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An Approach to Improve Direct Runoff Estimates and Reduce Uncertainty in the Calculated Groundwater Component in Water Balances of Large Lakes

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Abstract

Groundwater is important in the overall water budget of a lake because it affects the quantity and quality of surface water and the ecological health of the lake. The water balance equation is frequently used to estimate the net groundwater flow for small lakes but is seldom used to determine net groundwater flow components for large lakes because: 1) errors accumulate in the calculated groundwater term, and 2) there is an inability to accurately quantify the direct runoff component. In this water balance study of Lake Pyhäjärvi (155 km^2) in Finland, it was hypothesized a hydrograph separation model could be used to estimate direct runoff to the lake and, when combined with a rigorous uncertainty analyses, would provide reliable net groundwater flow estimates. The PART hydrograph separation model was used to estimate annual per unit area direct runoff values for the watershed of the inflowing Yläneenjoki River (a subwatershed of the lake) which were then applied to other physically similar subwatersheds of the lake to estimate total direct runoff to the lake. The hydrograph separation method provided superior results and had lower uncertainty than the common approach of using a runoff coefficient based method. The average net groundwater flow into the lake was calculated to be +43 mm per year (+3.0% of average total inflow) for the 38 water years 1971 to 2008. It varied from -197 mm to 284 mm over that time, and had a magnitude greater than the uncertainty for 17 of the 38 years. The average indirect groundwater contribution to the lake (i.e., the groundwater part of the inflowing rivers) was 454 mm per year (+32% of average total inflow) and demonstrates the overall importance of groundwater. The techniques in this study are applicable to other large lakes and may allow small net groundwater flows to be reliably quantified in settings that might otherwise be unquantifiable or completely lost in large uncertainties.

Keywords: groundwater, direct runoff, lake, water budget, uncertainty

1. Introduction

The flow of groundwater into lakes is important because it can affect: the quantity and quality of the surface water (LaBaugh et al., 1995; Winter, 1999; Dubrovsky et al., 2010; Fruh, 1967; Bruce et al., 2009); the ecosystem health (Hayashi and Rosenberry, 2002); the distribution of aquatic life (Baird and Wilby, 1999; Rosenberry et al., 2000); and the quality of the fish habitat (Power et al., 1999). Estimates of net groundwater discharge to a lake can indicate the relative importance of groundwater in the water budget, but accurately quantifying total discharge can be a challenge. Groundwater flows into and out of lakes can be estimated using: direct point measurements of flow (Cartwright et al., 1979; Cherkauer and Nader, 1989; Harvey et al., 1997 and 2000); water balance calculations (Winter, 1981; Sacks et al., 1998; Zacharias et al., 2003); isotopic tracers (Walker and Krabbenhoft, 1998; Stets et al., 2010), and numerical modeling of the lake and its watershed (Feinstein et al., 2010; Hoaglund et al., 2002; Mylopoulos et al., 2007). Point measurement techniques are useful but impractical to employ on a lake-wide basis, particularly when the lake is large and there is substantial spatial heterogeneity in lakebed deposits and flows. Likewise, geochemical methods are difficult to use in large lakes because of spatial variability in water quality and challenges in defining appropriate end member concentrations for calculating mixing ratios. Numerical models that quantify groundwater flow are potentially very useful and can handle considerable spatial and temporal complexities; however, the lack of field data to constrain and populate these models generally results in major simplifying assumptions which produce uncertainties and errors that are either unknown or not readily quantifiable. The water balance method requires the quantification of inflows (precipitation, direct runoff, surface water inflows), outflows (evaporation, surface water outflows), and change in lake storage to calculate net groundwater flow. If properly done, the water balance equation has the potential to provide accurate estimates of the net groundwater flow (i.e., groundwater inflow minus groundwater outflow, which represents a minimum value for groundwater discharge) with potentially less effort and uncertainty than is associated with the other techniques. Despite this potential, the water balance method tends not to be used to determine net groundwater discharges for large lakes (Quinn and Guerra, 1986; Neff and Killian, 2003; Lenters, 2004; Neff and Nicholas, 2005).

There are two main reasons why water balances performed on large lakes do not attempt to quantify groundwater-surface water exchanges and, instead, either assume groundwater contributions are insignificant (i.e., are zero) or simply lump them together with the direct runoff into a combined runoff term. The first reason is that net groundwater flow is usually solved for as an unknown in the water balance equation, which means all the uncertainty in other components translates to and accumulates in the uncertainty of the groundwater component. Even what appear to be small relative errors on large components (e.g., precipitation or evaporation) may result in errors of substantial absolute magnitude that are larger than the groundwater component being quantified (Winter, 1981; Thodal, 1997). Unfortunately, many studies do not perform the uncertainty analysis necessary to assess the reliability of results even though several studies discuss how to quantify uncertainties (Winter, 1981; Lee and Swancar, 1997; Winter and Rosenberry, 2009; Neff and Nicholas, 2005). Even in studies where the net groundwater flow in the water budget as a percent of total inflow appeared to be important (e.g., Zacharias et al., 2003; Demlie et al., 2007; and Ayenew and Gebreegziabher, 2006), uncertainty analysis of the groundwater term has not been included. Without the uncertainty analyses, it is not known if the calculated values of net groundwater flow are accurate and representative.

The second reason why net groundwater discharge is not calculated for lakes is because it requires the direct runoff component (i.e., non-channelized overland flow and interflow) be quantified and this is often neglected or cannot be done with confidence or certainty due to a lack of suitable methods. The direct runoff component is usually ignored for large lakes (Neff and Nicholas, 2005; Lenters, 2004; Neff and Killian, 2003), and little work has been done in the last three decades to specifically estimate non-channelized runoff to lakes despite its inclusion in data-intensive time-stepping models such as SWAT (e.g., Menking et al., 2003), MOD-HMS (e.g., Panday and Huyakorn, 2004), and WATLAC (e.g., Zhang, 2011). The few methods that have been applied have been for small lakes and were originally developed for streams. The methods include: the curve number (CN) method (Natural Resources Conservation Service, 2004; Motz et al., 2001), the use of coefficients associated with varying land use and permeability (Sacks et al., 1998; Dames and Moore, 1992), and the extrapolation of hydrograph separation results (Newbury and Beaty, 1980; Schindler et al., 1976). The hydrograph separation model approach is appealing because it represents an empirical relationship derived from and calibrated to a portion of that particular lake's watershed and takes into account the actual physical and climatological conditions at the site without relying on models that extrapolate and use empirical runoff relationships derived at other sites with different conditions. The hydrograph separation method has not been applied to large lakes, and there is a need to determine its applicability and accuracy when applied to large lakes.

An opportunity to examine these issues concerning quantification of net groundwater discharge and direct runoff to large lakes was presented when concerns were expressed regarding the current and future water quality of Lake Pyhäjärvi (155 km^2), located in glacial terrain near Säkylä, Finland. The concerns focused on the eutrophication of the lake resulting in part from the effects of the agricultural watershed around the lake, along with impacts on the fishing industry, recreational enjoyment, and overall ecological integrity of the lake (Kirkkala, 2014). Early studies of the lake (Hyvärinen et al., 1973; Kuusisto, 1975; Järvinen, 1978; Eronen et al., 1982) either insufficiently assessed the net groundwater component of the lake's water budget or assumed it was negligible (i.e., zero); however, recent work indicated significant groundwater discharge might occur through an esker that intersects Lake Pyhäjärvi and at other specific locations along the shoreline (Rautio, 2009; Korkka-Niemi et al., 2011; Rautio and Korkka-Niemi, 2011). Moreover, indirect groundwater discharge, where groundwater discharges to a river and then is transported into the lake by the river, can also influence the quantity and quality of water in large lakes (Holtschlag and Nicholas, 1998; Neff et al., 2005). It was hypothesized that using historical climatological and hydrological data, a carefully conducted water balance study could be used to successfully estimate the net groundwater flow into the lake, provided that a rigorous uncertainty analysis was performed to characterize potential errors and that a suitable method for determining direct runoff could be used. A specific objective of this study was to evaluate whether a hydrograph separation method that has been applied to streams and small lakes to estimate direct runoff could be successfully applied to a large lake. This study 1) provides the first rigorous water balance and estimates of net groundwater flow and indirect groundwater discharge for Lake Pyhäjärvi, 2) demonstrates the importance of uncertainty analyses, and 3) successfully tests the hypothesis that using a hydrograph separation method to estimate the direct runoff component to a large lake is a viable approach for water balances. This approach could be applicable to other large water bodies in various landscape settings.

2. Background

Lake Pyhäjärvi (60°54´-61°06´N, 22°09´-22°25´E) is the largest lake in southwestern Finland (155 km^2) and is a valuable fishery and recreational area (Ventelä et al., 2007; Ventelä et al., 2005). The lake is quite shallow (5.4 m on average) with a maximum depth of 26 m (e.g., Kirkkala, 2014), and it makes up a large percentage (25%) of its watershed (Figure 1). Lake Pyhäjärvi's watershed (616 km^2) is predominately agricultural land (Luoto, 2000; Häkkinen, 1996). The ground elevations in the watershed range from about 40 to 145 masl, and it is relatively flat with an average topographic slope of 2.8% (MML, 2009c; ESRI, 2010). Two rivers (Yläneenjoki and Pyhäjoki) are gauged, drain the agricultural lands in the south and east, and flow into the lake; while one river (Eurajoki, also gauged) flows from the northern end of the lake at Kauttua Falls and flows to the Baltic Sea. The remaining area (304 km^2) of the lake's watershed is ungauged and consists of four subwatersheds with single channels that drain water into the lake and another six subwatersheds that do not have significant drains or channels (Figure 1).

The landscape around Lake Pyhäjärvi has been sculpted by glacial erosion and deposition. The surficial geology around the lake is shown in Figure 2 and consists primarily of thin, discontinuous till layers, numerous granite and sandstone bedrock outcrops, and to a lesser extent clays, peats, and silts. Figure 3 shows that the watershed contains very few coarse grained aquifer deposits. Among these is the Kuivalahti-Säkylä esker, which is connected to the large Säkylänharju-Virttaankangas Glaciofluvial Complex that lies mostly outside the watershed and is on the eastern side of the Pyhäjoki River's subwatershed. The esker is found along 15 km of the lake's northeastern shoreline and contains several aquifers, including the Honkala Aquifer (Figure 3). Figure 2 shows that the Yläneenjoki River's subwatershed (234 km^2) contains more clay and bedrock and is less permeable than the Pyhäjoki River's subwatershed (78 km^2), which contains sands and coarse-grained materials of the Virttaankangas Glaciofluvial Complex (Eronen et al., 1982).

Little is known about the groundwater-surface water interactions for Lake Pyhäjärvi. Because the Eurajoki River drains water from the lake, the lake may be a gaining lake (i.e., gains groundwater), but for many lakes (referred to as flow-through lakes) groundwater can enter the lake in one area as groundwater discharge and surface water can leave the lake through the bottom sediments as groundwater recharge at another location. Groundwater discharge to the lake has been documented at specific locations along the shoreline (Rautio, 2009; Korkka-Niemi et al., 2011; Rautio and Korkka-Niemi, 2011; Artimo, 2002), but areas of groundwater recharge have not been documented. Hydraulic head data for wells in the watershed (Artimo, unpublished report, 1998; Artimo, 2002; Wiebe, 2012) show that groundwater hydraulic head gradients indicate flow toward the lake, even at the northern end of the lake where the Eurajoki River exits the lake and groundwater recharge conditions might be anticipated. It is not known if groundwater-surface water exchanges occur beneath the lake with the underlying Rapakivi granite and Satakunta sandstone, but the bedrock generally has low permeability and the exchange would likely need to involve regional or intermediate groundwater flow systems. This study was undertaken to provide insight regarding the importance of groundwater with regards to the lake by estimating the net groundwater flow component within the overall water balance of the lake.

3. Methods

3.1 Water Balance Approach

A water balance equation may be used either to solve for an unknown component such as evaporation or to verify that estimated input and output components balance (Järvinen, 1978; Sacks et al., 1998; Lenters, 2004; Trask, 2007). For the Lake Pyhäjärvi study, the net groundwater flow was estimated by solving the equation for a defined time period in which all other inputs and outputs, and the changes in storage volume, were known. Operating on a lake area basis over a single water year (October 1 to September 30), the water balance equation for the lake is:

$$, \qquad (1)$$

where: G represents the net groundwater flow into the lake, h_S is the vertical change in lake stage (increases being positive), E is the sum of evaporative losses from the lake, W is the amount withdrawn from the lake by pumping, P is the amount of direct precipitation on the lake, R is the normalized net river flow (sum of inflowing minus outflowing) plus channelized flow into the lake from ungauged subwatersheds, and DR is the normalized direct runoff contribution (non-channelized overland flow and interflow) from the watershed into the lake. Normalization of the R and DR components consisted of dividing their total volume of water for the year by the area of the lake to obtain values in mm per year. Unless otherwise stated in this paper, water balance components that are expressed in mm per year are normalized values equivalent to a volume per unit lake area per water year. Using this equation, the net groundwater flow was estimated for 38 water years (October to September) between 1971 and 2009, which allowed the method to be evaluated for a variety of different climatological conditions and to examine temporal trends in results. Resulting positive values of *G* represent net groundwater discharge conditions (i.e., on the whole the surface water is gaining groundwater) and negative values of *G* represent net groundwater recharge (i.e., on the whole the lake is losing surface water to groundwater), but in each case, there may be both gaining and losing portions of the lake.

3.2 **Quantifying Uncertainties**

The solving of the water balance equation for the net groundwater flow means that all the errors associated with each of the other components accumulate in the error associated with the net groundwater term. A standard method for calculating water balance uncertainty (described by Winter, 1981; Lee and Swancar, 1997; and Sacks et al., 1998) was used to determine the uncertainty for each water year:

$$/ \quad , \qquad (2)$$

where δG is the (absolute) uncertainty estimate for the net groundwater flow into the lake, and δh_S, δE, δW, δP, δR, and δDR are the absolute uncertainty estimates associated with the lake water level change and other terms in (1), respectively. The absolute uncertainty for each component may also be composed of multiple uncertainties (e.g., equipment measurement errors and data interpolation errors), which in turn are calculated using equations similar in form to Equation 2 (see Tyler, 1977; Ramette, 1981; Taylor, 1997; and Lee and Swancar, 1997). The common assumption (Winter, 1981) that the sources of uncertainty for each component are independent was made. Table 1 lists all the uncertainties that contribute to each individual component and shows the equations and data used to calculate the absolute or relative uncertainty value for each particular component.

3.3 Water Balance Components

3.3.1. Lake Storage

The change in lake storage (h_S) over the course of each water year was calculated from daily water level measurements (OIVA/HERTTA, 8 Sep 2010) taken at the staff gauge at the north end of the lake. The difference between the lake stage at the start of two consecutive water years constituted the change in storage for the water year thus bracketed (normalized by the lake area by default). The absolute uncertainty (δh_S) for the storage was calculated (Table 1) using an uncertainty of ± 5.0 mm associated with the staff gauge measurement and an uncertainty value of ± 25 mm to account for half of the possible lake stage fluctuations that could be caused by wind driven seiche effects (Hyvärinen et al., 1973). The δh_S value was representative of fluctuations observed during the days before the start of each water year.

3.3.2. Evaporation

The lake evaporation (E) was estimated using two different types of data, depending on the time of year. For the months of May to November, data from a Class A evaporation pan located at the Jokioinen meteorological station (about 60 km SE of the lake) was used (OIVA/HERTTA, 5 Jun 2010). An average pan coefficient of 0.80 was assigned, which was consistent with the only pan coefficients available in the region (i.e., three years of data for summer months at three meteorological stations within 60 km of the lake); these were between 0.76 and 1.25 (Järvinen, 1978). The value is consistent with coefficients from other studies of regions near oceans in the United States (Hounam, 1973; Kohler et al., 1959). The uncertainty in the evaporation estimates for each year was assigned to be 15%, based on the estimated accuracy range associated with a pan coefficient that accounts for this type of lake depth and climatic regime (Dingman, 1994; Harbeck et al., 1954). For the five months when pan data were not available due to freezing conditions (December to April), an evaporation rate of 8 mm per month was assigned based on work on Lake Pyhäjärvi that was performed by Kuusisto (1975), who employed the Dalton-type formula developed by Shuliakovski (1969) to obtain the value. The uncertainty associated with the Dalton type measurements was assumed to be 15% for consistency with the evaporation estimates for the other seven months of the year. The absolute uncertainty for evaporation (δE) was calculated as shown in Table 1.

3.3.3. Pumping Withdrawals

Few data were available regarding the total amount of water pumped from the lake each year (*W*). The Lohiluoma pumping station, which is located beside the lake near its northern extent, has a municipal well that reportedly extracted (by induced infiltration) 4700 m^3/d from the lake during 2010 (J. Reko, pers. comm., 2010). Historical data were not available, although the well has been in operation since 1965 (OIVA/HERTTA, 15 Jun 2011). Lake water extraction for irrigation was not included due to lack of data, nor were other known but minor withdrawals included. To account for the possible variations in annual pumping rate and for the minor withdrawals, an uncertainty of ± 500 m^3/d was assigned.

3.3.4. Precipitation

The direct precipitation on the lake (*P*) was estimated for each water year using data from eight nearby meteorological stations that are within 70 km of the lake (Finnish Meteorological Institute [FMI], 24 May 2011; OIVA/HERTTA, 13 Oct 2010). The isohyetal method (e.g., Dingman, 1994) was employed to spatially extrapolate and estimate precipitation over the lake for each water year using the available data. Surfer 8 (Golden Software, Inc., 2002) was used to contour the point precipitation sums for each water year via point kriging for a region 95 km E-W by 59 km N-S that encompassed the lake. ArcMap 10.0 GIS software (ESRI, 2010) was used to calculate the areas between 2 mm contour intervals. The absolute uncertainty of the precipitation (δP) was calculated (Table 1) using a baseline value of 5.0% for the instrument error (Winter, 1981) and a year to year spatial interpolation error term. The spatial interpolation term ranged from 0.2 to 15% and was the absolute value of the difference between the magnitude of total precipitation estimated for the lake using the isohyetal method and the magnitude obtained using a second spatial interpolation method (an areal average method using an arithmetic mean value; OIVA/HERTTA, 6 Oct 2010) for the watershed. This areal average approach is outlined in Winter (1981), who cites Linsley et al. (1958).

3.3.5. River Discharge

River discharge estimates were compiled from the net river discharge into the lake from the three gauged rivers, and from per unit area river flow extrapolations from the Yläneenjoki River for the four ungauged subwatersheds with single channel drainage (i.e., $R = R_{net} + R_{single\ chan}$). River discharge estimates were obtained for the Yläneenjoki, Pyhäjoki, and Eurajoki Rivers using gauging station flow estimates based on rating curves for daily water level measurements at weirs (OIVA/HERTTA, 23 Sep 2010). The sum of the two inflowing rivers minus the outflowing river yielded the net river discharge (R_{net}). Because the gauges for the two inflowing rivers were located a short distance upstream of the confluences of the rivers with the lake (Figure 1), the total flows for the rivers were corrected (adjusted upwards) to account for contributions from the ungauged part of the river’s watershed. In order to do this, river flow per unit gauged area was multiplied by the area of the ungauged portion and added to the flow for the river prior to calculating R_{net} (described by Wiebe, 2012). Similarly, Yläneenjoki flow per unit gauged area was applied to the areas of the four single drainage channel subwatersheds ($R_{single\ chan}$). Sums (R) for each water year were normalized by dividing values by the average lake area. Groundwater discharge into the two inflowing rivers was included in the flow volumes, and also in the per unit area flow volumes applied to the ungauged single drainage channel subwatersheds. Because stream discharge measurements may be accurate to within 5.0% for continuous monitoring of river stage (Winter, 1981; Herschy, 1973), an accuracy of 5.0% for each daily discharge estimate was assumed for each of the three rivers. An uncertainty of 9.0% was applied to $R_{single\ chan}$ based on the maximum difference observed by Devito and Dillon (1993) for this type of extrapolation. The uncertainty for each subwatershed area (other than the lake itself) was assumed to be ±1 km^2. The total uncertainty for the net river discharge (δR) was calculated as shown in Table 1.

All river and direct runoff flow volumes were normalized by the (average) lake area (155 km^2; OIVA/HERTTA, 10 Aug 2010). The variation in the area of the lake due to changes in lake stage was assessed using ArcMap in order to estimate the uncertainty related to the chosen value. Contour maps created from interpolation of topographic (MML, 2009c) and bathymetric (MML, 2009b) elevation data were used to calculate the maximum and minimum lake area based on the range of lake stages observed between 1960 and 2010. The areas varied by less than 1.6% (2.5 km^2) over that time (Wiebe, 2012). This value of uncertainty for the lake area was included in calculations of both δR and δDR (Table 1).

3.3.6. Direct Runoff

The direct runoff (DR) for the six subwatersheds of the lake where non-channelized flow occurs (Figure 1, “Direct Runoff” subwatersheds) was estimated in two ways: 1) the use of a runoff coefficient method to estimate runoff as a percentage of precipitation, and 2) the extrapolation of direct runoff values obtained by hydrograph separation analysis of a gauged river watershed within the lake’s watershed.

Runoff Coefficient Method

Several runoff coefficient methods were reviewed for use in this study (e.g., Natural Resources Conservation Service, 2004; Motz et al., 2001; Sacks et al., 1998; Dames and Moore, 1992; and Barazzuoli et al., 1989). To our knowledge, none have been used for estimating direct runoff for large lakes. The CN method (Natural Resources Conservation Service, 2004) is a well-known method and more commonly used in other scenarios; however, it could not be used because the soils of the Lake Pyhäjärvi watershed had not been classified according to the U.S. Natural Resources Conservation Service format. The method best suited to the data available at Lake Pyhäjärvi was the runoff coefficient map method developed by Kennessey (1930) and modified by Barazzuoli et al. (1989). The method calculates average annual direct runoff using the areal coverage of subcategories in three physiographic themes or “components” (surface soil permeability, vegetation types, and slope angles) for a watershed over a given time period. The components are summed to obtain a fraction of the precipitation that is direct runoff. ArcMap 10.0 was used to estimate the proportional coverage areas for the various categories of the method (Table 2), using surface geology (GTK, 2008), land cover (SYKE, 2004), and elevation (MML, 2009c) datasets with raster grid cells 25 m by 25 m in size. The appropriate set of coefficients for the method was selected based on the index of aridity calculated for the Jokioinen meteorological station using monthly and annual, daily-derived temperature and precipitation averages from Pirinen et al. (2012). The coefficient map method was applied to the six non-channelized direct runoff subwatersheds and also to the two gauged Yläneenjoki and Pyhäjoki River watersheds so that the results also could be directly compared to the hydrograph separation results for those two watersheds. Yearly precipitation values for these regions were estimated with the interpolation procedure described above for *P*.

Assigning errors to a runoff coefficient map method is problematic because to our knowledge no rigorous evaluations of its absolute effectiveness have been performed. The most applicable error estimate found in the literature was an absolute uncertainty of ± 0.16, which was the average error obtained from a study on the differences between observed and literature values of event runoff coefficients (Dhakal et al., 2012), and it was used in calculating δDR in Table 1. The uncertainty estimates for δP (described above) were also used for δDR calculations.

Hydrograph Separation Method

This method of determining direct runoff to the lake is based on the concept of determining the direct runoff and groundwater flow components for gauged rivers within the lake's watershed using hydrograph separation techniques and then applying the values to other (non-channelized) areas of the lake watershed that have similar physiographic characteristics. Despite the straight forward and intuitive nature of such an approach and the fact that hydrograph separation techniques have continued to improve in past decades, it appears only Newbury and Beaty (1980) and Schindler et al. (1976) have used a hydrograph separation approach to extrapolate direct runoff from gauged subwatersheds of a lake to those with only non-channelized flow.

In this study, the United States Geological Survey's PART automated hydrograph separation method was used (Rutledge, 2007). Several other techniques were considered but not used (e.g., HYSEP [Sloto and Crouse, 1996], UKIH [Piggott et al., 2005], BFLOW [Arnold and Allen, 1999], Eckhardt [Eckhardt, 2005]). The PART and HYSEP methods performed the best on average in an evaluation by Partington et al. (2012) and provided similar results in a study by

Eckhardt (2008). PART was selected because of its ease of use and because it is more commonly used. The data required for the PART program included: daily streamflow measurements (obtained from OIVA/HERTTA, 23 Sep 2010); the drainage area for the gauged region of each river (OIVA/HERTTA, 23 Sep 2010); and the starting and ending years for the data sets. Hydrograph records were processed using the following program settings: a threshold of 0.1 log cycles per day for the daily decline in streamflow (Rutledge, 1998) and a value of N of N-1. N is the number of days (as an integer) of impact that a rainfall event has after the peak flow when interflow and surface runoff components are significant (i.e., where $N=A^{0.2}$ and A is watershed area in square miles). The N-1 value was selected because it provided more accurate results for similar types of rainfall and hydrographs in a comprehensive evaluation of hydrograph separation methods performed by Partington et al. (2012). The PART method sums up daily calculated groundwater contribution values for the year and reports a base flow index (BFI) for the river, which is the fraction (i.e., 0 to 1.0) of total river flow that is groundwater for the year. The remainder of the fraction represents the streamflow that is contributed by direct runoff. Direct precipitation onto, and evaporation off of, the river surface are considered negligible.

Annual direct runoff estimates were obtained for the Pyhäjoki River from 1972 to 2009 and for the Yläneenjoki River from 1971 to 2009. The Yläneenjoki River results were used to estimate direct runoff in the six non-channelized direct runoff subwatersheds because the surficial geology of the subwatersheds adjacent to the lake were finer grained and a better match to the Yläneenjoki River watershed than to the coarser grained deposits of the Pyhäjoki River watershed (Figure 2; Table 2). The percentage of land area covered by bedrock, till, and clay in the Yläneenjoki watershed (76%) was slightly larger than the area for the direct runoff subwatersheds (71%), while the Pyhäjoki coverage area (36%) was much smaller. The percentage area covered by eskers, glaciofluvial materials, sands, and gravels in the Yläneenjoki watershed (7%) was smaller than the percentage area in the direct runoff subwatersheds (15%), while the Pyhäjoki coverage area (49%) was much larger. Direct runoff to the lake was estimated by multiplying the total Yläneenjoki River flow per unit gauged area by the area of the direct runoff subwatersheds and then by the direct runoff fraction (1 – BFI) obtained for the Yläneenjoki River using PART for the corresponding time period.

A main uncertainty associated with using the PART method centres around whether the automated graphical hydrograph interpolation method actually results in an accurate quantification of the true base flow. Partington et al. (2012) assessed the absolute accuracy of hydrograph separation techniques by simulating an artificial watershed and single precipitation events using HydroGeoSphere (Therrien et al., 2010), but the results could not be reliably scaled up to estimate the uncertainty for an entire year in our study. Therefore, lacking a comparison between PART and a true value of baseflow, the uncertainty for the BFI estimated by PART was derived from a study by Sanford et al. (2012), who compared results of PART with a hydrograph separation technique employing continuously measured specific conductivity values in rivers in Virginia over 18 months. Sanford et al. (2012) assessed two streams having average topographic slopes similar in magnitude to those in the Yläneenjoki and direct runoff subwatersheds (< 4%). These two streams yielded absolute percentage differences between the chemical hydrograph separation technique and PART (i.e., $|BFI_{chem} – BFI_{PART}| / BFI_{PART}$) of 8.9% and 7.5%, respectively. The average of these two values (8.2%) was assumed to be representative of the relative uncertainty for BFI values from the PART method. For the extrapolation of values from one subwatershed to another, the 9.0% uncertainty from Devito and Dillon (1993) was again applied (Table 1).

3.4 Quantifying Known Groundwater Discharges

Although the calculation of *G* using the water balance equation includes all groundwater inputs, groundwater discharge to the lake was calculated for an area of the shoreline where significant amounts of groundwater discharge were known to occur (Rautio, 2009; Korkka-Niemi et al., 2011; Rautio and Korkka-Niemi, 2011) and used as an independent value to compare to *G*. The groundwater discharge into the lake through the Honkala Aquifer in the Kuivalahti-Säkylä esker was estimated by Wiebe (2012) using Darcy's Law and a hydraulic conductivity value of $K = 1\times10^{-3}$ m/s ± one order of magnitude, which was chosen to represent flow in the coarse-grained esker core (Artimo, 2002). The cross-sections for the calculation are shown on Figure 3. Discharge estimates were normalized by the lake area, and an uncertainty estimate was developed according to the general procedures in Table 1. This approach to estimating groundwater discharge into sections of shorelines of large lakes is not new; Singer (1974) used the same approach to estimate flow into Lake Ontario in Canada.

4. Results

The water balance components and net groundwater (*G*) values calculated for each year using the runoff coefficient map method of obtaining direct runoff are shown in Figure 4. Near-equilibrium (i.e., near zero values) or net groundwater recharge occurs during the 1970s, a small amount of net discharge occurs during the 1980s, and mostly net groundwater recharge occurs from the 1990s until to the end of the study period. Fifteen of the 38 water years in the water balance appear to have net groundwater discharge conditions (of which only three are larger than the calculated uncertainty), while 23 of the 38 water years appear to have net groundwater recharge conditions (of which nine are larger than their associated uncertainty). Overall, the magnitude of *G* was less than the uncertainty during 26 water years. Table 3 summarizes the average value for each component of the water balance for the entire study period. The average total inflow and outflow for this water balance were 1481 mm and 1414 mm per year (not including the groundwater component). The average value of *G* was -24 mm (-1.7% of average total outflow) and indicates average net groundwater recharge conditions for the lake. During the study period, the magnitude of *G* ranged from -268 mm to 268 mm with a standard deviation of 117 mm. The average uncertainty was 119 mm. The estimates of direct runoff to the lake ranged from 83 mm to 167 mm per year and averaged 130 mm during the study period.

The water balance components and net groundwater values calculated for each year using the PART hydrograph separation method to estimate direct runoff are shown in Figure 5. The figure shows near equilibrium groundwater discharge conditions for the 1970s, groundwater discharge conditions for the 1980s, and near equilibrium conditions for the 1990s onward. The average value of *G* was +43 mm and indicates overall net groundwater discharge conditions. This *G* represents about 3.0% of the average total inflow for the lake (i.e., 1414 mm, not including the groundwater component). The magnitude of *G* ranged from -197 mm to 284 mm with an average standard deviation of 118 mm during the study period. Twenty-six of the 38 water years appear to have net groundwater discharge conditions, and the magnitude of *G* was greater than the calculated uncertainty for 12 of these years (Figure 6). The average uncertainty was 103 mm. The estimates of direct runoff to the lake ranged from 14 mm to 140 mm and averaged 63 mm during the study period.

Table 3 summarizes the average uncertainties associated with each component of the water balance equation over the 38 year study period for both the coefficient map and hydrograph separation methods for determining direct runoff. For both methods the component having the largest average absolute uncertainty was evaporation (67.5 mm). The second largest uncertainty of the PART derived method was precipitation (61.2 mm), while for the coefficient map method both the direct runoff term and precipitation had the second largest uncertainty (each equal to 61.2 mm). For the hydrograph separation method, components with the next largest average absolute uncertainties were the change in lake storage, net river flow, and then the direct runoff term (13 mm). As noted above, the uncertainty that accumulated in the net groundwater term was 119 mm for the coefficient map method and 103 mm for the hydrograph separation method. The difference between the two uncertainty values associated with G is a direct result of the accuracy of the direct runoff component because all the other components were calculated in the same way for both methods.

Relative and absolute uncertainties associated with each water balance component differed from year to year during the study, depending on the component. For the hydrograph separation method the ranges in uncertainties for the 38 year period were as follows. The absolute uncertainty of the net groundwater flow component ranged from 77 mm to 135 mm and the relative uncertainty ranged from 37% to 3800% (values greater than 100% mean the uncertainty is greater than the value of the component). For evaporation the absolute uncertainty ranged between 50 mm and 85 mm while the relative uncertainty was fixed at 15% (as described earlier). The absolute uncertainty for precipitation ranged between 24 mm and 112 mm, and relative uncertainty ranged between 5.2% and 20%. The relative net river inflow uncertainty ranged between 4.1% and 89%, and the relative direct runoff uncertainty ranged from 13% to 36%. The lake water level absolute uncertainty (36 mm) was constant as described earlier, and the relative uncertainty ranged between 9.5% and 361%. The estimated relative uncertainty on the pumping withdrawals at Lohiluoma was 11%.

The groundwater baseflow and direct runoff for the Yläneenjoki and Pyhäjoki Rivers also changed from year to year (Figure 7). The average baseflow index (BFI) obtained from the PART hydrograph separation model for the study period was 0.68 for the Yläneenjoki River and 0.84 for the Pyhäjoki River, while the standard deviations for the two were 0.069 and 0.043, respectively. The average indirect groundwater contributions from these rivers to the lake were 327 mm and 123 mm for the Yläneenjoki and Pyhäjoki, respectively. Overall, the average indirect groundwater contribution to Lake Pyhäjärvi was at least 454 mm (or about +32% of average total inflow when compared to the PART derived water balance). Based on these average BFI values, the corresponding average direct runoff values for the Yläneenjoki and Pyhäjoki Rivers during the study period were 32% and 16% of the river flow, respectively. Figure 7 also shows how the values of direct runoff per unit area for each of the rivers' watersheds varied during the study period. The finer grained deposits of the Yläneenjoki River watershed resulted in direct runoff values that ranged from 23 mm to 228 mm per unit gauged area of its watershed per year (with an average value of 103 mm per unit gauged area of its watershed per year), and were, on an annual basis, consistently 3.3% to 29% higher than those for the Pyhäjoki River watershed.

The coefficient map method of determining direct runoff to the lake during the study period resulted in values that were typically higher than those determined using the PART hydrograph separation method. In order to perform a direct comparison of the coefficient map and PART methods of determining direct runoff, both methods were applied to both of the gauged river watersheds. The average surface flow estimated by PART constituted 7.8% of the precipitation (estimated using the isohyetal method) in the gauged region of the Pyhäjoki River watershed, and 16% of the isohyetally derived precipitation estimate in the gauged region of the Yläneenjoki River watershed. The corresponding percentages for the coefficient map method were 29% and 38% for the gauged regions of the Pyhäjoki River and Yläneenjoki River watersheds, respectively (Table 2). The runoff coefficient map estimates were about 2.4 to 3.7 times higher than those estimated by the PART method. Higher direct runoff values result in lower amounts of precipitation entering the groundwater.

The amount of groundwater estimated to directly enter the lake through the Honkala Aquifer in the Kuivalahti-Säkylä esker appears to be significant and relatively constant but is subject to significant uncertainty. The groundwater discharge from the esker was estimated to be 22 mm or about 1.6% of the average total inflow for the PART derived water balance, with an uncertainty of ± one order of magnitude. The uncertainty in the hydraulic conductivity value of the geological materials was responsible for essentially all the uncertainty in this Darcian flow estimate. However, the average estimated Darcy flux of 6.81×10^{-6} m/s (per unit cross section area of the aquifer) is consistent with and in the 10^{-7} to 10^{-5} m/s range for groundwater discharge into the lake measured by Rautio (2009) using seepage meters where the esker and aquifer intersect the shoreline. The amount of groundwater entering the lake from the aquifer each year is likely relatively constant because the water levels (and hydraulic gradients) in the Honkala Aquifer are relatively constant (Artimo, 2002), as are the regulated lake levels that vary within a 1 m range. The average groundwater discharge from the aquifer appears equal to approximately half of the +43 mm average net groundwater component for the entire lake estimated using the PART method.

5. Discussion

An accurate and scientifically meaningful water balance for a large lake requires: collecting a considerable amount of data, implementing successful upscaling schemes, employing techniques for estimating components that are not easily measured, and understanding the uncertainty related to both the measurement methods and their spatial and temporal extrapolation methods. The lack of data or lack of good quality data, or an inability to properly quantify or reduce errors that accumulate in the calculation method has often led to water balances that do not even attempt to quantify net groundwater flow for large lakes (e.g., Kuusisto (1975) and Järvinen (1978) for Lake Pyhäjärvi). The main factor that prevents calculation of the net groundwater flow component is the inability to accurately separate out and quantify direct runoff contributions from a term that lumps all groundwater flow with non-channelized overland flow and interflow from subwatersheds with no streams. A second and almost equally large problem is that unless all uncertainties in the water balance equation components are accurately quantified, one will not know if the calculated net groundwater value is real (i.e., larger than the accumulated error) or not. This study of Lake Pyhäjärvi appears to have resolved these two problems by using the PART hydrograph separation method to estimate and minimize the uncertainty related to the direct runoff to the lake, and by employing the rigorous uncertainty analysis summarized in Table 1. The key to the success of this study was the opportunity to use stream flow gauging data for a river within the lake's watershed and then apply the results to non-channelized parts of the watershed.

The two main advantages to using the PART hydrograph separation method to estimate direct runoff for the lake is the site specific representativeness of the technique and the relatively low uncertainty associated with the method. Unlike the coefficient mapping technique, the PART method develops an empirical relationship between direct runoff and the actual rainfall events (magnitudes and intensities), antecedent conditions (i.e., moisture contents of soils), geology, vegetation types, topographic slopes, and groundwater flow processes actually occurring in the lake's watershed. In contrast, the runoff coefficient map method cannot appropriately deal with overland drainage to a low-lying area where water infiltrates or evaporates rather than flowing to the lake, and it does not account for rainfall intensity or antecedent soil conditions. Furthermore, the runoff coefficient map method is unable to produce different percentages of runoff versus rainfall for different water years and is unable to adapt to a climate having precipitation that varies over a range of several hundred millimetres per water year. These deficiencies in the coefficient mapping method resulted in: 1) estimates of direct runoff that were on average 2.4 to 3.7 times higher than those from the PART method, 2) an underestimation of direct groundwater contributions to the lake, and 3) the conclusion that on average during the 38 years the lake was losing surface water to (i.e., recharging) the groundwater at a rate (G) of at least -24 mm despite the fact that none of the field investigations have detected significant losing areas within the lake. Moreover, there do not appear to have been sufficient studies to definitively verify the accuracy of the coefficient map method developed by Kennessey (1930) and modified by Barazzuoli et al. (1989) or other similar methods on an annual basis. The CN method (Natural Resources Conservation Service, 2004) attempts to account for issues such as the antecedent moisture content of the soil and the threshold rainfall that will generate runoff, but it still does not incorporate rainfall intensity. In contrast, the PART method for determining the direct runoff provided a more realistic average groundwater discharge rate of +43 mm (overall a gaining lake condition) and the uncertainty associated with the method is smaller.

The magnitude of uncertainties assigned to the PART method itself and the calculated values of direct runoff appear to be reasonable and accurate, and the concept of assigning those values to adjacent non-channelized watersheds appears to be valid. The relative uncertainty of ± 8.2% from the comparison by Sanford et al. (2012) of PART results and a chemical hydrograph separation method was the average value for two watersheds in Virginia with average topographic slopes similar to those found in the Yläneenjoki River watershed. This uncertainty value corresponds well to the estimated uncertainty derived from a controlled numerical experiment by Partington et al. (2012). The average absolute difference of ± 0.023 per event from that study, when upscaled to the average number of similar events (12) in the Yläneenjoki River per water year, yields an absolute uncertainty of ± 8.0%. Further, annual absolute percent differences between PART and the HYSEP-sliding interval or HYSEP-fixed interval programs were also between 6.5 and 8.4% in a study by Risser et al. (2005). The concept of using PART to calculate stream baseflow in a gauged watershed and extrapolating results to physically similar watersheds was performed quite successfully in the Great Lakes watershed in Canada and the USA (Neff et al., 2005). The Neff et al. (2005) empirical approach for extrapolating the PART results to other watersheds was more sophisticated than that employed in this study and was based on data from hundreds of gauging stations and demonstrated the validity of this type of approach. In the Lake Pyhäjärvi watershed, extrapolating the PART derived direct runoff results from the Yläneenjoki River watershed to the adjacent non-channelized watersheds was clearly more appropriate than using the results for the Pyhäjoki River based on the geological considerations. For lake watersheds that have a sufficient number of streams within them, it should be possible to select the most appropriate ones for streamflow gauging and subsequent extrapolation to non-channelized portions of the lake watershed.

The comprehensive and detailed uncertainty analysis performed in this study provided the information necessary to confidently assess if small values of net groundwater discharge were real or still too uncertain to be determined reliable, and this methodology can be applied to other lake water balances. The uncertainty calculation methodology presented in this study builds on earlier work on uncertainties in lake water balances by Winter (1981), Lee and Swancar (1997), and Neff and Nicholas (2005). As shown in Table 1, most of the uncertainty values for individual components of the water balance cannot be obtained simply from these earlier publications or from a single literature value but instead must be calculated using knowledge of the site specific techniques used to collect and calculate each water balance component. The equations contained in Table 1 can be adapted and used at other sites, and Table 1 also provides specific values of uncertainty for components that have not previously been quantified for the purpose of a lake water balance (e.g., the uncertainty associated with using PART to calculate BFI values and estimate direct runoff). In this study, magnitudes of G as small as 8% of the total inputs could be reliably determined (depending on the particular water year) and the value of δG provides a meaningful upper boundary for what G can be for years when the uncertainty is larger than G. Over the 38 year period, the PART derived water balance results indicated that 17 values of G that were greater than δG ranged between 101 mm and 284 mm and represented 8% to 19% of the total inflows (without groundwater) for those years. The G values obtained in this study are significant enough to be measureable but still a very small part of the overall water balance, whereas indirect groundwater discharge to the lake (via the rivers) is very significant and on average accounted 454 mm or about 32% of the total inputs to the lake during the study period.

A water balance should be conducted in addition or as an alternative to numerical modelling of groundwater-surface water interactions involving large lakes. Because a water balance method can involve quantification of the uncertainties on the various individual components, it can clarify the reliability of the component estimates and present meaningful error bars. Numerical models inherently struggle with accurately defining boundary conditions and with appropriately representing the hydrogeological properties (often having several orders of magnitude variability and uncertainty) and other characteristics of the site. It can be argued that large numerical models based on sparse data sets may introduce more uncertainty and make uncertainties unquantifiable with respect to estimates of G or other components because of the large number of assumptions and wide range of possible values needed to populate such models (e.g., hydraulic conductivities, unsaturated zone flow characteristics). Performing a water balance (as shown here) is a necessary first step to providing the data, calibration targets, and reality checks needed for numerical models to provide meaningful predictions.

6. Conclusions

This study of Lake Pyhäjärvi and its watershed demonstrated that minimizing and carefully quantifying uncertainties in the components used in the lake's water balance calculations is the key to determining meaningful estimates of net groundwater flow for a large lake, especially if net groundwater contributions are a relatively small part of the water balance. The estimate of the direct runoff component of the lake water balance was improved by using the PART hydrograph separation derived estimates of runoff for a river within the lake's watershed and then applying those values to the non-channelized areas of the lake's watershed. The key is to use direct runoff estimates obtained from a river that has similar characteristics (e.g., geology, topography, vegetation, antecedent rainfall conditions, and rainfall intensities) to the non-channelized areas, which was why the direct runoff values for the Yläneenjoki River (average BFI = 0.68) were used instead of those derived for Pyhäjoki River (average BFI = 0.84). The main advantage of this approach to estimating direct runoff to the lake is that it represents an empirical relationship that is calibrated to an actual portion of the lake's watershed and climatic conditions, unlike other coefficient runoff methods or numerical models that rely on relationships developed elsewhere and for very different watershed conditions.

The average net groundwater flow and the indirect groundwater discharge for the lake were quantified in this study. The average net groundwater flow into Lake Pyhäjärvi over the 38 water years between October 1971 and September 2009 was calculated to be +43 mm (3.0% of average total inflow) using the PART derived direct runoff values (average: 63 mm) for the Yläneenjoki River. The uncertainty analysis showed that the magnitude of the net groundwater flow was greater than the overall uncertainty in 17 out of 38 water years. A positive net groundwater flow value represents the minimum possible value of direct groundwater discharge to the lake (i.e., when groundwater recharge is zero), and if parts of the lake are also losing surface water to groundwater (i.e., recharging groundwater), the direct discharges could be proportionally larger. It is not known if any areas of the lake are recharging the groundwater, but previous field investigations suggest that the lake is gaining groundwater rather than losing surface water (Rautio, 2009; Korkka-Niemi et al., 2011; Rautio and Korkka-Niemi, 2011). A significant amount of the direct groundwater discharge into the lake may occur through the Honkala Aquifer in the Kuivalahti-Säkylä esker, which was estimated using Darcy's Law to be 22 mm (about 1.6% of the average total inflow for the PART derived water balance), but that estimate has an uncertainty of ± one order of magnitude. Independent, field-based measurements of groundwater discharge provide an important check on the magnitude of the net groundwater flow values, and if larger than that value, they can be used to infer that part of the lake must be recharging the aquifer. Although direct groundwater discharges actually may be much larger than the net value calculated, it is clear that indirect discharges of groundwater to the lake play a major role in the water balance. The total average indirect groundwater contribution to the lake from the Yläneenjoki and Pyhäjoki River discharges was 454 mm (+32% of average total inflow), which indicates that the groundwater entering the rivers can have a large influence on the quantity and quality of the water in the lake.

The techniques used in this study are applicable to other large lakes with inflowing streams and rivers and may allow small net groundwater flows to be reliably quantified in situations that might otherwise be unquantifiable or cause values to be completely lost in large uncertainties.

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Table 1: Summary of equations and data used to calculate relative (rel) and absolute (abs) uncertainties for the Lake Pyhäjärvi water balance. Relative uncertainty = absolute uncertainty divided by the associated quantity.

Uncertainty Assigned in WB Equation	Equations	Description of Uncertainty	Assigned Uncertainty	Source / Reference for Assigned Uncertainty
Lake Storage (h_S) -- Lake level via staff gauge -- OIVA/HERTTA (8 Sep 2010)				
δh_S = ±36 mm	, , ,	Overall uncertainty on storage change [1]		
	, !"#,	Measurement precision plus lake seiche effects	δWL_{meas} = ± 5.0 mm	
	!"#, $%&'%/2 25 **	Lake seiche effects [2]	*seiche* = ± 50 mm	Hyvärinen et al. (1973)
Lake Evaporation (E) -- Class A Evaporation Pan (Jokioinen meteorological station, May - Sep) -- OIVA/HERTTA (5 Jun 2010)				
δE = ±15%	, +, -· /0 123!2, - 455!62,	Appropriateness of the assigned pan coefficient [3][4]	$\delta E_{rel} = [\delta E_{Jokio}]_{rel} = [\delta E_{Kuusisto}]_{rel}$ = 15%	Dingman (1994), Harbeck et al. (1954)
Pumping Withdrawals (W) -- Lohiluoma pumping station municipal records (J. Reko, pers. comm., 2010)				
δW = ±11%	+, 7 8!, 9 8!, : ; +,	Lack of historical records on variable rates [5][6]	δW_{daily} = ± 500 m^3/d	
Lake Area (A) -- Topographic and Bathymetric Elevations -- OIVA/HERTTA (10 Aug 2010), MML (2009b and 2009c)				
δA = ±2.5 km^2	; max ?@ ; AB ; AB ; !0 D	Lake area dependence on lake stage [7]	Range of water level variation	
Precipitation (P) -- Seven meteorological stations, areal interpolation method (OIVA/Hertta, 6 and 13 Oct 2010)				
δP = ±5.2% to ±20%	,+, 7 E5B !06+/2,6!20,,+, :	Instrument measurement errors plus spatial interpolation errors	δP_{Gauge} = 5.0%	Winter (1981)
	,!06+/2,6!20,+, | !2# 6,, +,, |	Spatial interpolation over lake area [8]	0.2% to 15%	
Net River Discharge (R) -- Daily flow rates from weirs on Yläneenjoki, Pyhäjoki, and Eurajoki Rivers (OIVA/HERTTA, 23 Sep 2010), subwatershed areas (SYKE, 2010; OIVA/HERTTA,23Sep2010)				
δR = ±4.1% to ±89%	, 7 ,06 : 7 G,,,H : 7 I,,H : 7 J5,,H : "#0, - "#0,,+,	Overall river uncertainty [9]		
	"#0, 7 G,,,+, : 7; +,B,+, : 7; "#0,+, : ; +, %K +,	Extrapolating Yläneenjoki gauged flows to single channels [5][10]	$[\delta Reg]_{rel}$ = ± 9.0% $[\delta A_{r,g}]_{abs}$ = ±1 km^2 $[\delta A_{chan}]_{abs}$= ±1km^2	Devito and Dillon (1993)
	+,,H,+, 7 +,,+, : 7; +,B,+, : 7; +,6,+, : ; +,	Each inflow river: final combined uncertainty [5][10][11]	$[\delta A_{r,g}]_{abs}$= ± 1 km^2 $[\delta A_{r,t}]_{abs}$ = ± 1 km^2	
	J5,,H,+, 7 J5,,+, : ; +,	Outflow river: final combined uncertainty [5][9]		
	+,, LM 0 !N 7 +,!, :	Uncertainty associated with daily gauge readings [12][13]	$[\delta R_{r,i}]_{rel}$ = ± 5.0%	Winter (1981), Herschy (1973)
Direct Runoff -- Yläneenjoki R. daily flows (OIVA/HERTTA, 23 Sep 2010), surface geology (GTK, 2008), land cover (SYKE, 2004), elevation (MML, 2009c), Barazzuoli et al. (1989) coefficients.				
δDR (Coeff. Map) = ±46% to ±50%	,+, 7O PQ,+, : 7 ,+, :	Estimating *DR* via Barazzuoli et al. (1989) runoff coefficients and watershed raster maps [14]	δC_{DR} = 0.16 C_{DR} = 0.35	Dhakal et al. (2012)
δDR (PART derived): ±13% to ±36%	626,,+, L 7RST'U&VWX,G,,,+, : %K +, Y ; ; Z Y ; PQ ; PQ Z	Total error on PART derived *DR* [5][9][10]	$[\delta Reg]_{rel}$ = ± 9.0%	Devito and Dillon (1993)
	RST'U&VWX,G,,,+, [\] +, - [\] G,, / RST'U&VWX,G,,	Error on surface water fraction from PART (i.e., 1 – BFI) [9]	$[\delta BFI]_{rel}$ = ± 8.2%	Sanford et al. (2012)

Comments:

[1] Combined uncertainty, where δWL_y is the uncertainty of the water level at start of water year beginning in year y. [2] Uncertainty assigned to be half of the seiche effect. [3] Coefficient within range observed by Järvinen (1978). Jokioinen (subscript: *Jokio*) is the closest station to the lake. [4] Kuusisto (1975) assumed constant value of 8 mm per month for October to April based on a Dalton-type formula. Value assumed to have same uncertainty as the pan data. [5] A is the average lake area from OIVA/HERTTA. [6] Assigned uncertainty addresses variation in pumping and known minor withdrawals. [7] Areas obtained from contour maps based on water levels observed between 1960 and 2010. [8] Differences between the isohyetal and areal average method estimates. [9] *Yl* denotes Yläneenjoki River, *Py* denotes Pyhäjoki River, and *Eu* denotes Eurajoki River; *f* denotes final. The three terms in the square root are relative uncertainties. [10] $A_{r,g}$ is area of river r, gauged watershed region only. A_{chan} is the total area of the four single drainage channel regions. δReg is the regionalization error for applying per unit area results from Yläneenjoki River to the *DR* areas. [11] $A_{r,t}$ is area of river r, total river watershed area. [12] $R_{r,i}$ indicates river r, daily reading i; n is the number of days in the year. [13] Uncertainty for a continuously monitored watershed. [14] C_{DR} is the runoff coefficient calculated for the *DR* regions.

Table 2: Application of the runoff coefficient map method (Barazzuoli et al., 1989; Kennessey, 1930).

	Direct Runoff Subwatersheds	Yläneenjoki River Watershed	Pyhäjoki River Watershed
Permeability	**Fraction of Subwatershed Area (Weight)[a]**		
Bedrock	0.16 (0.30)	0.31 (0.30)	0.07 (0.30)
Till	0.52 (0.25)	0.17 (0.25)	0.23 (0.25)
Moraine ridges and hummocks	0.01 (0.20)	0.00 (0.20)	0.01 (0.20)
Eskers and other glaciofluvial deposits	0.02 (0.05)	0.01 (0.05)	0.22 (0.05)
Sand and gravel deposits	0.13 (0.10)	0.06 (0.10)	0.27 (0.10)
Silt	0.06 (0.20)	0.05 (0.20)	0.06 (0.20)
Clay	0.03 (0.25)	0.28 (0.25)	0.06 (0.25)
Peat and organic matter	0.07 (0.20)	0.12 (0.20)	0.07 (0.20)
Slope Angle			
0 – 3.5%	0.71 (0.03)	0.67 (0.03)	0.79 (0.03)
3.5 – 10%	0.27 (0.05)	0.28 (0.05)	0.19 (0.05)
10 – 35%	0.02 (0.20)	0.05 (0.20)	0.02 (0.20)
> 35%	0.00 (N/A)	0.00 (0.30)	0.00 (0.30)
Vegetative Cover			
Discontinuous urban fabric	0.08 (0.27)	0.01 (0.27)	0.00 (0.27)
Industrial or commercial	0.01 (0.30)	0.00 (0.3)	0.00 (0.3)
Sport and Leisure facilities	0.00 (0.25)	0.00 (0.25)	0.00 (0.25)
Agriculture	0.13 (0.15)	0.34 (0.15)	0.27 (0.15)
Forest	0.76 (0.05)	0.59 (0.15)	0.62 (0.05)
Transitional woodland	0.02 (0.15)	0.06 (0.15)	0.10 (0.15)
Mineral extraction	0.00 (N/A)	0.00 (N/A)	0.01 (0.3)
Component[b]	**Coefficients based on an Index of aridity > 0.40**		
Permeability	0.23	0.25	0.16
Slope angle	0.04	0.04	0.04
Vegetation cover	0.09	0.09	0.09
Sum (Fraction of annual *P* that is *DR*)	**0.35**	**0.38**	**0.29**

Notes:

[a] The weights (runoff coefficients related to the geology/slope/vegetation types) are based on the coefficients listed by Barazzuoli et al. (1989), who provide values for four categories for each component.

[b] Component = physiographic theme

Table 3: Summary of water balance results for water years 1971 to 2008 for two different methods of calculating the direct runoff component.

Direct Runoff Calculation Method Used for Water Balance	**Water Balance Component** (mm per year)						
	h_S **Lake Level (Storage)**	**E** **Evaporation**	**W** **Pumping Withdrawals**	**P** **Precipitation**	**R** **Net River Flows**	**DR** **Direct Runoff**	**G** **Net groundwater Flow**
Component Average							
Coefficient Map Method	3.2	450.1	11.1	607.7	-249.4	130.3	-24.3
PART Method	3.2	450.1	11.1	607.7	-249.4	63.1	43.0
Standard Deviation							
Coefficient Map Method	171.1	45.4	0.0	91.5	211.8	19.6	116.8
PART Method	171.1	45.4	0.0	91.5	211.8	26.9	118.0
Absolute Uncertainty							
Coefficient Map Method	36.1	67.5	1.2	61.2	21.8	61.2	119.4
PART Method	36.1	67.5	1.2	61.2	21.8	12.5	102.9

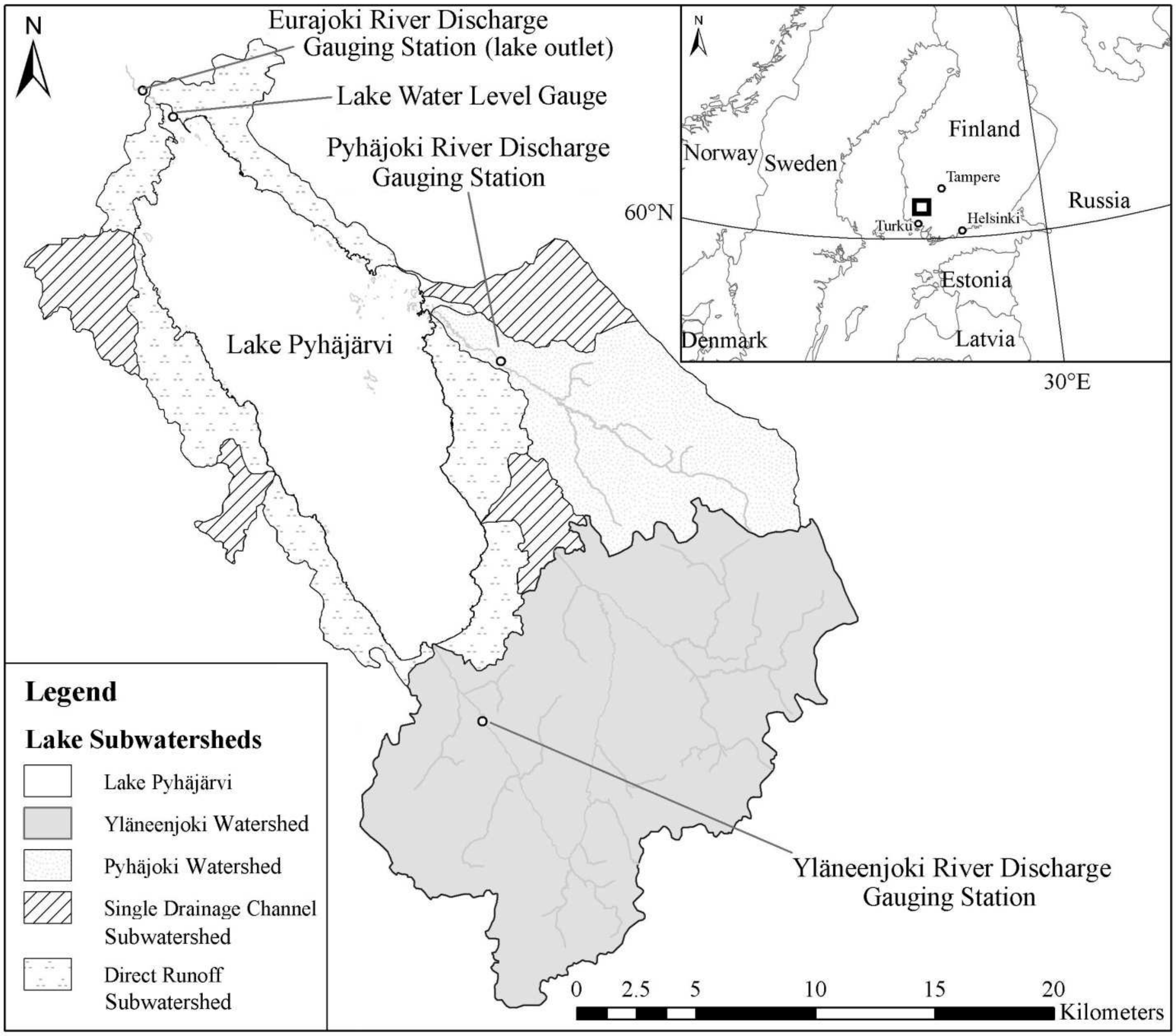

Figure 1: The Lake Pyhäjärvi watershed and its two gauged, four single channel (i.e., ungauged), and six direct runoff (i.e., non-channelized) subwatersheds (imagery from MML, 2009a; ESRI, 2009; SYKE, 2010; and OIVA/HERTTA, 4 Jan 2011).

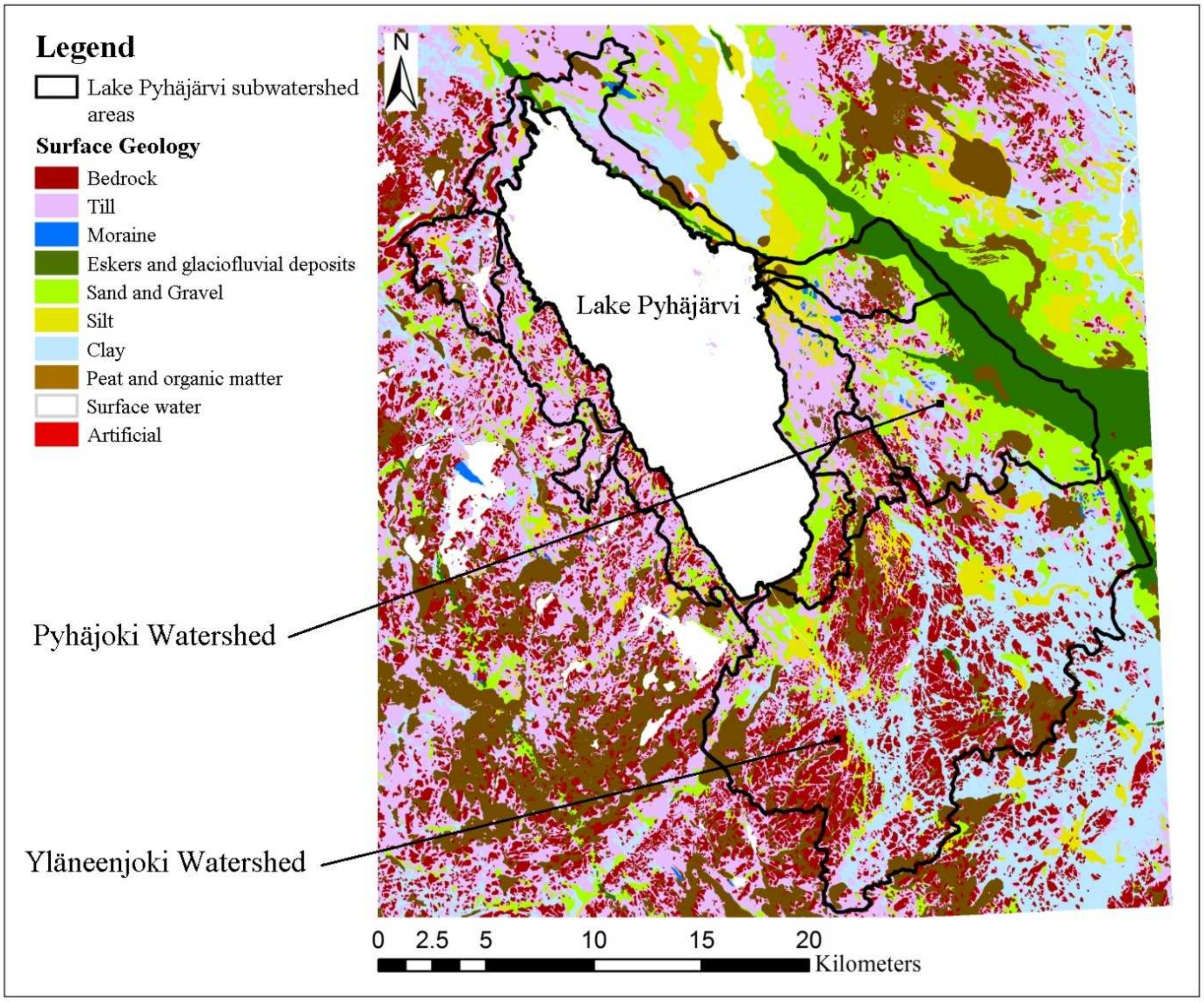

Figure 2: Surficial geology in the vicinity of Lake Pyhäjärvi, showing similarity between the Yläneenjoki River watershed and the non-channelized subwatersheds of the lake (after Wiebe, 2012; imagery from GTK, 2008; and SYKE, 2010).

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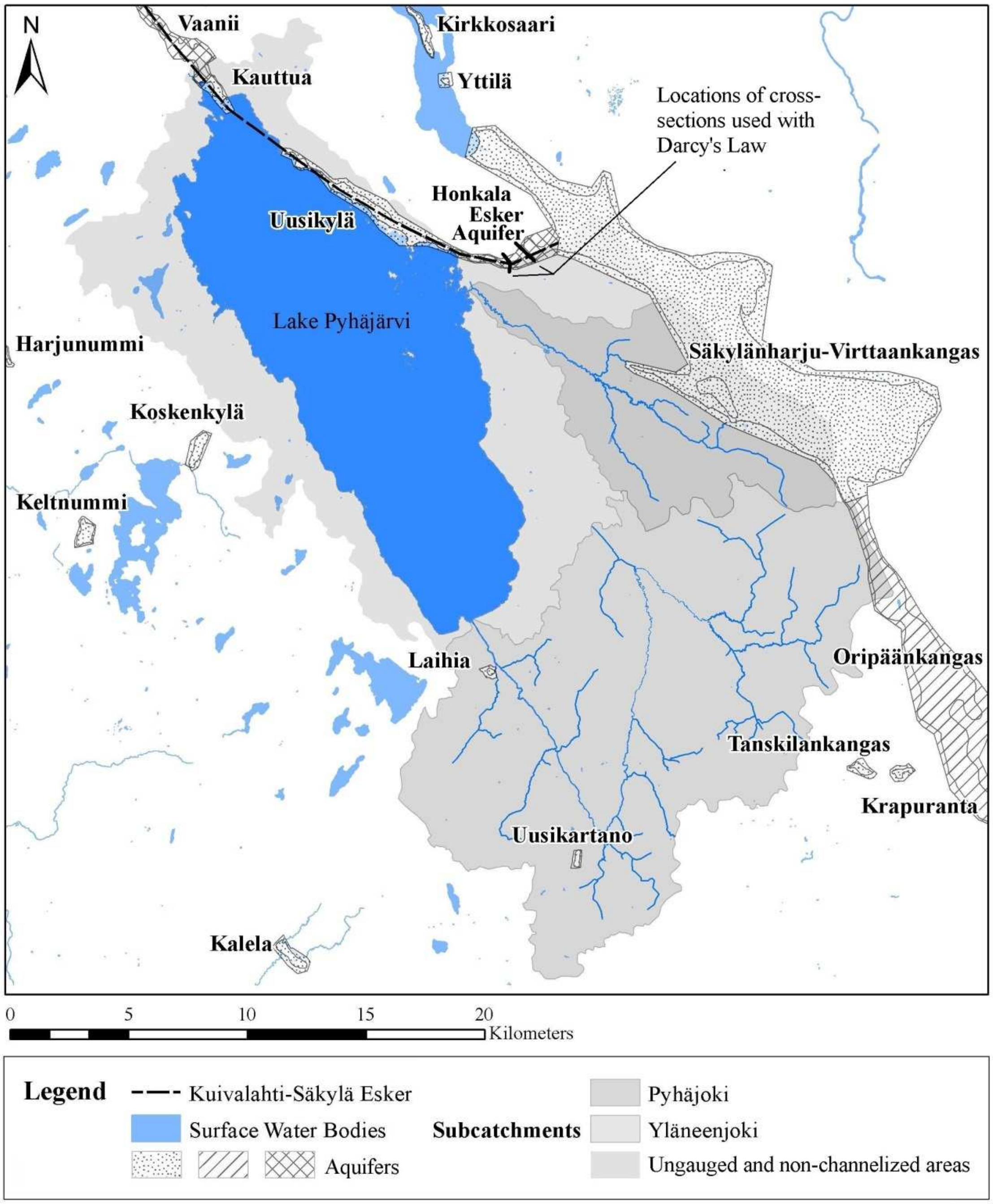

Figure 3: Locations and names of coarse grained overburden aquifers in the vicinity of the Lake Pyhäjärvi watershed (after Wiebe, 2012; imagery from GTK, 2008; MML, 2009a; SYKE, 2009; and SYKE, 2010).

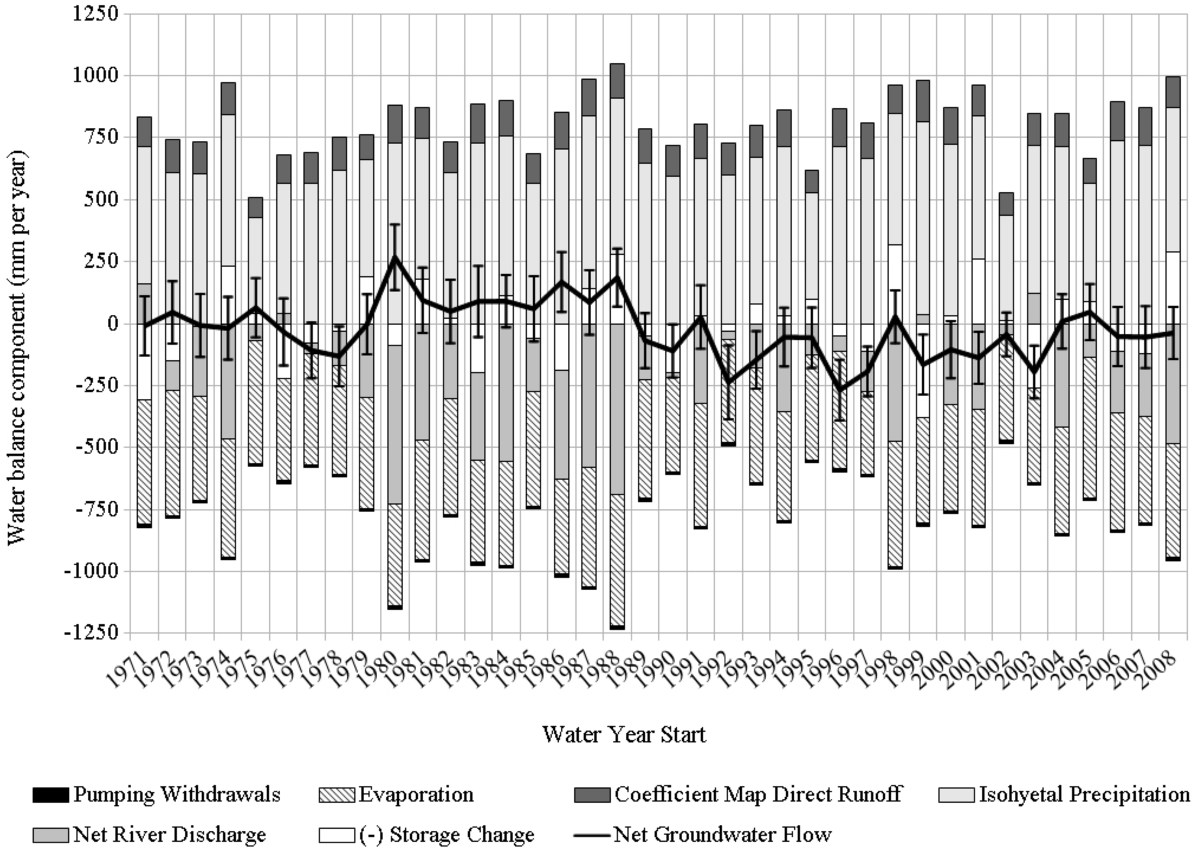

Figure 4: Water balance employing the coefficient map direct runoff estimate for water years 1971 to 2008. Components contributing a net gain to the lake are shown as positive; those exhibiting a net loss from the lake are shown as negative. Storage change is plotted above or below the zero line to indicate net gains or losses in storage. The error bars depict the annual absolute uncertainty values ($|\delta G|$) for the net groundwater flow.

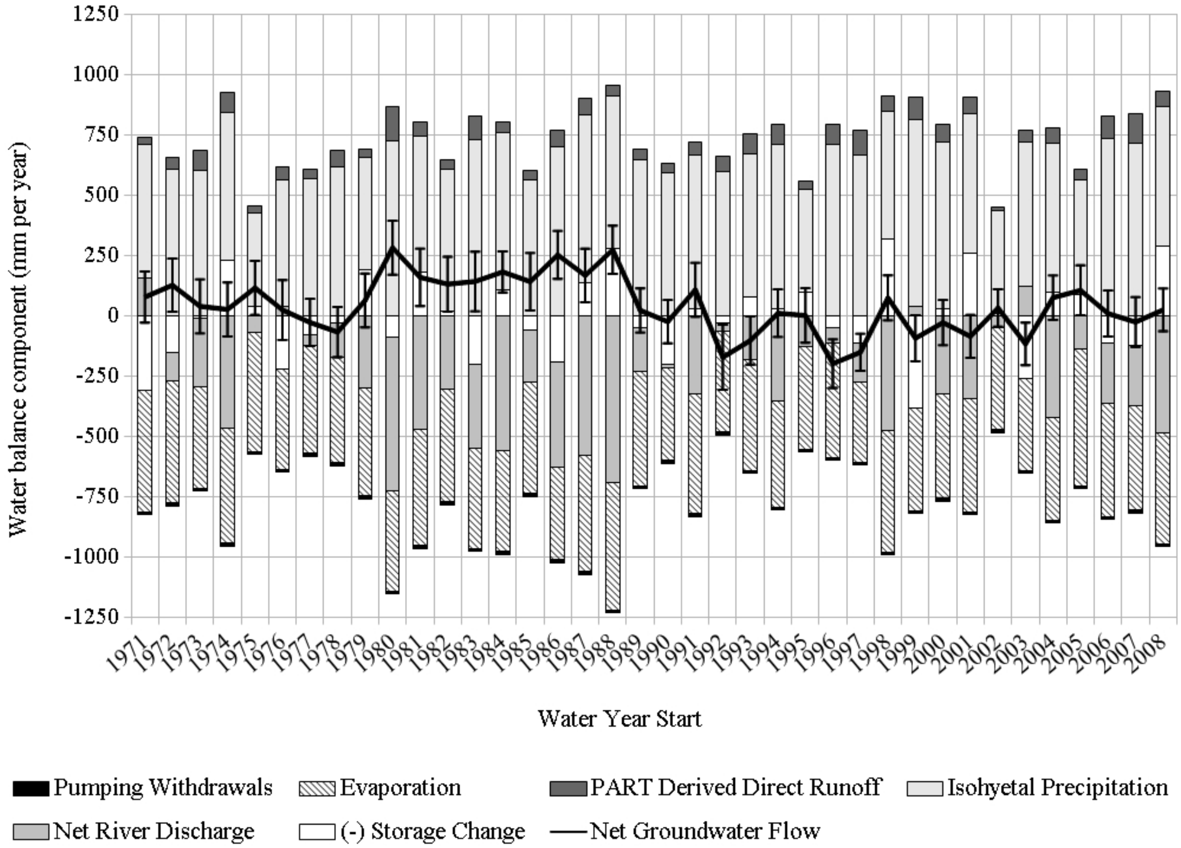

Figure 5: Water balance employing the PART derived direct runoff estimate for water years 1971 to 2008. Components contributing a net gain to the lake are shown as positive; those exhibiting a net loss are shown as negative. Storage change is plotted above or below the zero line to indicate net gains or losses in storage. The error bars depict the annual absolute uncertainty values ($|\delta G|$) for the net groundwater flow.

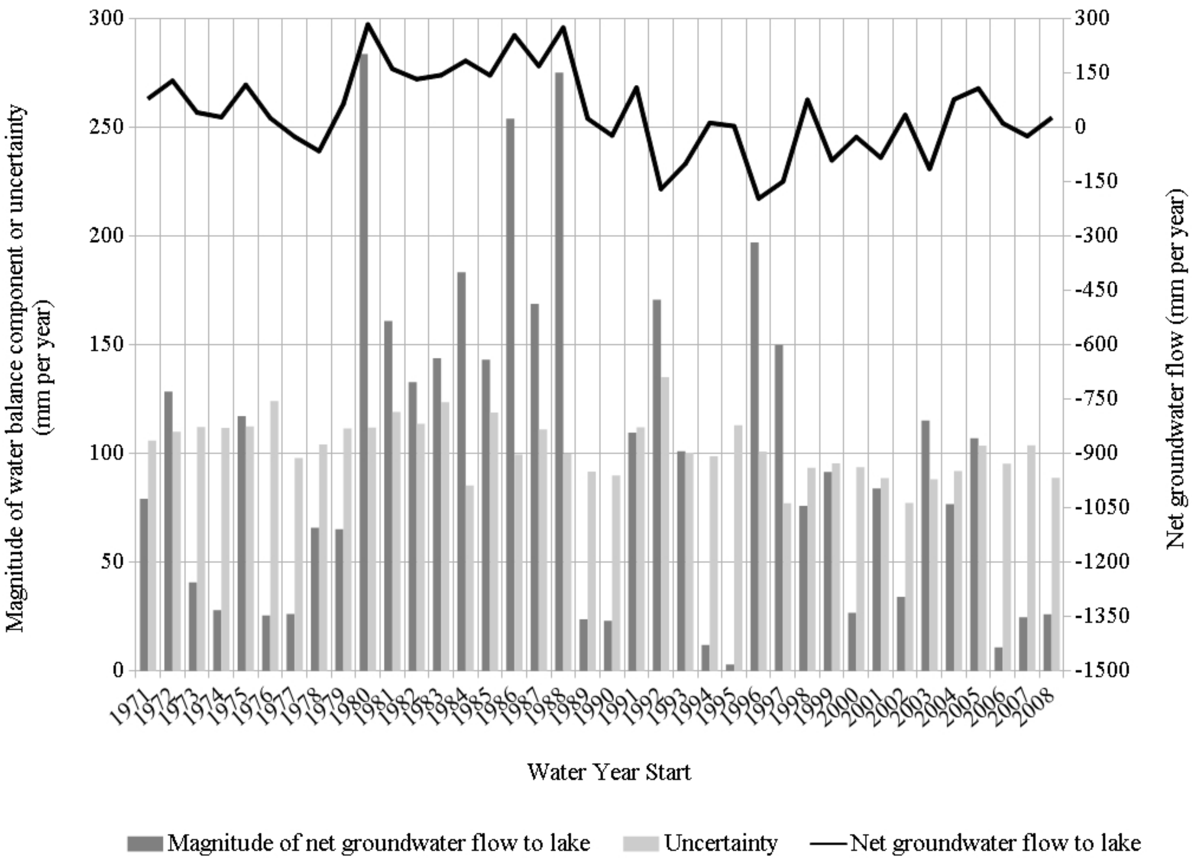

Figure 6: A comparison of the net groundwater component magnitudes (|G|, expressed as positive values) to the absolute uncertainty values (|δG|) obtained for the PART derived water balances for water years 1971 to 2008. The actual values of G are shown for reference.

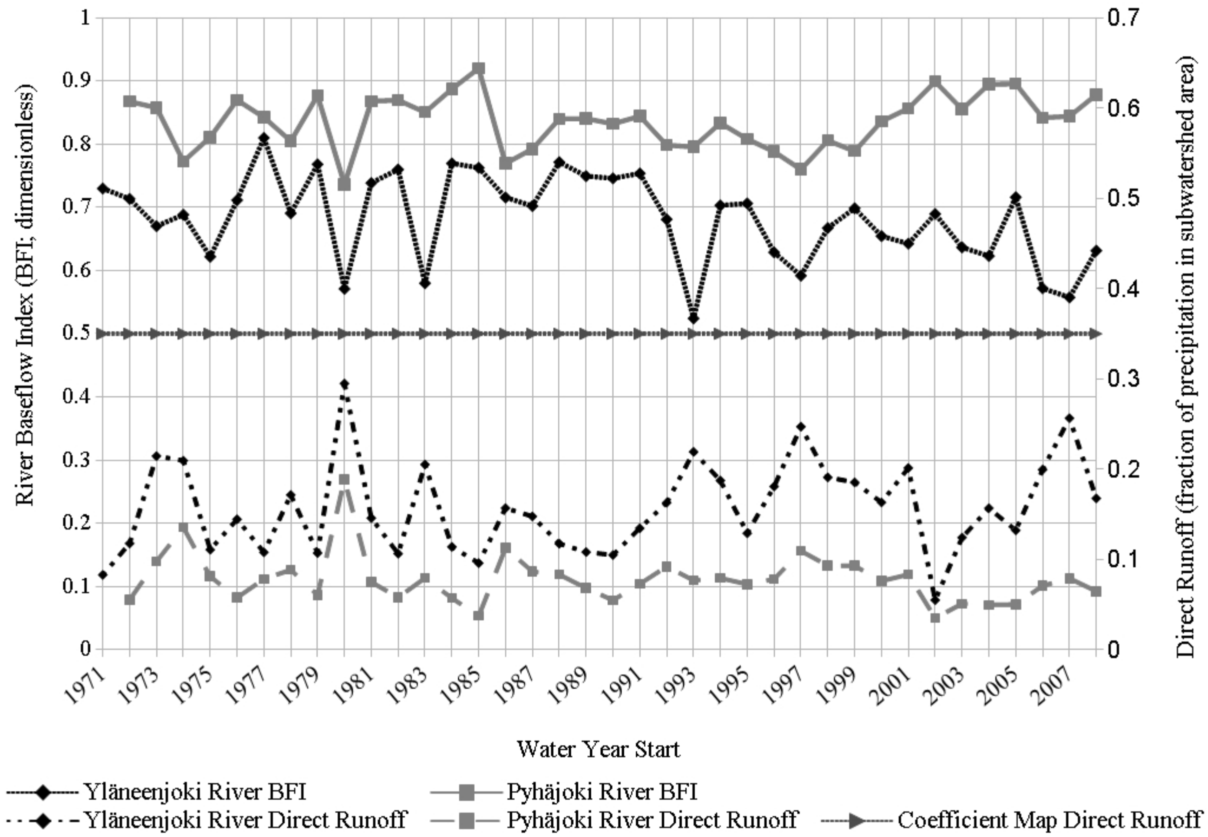

Figure 7: Baseflow index (BFI) values and corresponding direct runoff (*DR*) estimates calculated using PART for the Yläneenjoki and Pyhäjoki Rivers for water years 1971 to 2008. The *DR* value for Lake Pyhäjärvi from the coefficient map runoff method for the direct runoff subwatersheds is also shown for comparison.