Groundwater

New Methods to Estimate 2D Water Level Distributions of Dynamic Rivers

by Samuel Diem[1,2], Philippe Renard[2], and Mario Schirmer[2,3]

Abstract

River restoration measures are becoming increasingly popular and are leading to dynamic river bed morphologies that in turn result in complex water level distributions in a river. Disconnected river branches, nonlinear longitudinal water level profiles and morphologically induced lateral water level gradients can evolve rapidly. The modeling of such river-groundwater systems is of high practical relevance in order to assess the impact of restoration measures on the exchange flux between a river and groundwater or on the residence times between a river and a pumping well. However, the model input includes a proper definition of the river boundary condition, which requires a detailed spatial and temporal river water level distribution. In this study, we present two new methods to estimate river water level distributions that are based directly on measured data. Comparing generated time series of water levels with those obtained by a hydraulic model as a reference, the new methods proved to offer an accurate and faster alternative with a simpler implementation.

Introduction

Over the past few decades there has been a shift in focus from river corridor channelization toward restoration. It has been recognized that rivers need more space for the purpose of flood protection (Woolsey et al. 2007). Furthermore, restoration measures such as widening of the riverbed, re-meandering stream reaches, and constructing gravel bars, should increase the exchange between rivers and groundwater, which is essential for the ecological health of a river (Brunke and Gonser 1997). On the other hand, riverbed widening can decrease travel times between rivers and pumping wells, which may increase the risk of drinking water contamination by pollutants or bacteria (Hoehn and Scholtis 2011).

Groundwater flow and transport modeling is a valuable tool to obtain a process-based understanding of (restored) surface water-groundwater systems. Compared to tools developed to work with artificial and natural tracers, a calibrated model provides quantitative conclusions on flowpaths, mixing ratios, and travel times (Wondzell et al. 2009). It is well known that river bed morphology affects the river water level distribution, which in turn affects or drives the exchange with groundwater (Woessner 2000; Cardenas et al. 2004; Cardenas 2009). Therefore, one of the prerequisites for the set up of a groundwater flow model of a real river-groundwater system is an accurate description of the water level distribution in the river.

Restored river systems may have complex water level distributions that need to be characterized by their full spatial (i.e., two horizontal dimensions) and temporal extent (i.e., for any discharge condition). Past small-scale field and modeling studies (10 to 100 m) applied one or two sets of detailed water level measurements in one or two dimensions (Wroblicky et al. 1998; Storey et al. 2003; Lautz and Siegel 2006; Wondzell et al. 2009). On

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the scale in the order of kilometers, which is more relevant for practical problems, this approach is not applicable. Instead, the extraction of river water level information from a hydraulic model might be a proper solution (Doppler et al. 2007; Derx et al. 2010; Engeler et al. 2011). However, the setup of a hydraulic model is time consuming and requires a large amount of data input, that is, the river bed bathymetry and water level information for the calibration and validation process.

In this study, we present two alternative methods to estimate water level distributions of highly dynamic rivers in the context of modeling river-groundwater systems at scales in the order of kilometers. The two methods combine continuous and periodic water level measurements from different locations in order to account for the spatial and temporal variability of the water level distribution. We predicted water levels at several locations at a restored reach of the perialpine Thur River and compared them with water level predictions of a reference method, which is based on an existing hydraulic model. Finally, we point out the advantages and disadvantages of the different methods and discuss the optimal use of each method when assigning the water level distribution of restored and dynamic river systems to groundwater flow and transport models.

Interpolation Methods to Estimate River Water Levels

Problem Setup

Instrumentation of a 1 km river reach is typically composed of two water level gauges, one upstream and one downstream of the field site. For groundwater flow modeling, river water levels need to be estimated at each river boundary node between the two gauging stations. The simplest method assumes a constant gradient (linear interpolation) in the longitudinal direction and a zero gradient in the lateral (transverse) direction. Although this might be adequate for a channelized system, for restored corridors the dynamic river bed morphology might lead to a nonlinear longitudinal water level profile and to lateral water level gradients. Furthermore, the water level distribution might change as a function of the discharge.

The basic idea of the two new interpolation methods is to combine continuous water level records from water level gauges with water levels measured periodically under different discharge conditions. The latter are measured at "fixpoints," which are distributed throughout the river reach to refine the water level distribution at locations where the installation of a water level gauge is technically difficult or simply too expensive. A fixpoint is defined as a reference point in the river, for example, a point on a construction rock or a steel rod, whose altitude is known. By establishing a mathematical relationship between the water level data at the gauging stations and the fixpoint, the water level at the fixpoint can be estimated for any measured water level at the gauging station.

We consider the river as a two-dimensional (2D) domain, which is discretized by multiple lines parallel to the main flow direction of the river and several sections of support points (s) perpendicular to the flow direction (Figure 1). The key task of the interpolation methods is to estimate a water level h^{s} at each support point from any water level measured at the gauging station (i.e., for any discharge condition). Support points are placed at a location where a fixpoint (F) or a gauging station (G) exists. One fixpoint per section is enough unless lateral water level gradients are observed, in which case a fixpoint must be defined on both sides of the river (Figure 1). The periodic water level measurements at the fixpoints are denoted as h^{F}, while the continuous water level measurements at the gauging station are denoted as h^{G}. The estimation of h^{s} consists of three steps:

1. Establish a mathematical relationship ϕ between the measurements h^{F} and h^{G}.
2. Use ϕ to compute h^{F} for any time step or time series: $h^{\mathrm{F}} \approx \phi[h^{\mathrm{G}}]$.
3. Estimate the water levels or water level time series h^{s} for the support points located on the same section as the fixpoint F.

Different options for these steps have been considered for the two new methods, which are referred to as alternative methods. In order to compare these alternative methods with a reference, we developed a third method that is based on a hydraulic model and is referred to as the reference method. These three methods are described in the following sections. The explanation is descriptive in order to convey the main idea. A thorough mathematical development of the methods and their application to real data is presented in Appendix S1.

Once the set of lines and support points with water levels h^{s} has been obtained using one of the three methods, the final interpolation of the water levels from the support points to the river boundary nodes of the numerical model has to be performed. This step is identical for all three methods and is accomplished by a one-dimensional linear interpolation along the lines, which implies that each of the lines is mapped by a curvilinear system of longitudinal coordinates to account for different curvatures of the lines. More precisely,

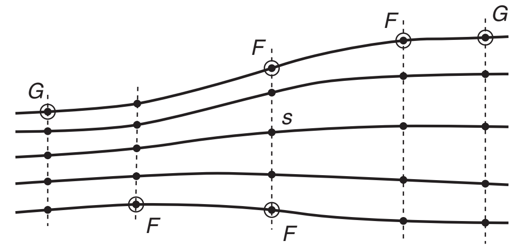

Figure 1. Schematic representation of a river system with multiple lines and sections of support points (s, black dots). Gauging stations (G) and fixpoints (F) are shown as black circles.

each of the river boundary nodes of the groundwater model is projected perpendicularly onto the closest line and the water level is linearly interpolated between the upstream and downstream support point. Some simulation codes (e.g., FEFLOW) offer tools to accomplish this final interpolation step.

Method 1: Regression of Measured Data (RM)

The first alternative method uses a polynomial regression technique to obtain the mathematical relationship ϕ. The method requires a continuous water level time series at one gauging station G and periodic water level measurements at a fixpoint F (Figure 2). Each time a measurement is made at the fixpoint, the corresponding water level at the gauging station can be extracted from the continuous water level time series. A polynomial equation is then fitted to the data pairs, which constitutes the mathematical relationship ϕ. The polynomial order has to be chosen according to the range and the characteristics of the data. A guideline to defining the polynomial order is given in Appendix S1. Applying ϕ to any water level time series measured at the gauging station produces predictions of corresponding water level time series at the fixpoint. If there are two fixpoints on the same section to capture lateral gradients, a separate polynomial equation is fitted to the corresponding data pairs. The gauging station G is denoted as determining gauging station G^{d}, as its water level uniquely defines the water level at the fixpoint.

The estimation of the water level at the support points from the water level at the fixpoint is made in the simplest possible manner. If no lateral gradients exist, the water level of the fixpoint is assigned to all of the support points located on the same section. However, if a second fixpoint was installed on the same section to capture a lateral gradient, assigning the water levels to the support points should be based on field observations. For example, if a discrete step forms the lateral gradient, the water level at each support point can be determined from the most representative fixpoint. Otherwise, a linear interpolation between the two fixpoints might be an appropriate solution.

Method 2: Interpolation of Measured Data (IM)

The second alternative method uses an interpolation approach that requires two gauging stations. One of them has to be defined as determining gauging station G^{d}. The fixpoints can be located between or outside of the two gauging stations.

This IM method is based on a conceptual behavior model. According to the model, the river bed morphology exerts a high influence on the river water level distribution under low-flow conditions, potentially leading to nonlinear longitudinal water level profiles or lateral water level gradients. As the water level rises, the influence of the river bed morphology on the water level distribution decreases, and at some point lateral water level gradients disappear and a linear longitudinal water level profile is reached. In general, the IM method describes the water level at a fixpoint by an interpolation function (ϕ) that

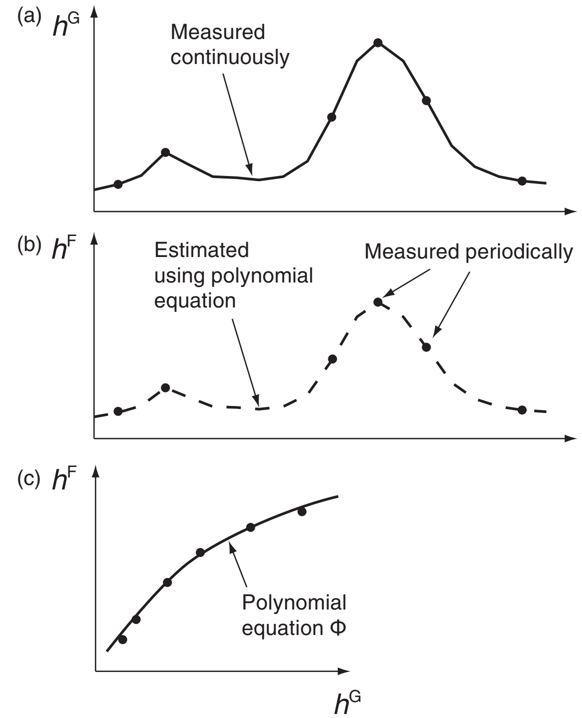

Figure 2. Illustration of the regression approach. (a) Continuous water level time series h^{G} measured at the gauging station. (b) Periodic water level measurements h^{F} at a fixpoint (black dots). For the same measurement times, water levels at the gauging station are extracted in (a). (c) ϕ is established between h^{G} and h^{F} by polynomial regression and is used to estimate the water level time series (dashed line) at the fixpoint in (b).

consists of a linear interpolation term l between the water levels at the two gauging stations (h^{G} and h^{Gd}) and a deviation D from the linear trend (Figure 3). Following the conceptual behavior model, the deviation D is at a maximum, $D = a$, when water levels at the determining gauging station h^{Gd} are smaller than h^{a}. On the other hand, the deviation goes to zero for water levels higher than a threshold water level h^{thresh}. Between h^{a} and h^{thresh}, the deviation D is assumed to decrease linearly (Figure 3). Similarly, we can describe the water level at a second fixpoint on the same section by the water level of the first fixpoint plus a lateral difference D^{L}. Again, D^{L} is at a maximum for low water levels and goes to zero above a threshold water level.

To estimate the parameters a, h^{a}, and h^{thresh}, the deviation from the linear trend has to be calculated for each measurement at a fixpoint and plotted against the corresponding water level at the determining gauging station. More details on how to estimate the parameters for the IM method are given in Appendix S1. The estimation of the water level at the support points is identical to the procedure described in the previous section for the RM method.

Method 3: Regression of Hydraulic Model Data (RH)

To compare the two alternative methods with an independent reference, we developed a third method

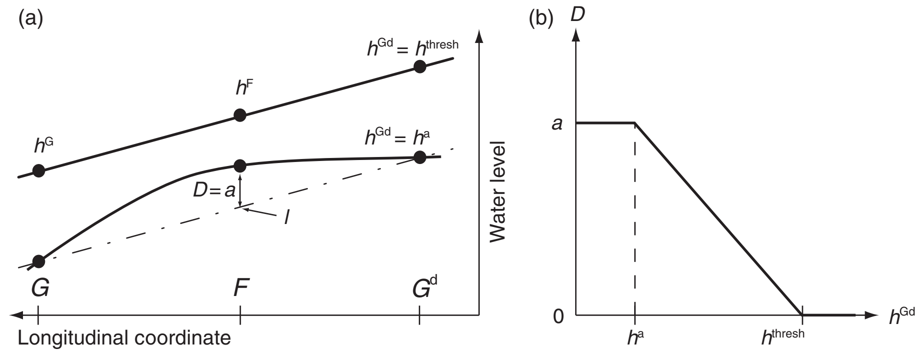

Figure 3. Illustration of the interpolation approach. (a) Two longitudinal water level profiles for the conditions $h^{Gd} = h^{a}$ and $h^{Gd} = h^{thresh}$. (b) Deviation from the linear trend (D) as a function of h^{Gd}.

that is based on data from an existing hydraulic model of the main river channel in the section of interest. The hydraulic model output consists of 2D water level distributions, each corresponding to a specific discharge condition. Similar to the RM method, the RH method applies a polynomial regression technique to obtain the mathematical relationship ϕ. However, the relationships are based on water levels extracted from the hydraulic model output at each support point and at the location of the determining gauging station, and not on measured water levels.

Application to the Field Site Niederneunforn

Field Site

The perialpine Thur River drains a catchment area of 1730 km^2 and originates in an alpine region that reaches its highest point on Mount Säntis (2502 meter above sea level, m asl). The Thur River is the largest river in Switzerland without a retention basin. This leads to a very dynamic discharge regime ranging from 3 to 1100 m^3/s with an average of 47 m^3/s. The field site (Figure 4) is located approximately 12 km upstream of the confluence with the Rhine River. In the western part of the field site, restoration measures were realized in 2002. Restoration measures were forbidden in the upstream section to protect the water quality at the nearby pumping station, where a pumping well supplies the community of Niederneunforn with drinking water. The field site was instrumented with more than 80 piezometers (2″) during the interdisciplinary RECORD project (Restored corridor dynamics, http://www.cces.ethz.ch/projects/nature/Record; Schneider et al. 2011). The aquifer has a thickness of 5.3 $\pm$ 1.2 m and its hydraulic conductivities were estimated to range from 4×10^{-3} to 4×10^{-2} m/s (Diem et al. 2010; Doetsch et al. 2012). The silty sand of the alluvial fines on top of the aquifer has a much lower hydraulic conductivity and can be regarded as the semi-confining unit, with a thickness of 0.5 to 3 m.

The Thur River has a width of 50 to 100 m (Figure 4). In the restored section a large gravel bar has evolved during the past few years. To the north, a disconnected branch of the river exists, which is only flooded at high river stages (>200 m^3/s). The longitudinal river water level profile does not have a linear shape for low-flow conditions. In the upstream 400 m of the river the gradient is 0.5‰ and in the downstream 800 m it is 2‰. In the middle of the river reach, lateral water level gradients occur during low-flow conditions. These lateral surface water level differences exist due to the asymmetrical river bed morphology and can reach up to 0.4 m. Two side channels (north and south) flow parallel to the river and have an average width of 4 to 8 m. Two beaver dams are located in the northern side channel. The upstream dam has a significant effect on water levels, creating differences of up to 0.5 m.

A 2D hydraulic model of the Thur River was developed based on the bathymetry measured in September 2009 (Pasquale et al. 2011). River bed cross sections with an average spacing of 50 m were interpolated using the technique presented by Schäppi et al. (2010). The modeling results comprise 19 steady-state simulations for flows ranging between 10 and 650 m^3/s, and provide water level altitudes at each raster cell (2×2 m). The hydraulic model does not include the side channels and the disconnected branch and is considered to be valid until the major flood events of June 2010.

Data Collection and Method Implementation

We installed two water level gauges in the main channel of the river, as well as in both side channels, and one in the river branch (Figure 4). Three of these water level gauges are maintained by the Agency for Environment of the canton Thurgau. The sensors (DL/N 70, STS AG, Switzerland) have been continuously measuring pressure, temperature, and electrical conductivity (EC) at 15-min intervals since April 2010 (error of single measurement: $\pm$0.1% for pressure, $\pm$0.25% for temperature, and $\pm$2% for EC, according to the manufacturer's manual). We installed sensors of the same type in selected piezometers (Figure 4). The raw data were processed in order to remove outliers, to subtract the barometric air pressure, and to transform the pressure data to absolute water levels (m asl). To have more information on the water level distribution between the water level gauges, we installed

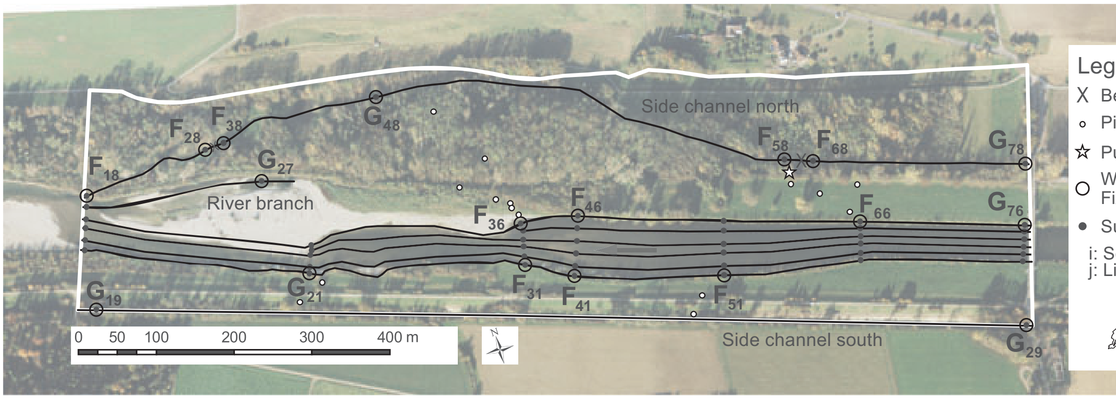

Figure 4. Field site located in Niederneunforn at the Thur River with indicated lines, support points, fixpoints, and gauging stations. Selected piezometers are shown as well. The white polygon shows the perimeter of method implementation.

several fixpoints along the river and the northern side channel. We chose the fixpoint positions according to the location of piezometer transects and visibly steep water level gradients (e.g., hydraulic jumps at beaver dams) or lateral gradients (central part of the river). We defined an indexing system that allows a distinct identification of each point. The first index refers to the section number and the second index to the line number. We leveled the absolute height of the fixpoints using a high-precision differential GPS (Leica GPS1200; Leica Geosystems AG, Switzerland) and a leveling device (Sprinter 100M, Leica Geosystems AG, Switzerland). We measured water levels at these fixpoints periodically between February and May 2011 covering a discharge range of 10 to 100 m^3/s.

Based on the resulting dataset, we implemented both alternative methods and the reference method, each of which covered the three steps described in the "problem setup" section. As the hydraulic model did not include the disconnected branch and the side channels, we coupled the RH method to the RM method. More details on the method implementation can be found in Appendix S1.

Comparison of the Interpolation Methods

To evaluate the performance of the two alternative methods, we applied each of the interpolation methods (RM, IM, and RH) to generate water levels for a 1-month period (May 26, 2010 to June 30, 2010, vertical lines in Figure 5) at each fixpoint and gauging station within the river domain. The upstream gauging station G_{76} was used as the determining gauging station for each of the methods. We plotted the generated time series of water levels (30-min intervals) of the reference method (RH) against the time series of both alternative methods (RM and IM). Figure 6 shows examples for one gauging station and two fixpoints. If the generated water levels were the same for each time step, the dots of the scatter plots would be located on a line with a slope of one. Deviations from this line correspond to deviations of the RM/IM methods from the RH method and were quantified by a mean and a standard deviation (Figure 6, Table 1).

The scatter plots for the gauging station G_{21} are identical for the RM and the IM method, which both used the gauging station data directly for this point. The water levels of the RH method, however, were generated based on the hydraulic model. Therefore, these two plots actually compare the measured water level data with the water level predictions of the hydraulic model. The match was good for lower water levels but deviations of up to 0.5 m occurred for the peak flows during flood events. As only a small portion of the data was subject to such large errors, the mean error of water level prediction with the RH method was small (1 cm). The uncertainty of the water level prediction is reflected in the standard deviation of 10 cm (Table 1).

The scatter plots of the fixpoint F_{41} revealed a difference in the behavior of the two alternative methods. The water level predictions for the study period (Figure 5) exceeded the range of water levels measured at F_{41} (371.6 to 372.5 m asl). Depending on the polynomial fit, the RM method is likely to fail for predictions outside of the measured range, as there are no data points constraining the polynomial equation. In our case, the parabola that was fitted to the data at the fixpoint F_{41} was obviously too narrow in order to reliably predict water levels beyond the measured range. Correspondingly, the deviations with respect to the RH method showed a mean error of

Table 1
Mean (μ_{error}) and Standard Deviations (σ_{error}) of the Residuals (m) Between the Time Series of Water Levels Generated by Both Alternative Methods and the Reference Method at the Gauging Stations and Fixpoints Within the River Domain

	F_{18}	G_{21}	F_{36}	F_{31}	F_{46}	F_{41}	F_{51}	F_{66}	G_{76}
RM									
μ_{error}	0.07	−0.01	0.05	0.06	−0.08	−0.11	−0.01	0.06	0.00
σ_{error}	0.12	0.10	0.04	0.04	0.42	0.42	0.07	0.05	0.00
IM									
μ_{error}	0.11	−0.01	0.16	0.17	0.12	0.06	−0.04	0.02	0.00
σ_{error}	0.16	0.10	0.09	0.08	0.08	0.09	0.04	0.02	0.00

−11 cm and a standard deviation of 42 cm. The IM method performed better for the peak flow water levels. As the water level predictions at a fixpoint are always bounded by two measured water levels at the upstream and downstream gauge, the stability beyond the measured range was better for the IM method. For the fixpoint F_{66}, both the RM and the IM method performed well despite the systematic positive offset for the RM method at high water levels. This offset might be attributed to the problem described above for the fixpoint F_{41}.

Most of the mean errors and their standard deviations for the fixpoints within the river domain varied between 1 and 10 cm (Table 1). We considered this level of error to be acceptable as the reference RH method itself had an accuracy of ±10 cm at G_{21}. As F_{76} was the determining gauging station for all fixpoints in the river, predictions were identical for all methods, explaining the values of zero in Table 1. The high standard deviations for the fixpoints F_{41} and F_{46} for the RM method can be blamed on the failure of the regression approach beyond the measured data.

The alternative methods—especially the IM method —tend to show higher deviations (>10 cm) from the RH method at the fixpoints on sections 1 to 4, which are located in the restored part of the river. We found evidence that the major flood event on June 17 through 24, 2010 (Figure 5) led to a change in the river bed morphology in the restored river section and to a corresponding change in the relationship between the water levels at the determining gauging station and at the fixpoints. First, the scatter plot for the gauging station G_{21} (Figure 6) reveals two distinct regimes at the lower end of the water level spectrum for the 1-month period that covers the major flood event. Second, after the flood event there is a clear and permanent decrease (∼10 cm) in the difference between the water levels at F_{46} predicted with the RH method, and groundwater heads measured at a nearby piezometer (Figure 5). The drop in the water level difference could be due to a change in the river bed morphology after the flood, resulting in an underprediction of water levels by the RH method, which assumed a constant morphology. Because the data for the RM and the IM method were collected after this morphologically active flood event, they might have captured the higher water levels and accordingly led to an overestimation of water levels compared to the RH method. Therefore, the differences in predicted water levels among the different methods were presumably not only caused by structural artifacts, but also by real differences due to morphology changes in the river.

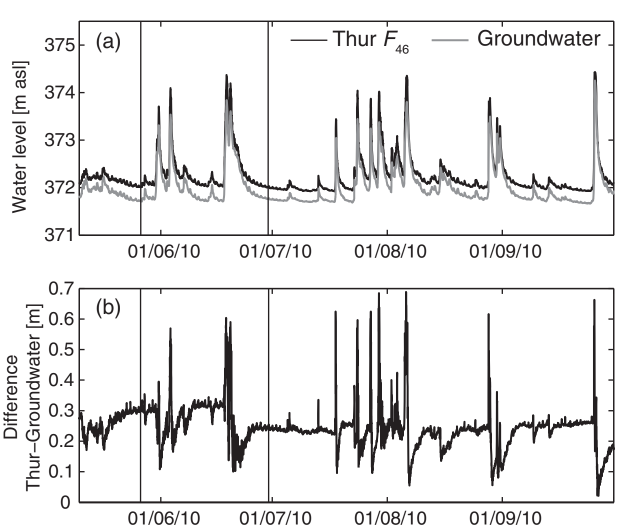

Figure 5. (a) Water level time series in the Thur River at fixpoint F_{46} predicted with the RH method and measured groundwater heads in a nearby piezometer between May and October 2010. (b) Calculated water level difference between the river and groundwater. The vertical lines indicate the 1-month period used for creating the scatter plots of Figure 6.

Discussion

A hydraulic model has the advantage of being physically based, which allows a wide range of discharge conditions to be simulated. This makes water level predictions by the RH approach very robust. However, setting up a hydraulic model is time consuming, and needs both a large dataset and to be thoroughly calibrated. At our field site, the hydraulic model was not able to include the water levels of the disconnected river branch, because during low-flow conditions this branch is fed by groundwater. Furthermore, the hydraulic model did not cover the side channels. The RH method had to be coupled to the RM method in order to include the full surface water level distribution required for the assignment of boundary conditions in a groundwater model.

Compared to the RH method, both alternative methods presented in this study (RM and IM) provide a more efficient way of predicting water level distributions in a hydraulically and morphologically varying environment. First, the accuracy of the water level predictions with the alternative methods was in the same range as the accuracy of the reference RH method itself. Second, the alternative methods require minimal data and computational effort, making them simpler and faster to implement than the hydraulic model.

In comparison to the IM method, the RM method benefits from the regression approach, which is fast and straightforward in its implementation and its application. However, the RM method does not provide reliable water level predictions when they exceed the range of water levels measured at the fixpoints. This in turn is the strength of the IM method, whose water level predictions are always bounded by water levels measured at two gauging stations. Furthermore, the IM method is based on a conceptual behavior model that is physically consistent. Even though the IM method is empirical and more complex in its implementation, the measured data at most of the fixpoints supported the method's underlying assumptions.

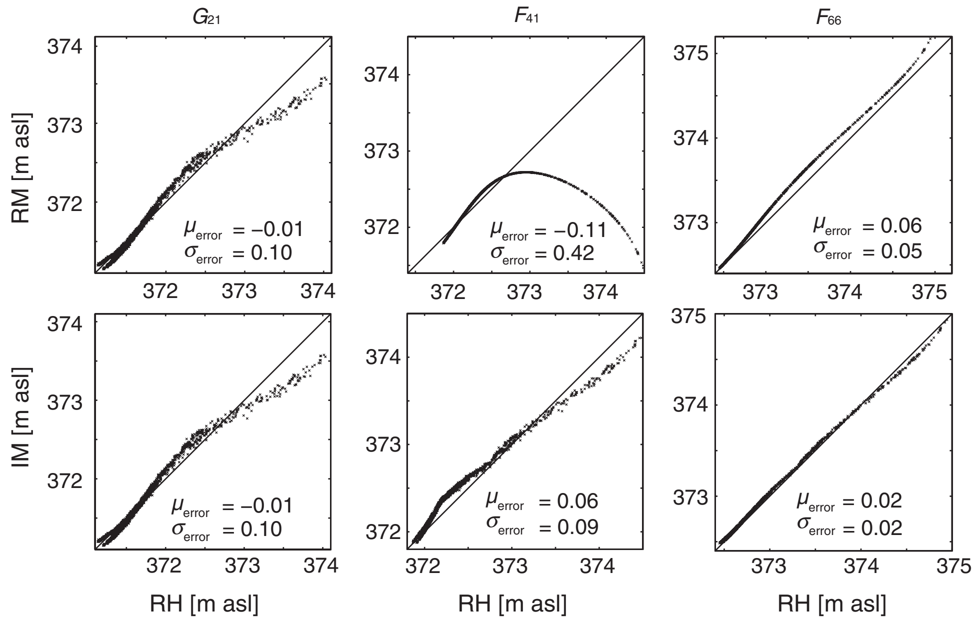

Figure 6. Scatter plots for one gauging station (G_{21}) and two fixpoints (F_{41} and F_{66}) within the river domain. The time series of water levels generated by the reference method RH are plotted against the time series of water levels generated by the RM (top 3 figures) and the IM method (bottom 3 figures). Mean (μ_{error}) and standard deviations (σ_{error}) of the residuals between the alternative methods and the reference method are indicated.

Conclusions

Several field and modeling studies have shown that the water level distribution in rivers exerts an important influence on the exchange between rivers and groundwater. In this study, we presented two new methods to define spatial and temporal river water level distributions for the purpose of modeling surface water-groundwater systems. The basic idea is to record water levels continuously at water level gauges and measure water levels periodically under different discharge conditions at fixpoints to refine the water level distribution at locations where it is technically difficult or too expensive to install a gauging station. The RM method applies a polynomial regression approach for the prediction of water levels at fixpoints as a function of the corresponding water levels at the determining gauging station, while the IM method uses an interpolation approach between two gauging stations. To compare these alternative methods to a reference method, we developed a third method, which is based on water level data from a 2D hydraulic model and also applies a regression approach (RH). The hydraulic model has the clear advantage of being physically based and covering a wide range of discharge. On the other hand, the alternative methods are simpler and faster in their implementation, while still being able to account for typical hydromorphological features of dynamic (restored or natural) river sections (e.g., nonlinear longitudinal water level distributions, lateral water level gradients, disconnected river branches, and hydraulic jumps).

We compared water level time series generated by both alternative methods with those generated by the reference method at all of the fixpoints located in the 1.2 km long river reach of our field site. For most cases, the accuracy of the water level predictions of the alternative methods was comparable to the accuracy of the reference method itself. In addition, we found evidence that the river bed, and hence the water level distribution for a given discharge condition, changed between the implementation of the reference and the alternative methods. This change in river bed morphology might have contributed to some of the larger deviations among water level predictions.

The results of this study allow us to recommend both alternative methods for the river water level assignment in future modeling studies of river-groundwater systems at scales in the order of kilometers. The RM method is straightforward in its implementation, but is limited to water level predictions within the range of measurements made at the fixpoints. If discharge conditions beyond the measured range have to be simulated, we recommend the use of the IM method instead.

Each of the presented methods has limitations in terms of accuracy in water level predictions. Even though we consider each of the methods to be accurate, water level predictions will differ for a specific discharge condition. The impact of the river water level uncertainty on key predictions of groundwater models as exchange flux or groundwater residence time, both in steady-state and transient conditions, could be the basis for future research.

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Supporting Information

Additional Supporting Information may be found in the online version of this article:

The following supporting information is available for this article:

Appendix S1. Detailed description of the interpolation methods and their implementation at the field site Niederneunforn.

Figure S1. Field site Niederneunforn with indicated lines, support points, fixpoints, and gauging stations.

Figure S2. Schematic representation of a curvilinear coordinate system.

Figure S3. Illustration of the notations used for describing the RM method.

Figure S4. Schematic representation of the regression approach.

Figure S5. Application of the regression approach according to the RM method.

Figure S6. Illustration of the notations used for describing the IM method.

Figure S7. Schematic illustration of the interpolation approach.

Figure S8. Application of the interpolation approach according to the IM method.

Figure S9. Illustration of the notations used for describing the RH method.

Figure S10. Application of the regression approach according to the RH method.

Table S1. Determining gauging stations of each fixpoint and gauging station.

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