a given time interval. Each cycle creates one data file. Time series has been generated by extracting data points for a common location and apparent depth from each file. The result is shown in figure 6 for one selected depth, indicating lateral variations of the resistivity by one order of magnitude due to different materials (compare with figure 4). In a second step all data points representing one depth but different locations has been averaged and normalized with respect to the first data point. The resulting time series (figure 7) show no longer variations due to lateral inhomogeneities, but a systematic variation of the time series with depth.



Figure 6: Time series over 35 days with constant spacing (depth) and varying location



Figure 7: Time series normalized and averaged over all locations, constant spacing (depth)

5 Results

The obtained time series over 35 days show a general increase of the apparent resistivity with some long period variations, and clear daily cycles. Both signals become weaker with depth. In terms of soil moisture, the increase of resistivity could be interpreted as a drying of the soil during these sunny autumn days. The daily cycles however are not likely to be a soil moisture effect. The minimum occurs in the evening hours, which should be the maximum in the case of soil moisture. This is obviously a temperature effect causing a higher ion mobility in the uppermost soil layer during the warm evening hours. There is no phase shift with increasing depth, indicating that also measurements with big spacing are influenced by surface effects.

As no soil temperature data were available to verify the temperature hypothesis, a synthetic temperature curve has been generated using a simple one-dimensional heat flow model and air temperature data as boundary condition. The comparison of the model temperature in 0.3 m depth with the resistivity in 0.5 m depth (figure 8) shows a clear anticorrelation of both curves, indicating that the temperature has a major impact on the measured resistivity. A rough estimate is 2% resistivity change per degree temperature change.



Figure 8: Comparison with 1-dimensional thermal conductivity model, basing on air temperature data



Figure 9: Comparison with other hydrological data and residual gravity time series When taking the temperature corrected conductivity (inverted resistivity) as a reading of soil moisture and comparing it with other hydrological data, there is a partial correlation

observable (figure 9). The general trend is present in all data sets, whereas some signatures are visible in the conductivity data only. The residual gravity time series however shows only a weak correlation with the conductivity data. This might have different reasons,

- the location is not representative,
- the residual gravity still contains other signals (atmosphere),
- the contribution of soil moisture to gravity is weak,
- the presented method is inadequate.

From these first results it cannot be stated if this method is adequate to obtain an integral information of the soil moisture, but they encourage to further investigations. This includes relocation of the test site closer to the gravimeter, monitoring of the soil temperature, and creation of longer time series.

6 References

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