Dear Editor,

First of all we'd like to thank you and the reviewers for handling our manuscript, entitled "SPHY v2.0: Spatial Processes in Hydrology". Based on the reviewers' comments we have modified the manuscript substantially, and as a result we think that the manuscript has improved a lot.

Please find our revised manuscript with highlighted changes on the next pages. As we've discussed through email, the marked-up version of the manuscript does not include inserted or deleted citations because of the issue related to the latexdiff package. Additionally, a point-by-point response was not necessary because we've already included a point-by-point response in the reviewers' comments section.

We are looking forward to your response.

Best Regards,

Wilco Terink and co-authors

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SPHY v2.0: Spatial Processes in HYdrology

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Abstract. This paper introduces and presents the Spatial Processes in HYdrology (SPHY) model (v2.0), its development background, its underlying concepts, and some typical applications. The SPHY model is developed using the best

- ⁵ components of existing and well-tested simulation models, and is developed SPHY has been developed with the explicit aim to simulate terrestrial hydrology at flexible scales, under various land use and elimate conditionsphysiographical and hydro-climatic conditions, by integrating key components
- ¹⁰ from existing and well-tested models. SPHY is a spatially distributed leaky bucket type of model, and is applied on a cell-by-cell basis. The model is written in the Python programming language using the PCRaster dynamic modelling framework. Compared to other hydrological models,
- that typically focus on the simulation of streamflow only, the SPHY model has several advantages: it SPHY i) integrates most relevant hydrological hydrologic processes, ii) is setup modular, iii) is easy adjustable and applicable, iii) can easily be linked to remote sensing databas the flexibility
- to study a wide range, and iii) scale of applications, and iv) can be applied for operational as well as strategic decision supporteasily be implemented. The most relevant hydrological processes that are integrated in the SPHY model are rainfall-runoff processes, cryosphere processes, evapotran-
- spiration processes, the simulation of dynamic vegetational cover, lake/reservoir outflow, and the simulation of rootzone moisture contents. Studies in which the SPHY model was successfully applied and tested are described in this paper, and range from including i) real-time soil moisture predic-
- tions to support irrigation management in lowland areas, to-ii) detailed climate change impact studies in snow and glacier-fed river basins, to and iii) operational flow forecasting in mountainous catchments.

1 Introduction

The number and diversity of water-related challenges are large and are expected to increase in the future (Wagener et al., 2010; Lall, 2014). Even today, the ideal condition of having the appropriate amount of good-quality water at the desired place and time, is most often not satisfied (Biswas and Tortajada, 2010; Droogers and Bouma, 2014). It is likely that climate variability and change will intensify food insecurity by water shortages (Wheeler and von Braun, 2013), and loss of access to drinking water (Rockström et al., 2012). Current and future water related challenges are location and time specific and can vary from impact of glacier dynamics (Immerzeel et al., 2011), economic and population growth (Droogers et al., 2012), floods or extended and more prolonged droughts (Dai, 2011), amongst others.

In response to these challenges hydrologist and water resources specialists are developing modeling tools to analyze, understand and explore solutions to support decision makers and operational water managers (Pechlivanidis et al., 2011). Despite difficulties to connect the scientific advances in hydrological modeling with the needs of decision makers and water managers, progress has been made and there is no doubt that modeling tools are indispensable in what is called good "water governance" (Droogers and Bouma, 2014; Liu et al., 2008).

The strength of hydrological models is that they can provide output on ultimate high temporal and spatial resolutions, and for hydrological processes that are difficult to observe sub-processes on the large scale that they are generally applied on (Bastiaanssen et al., 2007). The most important aspect of applying models, is in their use to explore different scenarios, expressing for example, possible effects of changes in population and climate on the water cycle (Droogers and Aerts, 2005). Such scenarios are often referred to as projections. Models are also applied at the operational level to explore interventions (management scenarios) to be

- used by water managers and policy makers. Examples of 125 70 this are changes in reservoir operation rules, water allocation between sectors, investment in infrastructure such as water treatment or desalination plants, and agricultural and irrigation practices. In other words: models enable hydrologists
- and water managers to change focus from a re-active towards 130 a pro-active approach.
 - Over the past decades the land surface and hydrologic communities have made substantial progress in understanding the spatial presentation of fluxes of water
- and energy (Abbott et al., 1986; Wigmosta et al., 1994; 135 VanderKwaak and Loague, 2001; Rigon et al., 2006). Their efforts have led to the development of well-known hydrological models, such as e.g. VIC (Liang et al., 1994, 1996), SWAT (Neitsch et al., 2009), TOPKAPI-ETH (Fin-
- ger et al., 2011; Ragettli and Pellicciotti, 2012; Ragettli 140 et al., 2013, 2014), LISFLOOD (Van Der Knijff et al., 2010), SWIM (Krysanova et al., 2015, 2000, 1998), HYPE (Lindström et al., 2010), mHM (Samaniego et al., 2010), PCR-GLOBWB (van Beek and Bierkens, 2008; Bierkens
- and van Beek, 2009; Wada et al., 2010; Sperna Weiland 145 et al., 2010), MIKE-SHE (Refshaard and Storm, 1995; Oogathoo et al., 2008; Deb and Shukla, 2011) and GEOtop (Rigon et al., 2006; Endrizzi et al., 2013, 2011), amongst others. The number of existing hydrological models is
- probably in the tens of thousands (Droogers and Bouma, 150 2014). Some existing model-overviews-model reviews cover a substantial amount of models: IRRISOFT (Irrisoft, 2014): 114, USGS (USGS, 2014): 110, EPA (EPA, 2014): 211, USACE (HEC, 2014): 18. Interesting is that in groundwater
- modelling ModFlow is the de-facto standard tool to use. For 155 100 hydrological models such a standard is lacking, although the SWAT model is increasingly becoming popular.

Traditionally, hydrologists have put a strong emphasis on streamflow analysis and forecasting, while ignoring

- other hydrological processes . Typical examples are All 160 105 these hydrological models are different with respect to i) the number and detail of hydrological processes that are integrated, ii) their field and iii) scale of application, and iv) the way they are implemented. Whereas for example the
- SWIM (Krysanova et al., 2015, 2000, 1998) and the so-called 165 110 distributed model inter-comparison projects (DMIP) which have been undertaken over the last decade . DMIP, and similar initiatives , have analysed the capacity of models to reproduce observed streamflow.It is however clear that the
- need of decision makers and water managers goes far beyond 170 115 streamflow only. The notion that not only water in streams is relevant, but a better understanding of the full hydrological eycle is required, has led to the popular division of "blueHYPE model (Lindström et al., 2010) both include all
- major hydrological processes, they are typically developed 175 for large-scale (large river basins to continental) applications

and therefore contain less detail as fully distributed models such as e.g. GEOtop (Rigon et al., 2006; Endrizzi et al., 2013, 2011) and "green" water . Moreover, there is a clear need that not only rainfall-runoff processes are relevant, but an integrated approach encompassing processes as glacier and snowmelt, evapotranspiration, reservoirs, changes in landcover and landuse should be better integrated in our models. Finally, the linkage with satellite remote sensing data has been shown to be very relevant to increase accuracy and usefulness of hydrological models TOPKAPI-ETH (Finger et al., 2011; Ragettli and Pellicciotti, 2012; Ragettli et al., 2013, 2014) that typically focus on the catchment and river basin scale. Models like e.g. MIKE-SHE (Refshaard and Storm, 1995; Oogathoo et al., 2008; Deb and Shukla, 2011) and LISFLOOD (Van Der Knijff et al., 2010) have the advantage of being flexible in terms of the spatial and temporal resolution, but their disadvantages are that they do not include glacier processes and that they are not open-source and therefore not available to the larger community.

It is clear that all these models have their pros and cons in terms of i) processes integrated, especially in data scarce areas .

Based on the discussions above, there is a clear need for a hydrological model that combines the strengths of existing modelling approaches such that: (i) most relevant hydrological processesare integrated, (ii) setup is-ii) field of application, iii) scale of application, and iv) implementation. Table 2 shows the pros and cons of some well-known hydrological models, including the Spatial Processes in HYdrology (SPHY) model. Over the last couple of years we have developed the SPHY model, and improved its usefulness by applying the model in various research projects. SPHY has been developed with the explicit aim to simulate terrestrial hydrology under various physiographical and hydro-climatic conditions by integrating key components from existing and well-tested models: HydroS (Droogers and Immerzeel, 2010), SWAT (Neitsch et al., 2009), PCR-GLOBWB (van Beek and Bierkens, 2008; Bierkens and van Beek, 2009; Wada et al., 2010; Sperna Weiland et al., 2010), SWAP (van Dam et al., 1997) and HimSim (Immerzeel et al., 2011). Based on Tab. 2 it is clear that SPHY i) integrates most hydrologic processes, including glacier processes, ii) has the flexibility to study a wide range of applications, including climate and land use change impacts, irrigation planning, and droughts, iii) can be used for catchment- and river basin scale applications as well as farm- and country-level applications, and has a flexible spatial resolution, and iv) can easily be implemented. Implementation of SPHY is relatively easy because it i) is open-source, ii) in- and output maps can directly be used in GIS, iii) is setup modular in order to switch on/off relevant/irrelevant processes and thus decreases model run-time , (iii) can relatively easily be adjusted and applied, (iv) can easily be linked to remotely sensed and

data requirements, iv) needs only daily precipitation and temperature data as climate forcing, v) can be forced with remote sensing data, and (v) can be applied for operational as 230

well as strategic decision support. vi) uses a configuration file that allows the user to change model parameters and choose the model output that needs to be reported.

Over the last couple of years we have developed the Spatial Processes in Hydrology (SPHY) model and improved 235

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its usefulness by applying the model in various research 185 projects. The objective of this publication is therefore to introduce and present the SPHY model, its development background, and demonstrate some typical applications. The model executable and source code are in the public domain 240 (open access) and can be obtained from our website free of

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charge (www.sphy.nl).

2 Model overview

IntroductionBackground 2.1

- The SPHY model has been developed by combining the best components of existing and well tested simulation models: 250 195 HydroS , SWAT , PCR-GLOBWB , SWAP and HimSim . SPHY was developed with the explicit aim to simulate terrestrial hydrology at flexible scales, under various land use and climate conditions. SPHY SPHY is a spatially dis-
- tributed leaky bucket type of model, and is applied on a cell- 255 200 by-cell basis. The main terrestrial hydrological processes are described in a physically consistent conceptual way so that changes in storages and fluxes can be assessed adequately over time and space. SPHY is written in the Python program-
- ming language using the PCRaster (Karssenberg et al., 2001; 260 Karssenberg, 2002; Karssenberg et al., 2010; Schmitz et al., 2009, 2013) dynamic modelling framework.

SPHY is grid-based and cell values represent averages over a cell (Fig. 1). For glaciers, sub-grid variability is taken

into account: a cell can be glacier-free, partially glacierized, 265 210 or completely covered by glaciers. The cell fraction not covered by glaciers consists of either land covered with snow or land that is free of snow. Land that is free of snow can consist of vegetation, bare soil, or open water. The dynamic vegeta-

tion module accounts for a time-varying fractional vegetation 270 215 coverage, which affects processes such as interception, effective precipitation, and potential evapotranspiration. Figure 2 provides a schematic overview of the SPHY model modeling concepts.

The land compartment is divided in soil column structure 275 220 is similar to VIC (Liang et al., 1994, 1996) with two upper soil stores and a third groundwater store, with their. Their corresponding drainage components are: surface runoff, lateral flow and base flow. SPHY simulates for each cell precip-

itation in the form of rain or snow, depending on the tempera- 280 225 ture. Any precipitation Precipitation that falls on land surface can be intercepted by vegetation and in part or in whole evaporated. The snow storage is updated with snow accumulation and/or snow melt. A part of the liquid precipitation is transformed in surface runoff, whereas the remainder infiltrates into the soil. The resulting soil moisture is subject to evapotranspiration, depending on the soil properties and fractional vegetation cover, while the remainder contributes to river discharge by means of lateral flow from the first soil layer, and baseflow from the groundwater layer.

Melting of glacier ice contributes to the river discharge by means of a slow and fast component, being (i) percolation to the groundwater layer that eventually becomes baseflow, and (ii) direct runoff. The cell-specific runoff, which becomes available for routing, is the sum of surface runoff, lateral flow, baseflow, snow melt and glacier melt.

If no lakes are present, then the user can choose for a simple flow accumulation routing scheme: for each cell the accumulated amount of water that flows out of the cell into its neighboring downstream cell is calculated. This accumulated amount is the amount of water in the cell itself plus the amount of water in upstream cells of the cell. For each cell, the following procedure is performed: using the local drain direction networkthe catchment of a cell is determined which is made up the cell itself and all cells that drain to the eell, and is calculated using the flow direction network. If lakes are present, then the fractional accumulation flux routing scheme is used:-; depending on the actual lake storage, a fraction of that storage becomes available for routing and is extracted from the lake, while the remaining part becomes the updated actual lake storage. The flux available for routing is routed in the same way as in the simple flow accumulation routing scheme.

As input SPHY requires data on state variables static data as well as dynamic variables data. For the state variables static data the most relevant are: Digital Elevation Model (DEM), land use type, glacier cover, lakes/reservoirs and soil characteristics. The main dynamic variables are climate datadata consists of climate data, such as precipitation, temperature, and reference evapotranspiration. Since SPHY is grid-based, optimal use of remote sensing data and global data sources can be made. For example, the Normalized Difference Vegetation Index (NDVI) (Tucker, 1979; Carlson and Ripley, 1997; Myneni and Williams, 1994) can be used to determine the Leaf Area Index (LAI) in order to estimate the growth-stage of land cover. For setting-up the model, data on streamflows streamflow data are not necessary. However, to undertake a proper calibration and validation procedure flow data are required. The model could also be calibrated using actual evapotranspiration, soil moisture contents, and/or snow coverage. covered area (SCA). Section 3.2 contains an example application in which the SPHY model has been calibrated using MODIS snow cover images. An overview of the adjustable SPHY model parameters is shown in Appendix A (Tab. 1).

The SPHY model provides a wealth of output data that can be selected based on the preference of the user. Spatial output can be presented as maps of all the hydrological processes. Maps that can be displayed as output include

- available hydrological processes, i.e. actual evapotranspiration, runoff generation (separated by its components), and groundwater recharge. These maps can be generated on daily base, but most users prefer to get those a daily basis, but can also be aggregated at monthly or annual aggregated time pe-
- 290 riods. Time-series can be generated for each location cell in 340 the study area. Time-series often used are stream flow under current and future conditionsstreamflow, actual evapotranspiration and recharge to the groundwater.

2.2 Modules

- ²⁹⁵ SPHY enables the user to turn on/off modules that are not required(processes) that are relevant/irrelevant for the area of interest. This concept is very useful if the user is studying hydrological processes in regions where not all hydrologi-³⁵⁰ cal processes are relevant. A user may for example be in-
- terested in studying irrigation water requirements in central Africa. For this region glacier and snow melting processes are less relevantirrelevant, and can thus be switched off. Another user may only be interested in simulating moisture ³⁵⁵ conditions in the first soil layer, allowing the possibility to
- 305 switch off the routing and groundwater modules. The advantages of turning off irrelevant modules are two-fold: (i) decrease model run-time, and (ii) decrease the amount of required model input data. It should be noted, however, that the hydrologic model structure should be specific to
- the catchment's characteristics (Pomeroy et al., 2007; Clark et al., 2008; Niu et al., 2011; Essery et al., 2013; Clark et al., ³⁶⁰ 2015a, b). It is therefore essential that the user knows which catchment characteristics and processes should be included in their modeling framework.
- Figure 3 represents an overview of the six modules available: glaciers, snow, groundwater, dynamic vegetation, sim-³⁶⁵ ple routing, and lake/reservoir routing. All modules can run independently from each other, except for the glaciers module. If glaciers are present, then snow processes are relevant as well (Verbunt et al., 2003; Singh and Kumar, 1997)₇
- meaning that the snow module is turned on automatically if ³⁷⁰ the glaciers module is turned on. Since melting glacier water percolates to the groundwater layer, the glaciers module cannot run with the groundwater module turned off. For
- routing two Two modules are available , being for runoff routing: (i) a simple flow accumulation routing scheme, and ³ (ii) a fractional flow accumulation routing scheme used when lakes/reservoirs are present. The user has the option to turn off routing, or to choose between one of these two routing modules. All hydrological processes incorporated in the
 - SPHY model are described in detail in the following sections.

2.3 Reference and potential evapotranspiration

Various methods exist to calculate the reference evapotranspiration (ET_r) . From the many existing methods, the FAO-56 application of the Penman-Monteith equation has been most widely used, and can be considered as a sort of standard. Despite the good physical underlying theory of this method, its major drawback is its relatively high data demand. The the Penman-Monteith equation requires air temperature, windspeed, relative humidity, and solar radiation data. In many regions around the world (e.g. Africa), these data are not readily measured at meteorological stations, and therefore not suitable. Another well-known method (Allen et al., 1998) to calculate the reference evapotranspiration $(ET_r is the Hargreaves)$ method . showed that in many regions around the globe, the Hargreaves and Penman-Monteith method showed comparable results. A lack of meteorological data brought), its major limitation is the high data demand for energy based methods. This brought Hargreaves and Samani (1985) to derive the modified Hargreaves equation that is only based on the average, minimum, and maximum daily air temperature based on temperature only. For this reason, this equation has also been implemented in the SPHY model, according to:

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$$ET_r = 0.0023 \cdot 0.408 \cdot Ra\left(T_{avg} + 17.8\right) \cdot TD^{0.5} \tag{1}$$

with RA $[MJm^{-2}d^{-1}]$ the extraterrestrial radiation, T_{avg} [°C] the average daily air temperature, and TD [°C] the daily temperature range, defined as the difference between the daily maximum and minimum air temperature. The constant 0.408 is required to convert the radiation to mmevaporation equivalentsuits to mm, and RA can be obtained from tables (Allen et al., 1998) or equations using the day of the year and the latitude of the area of interest.

According to Allen et al. (1998), the ET_r is the evapotranspiration rate from a reference surface , not short of water with access to sufficient water to allow evapotranspiration at the potential rate. The reference surface is a hypothetical grass reference crop with specific characteristics. The potential evapotranspiration ET_p has no limitations on crop growth or evapotranspiration from soil water and salinity stress, crop density, pests and diseases, weed infestation or low fertility. Allen et al. (1998) determined the ET_p by the crop coefficient approach, where the effect of various weather conditions are incorporated into ET_r , and the crop characteristics in the crop coefficient (Kc), using:

$$ET_{p,t} = ET_{r,t} \cdot Kc \tag{2}$$

with $ET_{p,t}$ [mm] the potential evapotranspiration on day t, $ET_{r,t}$ [mm] the reference evapotranspiration on day t, and Kc [-] the crop coefficient. The effect of both crop transpiration and soil evaporation are integrated into the Kc.

If the dynamic vegetation module in SPHY is not used, then the user can opt to i) use a single constant Kc throughout the entire simulation periodor use a , or ii) use a pre-defined time-series of crop coefficients as model in-430

- ³⁸⁵ pre-defined time-series of crop coefficients as model input. Plausible values for Kc can be obtained from literature (Allen et al., 1998; FAO, 2013). However, vegetation is generally very dynamic throughout the year, and therefore it is not very realistic to use a single constant Kc throughout the
- entire simulation period. It is therefore more realistic to use a <u>pre-defined</u> time-series of crop coefficients as model input, or use the dynamic vegetation moduleinstead, instead of a ⁴ <u>single constant Kc</u>. This can be adjusted according to the user's preferences.
- The Kc can be estimated using remotely sensed data (Rafn et al., 2008; Contreras et al., 2014). In the dynamic vegetation module the Kc is scaled throughout the year using the NDVI⁴⁴⁰ and the maximum and minimum values for Kc, which are crop specific. These values for Kc can easily be obtained from Allen et al. (1998). Then the Kc is calculated using:
- $_{400}$ from Allen et al. (1998). Then the K c is calculated using:

$$Kc = Kc_{min} + (Kc_{max} - Kc_{min}) * \frac{(NDVI - NDVI_{min})}{(NDVI_{max} - NDVI_{min})} \quad (3)$$

- with $NDVI_{max}$ [-] and $NDVI_{min}$ [-] the maximum and minimum values for the NDVI (vegetation type dependent). Using this approach in SPHY to calculate the Ke dynamically, using the NDVI as input, shows the strong linkage with the use of This approach shows the flexibility
- ⁴¹⁰ of SPHY to use remote sensing data (e.g. NDVI) as input to improve model accuracy.

2.4 Dynamic vegetation processes

2.4.1 Canopy Maximum canopy storage

SPHY allows the user to use the dynamic vegetation module in order to incorporate a time-variable vegetation cover and corresponding rainfall interception. In order to calculate the rainfall interception, the canopy storage needs to be calculated, using a time-series of the Normalized Difference Vegetation Index (NDVI) NDVI (Carlson and Ripley, 1997).

⁴²⁰ The first step involves the calculation of the fraction photosynthetically active radiation (FPAR). The FPAR can be calculated using a relation between the NDVI and the FPAR, which was found by Peng et al. (2012) and described by Sellers et al. (1996), according to:

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$$FPAR = min \left(\frac{(SR - SR_{min})(FPAR_{max} - FPAR_{min})}{(SR_{max} - SR_{min})} + FPAR_{min}, 0.95\right) \quad (4)^{47}$$

with:

$$SR = \frac{1 + NDVI}{1 - NDVI} \tag{5}$$

and $FPAR_{max}$ [-] and $FPAR_{min}$ [-] having values of 0.95 and 0.001, respectively. An FPAR of 0.95 is equivalent to the maximum LAI for a particular class, and an FPAR of 0.001 is equivalent to a minimum LAI. In order to calculate the FPAR, an NDVI time-series is required.

The second step is the calculation of the Leaf-Area-Index (LAI), which is eventually required to calculate the maximum canopy storage ($Scan_{max}$). According to Monteith (1973), the LAI for vegetation that is evenly distributed over a surface can be calculated using a logarithmic relation between the LAI and FPAR, according to:

$$LAI = LAI_{max} \cdot \frac{\log\left(1 - FPAR\right)}{\log\left(1 - FPAR_{max}\right)} \tag{6}$$

with LAI [-] the Leaf-Area-Index, and LAI_{max} [-] the maximum Leaf-Area-Index (vegetation type dependent). This means that the maximum and minimum LAI values are related to the maximum and minimum of FPAR. Table 3 shows the LAI_{max} values for a certain number of vegetation types.

For vegetation that is concentrated in clusters the linear relation from Goward and Huemmrich (1992) is often used. However, since SPHY is generally applied using grid cell resolutions between 250 m and 1 km, we can assume that the effect of having vegetation concentrated in clusters is neglectable. Therefore, the calculation of the LAI in SPHY is done using the logarithmic relation of Monteith (1973) (Eq. (6)).

The next step involves the calculation of the maximum canopy storage ($Scan_{max}$ [mm]). Many different relations between $Scan_{max}$ and the LAI can be found in literature depending on the vegetation type (de Jong and Jetten, 2010). The best results for crop canopies are shown by Kozak et al. (2007) and are archived by Von Hoyningen-Huene (1981) who derived the following relation between $Scan_{max}$ and the LAI:

$$Scan_{max} = 0.935 + 0.498LAI - 0.00575LAI^2 \tag{7}$$

Note that even with an LAI of zero, $Scan_{max}$ is still close to 1 mm.

2.4.2 Interception

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Interception is calculated on a daily basis if the dynamic vegetation module is used, and consists of the daily precipitation plus the intercepted water remaining in the canopy storage from the previous day. First of all the canopy storage is updated with the amount of precipitation of the current day:

$$Scan_t = Scan_{t-1} + P_t \tag{8}$$

with $Scan_t$ [mm] the canopy storage on day t, $Scan_{t-1}$ [mm] the canopy storage on day t-1, and P_t [mm] the amount of precipitation on day t. The portion of precipitation that cannot be stored in the canopy storage is known as precipitation throughfall, or effective precipitation, according to:

$$Pe_t = max\left(0, Scan_t - Scan_{max,t}\right) \tag{9}$$

with $Pe_t \ [mm]$ the effective precipitation on day t, and $Scan_t \ [mm]$ the canopy storage on day t. This equation shows that precipitation throughfall only occurs if the water stored in the canopy exceeds the maximum canopy storage. After the effective precipitation has been calculated, the canopy storage is updated as:

$$Scan_t = Scan_t - Pe_t \tag{10}$$

The remaining amount of water stored in the canopy is available for interception, and the amount of water that will be intercepted depends on the atmospheric demand for open water evaporation. A commonly used value for the atmospheric demand for open water evaporation is 1.5 (Allen et al., 1998), ⁵³⁰ which is derived from the ratio between one and the mean

⁴⁹⁵ pan evaporation coefficient Kp (~0.65). If desired this value can easily be altered in the model code. The interception can now be calculated using:

$$Int_t = \min\left(1.5ET_{r,t}, Scan_t\right) \tag{11}_{535}$$

with Int_t [mm] the intercepted water on day t, and $ET_{r,t}$ [mm] the reference evapotranspiration on day t. Finally, the canopy storage is updated by subtracting the interception:

$$Scan_t = Scan_t - Int_t \tag{12}$$

2.5 Snow processes

For each cell a dynamic snow storage is simulated at a daily time step, adopted from on the model presented by Kokkonen et al. (2006). The model keeps track of a snow storage, which is fed by precipitation and generates runoff from snow melt. 545 Refreezing of snow melt and rainfall within the snowpack are 510 simulated as well.

2.5.1 Snow and rainfall

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Depending on a temperature threshold, precipitation is defined to fall in either solid or liquid state. Daily snow accumulation, which is defined as solid precipitation, is calculated as:

$$P_{s,t} = \left\{ \begin{array}{cc} Pe_t & \text{if } T_{avg,t} \le T_{crit} \\ 0 & \text{if } T_{avg,t} > T_{crit} \end{array} \right\}$$
(13)

with $P_{s,t}$ [mm] the snowfall on day t, Pe_t [mm] the effective precipitation on day t, $T_{avg,t}$ [°C] the mean air temperature on day t, and T_{crit} [°C] a calibrated temperature threshold for precipitation to fall as snow. The precipitation that falls as rain is defined as liquid precipitation, and is calculated as:

$$P_{l,t} = \left\{ \begin{array}{cc} Pe_t & \text{if } T_{avg,t} > T_{crit} \\ 0 & \text{if } T_{avg,t} \le T_{crit} \end{array} \right\}$$
(14)

with $P_{l,t}$ [mm] being the amount of rainfall on day t.

2.5.2 Snow melt, refreezing, and storage

To simulate snow melt, the well-established and widely used degree day melt modeling approach is used (Hock, 2003). The application of degree-day models is widespread in cryospheric models and is based on an empirical relationship between melt and air temperature. Degree-day models are easier to set up compared to energy-balance models, and only require air temperature, which is mostly available and relatively easy to interpolate (Hock, 2005). Using a degreeday modeling approach, the daily potential snow melt is calculated as follows:

$$A_{pot,t} = \left\{ \begin{array}{cc} T_{avg,t} \cdot DDF_s & \text{if } T_{avg,t} > 0\\ 0 & \text{if } T_{avg,t} \le 0 \end{array} \right\}$$
(15)

with $A_{pot,t}$ [mm] the potential snow melt on day t, and DDF_s [mm °C⁻¹d⁻¹] a calibrated degree day factor for snow. The actual snow melt is limited by the snow store at the end of the previous day, and is calculated as:

$$A_{act,t} = \min(A_{pot,t}, SS_{t-1}) \tag{16}$$

with $A_{act,t}$ [mm] the actual snow melt on day t, and SS_{t-1} [mm] the snow store on day [t-1]. The snow store from day [t-1] is then updated to the current day t, using the actual snow melt $(A_{act,t})$ and the solid precipitation $(P_{s,t})$. Part of the actual snow melt refreezes freezes within the snow pack and thus does not runoff immediately. When temperature is below the melting point, melt water that has refrozen

frozen in the snow pack during [t-1] is added to the snow store as:

$${}_{550} \quad SS_t = \left\{ \begin{array}{ccc} SS_{t-1} + P_{s,t} + SSW_{t-1} & \text{if } T_{avg,t} \\ & < 0 \\ SS_{t-1} + P_{s,t} - A_{act,t} & \text{if } T_{avg,t} \\ & \geq 0 \end{array} \right\}$$
(17)

with SS_t the snow store on day t, SS_{t-1} the snow store on day [t-1], $P_{s,t}$ the solid precipitation on day t, $A_{act,t}$ the actual snow melt on day t, SSW_{t-1} the amount of refrozen frozen melt water on day [t-1]. The units for all terms are mm.

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The capacity of the snow pack to refreeze freeze snow melt is characterized by introducing a calibrated water storage capacity (SSC $[mm \cdot mm^{-1}]$), which is the total mm water equivalent of snow melt that can refreeze [mm] that that can freeze per mm water equivalent of snow in the snow

that can freeze per mm water equivalent of snow in the snow store. The maximum of melt water that can refreeze freeze $(SSW_{max} \ [mm])$ is thus limited by the thickness of the snow store:

$$SSW_{max,t} = SSC \cdot SS_t \tag{18}$$

⁵⁶⁵ Then the amount of melt water stored in the snowpack, and ⁶⁰⁰ that can refreeze freeze in the next time-step, is calculated as:

$$SSW_{t} = \left\{ \begin{array}{l} 0 & \text{if } T_{avg,t} < 0\\ min(SSW_{max,t}, SSW_{t-1} + & \text{if } T_{avg,t} \ge 0\\ P_{l,t} + A_{act,t}) & (19) \end{array} \right\}$$

with SSW_t the amount of melt water in the snow pack on day t, $SSW_{max,t}$ the maximum of melt water that can refreeze freeze on day t, SSW_{t-1} the amount of refrozen frozen melt water on day [t-1], $P_{l,t}$ the amount of rainfall on day t, and $_{610}$ $A_{act,t}$ the actual snow melt on day t. The units of all terms are in mm.

The total snow storage $(SST \ [mm])$ consists of the snow store and the melt water that has refrozen can freeze within it, according to:

$$SST_t = (SS_t + SSW_t) \cdot (1 - GlacF) \tag{20}_{615}$$

with (1 - GlacF) [-] the gridcell fraction not covered with glaciers. In SPHY it is therefore assumed that snow accumulation and melt can only occur on the gridcell fraction determined as land surface. Snow falling on glaciers is incorporated in the glacier module.

2.5.3 Snow runoff

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Runoff from snow $(SRo \ [mm])$ is generated when the air temperature is above melting point and no more melt water can be refrozen frozen within the snow pack, according to:

$$SRo_{t} = \left\{ \begin{array}{c} A_{act,t} + P_{l,t} - \Delta SSW & \text{if } T_{avg,t} > 0\\ 0 & \text{if } T_{avg,t} \le 0 \end{array} \right\}$$
(21)

with $\Delta SSW \ [mm]$ the change in melt water stored in the snowpack according to:

$$\Delta SSW = SSW_t - SSW_{t-1} \tag{22}$$

2.6 Glacier processes

Since the SPHY model usually operates at a spatial resolution between 250 m and 1 km, the dynamics of glaciers such as ice flow cannot be resolved explicitly. Therefore, glaciers in SPHY are considered as melting surfaces which can completely or partly cover a grid cell.

2.6.1 Glacier melt

Glacier melt is calculated with a degree day modeling approach as well (Hock, 2005). Because glaciers covered with debris melt at different rates than debris-free glaciers (Reid et al., 2012), a distinction can be made between different degree day factors for both types. The daily melt from debris free glaciers (A_{CI} [mm]) is calculated as:

$$A_{CI,t} = \left\{ \begin{array}{cc} T_{avg,t} \cdot DDF_{CI} \cdot F_{CI} & \text{if } T_{avg,t} > 0\\ 0 & \text{if } T_{avg,t} \le 0 \end{array} \right\}$$
(23)

with DDF_{CI} $[mm \circ C^{-1}d^{-1}]$ a calibrated degree day factor for debris free glaciers and F_{CI} [-] is the fraction of debris free glaciers within the fractional glacier cover (*GlacF*) of a grid cell. The daily melt from debris covered glaciers (A_{DC} [mm]) is calculated in a similar way, but with a different degree day factor:

$$A_{DC,t} = \left\{ \begin{array}{cc} T_{avg,t} \cdot DDF_{DC} \cdot F_{DC} & \text{if } T_{avg,t} > 0\\ 0 & \text{if } T_{avg,t} \le 0 \end{array} \right\}$$
(24)

where DDF_{DC} $[mm \circ C^{-1}d^{-1}]$ is a degree day factor for debris covered glaciers and F_{DC} [-] is the fraction of debris covered glaciers within the fractional glacier cover of a grid cell. The total glacier melt per grid cell $(A_{GLAC} [mm])$ is then calculated by summing the melt from the debris-covered and debris-free glacier types and multiplying by the fractional glacier cover, according to:

$$A_{GLAC,t} = (A_{CI,t} + A_{DC,t}) \cdot GlacF$$
(25)

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620 2.6.2 Glacier runoff

In SPHY a fraction of the glacier melt percolates to the $_{660}$ ground water while the remaining fraction runs off. The distribution of both is defined by a calibrated glacier melt runoff factor (*GlacROF* [-]) which can have any value ranging from 0 to 1. Thus the generated runoff *GRo* [*mm*] from

glacier melt is defined as:

$$GRo_t = A_{GLAC,t} \cdot GlacROF \tag{26}$$

2.6.3 Glacier percolation

The percolation from glacier melt to the groundwater $(G_{perc} [mm])$ is defined as:

$$G_{perc,t} = A_{GLAC,t} \cdot (1 - GlacROF) \tag{27}_{675}$$

The glacier water that percolates to the groundwater percolated glacier water is added to the water percolating that percolates from the soil layers of the non-glacierized part of the grid cell (Sect. 2.7.1 and Sect. 2.7.6) which will then recharge, which eventually recharges the groundwater.

2.7 Soil water processes

2.7.1 Soil water balances

The soil water processes in SPHY are modelled for three soil layers (Fig. 2), being (i) the first soil layer (rootzone), (ii) second soil layer (subzone), and (iii) third soil layer (groundwater layer). The water balance of the first soil layer is:

$$SW_{1,t} = SW_{1,t-1} + Pe_t - Ea_t - RO_t - LF_{1,t} - Perc_{1,t} + Cap_t$$
(28)

with $SW_{1,t}$ and $SW_{1,t-1}$ the water content in the first soil layer on day t and [t-1], respectively, Pe_t the effective precipitation on day t, Ea_t the actual evapotranspiration on day t, RO_t the surface runoff on day t, $LF_{1,t}$ the lateral flow from the first soil layer on day t, $Perc_{1,t}$ percolation from the first ⁶⁹⁵ to the second soil layer on day t, Cap_t capillary rise from the second to the first soil layer on day t. The second soil layer water balance is:

$$SW_{2,t} = SW_{2,t-1} + Perc_{1,t} - Perc_{2,t} - Cap_t$$
(29)

with $SW_{2,t}$ and $SW_{2,t-1}$ the water content in the second soil layer on day t and [t-1], respectively, and $Perc_{2,t}$ percolation from the second to the third soil layer on day t. The third ₇₀₅ soil layer water balance is given as:

$$SW_{3,t} = SW_{3,t-1} + Gchrg_t - BF_t$$
(30)

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with $SW_{3,t}$ and $SW_{3,t-1}$ the water content in the third soil layer on day t and [t-1], respectively, $Gchrg_t$ groundwater recharge from the second to the third soil layer on day t, and BF_t baseflow on day t. If the glacier module is used, then groundwater recharge consists of percolation from the second soil layer and percolated glacier melt, otherwise only percolation from the second soil layer is taken into account.

The user can opt to run SPHY without the third soil layer (groundwater). This may be desirable if the user for example is mainly interested in simulating soil moisture conditions in the rootzone, instead of evaluating e.g. for instance the contribution of baseflow to the total routed river flow. In that case only the two upper soil layers are used where the bottom boundary of soil layer two is controlled by a seepage flux (pos. outward), and instead of baseflow from the third soil layer, water leaves the second soil layer through lateral flow. With the groundwater module turned off, the water balance for the second soil layer is:

$$SW_{2,t} = SW_{2,t-1} + Perc_{1,t} - LF_{2,t} - Cap_t - Seep$$
 (31)

with $LF_{2,t}$ lateral flow from the second soil layer, and Seep seepage in or out of the second soil layer (pos. is outgoing). The units for all water balance terms are mm.

2.7.2 Actual evapotranspiration

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Evapotranspiration refers to both the transpiration from vegetation and the evaporation from soil or open water. As it was mentiond in Sect. 2.3, the Kc accounts for both the crop transpiration and soil evaporation. The additional use of the dynamic vegetation module accounts for a time-variable vegetation cover, meaning that the role of evaporation becomes more dominant as soons soon as vegetation cover decreases.

Many limiting factors (e.g. salinity stress, water shortage, water excess, diseases) can cause a reduction in potential evpotranspiration (ET_p) , resulting in the actual evapotranspiration rate (ET_a) . Since SPHY is a water-balance model, SPHY only accounts for stresses related to water shortage or water excess. If there is too much water in the soil profile, then the plant is unable to extract water because of oxygen stress (Bartholomeus et al., 2008). The calculation of evapotranspiration reduction due to water excess (oxygen stress) is quite complex and requires a vast amount of parameters that is substantial amount of plant and soil properties (e.g. soil temperature, root dry weight, plant respiration, and minimum gas filled soil porosity (Bartholomeus et al., 2008)) that are generally not available for the spatial scale that SPHY is applied on. Therefore, SPHY uses a-an evapotranspiration reduction parameter $(ETred_{wet})$ that has a value of 0 if the soil is saturated, and otherwise it will have a value of 1. This parameter is used in the following equation to calculate the actual evapotranspiration:

$$ET_{a,t} = ET_{p,t} \cdot ETred_{wet} \cdot ETred_{dry}$$
(32)

with $ET_{a,t}$ [mm] the actual evapotranspiration on day t, 755 $ET_{p,t}$ [mm] the potential evapotranspiration on day t, and 710 $ETred_{wet}$ and $ETred_{dry}$ being the reduction parameters for water excess and water shortage conditions, respectively. The $ETred_{dry}$ is calculated using the Feddes equation (Feddes et al., 1978), which assumes a linear decline in rootwater up- 760 take if the water pressure head drops below a critical value.

- This critical value can be determined using the soil water retention curve (pF-curve), which relates the soil type to its water moisture content of the soil with its binding capacity. This relation is unique for each soil type. The binding capacity is a suction force , and is (H) and is therefore often
- expressed in cm negative water column. The pF is simply $_{765}$ pF-value is simply a conversion of the suction force (*H*), and is calculated as:

$$pF = \log_{10} \left(-H\right) \tag{33}$$

with *H* [*cm*] the soil water pressure. Soils that are at field capacity generally have a pF of 2, meaning -100 cm of water ₇₇₀ column, and soils that are at permanent wilting point have a pF of 4.2, or -16000 cm of water column. The permanent wilting point is often referred to as the point where the crop dies. In SPHY it is assumed that the linear decline in root-

water uptake starts at a pF of 3 (-1000 cm water column). ₇₇₅ Therefore, the $ETred_{dry}$ [-] is calculated as:

$$ETred_{dry,t} = \frac{SW_{1,t} - SW_{1,pF4,2}}{SW_{1,pF3} - SW_{1,pF4,2}}$$
(34)

with $ETred_{dry,t}$ [-] the reduction in rootwater uptake due to water shortage on day t, $SW_{1,t}$ [mm] the actual soil water content in the first soil layer on day t, and $SW_{1,pF3}$ [mm] ₇₈₀ and $SW_{1,pF4}$ [mm] the soil water content in the first soil layer at pF3 and pF4, respectively. $ETred_{dry}$ can therefore have values ranging between zero and one, where a value of one represents optimal plant growing conditions, and zero means no rootwater uptake at all. Using $ETred_{dry}$

is eventually used in Eq. (32) to calculate the ET_{a} .

2.7.3 Surface runoff

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Within the field of hydrology two types of surface runoff can be distinguished: Hewlettian and Hortonian runoff.
745 Hewlettian runoff is often referred to as saturation excess overland flow, while Hortonian runoff is also known as infiltration excess overland flow. Hewlettian runoff occurs when the soil is saturated, and more rainfall is added, resulting in direct surface runoff. Hortonian runoff occurs 780

⁷⁵⁰ when the rainfall intensity exceeds the infiltration capacity of the soil: during wetting of the soil, the soil infiltration capacity decreases over time, and if at a certain moment the rainfall intensity exceeds the soil infiltration capacity, surface runoff occurs.- Since the SPHY model runs on a daily time-step, the model does not account for sub-daily variability in rainfall intensities, meaning that Hortonian runoff processes can be considered as. Therefore, the Hortonian runoff process (Beven, 2004; Corradini et al., 1998), which refers to infiltration excess overland flow, is considered less important. For this reason SPHY uses the Hewlettian runoff concept saturation excess overland flow process, known as Hewlettian runoff (Hewlett, 1961), to calculate surface runoff. Surface runoff is calculated from the first soil layer:

$$RO = \left\{ \begin{array}{cc} SW_1 - SW_{1,sat} & \text{if } SW_1 > SW_{1,sat} \\ 0 & \text{if } SW_1 \le SW_{1,sat} \end{array} \right\}$$
(35)

with $RO \ [mm]$ surface runoff, $SW_1 \ [mm]$ the water content in the first soil layer, and $SW_{1,sat} \ [mm]$ the saturated water content of the first soil layer.

2.7.4 Lateral flow

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Lateral flow is substantial in catchments with steep gradients and soils with high hydraulic conductivities (Beven, 1981; Beven and Germann, 1982; Sloan and Moore, 1984). In SPHY it is assumed that only the amount of water exceeding field capacity can be used for lateral flow. Therefore, first the drainable volume of water (excess water) needs to be calculated first:

$$W_{l,exc} = \left\{ \begin{array}{cc} SW_l - SW_{l,fc} & \text{if } SW_l > SW_{l,fc} \\ 0 & \text{if } SW_l \le SW_{l,fc} \end{array} \right\}$$
(36)

with $W_{l,exc}$ [mm] the drainable volume of water from soil layer l, SW_l [mm] the water content in soil layer l, and $SW_{l,fc}$ [mm] the field capacity of soil layer l. According to Sloan and Moore (1984), the lateral flow at the hillslope outlet can be calculated as:

$$LF_l^* = W_{l,excfrac} \cdot v_{lat,l} \tag{37}$$

with LF_l^* [mm] lateral flow from soil layer l, $W_{l,excfrac}$ [-] the drainable volume of water as fraction of the saturated volume, and $v_{lat,l}$ [mm $\cdot d^{-1}$] the flow velocity at the outlet. In SPHY, the drainable volume as fraction of the saturated volume is calculated as:

$$W_{l,excfrac} = \frac{W_{l,exc}}{SW_{l,sat} - SW_{l,fc}}$$
(38)

The velocity of flow at the outlet, $v_{lat,l} [mm \cdot d^{-1}]$, depends on both the saturated hydraulic conductivity $K_{sat,l} [mm \cdot d^{-1}]$ and the slope of the hill slp [-], and is defined as:

$$v_{lat,l} = K_{sat,l} \cdot slp \tag{39}$$

The slope (slp) in SPHY is calculated for each gridcell as the increase in elevation per unit distance.

According to Neitsch et al. (2009), only a fraction of lateral flow will reach the main channel on the day it is generated if the catchment of interest is large with has a time of 835 concentration is greater than 1 day. This concept is also im-

plemented in the SPHY model, and uses a lateral flow travel time $TT_{lag,l}$ [d] to lag a portion of lateral flow release to the channel:

$$LF_{l} = \left(LF_{l}^{*} + LF_{l,t-1}^{*}\right) \cdot \left(1 - \exp\left[\frac{-1}{TT_{lag,l}}\right]\right)$$
(40)

with LF_l [mm] the amount of lateral flow entering the channel on a given day, LF_l^* [mm] the lateral flow (Eq. (37)) generated within the cell on a given day, $LF_{l,t-1}^*$ [mm] the lateral flow lagged from the previous day. SPHY assumes the lateral flow travel time to be dependent on the field capacity $SW_{l,fc}$ [mm], saturated content $SW_{l,sat}$ [mm], and the saturated conductivity $K_{sat,l}$ [mm $\cdot d^{-1}$], according to:

$$TT_{lag,l} = \frac{SW_{l,sat} - SW_{l,fc}}{K_{sat,l}} \tag{41}$$

A larger lateral flow travel time will result in a smoother streamflow hydrograph.

2.7.5 Percolation

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⁸⁵⁵ If the groundwater module is used, then water can percolate from the first to the second soil layer, and from the second to the third soil layer. If the user decides to run SPHY without the groundwater module, percolation only occurs from the first to the second soil layer. In SPHY water can only
⁸²⁰ percolate if the water content exceeds the field capacity of that layer, and the water content of the underlying layer is ⁸⁶⁰ not saturated. A similar approach has been used in the SWAT model (Neitsch et al., 2009). The water volume available for percolation to the underlying layer is calculated as:

$$W_{l,exc} = \begin{cases} 0 & \text{if } SW_{l} \leq SW_{l,fc} \text{ or} \\ SW_{l+1} \geq SW_{l+1,sat} \\ SW_{l+1,sat} - SW_{l+1} & \text{if } SW_{l} - SW_{l,fc} > \\ SW_{l+1,sat} - SW_{l+1} & \text{else} \\ \end{cases} \begin{cases} 42 \end{cases}^{\text{gr}}$$

with $W_{l,exc}$ [mm] the drainable volume of water from layer l, SW_l [mm] the water content in layer $l, SW_{l,fc}$ [mm] field capacity of layer l, SW_{l+1} [mm] the water content in layer [l+1], and $SW_{l+1,sat}$ [mm] the saturated water content of layer [l+1]. Only a certain amount of $W_{l,exc}$ will percolate to the underlying soil layer, depending on the percolation travel time $TT_{perc,l}$ [d]. This approach follows the storage routing methodology, which is also implemented in the SWAT model (Neitsch et al., 2009):

$$w_{l,perc} = W_{l,exc} \cdot \left(1 - \exp\left[\frac{-1}{TT_{perc,l}}\right]\right) \tag{43}$$

with $w_{l,perc}$ [mm] the amount of water percolating to the underlying soil layer. The Since the speed in which water can move through the soil is mainly dependent on the saturated hydraulic conductivity (K_{sat}), the travel time for percolation is calculated in the same way as the travel time for lateral flow (Eq. (41)).

2.7.6 Groundwater recharge

Water that percolates from the second to the third soil layer will eventually reach the shallow aquifer. This process is referred to as groundwater recharge hereafter. If the glacier module is used as well, then also glacier melt that percolates contributes to the groundwater recharge. Groundwater recharge often does not occur instantaneousinstantaneously, but with a time lag that depends on the depth of the groundwater table and soil characteristics. SPHY uses the same exponential decay weighting function as proposed by Venetis (1969) and used by Sangrey et al. (1984) in a precipitation groundwater response model. This approach has also been adopted in the SWAT model (Neitsch et al., 2009), using:

$$Gchrg_t = \left(1 - \exp^{\frac{-1}{\delta_{gw}}}\right) \cdot w_{2,perc} + \exp^{\frac{-1}{\delta_{gw}}} \cdot Gchrg_{t-1}$$
(44)

with $Gchrg_t [mm]$ and $Gchrg_{t-1} [mm]$ the groundwater recharge on day days t and t-1, respectively. $\delta_{gw} [d]$ is the delay time and $w_{2,perc} [mm]$ is the amount of water that percolates from the second to the third layer on day t.

2.7.7 Baseflow

After groundwater recharge has been calculated, SPHY calculates baseflow which is defined as the flow going from the shallow aquifer to the main channel. Baseflow only occurs when the amount of water stored in the third soil layer exceeds a certain threshold (BF_{thresh}) that can be specified by the user. Baseflow calculation in SPHY is based on the steady-state response of groundwater flow to recharge (Hooghoudt, 1940) and the water table fluctuations that are a result of the non-steady response of groundwater flow to periodic groundwater recharge (Smedema and Rycroft, 1983). The SWAT model (Neitsch et al., 2009) assumes a linear relation between the variation in groundwater flow (baseflow) and the rate of change in water table height, according to:

$$\frac{dBF}{dt} = 10 \cdot \frac{K_{sat}}{\mu L_{gw}^2} \cdot (Gchrg - BF) = \alpha_{gw} \cdot (Gchrg - BF)$$
(45)

- with $BF \ [mm]$ the groundwater flow (baseflow) into the main channel on day t, $K_{sat} \ [mm \ d^{-1}]$ the hydraulic con-920 ductivity of the shallow aquifer, $\mu \ [-]$ the specific yield of the shallow aquifer, $L_{gw} \ [m]$ the distance from the subbasin divide for the groundwater system to the main channel,
- ⁸⁸⁰ Gchrg [mm] the amount of groundwater (Eq. (44)) recharge entering the shallow aquifer on day t, and α_{gw} [-] is the ⁹²⁵ baseflow recession coefficient. Equation 45 can be integrated and rearranged to calculate baseflow, according to:

$$BF_{t} = \left\{ \begin{array}{ll} 0 & \text{if } SW_{3} \leq BF_{thresh} \\ BF_{t-1} \cdot \exp^{-\alpha_{gw}} + & \text{if } SW_{3} > BF_{thresh} \\ Gchrg_{t} \cdot (1 - \exp^{-\alpha_{gw}}) & \end{array} \right\}^{\mathfrak{g}}$$

$$(46)$$

with BF_t [mm] the baseflow into the channel on day t, and BF_{t-1} [mm] the baseflow into the channel on day t-1. Since this equation has proven its succes in the SWAT model (Neitsch et al., 2009) throughout many applications worldwide, this equation has been adopted in the SPHY model as well.

The baseflow recession coefficient (α_{gw}) is an index that relates the baseflow response to changes in groundwater recharge. Lower values for α_{gw} therefore correspond to areas that respond slowly to groundwater recharge, whereas higher

values indicate areas that have a rapid response to groundwater recharge. The baseflow recession coefficient is generally used as a calibration parameter in the SPHY model, but a good first approximation of this coefficient can be calculated using the number of baseflow days (Neitsch et al., 2009):

$$\alpha_{gw} = \frac{2.3}{BFD} \tag{47}$$

with BFD [d] the number of baseflow days, which is de- 950 fined as the number of days required for baseflow recession to decline.

2.8 Streamflow routingRouting

- After calculating the different runoff components, the cell specific total runoff (QTot) is calculated by summarizing these different runoff components. Depending on the modules being switched on, the different runoff components are i) rain-rainfall runoff (RRo), ii) snow runoff (SRo), iii) glacier 960
- ⁹¹⁰ runoff (*GRo*), and iv) baseflow (*BF*). If the groundwater module is not used, then baseflow is calculated as being the lateral flow from the second soil layer. Rainfall runoff is the sum of surface runoff (*RO*, Sect. 2.7.3) and lateral flow from the first soil layer (*LF*₁, Sect. 2.7.4). Then If the groundwater ⁹⁶⁵
- ⁹¹⁵ module is not used, then baseflow is calculated as being the lateral flow from the second soil layer. QTot is eventually calculated according to:

$$QTot = RRo + SRo + GRo + BF \tag{48}$$

with $QTot \ [mm]$ the cell specific total runoff, $RRo \ [mm]$ rainfall runoff, $SRo \ [mm]$ snow runoff, $GRo \ [mm]$ glacier runoff, and $BF \ [mm]$ being baseflow from the third soil layer or latereral lateral flow from the second soil layer. In order to obtain river discharge, QTot needs to be routed through a flow direction network. SPHY allows the user to opt between the use of a simple routing scheme (Sect. 2.8.1), or a more complex routing scheme (Sect. 2.8.2) that involves the calculation of lake outflow through Q(h)-relations. Both methods require a flow direction network map, which can be obtained by delineating a river network using PCRaster or GIS software in combination with a Digital Elevation Model (DEM).

2.8.1 RoutingRunoff routing

In hydrology streamflow routing is referred to as being the transport of water through an open channel network. Since open-channel flow is unsteady, streamflow routing often involves the solving of solving complex partial differential equations. The St. Venant equations (Brutsaert, 1971; Morris and Woolhiser, 1980) are often used for this, but these have high data requirements related to the river geometry and morphology, which is are unavailable for the spatial scale SPHY is generally applied on. Additionally, solving these equations require requires the use of very small time steps, which result in large model calculation times. The use of very small time steps in the St. Venant equations is required to provide numerical stability. Other models, such as e.g. SWAT (Neitsch et al., 2009), use the Manning equation (Manning, 1989) to define the rate and velocity of river flow in combination with the variable storage (Williams, 1975) or Muskingum (Gill, 1978) routing methods to obtain river streamflow. But also the Manning equation requires river bed dimensions, which are generally unknown on the spatial scale that SPHY generally is applied on.

Therefore, SPHY calculates for each cell the accumulated amount of water that flows out of the cell into its neighbouring downstream cell. This can easily be obtained by using the accuflux PCRaster builtin functionbuilt-in function, which calculates for each cell the accumulated specific runoff from its upstream cells, including the specific runoff generated within the cell itself. If only the accuflux function would be is used, then it is assumed that all the specific runoff generated within the catchment on one day will end up at the most downstream location within one day, which is not plausible. Therefore, SPHY implements a flow recession coefficient (kx [-]) that accounts for flow delay, which can be a result of channel friction. Using this coefficient, river flow in SPHY is calculated using the three equations shown below:

$$QTot_t^* = \frac{QTot_t \cdot 0.001 \cdot A}{24 \cdot 3600} \tag{49}$$

$$Q_{accu,t} = accuflux \left(F_{dir}, QTot_t^*\right)$$
(50)

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$$Q_{rout,t} = (1 - kx) \cdot Q_{accu,t} + kx \cdot Q_{rout,t-1}$$
(51)

- with $QTot_t^*$ $[m^3s^{-1}]$ the specific runoff on day t, $QTot_t$ the specific runoff in mm on day t, A $[m^2]$ the grid cell area, $Q_{accu,t}$ $[m^3s^{-1}]$ the accumulated streamflow on day t with-¹⁰²⁵ out flow delay taken into account, $Q_{rout,t}$ $[m^3s^{-1}]$ the routed streamflow on day t, $Q_{rout,t-1}$ $[m^3s^{-1}]$ the routed streamflow on day t-1, F_{dir} the flow direction network, and kx [-]
- the flow recession coefficient. kx has values ranging between zero and one, where values close to zero correspond to a fast responding catchment, and values approacing approaching one correspond to a slow responding catchment.
- The user can opt to rout each of the four streamflow contributors separately, which may be useful if one wants to¹⁰³⁰ evaluate for example the contribution of glacier melt or snow melt to the total routed runoff. However, this increases model run-time substantially, because the accuflux function, which
- ⁹⁸⁵ is a time consuming function, needs to be called multiple times depending on the number of flow contributors to be routed.

2.8.2 Lake/reservoir routing

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Lakes or reservoirs act as a natural buffer, resulting in a delayed release of water from these water bodies. SPHY al-¹⁰⁴⁰ lows the user to choose a more complex routing scheme if lakes/reservoirs are located in their basin of interest. The use of this more advanced routing scheme requires a known relation between lake outflow and lake level height (Q(h)relation) or lake storage.

To use this routing scheme, SPHY requires a nominal map with cells identified as lakes each have the lake cells having a unique ID, while and the non-lake cells have having a value of zero. The user can supply a boolean map with True for cells that have measured lake levels, and False for lake cells that do not have measured lake levels. This specific application of SPHY is discussed in detail in Sect. 3.3.

Four different relations can be chosen to calculate the lake outflow from the lake level height or lake storage, being: i) an exponential relation, ii) a first-order polynomial function, iii) a second-order polynomial function, and iv) a third-order polynomial function. The user needs to supply maps containing the coefficients used in the different functions.

The lake/reservoir routing scheme simply keeps track of the actual lake storage, meaning that an initial lake storage should be supplied. Instead of the simple accuflux function described in the previous section, the lake/reservoir routing scheme uses the PCRaster functions accufractionstate and accufractionflux. The accufractionflux calculates for each1060 cell the amount of water that is transported out of the cell, while the accufractionstate calculates the amount of water that remains stored in the cell. For non-lake cells the fraction that is transported to the next cell is always equal to one, while the fraction that is transported out of a lake/reservoir cell depends on the actual lake storage. Each model timestep the lake storage is updated by inflow from upstream. Using this updated storage, the lake level and corresponding lake outflow can be calculated using one of the four relations mentioned before. The lake outflow can then be calculated as a fraction (Q_{frac} [-]) of the actual lake storage. Instead of using Eq. (50), the Q_{frac} is then used in Eq. (52) and Eq. (53) to calculate the accumulated streamflow and updated storage, respectively:

$$Q_{accu,t} = accufraction flux (F_{dir}, S_{act,t}, Q_{frac,t})$$
(52)

$$S_{act,t+1} = accufractionstate(F_{dir}, S_{act,t}, Q_{frac,t})$$
(53)

with $S_{act,t}$ [m^3] and $S_{act,t+1}$ [m^3] the actual storage and updated storage to be used in the next time-step, respectively, and $Q_{accu,t}$ [m^3d^{-1}] the accumulated streamflow on day t, without flow delay taken into account. Since Q_{frac} is always equal to one for the non-lake cells, the accufractionflux function becomes equal to the accuflux function used in the previous section. This actually means that for the river network the same routing function from Sect. 2.8.1 is used, and that Eq. (52) and Eq. (53) only apply to lake/reservoir cells.

In order to account for non-linearity and slower responding catchments, the same kx coefficient is used again. This involves applying Eq. (51) as a last step after Eq. (52) and converting the units from m^3d^{-1} to m^3s^{-1} . Since the accufractionflux and accufraction state functions are more complex to compute, the use of these functions increases model run-time.

3 Applications

The SPHY model has been applied and tested in various studiesranging from-, including real-time soil moisture predictions in lowlands, to operational reservoir inflow forecasting applications in mountainous catchments, to irrigation scenarios in the Nile Basin, to and detailed climate change impact studies in the snow-glacier-rain dominated Himalayan region. Some typical applications will be summarized in the following sections.

3.1 Irrigation management in lowland areas

As SPHY produces spatial outputs for the soil moisture content in the root zone and the potential and actual evapotranspiration (ET), it is a useful tool for application in agricultural water management decision support. By facilitating easy integration of remote sensing data, crop growth stages

can be spatially assessed at different moments in time. The SPHY dynamic vegetation module ensures that all relevant soil water fluxes correspond with crop development stages throughout the growing season. Spatially distributed maps¹¹²⁰ of root water content and ET deficit can be produced, enabling both the identification of locations where irrigation is required and a quantitative assessment of crop water stress.

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SPHY has been applied with the purpose of providing field-specific irrigation advice for a large-scale farm in west-1125 ern Romania, comprising 380 individual fields and approximately ten different crops. Contrary to the other case studies highlighted in this paper, a high spatial resolution is very relevant for supporting decisions on variable-rate irrigation.

- ¹⁰⁷⁵ The model has therefore been set up using a 30 m resolu-1130 tion, covering the 2013 and 2014 cropping seasons on a daily time step. Optical satellite data from Landsat 8 (USGS, 2013) were used as input to the dynamic vegetation module. Soil properties were derived from the Harmonized World Soil
- Database (Batjes et al., 2012), which for Romania contains data from the Soil Geographical Database for Europe (Lambert et al., 2003). Using the Van Genuchten equation (Van¹¹³⁵ Genuchten, 1980), soil saturated water content, field capacity, and wilting point were determined for the HWSD classes
 occurring at the study site. Elevation data was obtained from
- ¹⁰⁸⁵ occurring at the study site. Elevation data was obtained from the EU-DEM dataset (EEA, 2014), and air temperature was measured by two on-farm weather stations.

In irrigation management applications like these, a model should be capable of simulating the moisture stress expe-

rienced by the crop due to insufficient soil moisture contents, which manifests itself by an evapotranspiration deficit (potential ET – actual ET > 0). Field Figure 4 shows the spatial distribution of ET deficit, as simulated by the SPHY model for the entire farm on April 3rd, 2014. When SPHY is

run in an operational setting, this spatial information can be included in a decision support system that aids the farmer in irrigation planning for the coming days.

For calibration purposes, field measurements of soil mois-1150 ture and/or actual ET are therefore desired for calibration purposes desired. In this case study, a-one capacitance soil

moisture sensor was installed on a soybean field to monitor rootzone water content shortly after May 1, 2014, which is the start of the soybean growing season. The sensor mea-1155 sures volumetric moisture content for every 10 cm of the soil

¹¹⁰⁵ profile up to a depth 60 cm. It is also equipped with a rain gauge measuring the sum of rainfall and applied irrigation water, which was used as an input to SPHY. Soil moisture measured over the extent covered by the crop root depth was₁₁₆₀ averaged and compared to simulated values (Fig. 5).

Since this study was a demonstration project, only an initial model calibration was performed. The model was in this case most sensitive for the crop coefficient (Kc), affecting the evaporative demand for water. As can be seen 1165 in Fig. 5, the temporal patterns as measured by the soil mois-

ture sensor are well simulated by the SPHY model. Based on daily soil moisture values, a Nash-Sutcliffe (Nash and Sutcliffe, 1970) model efficiency coefficient of 0.6 was found, indicating that the quality of prediction of the uncalibrated SPHY model is "good" (Foglia et al., 2009).

The performance of the SPHY model can be improved by calibration using remotely sensed evapotranspiration Soil moisture simulations could be further improved by conducting a full model calibration, adjusting the soil physical parameters $K_{sat,1}$, $SW_{1,fc}$, $SW_{1,pF3}$, and $SW_{1,pF4,2}$. Remotely sensed sensed evapotranspiration can be used in the calibration process (Immerzeel and Droogers, 2008), although such data is often not available on these small scales as ET is a very complex parameter to assess (Samain et al., 2012). It should also be noted that soil moisture content is typically highly variable in space; a very high correlation between point measurements and grid cell simulations of soil moisture may therefore not always be feasible (Bramer et al., 2013).

3.2 Snow and glacier-fed river basins

SPHY is being used in large Asian river basins with significant contribution of glacier- and snow melt to the total flow (Immerzeel et al., 2012; Lutz et al., 2012, 2014b). The major goals of these applications are two-fold:

- Assess the current hydrological regimes at highresolution; e.g. assess spatial differences in the contributions of glacier melt, snow melt and rainfall-runoff to the total flow
- Quantify the effects of climate change on the hydrological regimes in the future and how these affect the water availability

Rivers originating in the high mountains of Asia are considered to be the most melt-water dependent river systems on Earth (Schaner et al., 2012). In the regions surrounding the Himalayas and the Tibetan Plateau large human populations depend on the water supplied by these rivers (Immerzeel et al., 2010). However, the dependency on melt water differs strongly between river basins as a result of differences in climate and differences in basin hypsometry (Immerzeel and Bierkens, 2012). Only by using a distributed hydrological modelling approach that includes the simulation of key hydrological and cryospheric processes, and inclusion of transient changes in climate, snow cover, glaciers and runoff, appropriate adaptation and mitigation options can be developed for this region (Sorg et al., 2012). The SPHY model is very suitable for such an approach goals, and has therefore been widely applied in the region.

For application in this region, SPHY was setup at a 1 km spatial resolution using a daily time-step, and forced with historical air temperature (Tavg, Tmax, Tmin) and precipitation data, obtained from global and regional datasets (e.g. APHRODITE (Yatagai et al., 2012), Princeton (Sheffield et al., 2006), TRMM (Gopalan et al., 2010)) or interpolated WMO station data from a historical reference period.

For this historical reference period SPHY was calibrated and

- validated using observed streamflow. For the future period, SPHY was forced with downscaled climate change projections obtained from General Circulation Models (GCMs),1225 as available through the Climate Model Intercomparison Projects (e.g. CMIP3 (Meehl et al., 2007), CMIP5 (Taylor
- et al., 2012)), and which were used as basis for the Assessment Reports prepared by the Intergovernmental Panel on Climate Change (IPCC).

In Central Asia, SPHY was successfully applied in a study (ADB, 2012; Immerzeel et al., 2012; Lutz et al., 2012) that

- focused on the impacts of climate change on water resources in the Amu Darya and Syr Darya river basins. SPHY was used to quantify the hydrological regimes in both basins, and 1235 subsequently to project the outflow from the upstream basins to the downstream areas by forcing the model with an ensem-
- ¹¹⁸⁵ ble of 5 CMIP3 GCMs. The SPHY model output fed into a water allocation model that was setup for the downstream parts of the Amu Darya and Syr Darya river basins. ¹²⁴⁰

In the Himalayan Climate Change Adaptation Programme (HICAP), led by the International Centre for Integrated

- Mountain Development (ICIMOD), SPHY has been successfully applied in the upstream basins of the Indus, Ganges, Brahmaputra, Salween and Mekong rivers (Lutz et al., 2013,1245 2014b). In this study the hydrological regimes of these five basins have been quantified and the calibrated and validated
- ¹¹⁹⁵ model (Fig. 6) was forced with an ensemble of eight GCMs to create water availability scenarios until 2050. Table 4 lists the calibration and validation results. Based on the validation¹²⁵⁰ results, we concluded that the model performs satisfactory given the large scale, complexity and heterogeneity of the
- modeled region and data scarcity (Lutz et al., 2014b).
 We use one parameter set for the entire domain, which inherently means some stations perform better than others. In the particular case of the upper Indus, another possible explanation could be uncertainty in air temperature forcing in1255
- the highest parts of the upper Indus basin (locations Dainyor bridge, Besham Qila and Tarbela inflow in Tab. 4), since especially in this area, the used forcing datasets are based on very sparse observations. SPHY allowed the assessment of current and future contribution of glacier melt and snow₁₂₆₀ melt to total flow (Fig. 7), and how total flow volumes and
- the intra-annual distribution of river flow will change in the future.

3.3 Flow forecasting

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In harsh environments like Chile, hydropower companies have been using forecasting solutions such as statistical regressions of flowobservations, or mathematical simulations with a minimum content of real-time physical information.1270 These solutions are generally based on long historical data sets of flowobservations, and do not take into account the current hydrological state of the basin and For basins

the current hydrological state of the basin and For basins with snow melt being an important contributor to the

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flow, besides calibration to observed flow, the elimate for snow-related parameters in the SPHY model can also be calibrated to observed snow cover. For the Upper Indus basin the coming weeks/months. During the INTOGENER project, the snow-related parameters degree-day factor for snow (DDF_s) and snow water storage capacity (SSC)were calibrated independently using MODIS snow cover imagery (Lutz et al., 2014a). The same MODIS dataset was used as in Immerzeel et al. (2009). From the beginning of 2000 until halfway 2008, the snow cover imagery was averaged for 46 different periods of 8 days (5 days for the last period) to generate 46 different average snow cover maps. E.g. period 1 is the average snow cover for 1-8 January for 2000 until 2008, whereas period 2 is the average snow cover for 9-16 January for 2000 until 2008, etc. The SPHY model was used to demonstrate the operational capabilities of a water flow monitoring and prediction system aimed at hydropower production and water management organizationsrun for 2000-2007 at a daily time step and for each 1x1 km grid cell the average snow cover was calculated for the same 46 periods as in the MODIS observed snow cover dataset. Subsequently, these simulated snow cover maps were resampled to 0.05° spatial resolution, which is the native resolution of the MODIS product. Figure 8 shows the basin-average observed and simulated fractional snow cover for the 46 periods during 2000-2008 and Fig. 9 shows the same at the 0.05° grid cell level. As a final step, the baseflow recession coefficient (α_{qw}) and routing coefficient (kx) were calibrated to match the simulated streamflow with the observed streamflow.

The INTOGNER systemuses-

3.3 Flow forecasting

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In data-scarce environments and inaccessible mountainous terrain, like in the Chilean Andes, it is often difficult to install instrumentation and retrieve real-time physical data from these instruments. This real-time data can be useful to capture the hydro climatic variability in this region, and improve the forecasting capability of hydrological models. Although statistical models can provide skillful seasonal forecasts, using large scale climate variables and in-situ data (Piechota and Chiew, 1998; Grantz et al., 2005; Regonda et al., 2006; Bracken et al., 2010), a particular hydropower company in Chile was mainly interested in the potential use of an integrated system, using measurements derived from both Earth Observation (EO) satellites and in-situ sensors-Data that were, to force a hydrological model to forecast seasonal streamflow during the snow melting season. The objective of the INTOGENER project was therefore to demonstrate the operational forecasting capability of the SPHY model in data-scarce environments with large hydro climatic variability.

During INTOGENER, data retrieved from EO satellites consisted of a DEM map and a map-series and a time-series

- of snow cover maps. Snow cover images were retrieved on a_{1330} weekly basis, using RADARSAT and MODIS (Parajka and Blöschl, 2008; Hall et al., 2002) imagery. These images were used to update the snow store ($SS \ [mm]$) in the model in order to initialize the model for the forecasting period. Figure
- 1280 10 shows the snow store as simulated by the SPHY model+335 during the snow melting season in the Laja Basin. These maps clearly show the capability of SPHY to simulate the spatial variation of snow storage, with more snow on the higher elevations, and a decrease in snow store throughout
- the melting season. Discharge, precipitation and temperature data were collected using in-situ meteorological stations. In order to calculate the lake outflow accurately, the SPHY model was initialized with water level measurements retrieved from reflected Global Navigation Satellite System
- (GNSS) signals in Laja Lake. Static data that was used in the1345
 SPHY model consisted of soil characteristics derived from the Harmonized World Soil Database (HWSD) (Batjes et al., 2009) and land use data obtained from the GLOBCOVER (Bontemps et al., 2011) product. The SPHY model was setup
 to run at a spatial resolution of 200 m.
- Figure 11 shows the observed vs. simulated daily streamflow for two locations within the Laja River Basin for the historical period 2007-2008. It can be seen that model performance is quite satisfactory for both locations, with volume
- errors of -4% and -9.4% for Canal Abanico (downstream of Laja Lake) and Rio Laja en Tucapel, respectively. The NScoefficient, which is especially useful to assess the simulation of high discharge peaks, is less satisfactory for these lo-1355 cations. Hydropower companies, however, have more inter-
- est in expected flow volumes for the coming weeks/months than for accurate day-to-day flow simulations, and therefore the NS-coefficient is less important in this case. If the NScoefficient is calculated for the same period on a monthly ba-1360 sis, then the NS-coefficients are 0.53 for Canal Abanico and
- ¹³¹⁰ 0.81 for Rio Laja en Tucapel. It is likely that SPHY model performance would even be better if a full model calibration would have been performed.

The hydropower company's main interest is the model's₁₃₆₅ capacity to predict the total expected flow for the coming

- weeks during the melting season (October 2013 through March 2014). To forecast streamflow during the snow melting season, the SPHY model was forced with gridded temperature and precipitation data from the European¹³⁷⁰ Centre for Medium-range Weather Forecasts (ECMWF)
- HRES (High-Resolution) Model (Andersson, 2013), and the Seasonal Forecasting System (SEAS) (Andersson, 2013). The HRES model provided short-term (7-day) meteorological forecasts on a spatial resolution of 0.125°, 1375 whereas the SEAS model provided daily forecasts on a
- spatial resolution of 0.75°, 7 months ahead, and was used to forecast streamflow up till the end of the melting season. Figure 12 shows the bias between the total cumulative forecasted flow and observed flow for the 23 model runs that₁₃₈₀ were executed during operational mode. Although there are

some bias fluctuations in the Rio Laja en Tucapel model runs, it can be concluded that the bias decreases for each next model run for both locations, which is a logical result of a decreasing climate forcing uncertainty as the model progresses in time. It can be seen that the SPHY model streamflow forecasts for Canal Abanico, which is downstream of Laja Lake, are substantially better than for Rio Laja en Tucapel (most downstream location). The reason for this has not been investigated during the demonstration study, but since model performance for these two locations was satisfactory during calibration, a plausible explanation could be the larger climate forecast uncertainty in the higher altitude areas (Hijmans et al., 2005; Rollenbeck and Bendix, 2011; Vicuña et al., 2011; McPhee et al., 2010; Mendoza et al., 2012; Ragettli and Pellicciotti, 2012; Ragettli et al., 2014) in the northeastern part of the basin that contributes to the streamflow of Rio Laja en Tucapel. Additionally, only two in-situ meteorological stations were available during operational mode, whereas during calibration 20+ meteorological stations were available. Moreover, these operational meteorological stations were not installed on higher altitudes, where precipitation patterns tend to be spatially very variable (Wagner et al., 2012; Rollenbeck and Bendix, 2011).

4 Future outlook

Further development and refinement of the SPHY model are foreseen in six areas, being i) seven areas, including i) the implementation of time steps smaller than 1 day, ii) evapotranspiration processes, iiiii) depression storage, iiiiv) representation of lakes/reservoirs, ivv) streamflow routing, vvi) cryospheric processes, and vivi) the development of a graphical user-interface.

Currently, the SPHY model can only run at a daily time step. The implementation of hourly model time steps is foreseen for future SPHY versions, which allows other processes, like e.g. Hortonian runoff (Beven, 2004; Corradini et al., 1998) to be included as well. Using hourly instead of daily climate forcings will improve other processes as well, such as snow and glacier melt.

The current version of the SPHY model calculates the reference evapotranspiration using the modified Hargreaves equation (Hargreaves and Samani, 1985), which requires three meteorological variables: the average, maximum and minimum daily air temperature. Although this method is less data demanding than the well-known Penman–Monteith (Allen et al., 1998) equation, additional methods for the calculation of the reference evapotranspiration that are even less data demanding should be explored for future versions of the SPHY model.

Besides the dynamic simulation of a time-varying fractional vegetation coverage, using NDVI time-series as input, it is currently assumed that the root depth of a growing crop is constant throughout the simulation period, and that it is

equal to the depth of the first soil layer. In reality a crop is seeded, starts to develop its root system, and finally gets harvested. Although this root development process is less dynamically dynamical for forests and therefore more relevant for agricultural crops, it can be seen as an improvement¹⁴⁴⁰ to be included in the SPHY model. A similar approach as in the SWAP model (van Dam et al., 1997) could be used, where the user needs to define the root depths for each grow-

ing stage. However, it should be taken into account that the SWAP model is a field-scale model for which crop-specific1445 root depth information is generally available, and that SPHY is generally applied on larger spatial scales where cropping calendars, and thus corresponding root depth information is harder to obtain.

Surface runoff in SPHY occurs whenever excess rainfall₁₄₅₀ results in saturation of the first soil layer. Subsequently, this excess rainfall leaves the grid-cell as surface runoff without any obstacles or friction, and enters the river system within the same day. In reality, surface runoff can be delayed be-

- the same day. In reality, surface runoff can be delayed because of surface friction, or stored in local depressions or₁₄₅₅ man-made ponds (surface irrigation), before it is released to the river system. The concept of having a ponding layer is mainly beneficial for i) agricultural applications where surface irrigation in combination with a ponding layer plays a
- face irrigation in combination with a ponding layer plays a major role, e.g. rice irrigation, and/or ii) if the model is run1460 on very high spatial resolution where the role of local depressions becomes more prominent; water can be stored for hours/days in these depressions, and water may be partly or
- ¹⁴¹⁰ completely evaporated before it has the chance to flow to the channels. The effect of delayed surface runoff can be re-1465 lated to dense vegetation or agricultural land management practices (Hunink et al., 2013; Kauffman et al., 2014), with bench terraces and contour tillage being examples of these practices. The effect of delayed or stored surface runoff is

currently not implemented in the <u>SPHY</u> model, but should₁₄₇₀ be further explored for future <u>SPHY</u> versions.

As was mentioned in Sect. 2.8.2, each individual lake is represented by maximum one grid-cell in SPHY. This is related to the PCRaster flow direction calculation: each cell needs to flow to a downstream cell, and the total flow in₁₄₇₅ the most downstream cell should be equal to the accumulated flow of all upstream cells plus the flow generated within that cell. This means that large lakes, consisting of multiple

- neighboring cells, can be seen as an enormous pit in the flow direction network, meaning that streamflow routing from the₁₄₈₀ upstream lake cells is interrupted at these pit cells. Since large lakes can evaporate a substantial amount of water (Gat et al., 1994), a lake-cell merging procedure should be devel-
- oped for future versions of the SPHY model, that allows to automatically re-create the flow direction network based on₁₄₈₅ the locations of the lake inflow cells and outflow cell, such that all the accumulated water of the inflow cells is allocated and stored in the downstream cell, where it can continue its
 way downstream.

Although the currently implemented routing scheme in SPHY has proven its success during various applications (see Sect. 3.2 and Sect. 3.3), improvements in the routing scheme are foreseen for future SPHY versions. Using the current approach it is assumed that the open channel surface area equals the grid cell area. If coarse model resolutions are used, then this assumption becomes less plausible, because in reality the surface area of a river would only cover a small fraction of this grid cell area. Therefore, it would be interesting to explore using more advanced routing schemes that take into account the river's cross section dimensions and corresponding velocity flow rates. An example of a more advanced routing scheme, that takes into account the channel dimensions, and has been found to yield good results under a wide range of conditions in natural rivers (Brutsaert, 2005) is the Muskingum method (Gill, 1978). Considering the larger spatial resolutions that SPHY is generally applied for, it seems almost infeasible to determine the channel dimensions. However, some emperical empirical relations between the bankfull dischargen discharge and channel dimensions are available (Park, 1977), and should be explored for implementing in SPHY.

Sublimation of snow is an important component of the water balance in areas experiencing snow cover during significant time of the year (Strasser et al., 2008). Different approaches have been developed to estimate this flux in hydrological models (Bowling et al., 2004; MacDonald et al., 2009; Lenaerts et al., 2010; Groot Zwaaftink et al., 2013). Development of a parameterization of sublimation using the low data requirement of SPHY is foreseen.

In SPHY, glaciers are considered as entities generating melt using a temperature-index model. Improvements in the representation of glaciers are foreseen to make them massconserving, e.g. considering the precipitation falling in their accumulation areas. In the current SPHY version, all surface runoff that is generated on glacier covered areas is considered as glacier melt, and inclusion of further specification to seasonal snow melt and glacier ice melt is foreseen. Besides, the temperature-index model can be improved by including the incoming radiation in addition to air temperature in the temperature-index model (Hock, 2003; Pellicciotti et al., 2005; Heynen et al., 2013). Additionally, the glacier module will be extended with a parameterization for modeling glacier dynamics, which would enable quantifying the retreat or advance of glaciers as a result of climate perturbations.

When SPHY is run at a spatial resolution of 1 km or coarser for mountainous regions, improvement of the respresentation representation of sub-grid processes becomes useful. For example, when the daily average air temperature for a given 1 km grid cell is 0 °C, no melt will be simulated for that grid cell, although the subgrid variability in elevation shows that part of the grid cell is lower than the average elevation of the grid cell and thus has temperatures above 0 °C, resulting in the generation of melt water.

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Although the model is relatively easy to understand and applicable by hydrologists and scientists with basic skills, it would even be better if the model could be used by a wider₁₅₄₅ group of users with basic hydrological and computer skills. Therefore, the objective of an ongoing project is to develop a Graphical User Interface (GUI) for the SPHY model. This

GUI will be developed as plugin for the open-source QGIS environment, making it easy setup the model and analyse₁₅₅₀ model input and output spatially and temporally.

1500 5 Conclusions

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In response to the diversity of water-related challenges, hydrologists have been developing tools to analyze, understand and explore solutions to support decision makers and operational water managers. Although progress has been made, hydrologists have put a strong emphasis on streamflow1560 analysis and forecasting, while ignoring other hydrological processes. It is clear that not only rainfall-runoff processes are relevant, but that an integrated approach where processes as glacier and snowmelt, evapotranspiration, reservoirs, changes in landcover and landuse should be integrated in our1565 models.

The objective of this paper is to introduce and present the SPHY model, its development background, and demonstrate some typical applications. Compared to other hydrological

- models, that typically focus on the simulation of streamflow1570 only, the SPHY model has several advantages: it SPHY has been developed with the explicit aim to simulate terrestrial hydrology under various physiographical and hydro-climatic conditions by integrating key components from existing
- and well-tested models. SPHY i) integrates most relevant₁₅₇₅ hydrological hydrologic processes, ii) is setup modular, iii) is easy adjustable and applicable, iii) can easily be linked to remote sensing datahas the flexibility to study a wide range, and iii) scale of applications, and iv) can be applied
- ¹⁵²⁵ for operational as well as strategic decision support easily be 1580 implemented.

The most relevant hydrological processes that are integrated in the SPHY model are rainfall-runoff processes, cryosphere processes, evapotranspiration processes, the sim-

- ¹⁵³⁰ ulation of dynamic vegetational cover, lake/reservoir outflow, and the simulation of rootzone moisture contents. The capability of SPHY to successfully simulate rainfall-runoff and cryosphere processes was proven during its applications in the snow and glacier-fed river basins in Asia, and
- ¹⁵³⁵ in Chile, where it was used to forecast streamflow during the snow melting season. Both the applications in Chile and in Romania show the <u>strength easy implentation</u> of SPHY to include <u>remotely sensed remote sensing</u> data as¹⁵⁹⁰ dynamic model input: in Chile remotely sensed snow cover
- ¹⁵⁴⁰ was used to implement a time-variable fractional snow coverage, and in Romania the NDVI was used to simulate dynamic vegetational cover. The glacier-fed river basins in

Asia application also showed the potential use of remote sensing data, where remotely sensed snow cover was used to calibrate the simulated snow cover by the SPHY model. Whereas the application in the glacier-fed river basins in Asia to study the effects of climate change on the available water resources is a typical example of the use of SPHY to support strategic decision makers decision-making, the application in Romania for real-time irrigation support, and the application in Chile to support hydropower companies for their reservoir management, are typical examples of the use of SPHY for operational purposes. The application of real-time irrigation support in Romania emphasizes the strength of SPHY to succesfully simulate other hydrological processes than rainfall-runoff as well, being evapotranspiration and rootzone soil moisture simulations. The different spatial resolutions and orograhically and climatologically physiographical and hydro-climatic regions in which SPHY was applied , demonstrate its flexibility in scalingand applicability: in Romania it was applied at the farm-field levelfarm-level, requiring a spatial resolution of 30 m, whereas in Chile and Asia it was applied at the river basin scale at resolutions varying from 200 m to 1 km.

The modular setup The modular framework of SPHY enables it to be applied over a broad range of orographically and elimatologically oreographical and climatological conditions as demonstrated in this paper: the agricultural focus in Romania allowed switching off the glaciers, snow, and routing modules, whereas these modules were switched on for the Asian and Chilean applications. Decreased model run-time, and minimzed minimized input data requirements were the resulting benefits in cases where modules are switched off.

In summary , it can be concluded that the SPHY model is a suitable tool to be used for strategic decision SPHY is a model that can easy be implemented for strategic decision-making as well as for operational support, having the flexibility to be applied at different spatial scales, and including the use of remote sensing data and hydrological processes integrate the hydrological processes that are relevant for the area of application, and have the flexibility to be applicable under a wide range and scale of applications.

6 Code availability

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The SPHY model is available as executable (sphy.exe) and as source code, where the source code consists of the following files:

- sphy_config.cfg
- sphy.py
- hargreaves.py
- dynamic_veg.py
- snow.py
- rootzone.py
- ET.py
- subzone.py

- 1595 glacier.py
 - groundwater.py
 - routing.py
 - reservoir_routing.py
 - reporting.py
- 1600 timecalc.py

The order in which the model algorithms are executed are defined in the sphy.py file, in which the required modules are imported depending on the settings in sphy_config.cfg. Model settings (parameters, input and output) can be mod-₁₆₅₅ ified in the sphy_config.cfg configuration file. It is manda-

- ified in the sphy_config.cfg configuration file. It is mandatory to have this file and the source code (or model executable) in the same folder on the pc's harddrive. If the user opts to run the SPHY model using the model's source code, then the SPHY model is executed by entering "python.exe¹⁶⁶⁰
- ¹⁶¹⁰ sphy.py" in the windows command prompt. Otherwise, the model can be executed by entering "sphy.exe" in the windows command prompt. Both the model source code and it's executable can be obtained from the SPHY model website (http://www.sphy.nl).
- ¹⁶¹⁵ In order to run the SPHY model v2.0, it is required to have the following software installed on your pc:

-	Python 2.7.6 (32 bit)
	N D 100(201:0

- NumPy 1.8.0 (32 bit)
- PCRaster 4.0 (32 bit)

1620 Appendix A: SPHY model parameters

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Figure 1. Illustration of SPHY sub-grid variability. A grid-cell in SPHY can be a) partially covered with glaciers, or b) completely covered with glaciers, or c1) free of snow, or c2) completely covered with snow. In case of c1, the free land surface can consist of bare soil, vegetation, or open water.

 Table 1. Overview of SPHY model parameters. The last column indicates if this parameter is easy (+++), moderate (++), or difficult (+) to observe.

Acronym	Description	Units	Observable
Kc	Crop coefficient	-	++
Kcmax	Maximum crop coefficient	~	<u>++</u>
Kc_{min}	Minimum crop coefficient	-	++
NDVImax	Maximum NDVI	-	+++
NDVImin	Minimum NDVI	-	+++
FPARmax	Maximum Fraction of Absorbed Photosynthetically Active Radiation	- 	*** ~
$\underline{FPAR_{min}}$	Minimum Fraction of Absorbed Photosynthetically Active Radiation	.≂	
$\underline{T_{crit}}$	Temperature threshold for precipitation to fall as snow	°C	* ~
DDF_s	Degree Day Factor for snow	$\mathrm{mm}^{\circ}\mathrm{C}^{-1}\mathrm{d}^{-1}$	+
\underline{SSC}_{\sim}	Water storage capacity of snowpack	$\underline{\mathrm{mm}}\mathrm{mm}^{-1}$	+
\underline{GlacF}	Glacier fraction of grid-cell	.≂	++++
$\underline{DDF_{CI}}$	Degree Day Factor for debris free glaciers	$\underline{mm}^{\circ}\underline{C}^{-1}\underline{d}^{-1}$	+
$\underline{DDF}_{\underline{DC}}$	Degree Day Factor for debris covered glaciers	$\underline{mm}^{\circ}\mathbf{C}^{-1}\mathbf{d}^{-1}$	+
F_{CI}	Fraction of GlacF that is debris free	$\overline{\sim}$. <u>+++</u> +
F_{DC}	Fraction of GlacF that is covered with debris	$\ddot{\sim}$	***
$\underline{GlacROF}$	Fraction of glacier melt that becomes glacier runoff	$\overline{\sim}$	*
$\underbrace{SW_{1,sat}}$	Saturated soil water content of first soil layer	mm	+++
$\underline{SW}_{1,fe}$	Field capacity of first soil layer	mm	****
$SW_{1,pE3}$	Wilting point of first soil layer	mm	****
$\underline{SW}_{1,pF4,2}$	Permanent wilting point of first soil layer	mm	****
$\underbrace{K_{sat,1}}_{\sim}$	Saturated hydraulic conductivity of first soil layer	$\operatorname{mm} d^{-1}$	+++
$\widetilde{SW}_{2.sat}$	Saturated soil water content of second soil layer	mm	+++
$\widetilde{SW}_{2.fc}$	Field capacity of second soil layer	mm	+++
$K_{sat,2}$	Saturated hydraulic conductivity of second soil layer	$\operatorname{mm} d^{-1}$	****
$\underbrace{SW_{3,sat}}$	Saturated soil water content of groundwater layer	mm	+
\widetilde{slp}	Slope of grid-cell	$\underline{m} \underline{m}^{-1}$	+++
Saw	Groundwater recharge delay time	d.	+ ~
$\widetilde{\alpha_{gw}}$	Baseflow recession coefficient	$\overset{\mathrm{d}^{-1}}{\longrightarrow}$	*++
$\underbrace{BF_{tresh}}$	Threshold for baseflow to occur	mm	+
$\underbrace{kx}{\underbrace{kx}}$	Flow recession coefficient	\sim	+

	SPHY	TOPKAPI- ETH	SWAT	VIC	LIS- FLOOD	SWIM	НҮРЕ	mHM	MIKE- SHE	PCRGLOB- WB	GEO- top
Processes integrated											
Rainfall-runoff	+	+	+	+	+	+	+	+	+	+	+
Evapotranspiration	+	+	+	+	+	+	+	+	+	+	+
Dynamic vegetation	+	-	+	+	+	+	+	NA	+	+	-
growth											
Unsaturated zone	+	+	+	+	+	+	+	+	+	+	+
Groundwater	+	-	+	+	+	+	+	+	+	+	+
Glaciers	+	+	-	-	-	+	+	-	-	-	+
Snow	+	+	+	+	+	+	+	+	+	+	+
Routing	+	+	+	+	+	+	+	+	+	+	+
Lakes incorporated	+	-	+	+	+	+	+	NA	+	+	-
in routing scheme											
Reservoir management	-	-	+	-	-	+	+	NA	-	+	-
Field of annihoodien											
Climate change impacts											
L and use change impacts	+	+	+	+	+	+	+	+	+	+	+
Land use change impacts	- -	+	- -	+	+	+	+	Ŧ	+	+	+
Flasses	+	-	+	+	*3	+	+	-	+	-	+
Floods	-	-	-	-		-	-	-	+	+	+
Water supply and demand	+	+	+	+	+	+	+	+ NA	+	+	+
water suppry and demand	-	-	+	-	-	-	-		-	-	-
Scale of application											
Catchment-scale	+	+	+	+	-	-	-	-	+	-	+
River basin scale	+	+	+	+	+	+	+	+	+	-	-
Mesoscale river basins	+	-	+	+	+	+	+	+	+	+	-
Global-scale	-	-	-	+	-	-	+	-	-	+	-
Farm-level	+	-	-	-	-	-	-	-	-	-	-
Country-level	+	-	-	-	-	-	-	-	-	-	-
Fully distributed	+	+	-	+	+	-	-	+	+	+	+
Sub-grid variability	+	-	-	+	-	-	-	+	-	+	+
Felxible spatial	+	+	-	+	+	-	-	+	+	+	+
resolution											
Hourly resolution	-	+	+	-	+	-	-	+	+	-	+
Sub-daily resolution	-	-	-	+	+	-	-	NA	+	-	-
Daily resolution	+	+	+	+	+	+	+	NA	+	+	-
Implementation											
Open-source	+	-	+	+	-	-	+	-	-	-	+
Forcing with	+	+	-	+	+	-	-	NA	-	-	+
remote sensing	•			•							·
GIS compatibility	+	+	+	_	+	+		+	+	+	+
Modular setup	+	-	-	+	+	+	+	+	+	-	
Computational	+	+	+	-	+	+	+	+	-	+	+
efficient	·		·					•			
Climate forcing	+	+	_	*2	_	_	+	+	_	_	_
requirements	'						,				
Flexible output	+	+	_	+	+	+	+	NA	+	_	+
renoring options	'					i.	,				'
Graphical user-	*1	_	+	_	_	+	_	_	+	_	_
interface in GIS									•		

Table 2. Pros (+) and cons (-) of some well-known hydrological models, including the SPHY model. A categorization is made between i) processes that are integrated, ii) field of application, iii) scale of application, and iv) implementation.

*1 Currently in development.
 *2 More climate variables are required if model is run in energy balance mode.
 *3 Only if run in combination with LISFLOOD-FP.
 NA Information not available.



Figure 2. SPHY model modeling concepts. The fluxes in grey are only incorporated when the groundwater module is not used. Abbreveations are explained in the text.

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Table 3. LAI_{max} values for different vegetation types (Sellers et al., 1996).

Vegetation type Vegetation type	$\underline{LAI_{max}[-]} \underbrace{LAI_{max}[-]}$
Broadleaf evergreen trees	7
Broadleaf deciduous trees	7
Mixed trees	7.5
Needleleaf evergreen trees	8
High latitude deciduous trees	8
Grass with 10 - 40% woody cover	5
Grass with <10% woody cover	5
Shrubs and bare soil	5
Moss and lichens	5
Bare	5
Cultivated	6

Table 4. Station locations used for calibration and validation of the SPHY model in HICAP (Lutz et al., 2014b). Three stations were used for calibration for 1998-2007. Five stations were used for an independent validation for the same period. The Nash-Sutcliffe efficiency (NS) and bias metrics were calculated at a monthly time step.

Location	<u>NS</u> [-]	Bias [%]	Validation/Calibration
Dainyor bridge	0.39	58.2	validation
Besham Qila	0.66	24.7	validation
Tarbela Inflow	0.63	34.6	calibration
Marala Inflow	0.65	12.0	validation
Pachuwarghat	0.90	-1.6	validation
Rabuwa Bazar	0.65	-22.5	validation
Turkeghat	0.87	-5.4	calibration
Chatara	0.87	7.9	calibration



Figure 3. Modules of the SPHY model that can be switched on/off.



Figure 4. Spatial distribution of evapotranspiration (ET) deficit, as simulated by the SPHY model for a Romanian farm on April 3rd, 2014. Transparency means no ET deficit.



Figure 5. Measured and simulated daily rootzone soil moisture content during the 2014 growing season. Rainfall + irrigation has been measured by the rain gauge that was attached to the moisture sensor.



Figure 6. Average monthly observed and SPHY-simulated flow (1998-2007) for the Chatara major discharge measurment measurement location in the Ganges basin (Lutz et al., 2014b). Metrics are calculated based on monthly time steps.



Figure 8. Laja River Basin in Chile with Laja Lake Observed and simulated average fractional snow cover in the streamflow stations Canal Abanico upper Indus basin. The values represent the 9-year average for 46 (ID 198-day) and Rio Laja en Tucapel (ID 23)periods during 2000-2008.



Figure 7. The contribution of glacier melt (a), snow melt (b), and rainfall (c) to the total flow for major streams in the upstream basins of the Indus, Ganges, Brahmaputra, Salween and Mekong during 1998-2007 (Lutz et al., 2014b).

Figure 9. a) SPHY simulated snow cover 2000-2007 and b) MODIS observed snow cover 2000-2007.



Figure 10. Snow storage [mm] as simulated by the SPHY model on the 12th of August (left) and the 1st of October (right) during the snow melting season of 2013 in the Laja River Basin.





Figure 11. Daily observed vs. SPHY simulated streamflow (period 2007-2008) for the streamflow stations Canal Abanicao (ID 19) and Rio Laja en Tucapel (ID 23). The Nash-Sutcliffe (NS) and bias model performance indicators are shown as well.

Figure 12. Bias between total cumulative forecasted flow and observed flow for the 23 model runs that were executed between the end of September 2013 and March 2014. Results are shown for the locations Canal Abanicao (ID 19) and Rio Laja en Tucapel (ID 23).