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# Comparing the ISBA and J2000 approaches for surface flows modelling at the local scale in the Everest region

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## Abstract

This paper compares the hydrological responses at the local scale of two models using different degrees of refinement to represent physical processes in sparsely instrumented mountainous Himalayan catchments. This work presents the novelty of applying, at a small spatio-temporal scale and under the same forcing conditions, a fully distributed surface scheme based on mass and energy balance equations (ISBA surface scheme), and a semi-distributed calibrated model (J2000 hydrological model). A new conceptual module coupled to the ISBA surface scheme for flow routing is presented. Two small catchments located in mid- and high- mountain environments were chosen to represent the very different climatic and physiographic characteristics of the Central Himalayas in the Everest region of eastern Nepal. The results show that both models globally represent the dynamic of the processes for evaporation, quick runoff and discharge in a similar way. The differences in the model structures and results mainly concern the snow processes and the soil processes. In particular for the high-mountain catchment, the snow-pack simulation is shown to be the main driver of the discrepancy between the two models. The sub-daily variations of snow processes are shown to significantly influence the estimation of the snow-melt contribution to discharge.

*Keywords:* Central Himalayas, ISBA surface scheme, J2000 model, water budget at the local scale, structural uncertainty;

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## 1 Introduction

2 Modelling hydro-climatic systems for a Himalayan catchments is particularly challenging  
3 because of the double-edged situation of highly heterogeneous and sparsely instrumented  
4 catchments. On the one hand, sharp topographic variations in this region result in extreme  
5 climatic heterogeneities (Barros *et al.*, 2004; Anders *et al.*, 2006) and on the other hand, the  
6 high-altitude areas have limited hydro-meteorological monitoring devices. A combination of  
7 these issues critically limits the representation of hydrological responses at regional scales in  
8 the Himalayan region.

9  
10 The central part of the Hindu Kush Himalaya region ranges from the Terai agricultural  
11 plain in the South to the highest peaks in the world to the north (FIGURE 1). The two main  
12 driving climatic processes are the summer Indian monsoon, which contributes approximately  
13 80% of the total annual precipitation over the central Himalayan range (Bookhagen and  
14 Burbank, 2006; Dhar and Rakhecha, 1981), and winter precipitation arising from westerlies  
15 (Lang and Barros, 2004).

16  
17 Limited access and physical constraints stemming from the region's steep topography ex-  
18 plain that the density of meteorological stations is particularly low in the Himalayan region.  
19 Recorded time series are more often short in duration and associated with significant uncer-  
20 tainties (Salerno *et al.*, 2015). Moreover, most of the stations are located in river valleys,  
21 which may not represent the spatial variation of precipitation in nearby mountain ranges.  
22 The gridded climate products from regional and global data sets provide a good deal of un-  
23 certainty due to interpolation approaches and a trade-off between resolution and availability  
24 of observed data (Li *et al.*, 2017).

25  
26 Various hydrological modelling approaches have been set up for several basins of the cen-  
27 tral Himalayas, at different spatio-temporal scales, from physically-oriented representations  
28 of processes, such as TOKAPI by Pellicciotti *et al.* (2012) or SWAT by Bharati *et al.* (2016),  
29 to more conceptual ones, such as SRM by Immerzeel *et al.* (2010), GR4J by Andermann  
30 *et al.* (2012) and Pokhrel *et al.* (2014), GR4JSG by Nepal *et al.* (2017a), SPHY by Lutz

31 *et al.* (2014), HDSM by Savéan *et al.* (2015) and J2000 by Nepal *et al.* (2014, 2017b). How-  
32 ever, large discrepancies remain in the representation of hydrological processes among several  
33 studies at a regional scale stemming from the variation in modelling applications, input data  
34 and the processes taken into account.

35

36 For instance, for the Dudh Koshi River basin, annual actual evapotranspiration is es-  
37 timated at 14%, 20% and 52%, respectively, of annual precipitation by Andermann *et al.*  
38 (2012); Nepal *et al.* (2014) and Savéan *et al.* (2015). Estimations of the snow melt contri-  
39 bution to annual stream flow at the outlet of the Dudh Koshi River basin range from 6%  
40 (Andermann *et al.*, 2012) to 27% (Nepal *et al.*, 2014); estimations of the glacial melt contri-  
41 bution to annual stream flow range from 4% (Andermann *et al.*, 2012) to 19% (Lutz *et al.*,  
42 2014). Moreover, estimations of the contribution of underground water to surface flow are  
43 still very divergent because of the variation in methodological approaches. The contribu-  
44 tion of groundwater flows to annual stream flows is estimated at about 60%, 20% and 12%,  
45 respectively, by Andermann *et al.* (2012); Nepal *et al.* (2014) and Lutz *et al.* (2014). The  
46 variation is mainly due to the conceptualization of groundwater processes in different models,  
47 for example J2000 represents two compartments for groundwater storage, whereas SPHY has  
48 one and GR4J has a conceptual representation of groundwater.

49

50 Taking into account this difficult context, the aim of this paper is to initiate a model  
51 inter-comparison work by comparing two approaches that have been previously applied to  
52 sparsely instrumented catchments in the Himalayas, namely the work of Eeckman (2017)  
53 that uses the ISBA (Interaction Sol-Biosphere-Atmosphere) surface and the work of Nepal  
54 *et al.* (2014) that uses the J2000 distributed hydrological model. The ISBA surface scheme  
55 (Noilhan and Planton, 1989; Noilhan and Mahfouf, 1996) allows to simulate the interaction  
56 between the hydrosphere, the biosphere and the atmosphere taking into account both the  
57 mass and the energy balance at the surface and its propagation into the soil. In this study,  
58 an additional conceptual module is coupled to ISBA to represent the flow routing, which  
59 was not originally included in the surface scheme. The J2000 model applies a process-based  
60 approach through calibration parameters and is distributed based on Hydrological Response

61 Units (HRUs). The J2000 model has been applied in Himalayan catchments at meso-scale  
62 catchments such as the Dudh Koshi and and Tamor river basin (Nepal *et al.*, 2017b).

63

64 Two small catchments were chosen to represent different climatic and physiographic char-  
65 acteristics of the Central Himalayas: the Kharikhola ( $18.2km^2$ ) and the Tauche catchment  
66 ( $4.6km^2$ ) which represent middle mountains and headwaters of high mountains respectively  
67 of the Nepalese Himalayas. Input uncertainties associated with both climatic variables and  
68 static spatial parametrization for topography, soil and vegetation were minimized as much  
69 as possible using data sets that have been locally validated based on in situ measurements  
70 of both meteorological variables and surface properties. The same data sets are used in both  
71 models, not only for the meteorological forcing but also for the soil and surface descriptions.

72

73 The novelty of the study is to apply in this sparsely instrumented region, at a small spatio-  
74 temporal scale and under the same forcing conditions, two models that deeply differs in their  
75 degrees of conceptualization: on the one side, a fully distributed surface scheme based on  
76 mass and energy balance equations and, on the other side, a semi-distributed calibrated  
77 model.

## 78 1. Study area

79 The Kharikhola and Tauche sub-catchments are part of the Dudh Koshi River basin in  
80 Eastern Nepal. This basin has a steep topography and high mountain peaks including Mt  
81 Everest, (8848, m a.s.l), dominated by a sub-tropical climate in lower areas and an alpine  
82 climate in high-altitude areas (see FIGURE 1). These two sub-catchments present different  
83 climatic and physiographic characteristics.

84

85 The elevation of the Kharikhola catchment varies from from 1980 m a.s.l. to 4660 m a.s.l.  
86 with an area of  $18.20 km^2$ . This catchment is covered by extensive agricultural areas (below  
87  $2500 m.a.s.l$ ), forests ( between  $2500 m.a.s.l$  and  $3500 m.a.s.l$ ) and sparce vegetation areas  
88 (above  $3500 m.a.s.l$ ). The glaciary coverage on the Kharikhola catchment is nil. The elevation  
89 of the Tauche catchment varies from from 3980 m a.s.l. to 6110 m a.s.l. with an area of 4.6

90  $km^2$ . This catchment is sparsely vegetated, mainly covered by shrublands or alpine steppes.  
 91 On the Tauche catchment, the Tauche peak glacier is suspended upstream of the catchment  
 92 and accounts for about 0.37% of the basin's total area, according to Racoviteanu *et al.* (2013)  
 93 up-to-date glacial inventory. The glacial contribution to the flow for the Tauche catchment is  
 94 therefore considered to be negligible and is not included in the modelling applications. The  
 95 main morphological characteristics of the two catchments studied are summarized in TABLE  
 96 1.

97

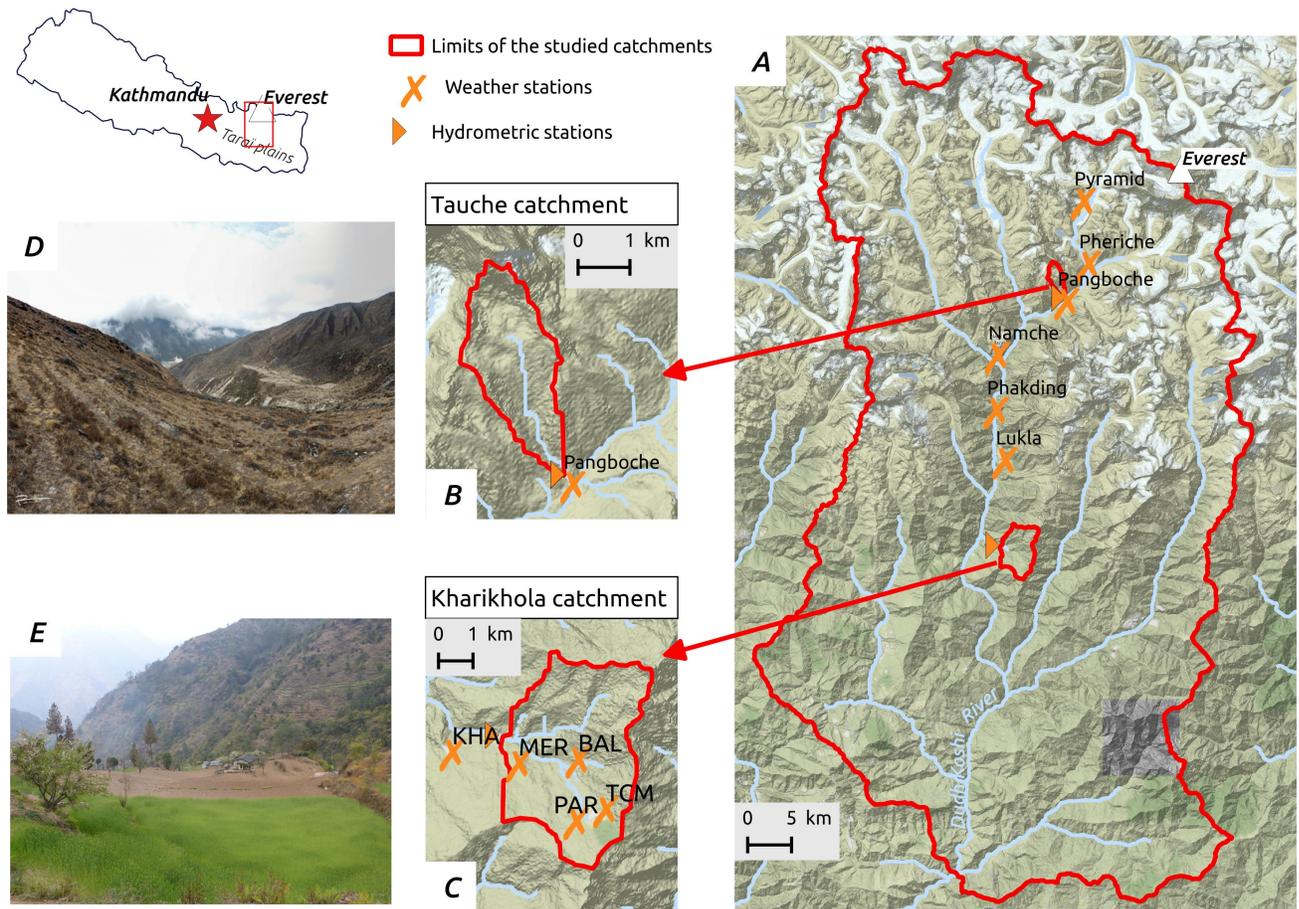


Figure 1: Map of the studied area: (A) the Dudh Koshi River basin at the Rabuwabazar gauging station, managed by the Department of Hydrology and Meteorology of the Nepal Government. The (B) Tauche and (C) Kharikhola sub-catchments are defined by the corresponding gauging stations. Source: OpenStreetMaps, photos by Rémi Muller (D) and Judith Eeckman (E).

Table 1: Summary of the main morphological characteristics of the two catchments studied: : Kharikhola catchment and Tauche catchment (Nepal), which represents mid-altitude mountains and high-mountain headwaters, respectively.

	Kharikhola	Tauche	unit
Area	18.2	4.6	$km^2$
Elevation range	1980 - 4660	3980 - 6110	m.a.s.l.
Glaciarized area	0%	0.37%	-
Discharge data	from 2014-05-03 to 2016-05-20	from 2014-05-07 to 2016-05-09	

## 98 2. Modelling approaches

99 The implementation choices are summarized for both models in TABLE 2.

### 100 2.1. The ISBA surface scheme and the HDSM routing module

101 The ISBA (Interaction Soil Biosphere Atmosphere) surface scheme (Noilhan and Planton,  
 102 1989; Noilhan and Mahfouf, 1996) is implemented in the SURFEX platform (Masson *et al.*,  
 103 2013) to represent the nature land tile. The latest version 8 of SURFEX is used for this  
 104 work. The ISBA surface scheme simulates vertical fluxes between the soil, vegetation and  
 105 the atmosphere at a sub-hourly time step (SVAT model). Different implementations of  
 106 soil transfers, vegetation, sub-grid hydrology and snow processes are available in SURFEX.  
 107 Implementations of ISBA functions described in TABLE 2 are used in this study. The  
 108 explicit multilayer version of ISBA (ISBA-DIF) uses a diffusive approach (Boone *et al.*, 2000;  
 109 Decharme *et al.*, 2011): surface and soil water fluxes are propagated from the surface through  
 110 the soil column. Transport equations for mass and energy are solved using a multilayer  
 111 vertical discretization of the soil. The explicit snow scheme in ISBA (ISBA-ES) (Boone and  
 112 Etchevers, 2001; Decharme *et al.*, 2016) uses a twelve-layer vertical discretization of snow pack  
 113 and provides a mass and energy balance for each layer. Snow-melt and snow sublimation are  
 114 taken into account in balance equations.

115 The Dunnes flow (Dunne, 1983) and Hortons flow (Horton, 1933) are separately modeled  
 116 in ISBA. The Dunnes flow is the saturation excess runoff i.e. the fraction of the precipitation  
 117 that flows at the surface when the soil is saturated. The Hortons flow is the infiltration excess  
 118 runoff i.e. the fraction of precipitation that flows at the surface when the intensity of the  
 119 precipitation is greater than the soil capacity of infiltration. The Horton's and Dunne's flow  
 120 mechanisms are modeled using a sub-grid parameterization described in Habets *et al.* (1999):  
 121 The Dunne runoff for each grid cell depends on the fraction of the cell that is saturated. The

122 fraction of the cell that is saturated depends on the total soil water content within the cell.  
123 Considering the different physiography of the two studied catchments, the shape parameter  
124  $\beta$  for the relation between soil water content and fraction of saturated area of the cell is  
125 2.0 for the Kharikhola catchment and 0.3 for the Tauche catchment. ISBA is set up for the  
126 Tauche and Kharikhola catchments on a regular grid at a 400-m spatial resolution and with  
127 an hourly time step.

128

129 Since the dependency between mesh cells is not initially implemented in the SURFEX  
130 platform, an additional routing module was implemented and coupled to ISBA offline sim-  
131 ulations. This module is adapted from the HDSM (Hydrological Distributed Snow Model)  
132 model, and it has been implemented and used by Savéan *et al.* (2015) on the Dudh Koshi  
133 River basin. The advantage of using the ISBA-HDSM coupling is to use both a non cali-  
134 brated surface scheme for production function and a routing module that has been previously  
135 applied on the same area. The structure of the module is extensively described in Savéan  
136 (2014). For each cell, surface runoff (given by the sum of Dunne runoff and Horton runoff)  
137 and the drainage at the bottom of the soil column are directed toward two simple linear  
138 reservoirs,  $R_s$  and  $R_d$  respectively. Residence times in  $R_s$  and  $R_d$  (respectively,  $t_s$  and  $t_d$ )  
139 are calibrated as uniform parameters over the catchment. The sum of the output flows of  
140  $R_s$  and  $R_d$  is then directed toward the transfer reservoir, which allows propagating the flows  
141 according to terrain orography. The residence time in the transfer reservoir is defined for  
142 each mesh point as the ratio between the flow velocity and the distance from the centre of  
143 the mesh point to the centre of the previous upstream mesh point. The flow velocity is calcu-  
144 lated as the ratio of the mesh point slope and a reference slope, taken equal to the catchment  
145 median slope. This ratio is weighted by a  $c_{vel}$  transfer coefficient.  $c_{vel}$  is calibrated as a  
146 uniform parameter. The main driving equations of this routing module and its calibration  
147 are reported in Appendix A. The code for this routing module is implemented in fortran90  
148 language and available at [www.papredata.org](http://www.papredata.org).

## 149 2.2. J2000 modelling system

150 The J2000 hydrological model is a process-oriented hydrological model (Krause, 2001).  
151 The model is implemented in the Jena Adaptable Modelling System (JAMS) framework

152 (Kralisch and Krause, 2006; Kralisch *et al.*, 2007), which is a software framework for component-  
153 based development and application of environmental models. The J2000 model includes the  
154 main hydrological processes of high-mountain catchments. A short description of the main  
155 processes has been provided in TABLE 2. A more detailed description is provided by Krause  
156 (2001) and Nepal (2012). The J2000 model has already been applied to Himalayan catch-  
157 ments (Nepal *et al.*, 2014, 2017a).

158 To optimize the J2000 model parameters for the KhariKhola and Tauche catchments,  
159 we used the base parameter set from a previous study by (Nepal *et al.*, 2014), which was  
160 defined for the Dudh Koshi River basin at the Rabuwabazaar gauging station (3712 km<sup>2</sup>).  
161 Similarly, (Nepal *et al.*, 2017a) also used the same parameter sets for nearby Tamor sub-  
162 catchment (4005 km<sup>2</sup>) to argue that spatial transferability of the J2000 model parameters  
163 is possible in neighbouring catchments with physical and climatic similarities. Out of 30  
164 parameters, six parameters were optimized further to match the catchment responses in  
165 the KhariKhola and Tauche catchments: the groundwater recession coefficient for baseflow  
166 (gwRG2Fact), the coefficient for the distribution of water between the upper and lower zone of  
167 groundwater (gwRG1RG2dist), the recession coefficient for RD1 and RD2 (soilConcRD1 and  
168 soilConcRD2), maximum percolation (soilMaxPerc) and snowmelt threshold (baseTemp) and  
169 the parameter to distribute precipitation into rainfall and snow (trs). The recession coefficient  
170 for floods from (Nepal *et al.*, 2014) is not applied here because of the local scale catchments.  
171 Because of the basin size and climatic variability within the catchment and related scale  
172 issues, optimization of parameters is suggested. The description of these parameters along  
173 with their dimensions are available in Nepal *et al.* (2017a).

### 174 2.3. Spatial discretization methods

175 The SPOT DEM (Gardelle *et al.*, 2012), as well as soil and land cover maps are provided  
176 for both catchments at the 40-m resolution. In ISBA, the catchments are discretized over  
177 a regular grid at the 400-m resolution. Sixty-nine grid cells are defined for the Kharikhola  
178 catchment and 28 grid cells are defined for the Tauche catchment. In J2000, the catchments  
179 are discretized into 346 and 132 HRUs, respectively. The minimum size of HRUs is forced  
180 to be larger than 5 DEM pixels, i.e. 0.008 km<sup>2</sup>. TABLE 3 summarizes the results of the  
181 spatial discretization for both modelling applications. FIGURE 2 shows the hypsometric

182 information of the land surface area in different elevation zones. Although the overall pattern  
 183 of hypsometry is similar in both models, they tend to show fairly opposite area coverage above  
 184 and below about 3000 *m.a.s.l.* for Kharekhola and 5000 *m.a.s.l.* for Tauche.

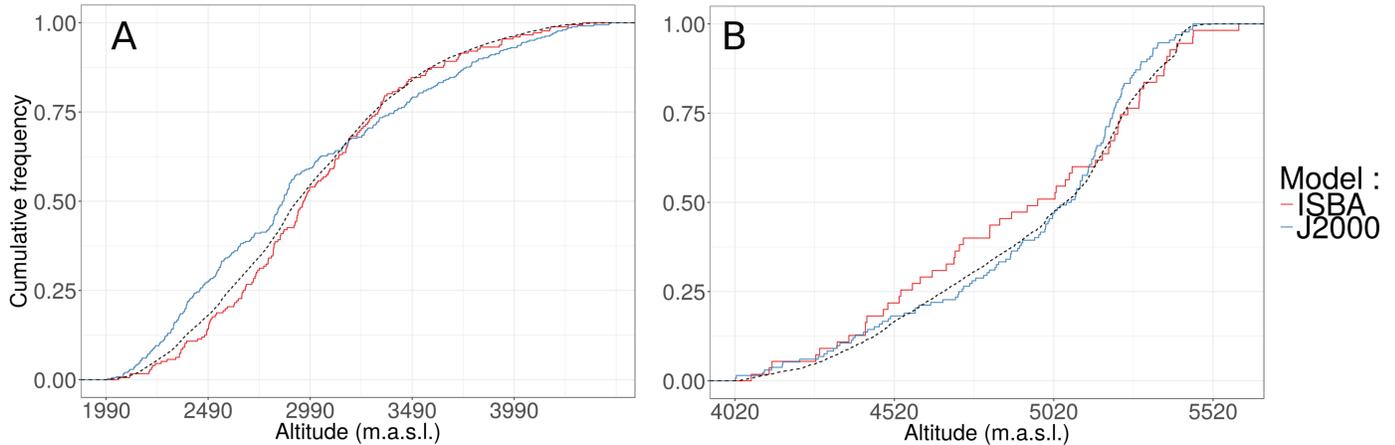


Figure 2: Hypsometric curve for Kharikhola catchment (A) and for the Tauche catchment (B), provided by the ISBA discretization on a regular grid at the 400-m resolution (red curve) and by the J2000 discretization into HRUs (blue curve). The dotted line is the hypsometric curves given by the 40-m SPOT DEM (Gardelle *et al.*, 2012).

#### 185 2.4. Soils and vegetation patterns

186 In order to enhance the local accuracy of soil and vegetation descriptions that are cur-  
 187 rently available, a classification of surfaces into nine categories is defined based on field  
 188 observations and soil characteristics measurements. This classification is spatially extrapo-  
 189 lated using a semi-supervised classification of two Sentinel 2 images (Drusch *et al.*, 2012) at  
 190 a 10-m resolution for the two catchments studied. The values for soil depth and texture,  
 191 root depth, vegetation type and vegetation fraction for each of the nine classes are shown on  
 192 TABLE 4. In addition, other parameters needed for the parametrization of the models (e.g.  
 193 leaf area index, surface albedo and surface emissivity) are taken from the ECOCLIMAP1  
 194 classification (Masson *et al.*, 2003) for the representative ecosystems presented on TABLE  
 195 4. This parametrization of soil and vegetation is used in both models. The classification  
 196 method and the characteristics of each class are described in detail by Eeckman *et al.* (2017).  
 197 The surface classification established at the 10-m resolution is aggregated at the resolution  
 198 of each model. The classification maps used to parameterize soil and vegetation in both  
 199 models, for both the Kharikhola and Tauche catchments, are presented in FIGURE 3. The  
 200 overall location of each class is consistent in both models, although the two different spatial

201 aggregation methods necessarily induce local differences in these maps.

202

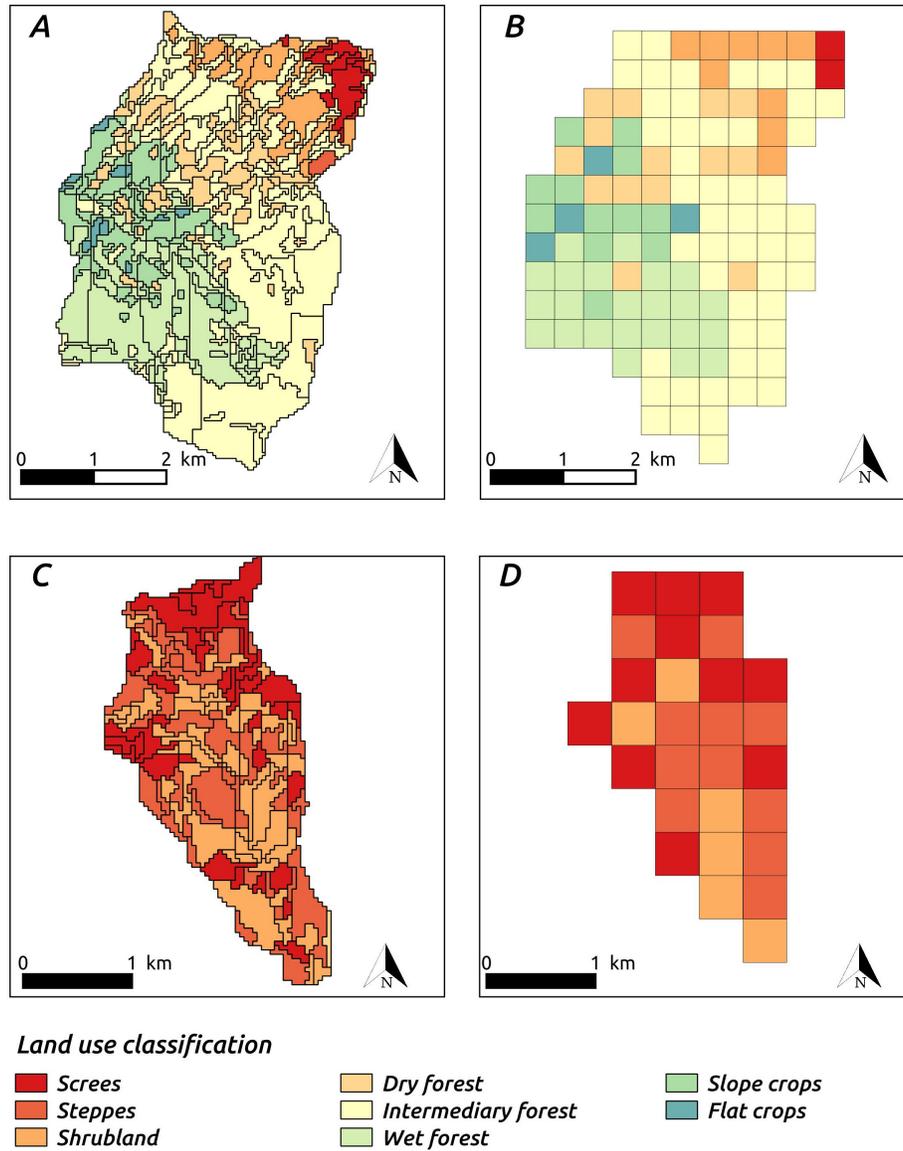


Figure 3: Land cover classification defined for each HRU in the J2000 model: (A) on the Kharikhola catchment, (C) on the Tauche catchment; and on a regular 400-m resolution grid in ISBA: (B) on the Kharikhola catchment, (D) on the Tauche catchment. Each land cover class provides soil and vegetation characteristics established from in situ measurements.

### 203 2.5. Climatic input

204 Temperature and total precipitation are measured at 11 weather stations installed within  
205 the Dudh Koshi basin (see FIGURE 1). Reliable measurements for short- and long-wave  
206 radiation, atmospheric pressure, relative air humidity and wind speed are available at the  
207 Pyramid station, located at 5035 m.a.s.l., within the Sagarmatha National Park, Khumbu  
208 region, Nepal, and managed by the association Ev-K2-CNR, Bergamo-Italy (see FIGURE

209 1). Hourly measurements for these variables are available at the Pyramid station from Oc-  
210 tober 2002 to December 2004 ( at *www.evkl2.isac.cnr.it/*). Hourly time series are computed  
211 from measurements over the three hydrological years 2013-2012, 2014-2015 and 2015-2016.  
212 The hydrological year is considered to start on April 1, as decided by the Department of  
213 Hydrology and Meteorology of the Nepalese Government and in general use (Savéan *et al.*,  
214 2015). Two seasons are defined: the summer season, from April 1 to October 30, that includes  
215 the monsoon and pre-monsoon periods, and the winter season, from November 1 to March 31.

216

217 Climatic variables are spatially interpolated according to the methods and values detailed  
218 in Eeckman *et al.* (2017):

219 - Air temperature measurements are spatially interpolated using a multi-linear method  
220 weighed by the inverse distance (IDW method), coupled with a seasonal altitudinal  
221 lapse rate. The altitudinal lapse rate is computed from the observation :  $-5.87^{\circ}C.km^{-1}$   
222 for winter and  $-5.64^{\circ}C.km^{-1}$  for summer.

223

224 - Total precipitation is interpolated using the method proposed by Valery *et al.* (2010):  
225 the IDW method is coupled to a multiplicative altitudinal factor  $\beta$ . The altitudinal  
226 factor  $\beta$  is represented as a piecewise linear function of altitude. Altitudinal thresholds  
227 and lapse rates are optimized to provide optimal bias on annual discharge for both the  
228 Kharikhola and Tauche catchments. During the summer season, precipitation is con-  
229 sidered to increase up to an altitudinal threshold of 3470 m.a.s.l. (3113 m.a.s.l. during  
230 winter) at a rate of  $0.032 km^{-1}$  ( $1.917 km^{-1}$  during winter), then to decrease at a rate  
231 of  $-1.382 km^{-1}$  ( $-1.83 km^{-1}$  during winter) up to 3709 m.a.s.l. (4943 m.a.s.l. during  
232 winter). For higher altitudes, precipitation is considered to decrease at a rate of  $-0.283$   
233  $km^{-1}$  ( $-0.191 km^{-1}$  during winter).

234

235 - Long-wave radiation, atmospheric pressure and specific air humidity measurements  
236 at the Pyramid station are spatialized as a function of altitude, using the method  
237 proposed by Cosgrove *et al.* (2003). The hourly temperature is used to interpolate  
238 the atmospheric pressure based on the ideal gases law. The specific air humidity is

239 deduced from the relative air humidity by combining the Wexler law and the definition  
240 of the saturating vapor pressure. The long wave radiation emitted is computed based  
241 on the air temperature using the Stefans law. Since short-wave radiation and wind  
242 speed have a quite low sensivity in the models in comparison with the other variables,  
243 these two variables are not spatially interpolated and are considered to be equal to the  
244 measurements at the Pyramid station for the two catchments studied.

245 This interpolation method for precipitation provides optimal precipitation fields for both  
246 the Kharikhola and Tauche catchments, for the two hydrological years 2014–2015 and 2015–  
247 2016, according to the discharges. However, the interannual variability is hardly represented  
248 in this interpolated data set. Indeed, these 2 years are very different. For the Kharikhola  
249 catchment, observed discharge at the outlet reached  $48.3 \text{ mm/day}$  in July 2014, whereas it  
250 did not exceed  $24.5 \text{ mm/day}$  in 2015–2016 (see FIGURE 4). For the Tauche catchment,  
251 the rainfall-runoff ratio was 53% in 2014–2015 and 82% in 2015–2016, considering interpo-  
252 lated precipitation and observed discharge (TABLE 5). These variations can be due to the  
253 combined effects of (i) the effective interannual variability of climatic variables, (ii) errors  
254 in precipitation measurements, in particular concerning snowfall underestimation (Sevruk  
255 *et al.*, 2009), (iii) errors in water level measurements or in the interpolation of discharge  
256 based on the rating curve. In particular, high discharge peaks might be overestimated when  
257 interpolated from the rating curve, because only a few gauging points are available for high  
258 water levels.

259  
260 However, since the aim of this paper is to compare the hydrological responses of two  
261 models when using the same input data set, the choice was made not to consider uncertainties  
262 in hydro-climatic input data, but to focus on comparing the simulated responses of the two  
263 models.

## 264 *2.6. Discharges*

265 Hourly discharge time series are available at the hydrometric stations located at the  
266 Kharikhola outlet and at the Tauche outlet, from 2014-05-03 to 2016-05-20 and from 2014-  
267 05-07 to 2016-05-09, respectively (see TABLE 1). Two hydrometric stations were equipped

268 with Campell® hydrometric sensors. The rate curves for the two stations have been defined  
269 using 25 measurements in Kharikhola (from 0.020 to 7.48  $m^3/s$ ) and 19 measurements in  
270 Tauche (from 0.003 to 0.202  $m^3/s$ ). The time series at Kharikhola station contains 34%  
271 missing data in 2014-2015, due to a high monsoon flood which damaged the sensor. The  
272 time series at Tauche station contains no missing data, but additional observations made by  
273 a local observer indicated that the river was frozen from 2015-01-22 to 2015-02-28 and from  
274 2016-01-08 to 2016-02-23. Discharge is considered as null during the frozen periods.

275

276 A particular attention has to be paid to the discharge peak happening in June 2015  
277 for the Tauche bassin. Indeed, the observed hydrograph increased from 0.06  $m^3/sec$  on 22  
278 June to 0.3  $m^3/sec$  on 26 June (the highest peak of 2015). The precipitation is below 7 mm  
279 and remains throughout the period (the discharge event may not be due to precipitation  
280 events). The maximum temperature increased from 7°C to 8°C from 22 to 24 June and then  
281 decreased to 5.5°C in 25 June. The discharge event may then be due to either snow-melt  
282 fluxes or instrument error.

### 283 2.7. Snow cover area

284 The MOD10A2 product (Hall *et al.*, 2002) provides the maximum snow cover extent over  
285 a 500-m resolution grid, at an 8-day time scale since 2000-02-26 to present. MOD10A2 is  
286 derived from the MODIS/Terra Snow Cover Daily product (MOD10A1). To compute the  
287 MOD10A2 maximum snow cover extent from MOD10A1 snow cover, the following condition  
288 is applied: if a pixel is considered as covered by snow at least once within each 8-day time lapse  
289 in the MOD10A1 product, this pixel is considered as covered by snow for the corresponding  
290 8-day period in MOD10A2. MOD10A2 is commonly used in glaciological and hydrological  
291 studies in the western Himalayas (Shrestha *et al.*, 2011; Panday *et al.*, 2014; Pokhrel *et al.*,  
292 2014; Savéan *et al.*, 2015; Nepal *et al.*, 2017a). Moreover, the accuracy of this product  
293 was assessed in mountainous areas by various studies (Jain *et al.*, 2008). In particular,  
294 Chelamallu *et al.* (2014) concluded that the MODIS products were more accurate in regions  
295 with substantial snow cover than in regions with low snow cover.

297 Observed discharges were available for only 1 complete hydrological year (2015-2016) at  
 298 the Kharikhola catchment and for 2 hydrological years at the Tauche catchment (2014-2016).  
 299 The ISBA and J2000 simulations over these catchments were run separately from 2013-01-  
 300 01 to 2016-03-31. The 2013–2014 year was used as a spin-up period and the results were  
 301 observed for the 2014–2016 hydrological years. The ISBA was run at an hourly time scale  
 302 and hourly model outputs were aggregated to the daily level. The ISBA routing module  
 303 was calibrated over the whole period of available discharge observations. No independent  
 304 validation period was then considered here due to the short period of observed data.  
 305 The choice has been made to apply the J2000 model at a daily time step, in order to be  
 306 consistent with the work of Nepal *et al.* (2011) (see Section 2.2). Indeed, the calibrations of  
 307 J2000 provided by Nepal *et al.* (2014) might not be valid at the hourly time step. Consider-  
 308 ing the very short period of availability of the water level measurements for the two studied  
 309 catchments, a new calibration of J2000 in this work would not be feasible. For this reason,  
 310 the choice has been made to keep using a daily time step to run J2000.

311

312 Model performance was assessed against observed discharge data using the four efficiency  
 313 criteria : coefficient of determination  $r^2$ , Nash-Sutcliffe Efficiency ( $NSE$ ),  $NSE$  for the  
 314 square root of discharges ( $NSE_{sqr}$ ) and relative bias ( $Bias_r$ ), computed at the daily time  
 315 scale. The  $NSE_{sqr}$  has the property of flattening flow peaks and therefore it is used to assess  
 316 performance for low-flow periods (Zhang *et al.*, 2007). To assess performance for high-flow,  
 317 the  $NSE$  criteria is also computed separately for the high flow periods, i.e. from June 1 to  
 318 September 30. This criterion is noted  $NSE_{high}$ .

### 319 3. Results and discussion

320 TABLE 5 presents annual volumes for total precipitation, solid precipitation, evapotran-  
 321 spiration, discharge and snow-melt contribution, in annual average over each of the two  
 322 catchments studied. FIGURE 4 and FIGURE 5 present the dynamics of simulated variables  
 323 in both models, respectively for the Kharikhola and Tauche catchments. This section aims  
 324 to investigate in detail the differences and the similarities between the two modelling ap-

325 proaches in order to better describe the uncertainties associated with model structure in the  
326 estimation for the annual water budgets provided in the literature.

### 327 3.1. Evaluation against observed discharge

328 FIGURE 4 and FIGURE 5 show the simulated and observed hydrographs in the Kharikhola  
329 and Tauche catchments, respectively. TABLE 5 presents the performance of both models for  
330 four different efficiency criteria computed at the daily time scale.

331

332 Annual relative bias on discharge for the Kharikhola catchment is satisfactory for 2015–  
333 2016 for both models (-9.7% for ISBA and 0.04% for J2000), but the discharge at the  
334 Kharikhola outlet is strongly under-estimated for 2014–2015 for both models ( $Bias_r$  is -  
335 45.7% for ISBA and -37.1% for J2000). On the Tauche catchment, the observed discharges  
336 are under-estimated for both years, for both models, with average  $Bias_r$  values of -15.6% for  
337 ISBA and -20.90% for J2000. These under-estimations are due to the under-estimation of  
338 total precipitation for the corresponding years, as presented section 2.5.

339

340 For the two hydrological years, the dynamics of the observed discharges is accurately rep-  
341 resented by the two models for the two catchments, with the annual average of  $r^2$  greater than  
342 0.72 and  $NSE$  values greater than 0.70. During the summer season, the discharge dynamics  
343 is driven by precipitation, with a quick response of the surface runoff for both catchments.  
344 These quick flow variations are satisfactorily represented by both models.

345

346 Low flows including rising and recession periods are accurately captured for both years  
347 by both models for the Kharikhola catchment: on average over the 2 years,  $NSE_{sqr t}$  is 0.77  
348 for IBSA and 0.82 for J2000. For the Tauche catchment, low flows are clearly represented for  
349 2014–2015 ( $NSE_{sqr t}$  is 0.78 for ISBA and 0.76 for J2000). The representation of high-flow  
350 peaks in the summer season is very satisfactory for the Kharikhola catchment in 2015–2016  
351 and for the Tauche catchment in 2014–2015 ( $NSE_{high}$  values greater than 0.66).

352

353 An interesting period of discordance between the two models occurs between March 2015  
354 and June 2015 (pre-monsoon period) on the Tauche catchment. During this period, the

355 ISBA model simulates a fast-responding discharge, whereas the seasonal increasing of the  
356 discharge simulated with the J2000 model is slower. Despite both models under-estimate the  
357 observed discharge, the discharge simulated with ISBA during this pre-monsoon period are  
358 greater than the discharge simulated with J2000. In order to better understand this period  
359 of discrepancies between the two models, each of the components of the water budget are  
360 described in the following section, and in particular their behavior during this pre-monsoon  
361 2015 period.

## 362 *3.2. Components of annual water budgets*

### 363 *3.2.1. Precipitation*

364 Slight differences exist for total precipitation between the J2000 and ISBA models, al-  
365 though the same precipitation input is provided for both models. These differences stem from  
366 the spatial discretization methods used in both models for precipitation spatial interpolation  
367 (see section 2.3). Indeed, even though the input grid data provided for precipitation are the  
368 same for both models, precipitation is further interpolated by J2000 from the grid scale to the  
369 HRU scale, using the inverse distance weighting method. However, for both catchments, the  
370 difference in total precipitation between the two models represents less than 1% of the annual  
371 volume (0.98% for the Kharikhola catchment and 0.45% for the Tauche catchment). The dif-  
372 ference between both models for total precipitation can then be considered as negligible.

373

374 Regarding the solid precipitation, the annual volumes considered in the two models also  
375 differ. For the Kharikhola catchment, this difference is about 7 *mm* (representing 1.1%  
376 of annual average solid precipitation) and can then be considered as negligible. But, for  
377 the Tauche catchment, the average solid precipitation is 219 *mm* higher for ISBA than for  
378 J2000. This difference represents about 40% of the annual volumes of solid precipitation. This  
379 significant difference is mainly due to the difference in the time step used for precipitation  
380 phase distribution in both models. Indeed, despite both total precipitation and temperature  
381 fields are provided for both models at the hourly time step, the precipitation phase in J2000  
382 is computed at the daily time step, whereas it is computed at the hourly time step in ISBA.  
383 However, as specified in section 2.8, the purpose of this work is to stay consistent with the  
384 previous studies that uses these models. The infra-daily variations of solid precipitation is

385 then missed in J2000 (see section 3.3 for further analysis). This point shows that the time  
386 step used for the partition of precipitation phase strongly influences the simulation results.  
387 The propagation of this difference toward the simulated water budgets in the two models is  
388 further investigated.

### 389 3.2.2. Discharge components

390 In ISBA, the surface overland flow is considered as the sum of the simulated Dunnes flow  
391 and Hortons flow (see section 2.1). In J2000, the Dunne and Horton flows mechanisms are  
392 not separated and the simulated surface runoff comprises both saturation and infiltration  
393 excess runoff. The drainage flow at the bottom of the soil column in ISBA is comparable  
394 to the sum of the three flows in the soil in J2000 ( $RD2$ ,  $RG1$  and  $RG2$ ). The following  
395 comparison is given on average over the two hydrological years studied.

396  
397 For ISBA, the Hortonian runoff represents less than 1% of discharge on the Tauche catch-  
398 ment and about 5% of discharge on the Kharikhola catchment. While in ISBA, this means  
399 than the surface flows occur mainly (on the Kharikhola catchment) or almost only (on the  
400 Tauche catchment) by the saturation of soil reservoirs rather than by excess infiltration ca-  
401 pacity.

402  
403 For the Kharikhola catchment, the annual volume of drainage (i.e. sub-surface flow)  
404 represents 77% of the discharge at the outlet for ISBA (drainage flow at the bottom of the  
405 soil column) and 87% for J2000 (sum of the  $RD2$ ,  $RG1$  and  $RG2$  flows). For the Tauche  
406 catchment, this volume is 70% for ISBA and 85% for J2000 of the annual discharge. The  
407 annual overland flow represent 30% of the annual discharge for ISBA and 13% for J2000 on  
408 the Kharikhola catchment. For the Tauche catchment, this volume is 23% for ISBA and 15%  
409 for J2000. These figures highlight the significant contribution of soil water to discharge for  
410 both middle- and high-mountain catchments.

411 Therefore, this model intercomparison reveals that most discharge at the outlet is provided  
412 by drainage. This result is consistent with the description of soils for the two catchments:  
413 sandy soils allow fast infiltration, resulting in a larger fraction of the flow occurring in the soils  
414 than on the surface. However, the definition of drainage strongly differs between the two

415 models. Indeed, in ISBA, the drainage represents the vertical flow at the bottom of the soil  
 416 column (without routing nor delays), whereas in J2000, it represents the sum of the outflows  
 417 from the soil water module.

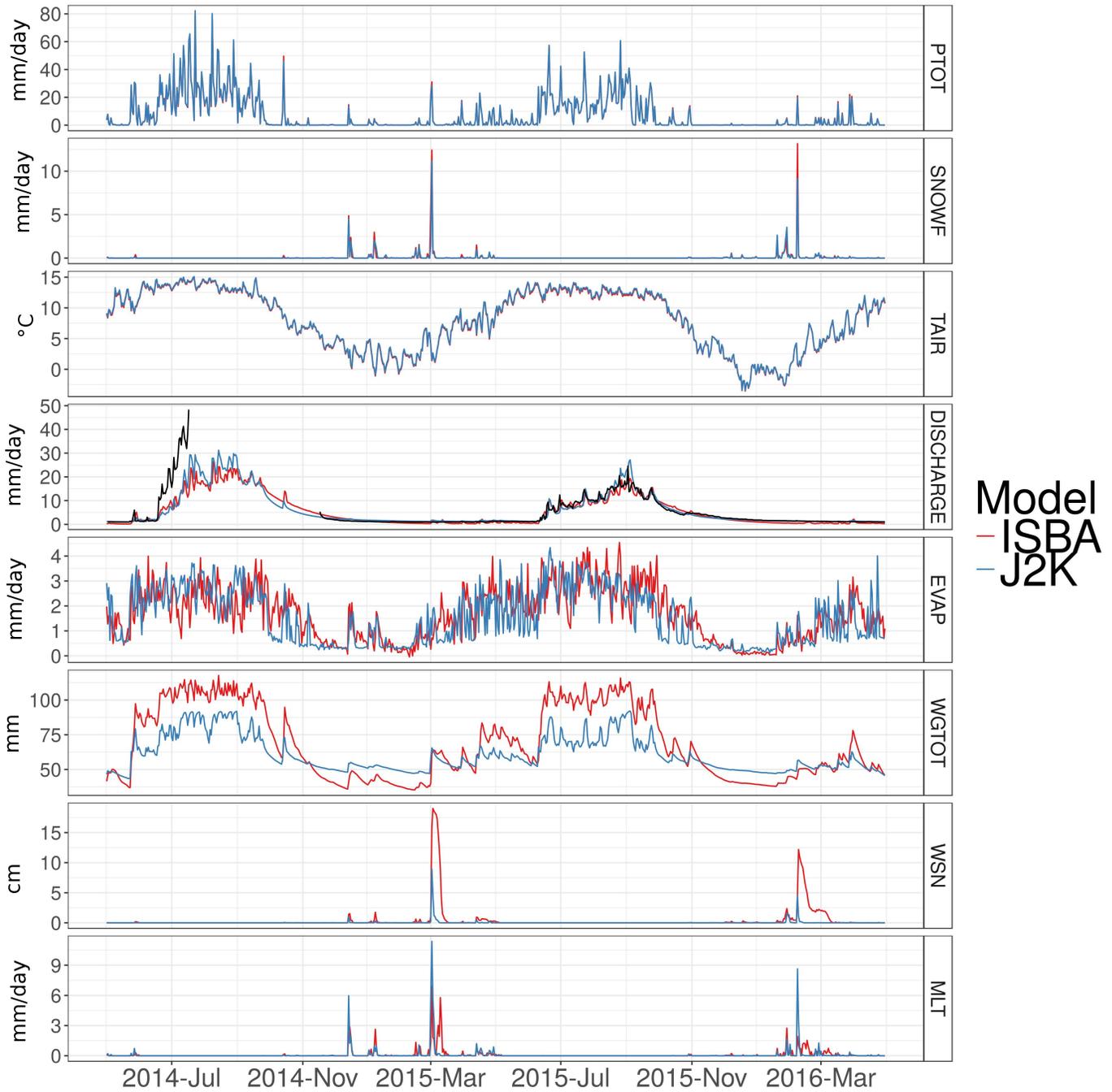


Figure 4: Daily time series for input variables : total precipitation (PTOT), solid precipitation (SNOWF) and air temperature (TAIR) and for variables simulated by ISBA and J2000 models at the daily time scale : discharge at the outlet (DISCHARGE), actual evapotranspiration (EVAP), soil water content (WGTOT), snow water equivalent (WSN) of the snow pack and snow-melt (MLT), for the 2014–2015 and 2015–2016 hydrological years, for the Kharikhola catchment. Black line is the daily observed discharge at the outlet.

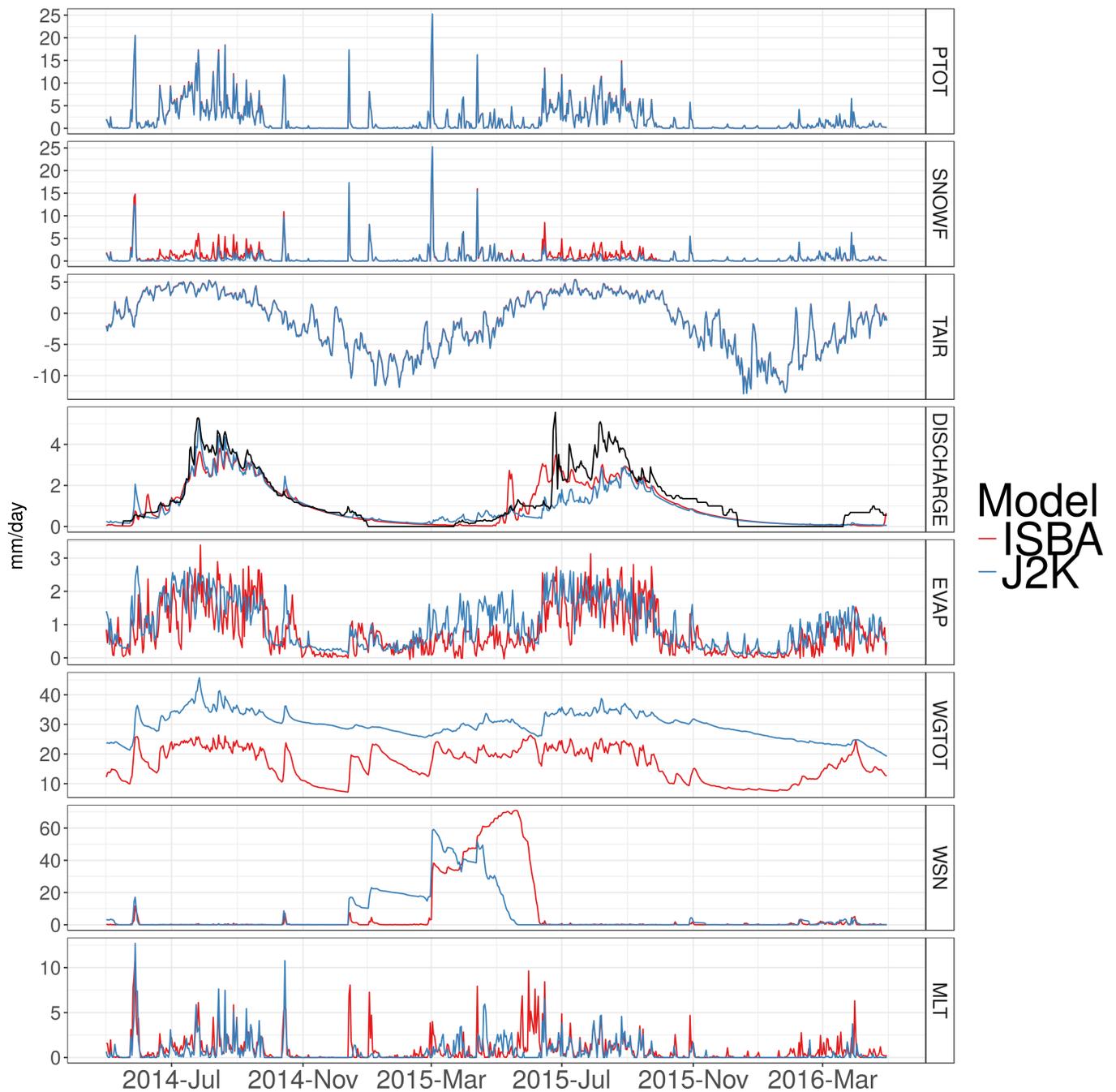


Figure 5: Daily time series for input variables : total precipitation (PTOT), solid precipitation (SNOWF) and air temperature (TAIR) and for variables simulated by ISBA and J2000 models at the daily time scale : discharge at the outlet (DISCHARGE), actual evapotranspiration (EVAP), soil water content (WGTOT), snow water equivalent (WSN) of the snow pack and snow-melt (MLT), for the 2014–2015 and 2015–2016 hydrological years, for the Tauche catchment. Black line is the daily observed discharge at the outlet.

### 418 3.2.3. Soil water content

419 The conceptualizations of the soil water storage in both models are very different (see  
 420 TABLE 2). Considering these structural discrepancies, the total water content of the soil  
 421 column simulated in ISBA can be compared to the sum of the volumes stored in MPS and  
 422 LPS reservoirs in J2000.

424 During the high-flow periods (between June and October), the dynamic of the soil wa-  
 425 ter content simulated with both models appear to be fast-responding to precipitation. The  
 426 timing of the simulated soil water content also matches well between both models. However,  
 427 on the Kharikhola catchment, the soil water content simulated with ISBA is lower than with  
 428 J2000 during low flow period, but this behavior is reversed during high flow period, with  
 429 greater soil water content simulated in J2000 than in ISBA. On the Tauche catchment, the  
 430 soil water content is permanently higher in J2000 than in ISBA.

431

432 This behavior can be explained by the fact that, in ISBA, the simulated soil water con-  
 433 tent is limited by the soil humidity at saturation ( $w_{sat}$ , in  $mm$ ).  $w_{sat}$  values are calculated  
 434 according to Clapp and Hornberger (1978), as a function of soil texture. In J2000, the vol-  
 435 ume stored in each reservoir MPS and LPS is limited by maximum volumes  $maxMPS$  and  
 436  $maxLPS$ , respectively.  $maxMPS$  and  $maxLPS$  are computed according to soil texture  
 437 for each HRU. The TABLE 6 presents the average values of  $w_{sat}$ ,  $maxMPS$  and  $maxLPS$   
 438 for the Kharikhola and Tauche catchment. Provided value for  $w_{sat}$  is greater than the sum  
 439  $maxMPS + maxLPS$  for the Kharikhola catchment, but it is lower for the Tauche catch-  
 440 ment. This parametrization can explain that the soil water content is globally higher in ISBA  
 441 than in J2000 for the Kharikhola catchment, but lower in ISBA than in J2000 the the Tauche  
 442 catchment. This point illustrates the fact that the representation of soil water content sig-  
 443 nificantly varies between these two models, as well as in other studies (see Introduction).

444

#### 445 3.2.4. Evapotranspiration

446 On average over the 2 hydrological years, the estimation of annual actual evapotranspi-  
 447 ration (actET) on the Kharikhola catchment was 22.6% of total annual precipitation with  
 448 ISBA and 19.8% with J2000. On the Tauche catchment, it was 34.4% with ISBA and 50.6%  
 449 with J2000 of the total annual precipitation. These values include bare soil evaporation, veg-  
 450 etation transpiration and snow sublimation. ActET for the two models in both catchments  
 451 correlated acceptably at the daily time scale, with  $r^2 = 0.48$  for the Kharikhola catchment  
 452 and  $r^2 = 0.38$  for the Tauche catchment. However, a major difference can be seen in the pre-

453 monsoon period (March-June) in the Tauche catchment where simulated actET is higher in  
454 J2000 (up to 2 *mm/day*) than in ISBA (less than 0.5 *mm/day*). This delay in the increasing  
455 of actET in ISBA is due to late simulated snow-melt in ISBA. Indeed, the simulated snow  
456 pack, that limits the evaporation simulated over bare ground, remains in ISBA until June  
457 2015, whereas it melts from March 2015 in J2000. This point indicates that the simulation  
458 of the snow-melt contribution significantly influences the simulation of both the discharge at  
459 the outlet and the evapotranspiration.

### 460 3.2.5. *Snow-melt contribution*

461 For both models, the contribution of snow-melt to discharge is less than 1.5% for the  
462 Kharikhola catchment. This point can then be considered as a robust result and it allows to  
463 enhance the actual understanding of the hydrological cycle for a middle-mountain catchment.  
464 For the Tauche catchment, the contribution of snow-melt accounts for 45.3% of the annual  
465 simulated discharge in ISBA results, and 33.2% of the annual simulated discharge in J2000  
466 results. For the Tauche catchment (see FIGURE 5), both models provide the majority (73%  
467 in ISBA and 82% in J2000) of snow-melt during the summer season. The timing of snow-  
468 melt between July and November (monsoon and post-monsoon periods) are similar in both  
469 models. However, the dynamic of the simulated snow-melt occurring between March and  
470 July 2015 significantly differs between the two models: On March, 2nd. a snow precipitation  
471 (25 *mm* in one day) leads to a sharp increasing of the snow-pack water equivalent in both  
472 models. However, in J2000, the snow-pack starts melting after this sudden snow fall, with  
473 snow-melt variations concomitant with snow falls. On the contrary, the snow-pack simulated  
474 with ISBA keeps accumulating the subsequent snow falls until May, 20th. These different  
475 dynamics of the simulated snow-pack in the two models explain the discrepancy not only of  
476 the simulated discharge but also of the simulated evapotranspiration between both models  
477 during this period. In order to further criticize these simulated snow pack, the MOD10A2  
478 maximum snow extent product is compared to the simulation results (see section 3.4).

479

480 Note that occasionally ISBA simulates snow melt for air temperatures below freezing,  
481 but only during winter, when the temperatures are low but also the cumulated snow packs  
482 are thin. This arises for essentially two reasons; i) solar radiation transmitted through the

483 snowpack when it is fairly thin and radiation is fairly high (over  $800 W.m^{-2}$  for the melting  
484 events in question) can reach the soil below thereby heating it sometimes substantially and  
485 to values above freezing thereby causing melt from below, and ii) when the snow is shallow  
486 the snow fraction tends to be low therefore the non-snow covered fraction of the grid cell  
487 warms considerably owing to the large solar radiation (well above freezing). This leads to  
488 an over-estimation of the soil temperature, that provokes snow melt at the bottom of the  
489 snow pack. The second effect can be argued to be not very physically realistic, but the  
490 goal of this study is not to develop a new physic parameterization but to use the model  
491 as-is. This reveals one the limitations of the sub-grid parameterization of the snow pack in  
492 ISBA when snow is concomitant with very high solar radiation at very high altitudes. Since  
493 ISBA's snow fraction parameterization is quite standard among large scale models, this study  
494 underscores that improvements should be made for the specific geographic context in this  
495 study. Moreover, for ISBA, the infra-daily variations of the air temperature significantly  
496 influence snow-melt (see section 3.3).

### 497 *3.3. Sub-daily variation of snow processes*

498 In order to better understand the difference between the two models for snow processes  
499 representation, the sub-daily variations of snow processes are investigated for the Tauche  
500 catchment. The FIGURE 6 presents the hourly dynamics of input variables of the ISBA and  
501 J2000 models (total precipitation, solid precipitation and air temperature) and the simulated  
502 snow melt in both models, for the two seasons. It can be observed that, at the hourly time  
503 step, the dynamics of solid precipitation reproduce the dynamics of total precipitation for  
504 both seasons. However, the typical sub-daily dynamics for total and solid precipitation and  
505 for snow melt significantly differ between the two seasons. During the summer season, solid  
506 precipitation is maximum at around 5 am (usually before sunrise) and minimal in the after-  
507 noon. This can be explained by the fact that, during the monsoon period, the air humidity  
508 is permanently close to saturation. The limiting factor for water condensation is then the air  
509 temperature. Consequently, precipitation happens when the air temperature is low enough  
510 for the dew point to be reached. In winter, solid precipitation is more important between  
511 18 pm and 3 am than during the day. As during the monsoon period, total precipitation in-  
512 creases around 18 pm when the air temperature starts decreasing. However the air humidity

513 being globally low during winter, the air get quickly too dry for condensation to happens.  
514 Consequently, total (and solid) precipitation remain low during the day. The daily dynamic  
515 of solid precipitation during winter is then not particularly significant.

516

517 The hourly dynamic of the snow melting is a resultant of both the solid precipitation  
518 dynamic and the air temperature dynamic. During the summer season, the snow melting is  
519 driven by both the increasing of the air temperature in the afternoon and the increasing of  
520 solid precipitation during the night. This leads to a bimodal hourly dynamic for the snow  
521 melting simulated in ISBA. However, since these snow melt peaks are not compensated by  
522 a decreasing during the night, the daily average of the snow melting simulated with ISBA is  
523 greater than with J2000. During the winter season, the hourly dynamics of the snow melting  
524 is mainly driven by the air temperature. The snow pack mainly melts during the afternoon  
525 and it is refilled by snow melt happening during the night. This leads to a daily average of  
526 the snow melting simulated with ISBA greater than with J2000 during the winter season.

527

528 This analysis of the sub-daily dynamics of the snow pack simulated at the hourly time  
529 step in ISBA explains the difference between both models for the simulation of the snow  
530 processes and in particular the fact that the snow-melt contribution is more important in  
531 ISBA than in J2000. This kind of analysis based on hourly simulation is actually pretty rare  
532 in the litterature. In particular, the analysis of the sub-daily variation of the air temperature  
533 in a high-elevation himalayan catchment presented by Heynen *et al.* (2016) is consistent with  
534 the behavior presented here.

#### 535 3.4. Comparison with MOD10A2 maximum snow extent

536 The simuled snow cover area (SCA) is computed by applying a threshold condition on the  
537 simulated snow depth (in ISBA) and on the simulated snow water equivalent (in J2000). For  
538 each time step (hourly time step in ISBA, daily time step in J2000), each unit of the model  
539 (grid cell for ISBA, HRU for J2000) is considered as covered by snow if the snow depth is  
540 greater than 60 *mm* (in ISBA) or if the snow water equivalent is greater than 40 *mm* (for  
541 J2000). These values are consistent with values used by Biskop *et al.* (2016) on the Tibetan  
542 Plateau and by Gascoin *et al.* (2015) in the Pyrenees.

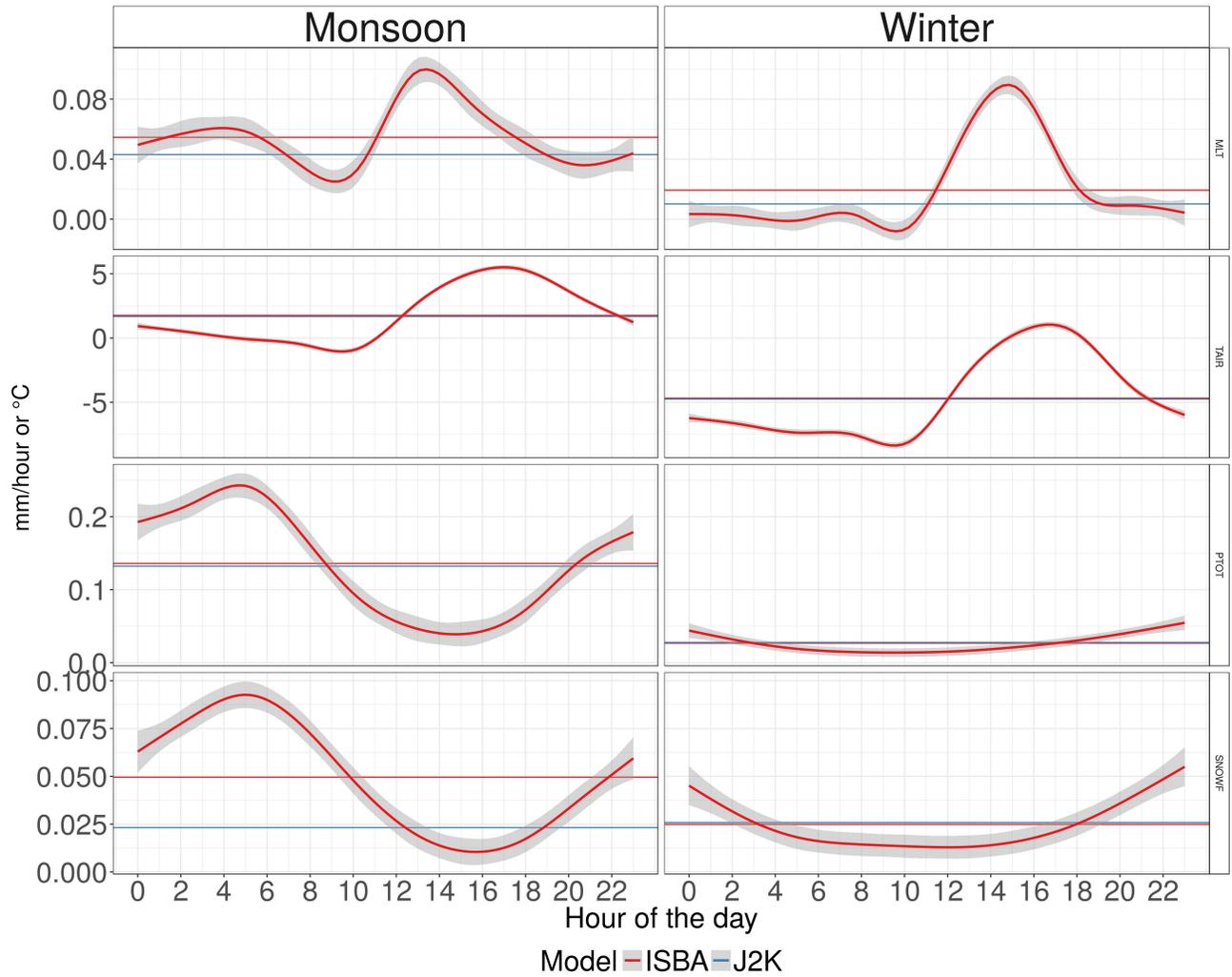


Figure 6: Hourly dynamics of input variables of the ISBA and J2000 models: total precipitation (PTOT), solid precipitation (SNOWF) and air temperature (TAIR) and the simulated snow melt in both models (MLT), for the two seasons, on average over the Tauche catchment. The continuous line represents the average hourly value (in mm per hour) for input variables or for simulated variables in ISBA. The grey interval represent the associated 95% confidence interval. The horizontal line represents the daily means (in mm per hour) of these variables in the two models.

544 Daily simulated SCA and MOD10A2 maximum snow extent is compared in FIGURE 7,  
545 on spatial average over the Tauche catchment, for the 2 hydrological years 2014–2015 and  
546 2015–2016. The overall timing of the MOD10A2 SCA is well reproduced by both models,  
547 with a significant snow period occurred between December 2014 and June 2015 and no  
548 significant snow pack was simulated or observed between July 2015 and March 2016. The  
549 maximum value of SCA (66%) is reached for both model on March, 2nd 2015. This value  
550 and timing is also consistent with the MOD10A2 values. Moreover, this comparison leads  
551 to two main analyzes. First, the short-duration peaks of SWA are better represented with  
552 ISBA than with J2000, despite they remain underestimated by about one-third compared to  
553 the MOD10A2 values. This point highlights the fact that using a hourly time step allows  
554 to better represent infra-daily processes. This processes are further described in section  
555 3.3. Second, snow-melt occurring during the pre-monsoon period (between March 2015 and  
556 May 2015) was faster in J2000 than in ISBA. In particular the SCA simulated with ISBA is  
557 bound around 0.5% over 2 months. The snow pack parametrization in J2000 being calibrated  
558 according to the MOD10A2 SCA and the six snow parameters being particularly adapted  
559 for the Tauche catchment, the simulated SCA is forced to reproduced the MOD10A2 values  
560 in J2000. However, the snow pack accumulation simulated with ISBA appears to over-  
561 estimate the MOD10A2 values during this period. This over-estimation could in particular  
562 be explained by the fact that the land-aspect is not parametrized in ISBA, despite it can  
563 significantly influence the snow pack simulation for such contrasted relief.

#### 564 4. Conclusion

565 This paper aims to assess the impact of using a different degree of refinement to model  
566 hydrological processes at the local scale in sparsely instrumented mountainous catchments.  
567 The methods and results of two approaches that have been previously applied in this re-  
568 gion are compared, namely the work of Eeckman (2017) that uses the ISBA (Interaction  
569 Sol-Biosphere-Atmosphere) surface and the work of Nepal *et al.* (2014) that uses the J2000  
570 distributed hydrological model. The ISBA and J2000 models are applied to two small catch-  
571 ments located in mid- and high mountain environment within the Everest region. In this

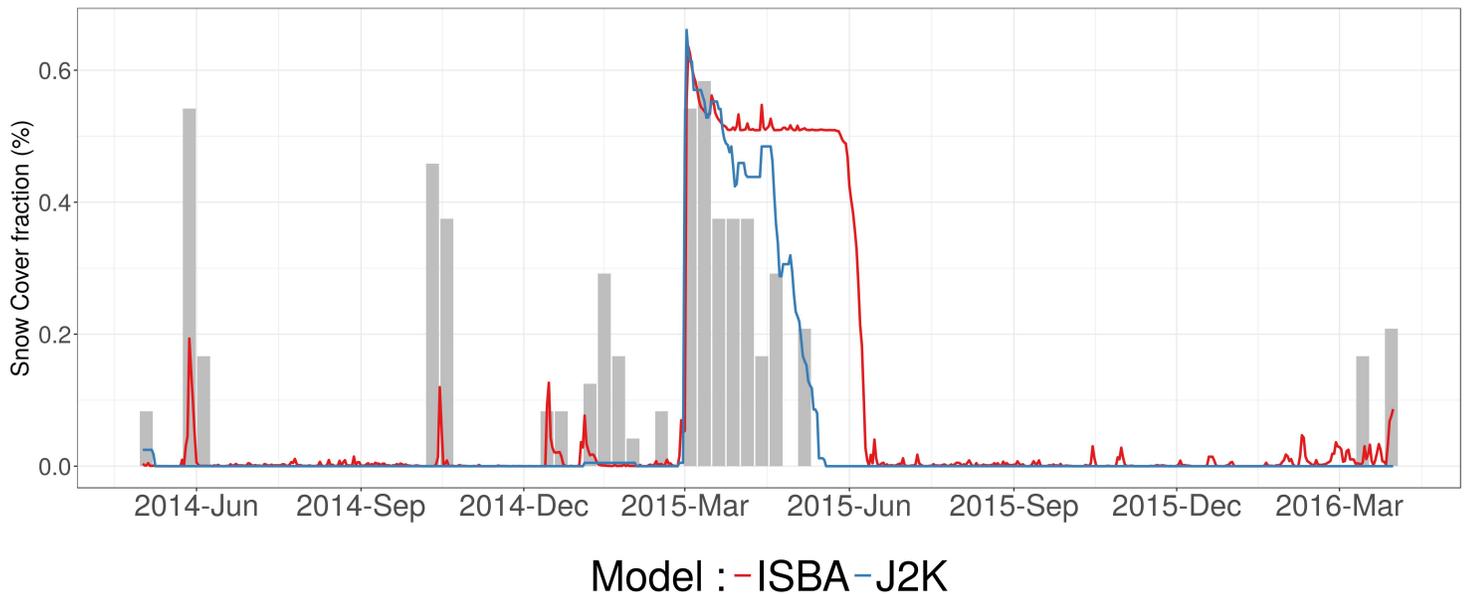


Figure 7: Snow cover area simulated by ISBA and J2000, aggregated at the daily time scale, on average over the Tauche catchment, for the 2014–2015 and 2015–2016 hydrological years. Grey bars are MOD10A2 maximal snow cover extend, on average over the Tauche catchment, at a 8 days time scale.

572 framework, several points should be underlined:

- 573 1. Since conceptual models rely more on calibration data, the reliability of a calibrated  
574 approach was tested by comparing it to an approach based on energy balance solv-  
575 ing in an environment where data quality and quantity is relatively low. One of the  
576 main results of this study is that both models overallly represent the dynamics of the  
577 processes for evaporation, quick runoff and discharge in a similar way. Estimations of  
578 annual volumes for these components of the water budget are provided and they can  
579 be considered as a contribution to the current knowledge of the hydro-systems in this  
580 region.
- 581 2. An interesting period of discrepancy between the two models is found for the high  
582 mountain catchment at the beginning of the monsoon 2015. In this case, the snow-pack  
583 simulation is shown to be the main driver of the discrepancy between the two models.  
584 This work also analyses in details the sub-daily variation of snow processes. The time  
585 step used in the model (daily or hourly time step) is shown to strongly influence the  
586 precipitation phase partition and consequently the snow-melt contribution to discharge,  
587 in particular during the summer season. This work leads then to suggest for upcoming  
588 researches that modelling approaches should be set up at a sub-daily time step in order  
589 to represent the significant sub-daily variations of snow-melt processes.

- 590 3. The contribution of drainage in the soil are shown to be more important than the runoff  
591 contribution to discharges for both models and for the two catchments. This result is  
592 consistent with the physical descriptions of soils based on in-situ measurements, that  
593 describe fast-infiltrating, mostly sandy soils. However, this work exposes that the  
594 structural hypotheses made in the two models significantly influence the simulation of  
595 water storages and flows in the soil, in particular for the high mountain catchment.  
596 This point illustrates the fact that, since the conceptualization of the soil processes  
597 differs in the different studies in the litterature, the estimations of these processes are  
598 associated with important uncertainties due to the model structure.
- 599 4. Finally, this model comparison work performed at a local scale allows to analyse the  
600 local accuracy of the distributed modelling approach proposed by Nepal *et al.* (2011),  
601 that is available at a larger scale over the Dudh Koshi basin. This work applies then  
602 the idea that, in a context where a very few data is available for the estimation of  
603 the performances of the simulations, model inter-comparison can then be used as an  
604 alternative way to estimate simulation robustness.

605 Based on this research, both models present equivalent results for discharge and evapo-  
606 transpiration simulation in middle- and high-mountain environments. They could be used  
607 for operational purposes in two complementary ways: (i) the assessment of water availability  
608 considering new scenarios of climate forcing or land use and land cover change and (ii) the  
609 sizing of hydraulic installations for agriculture, domestic water supply or hydropower, on the  
610 request of the local water users.

## 611 **Appendix A. The HDSM routing module coupled to ISBA**

612 The routing module coupled to the ISBA surface scheme is taken from the HDSM hy-  
613 drological model (Delclaux *et al.*, 2008). The transfer function of HDSM is derived from the  
614 THMB model (Coe, 2000). The HDSM model has already been applied for various studies,  
615 but this study is the first application of the routing module coupled to the ISBA surface  
616 scheme. The HDSM model has been applied by Savean, 2015 on the Dudh Koshi basin and  
617 showed good performances for discharge and snow cover area modelling.  
618

620 For each cell, surface runoff ( $R_s$ ), given by the sum of Dunne runoff and Horton runoff,  
 621 and the drainage at the bottom of the soil column ( $R_d$ ) are directed toward two simple linear  
 622 reservoirs,  $W_s$  and  $W_d$  respectively.  $W_s$  and  $W_d$  are characterized by a residence time ( $T_s$  and  
 623  $T_d$  (s), resp.) and a volume ( $V_s$  and  $V_d$  ( $m^3$ ), resp.).  $W_s$  and  $W_d$  are governed by EQUATION  
 624 A.1.

$$\begin{cases} \frac{dV_s}{dt} = R_s - \frac{V_s}{T_s} \\ \frac{dV_d}{dt} = R_d - \frac{V_d}{T_d} \end{cases} \quad (\text{A.1})$$

625 At each time step, the sum of the outflows from  $W_s$  and  $W_d$  is directed toward a linear  
 626 routing reservoir  $W_r$ .  $W_r$  is characterized by a volume  $V_r(m^3)$  and a residence time  $T_r$ (s).  
 627 The outflow of  $W_r$  ( $R_{out}$ , in  $m^3/s$ ) is computed as in EQUATION A.2.

$$R_{out} = \frac{V_r}{T_r} \quad (\text{A.2})$$

628 The residence time in the transfer reservoir is defined for each mesh point as the ratio between  
 629 the flow velocity ( $u$ , in  $m.s^{-1}$ ) and the distance from the center of the mesh point to the  
 630 center of the previous upstream mesh point ( $d$ , in m). The flow velocity is calculated as  
 631 the ratio of the mesh point slope ( $i_c$  in  $m.m^{-1}$ ) and a reference slope ( $i_0$  in  $m.m^{-1}$ ), taken  
 632 equal to the catchment median slope. This ratio is weighted by a  $c_{vel}$  transfer coefficient (see  
 633 EQUATION A.3).  $c_{vel}$  is calibrated as a uniform parameter.

$$\begin{cases} T_r = \max\left(\frac{d}{u}, \Delta t\right) \\ u = c_{vel} \cdot \sqrt{\frac{i_c}{i_0}} \end{cases} \quad (\text{A.3})$$

634 This coupling then requires the calibration of the parameters  $T_r$ ,  $T_d$  and  $c_{vel}$ . These  
 635 parameters are calibrated against the observed discharges at the outlets, according to the  
 636 NSE criteria, the bias computed on daily discharges and to the  $NSE$  criteria computed on  
 637 the square root of the daily discharge ( $NSE_{sqrt}$ ). Optimum parameter sets are computed  
 638 using the Pareto's optimum method. Initial ranges of parameter values for calibration are  
 639 taken from Savéan (2014). Considering the few available discharge measurements, the entire  
 640 observation period is used for calibration.

642 The calibration is run out independently for the Kharikhola and Tauche catchments. The  
643 optimal parameter set for each basin, as well as the calibrated values for the Dudh Koshi  
644 basin by Savéan *et al.* (2015), are presented in TABLE A.7. The performances according  
645 to the three criterias are satisfactory for both basins. Low discharges are better simulated  
646 for the Kharikhola basins ( $NSE_{sqrt} = 0.80$ ) than for the Tauche Basin ( $NSE_{sqrt} = 0.77$ ).  
647 Residence times  $t_s$  and  $t_d$  are shorter for the Tauche Basin than for the Kharikhola Basin.  
648 Residence times calibrated by Savéan *et al.* (2015) for the Dudh Koshi Basin are shorter  
649 than for the two sub-basins. In addition, the *cvel* transfer coefficient is significantly higher  
650 for the Dudh Koshi Basin than for the two sub-basins. More physical interpretations for  
651 these calibrated values can be found in Eeckman (2017).

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## 673 **References**

674 Andermann, C., *et al.*, 2012. Impact of transient groundwater storage on the discharge of  
675 himalayan rivers. *Nature Geoscience*, 5 (2), 127–132.

676 Anders, A.M., *et al.*, 2006. Spatial patterns of precipitation and topography in the himalaya.  
677 *Geological Society of America Special Papers*, 398, 39–53.

678 Barros, A., *et al.*, 2004. Probing orographic controls in the himalayas during the monsoon  
679 using satellite imagery. *Natural Hazards and Earth System Science*, 4 (1), 29–51.

680 Bharati, L., *et al.*, 2016. Past and future variability in the hydrological regime of the koshi  
681 basin, nepal. *Hydrological Sciences Journal*, 61 (1), 79–93.

682 Biskop, S., *et al.*, 2016. Differences in the water-balance components of four lakes in the  
683 southern-central tibetan plateau. *Hydrology and Earth System Sciences*, 20 (1), 209–225.

684 Bookhagen, B. and Burbank, D.W., 2006. Topography, relief, and trmm-derived rain-  
685 fall variations along the himalaya. *Geophysical Research Letters*, 33 (8). Available from:  
686 <http://dx.doi.org/10.1029/2006GL026037>.

687 Boone, A., *et al.*, 2000. The influence of the inclusion of soil freezing on simulations by a soil-  
688 vegetation-atmosphere transfer scheme. *JOURNAL OF APPLIED METEOROLOGY*, 39  
689 (9), 1544–1569.

690 Boone, A. and Etchevers, P., 2001. An intercomparison of three snow schemes of  
691 varying complexity coupled to the same land surface model: Local-scale evalua-  
692 tion at an alpine site. *Journal of Hydrometeorology*, 2 (4), 374–394. Available from:  
693 [http://dx.doi.org/10.1175/1525-7541\(2001\)002j0374:AIOTSSj2.0.CO;2](http://dx.doi.org/10.1175/1525-7541(2001)002j0374:AIOTSSj2.0.CO;2).

- 694 Chelamallu, H.P., Venkataraman, G., and Murti, M., 2014. Accuracy assessment of  
695 modis/terra snow cover product for parts of indian himalayas. *Geocarto International*,  
696 29 (6), 592–608.
- 697 Clapp, R.B. and Hornberger, G.M., 1978. Empirical equations for some soil hy-  
698 draulic properties. *Water Resources Research*, 14 (4), 601–604. Available from:  
699 <http://dx.doi.org/10.1029/WR014i004p00601>.
- 700 Coe, M.T., 2000. Modeling terrestrial hydrological systems at the continental scale: Testing  
701 the accuracy of an atmospheric gcm. *Journal of Climate*, 13 (4), 686–704.
- 702 Cosgrove, B., *et al.*, 2003. Real-time and retrospective forcing in the North American  
703 Land Data Assimilation System (NLDAS) project. *JOURNAL OF GEOPHYSICAL*  
704 *RESEARCH-ATMOSPHERES*, 108 (D22).
- 705 Decharme, B., *et al.*, 2011. Local evaluation of the interaction between soil biosphere at-  
706 mosphere soil multilayer diffusion scheme using four pedotransfer functions. *Journal of*  
707 *Geophysical Research: Atmospheres*, 116 (D20).
- 708 Decharme, B., *et al.*, 2016. Impacts of snow and organic soils parameterization on north-  
709 ern eurasian soil temperature profiles simulated by the isba land surface model. *The*  
710 *Cryosphere*, 10 (2), 853–877.
- 711 Delclaux, F., *et al.*, 2008. Confronting models with observations for evaluating hydrological  
712 change in the lake chad basin, africa. *In: XIIIth World Water Congress*.
- 713 Dhar, O. and Rakhecha, P., 1981. The effect of elevation on monsoon rainfall distribution in  
714 the central himalayas. *Monsoon Dynamics*, 253–260.
- 715 Dickinson, R.E., 1984. Modelling evapotranspiration for three-dimensional global climate  
716 models. *In: Climate Processes and Climate Sensitivity Geophysical Monograph, Hansen,*  
717 *J. E. Takahasi, T. (Eds.), Series 29, Washington*.
- 718 Drusch, M., *et al.*, 2012. Sentinel-2: Esa’s optical high-resolution mission for gmes operational  
719 services. *Remote Sensing of Environment*, 120, 25–36.

- 720 Dümenil, L. and Todini, E., 1992. A rainfall-runoff scheme for use in the hamburg climate  
721 model. *In: Advances in theoretical hydrology: a tribute to james dooge*. Elsevier Science  
722 Publishers BV, 129–157.
- 723 Dunne, T., 1983. Relation of field studies and modeling in the prediction of storm runoff.  
724 *Journal of Hydrology*, 65 (1-3), 25–48.
- 725 Eeckman, J., *et al.*, 2017. Providing a non-deterministic representation of spatial variability  
726 of precipitation in the Everest region. *Hydrology and Earth System Sciences Discussions*,  
727 2017, 1–21. Available from: <http://www.hydrol-earth-syst-sci-discuss.net/hess-2017-137/>.
- 728 Eeckman, J., 2017. *Caractérisation des systèmes hydro-climatiques à l'échelle locale dans*  
729 *l'himalaya népalais*. Thesis (PhD). Université Montpellier.
- 730 Gardelle, J., Berthier, E., and Arnaud, Y., 2012. Impact of resolution and radar penetration  
731 on glacier elevation changes computed from dem differencing. *Journal of Glaciology*, 58  
732 (208), 419–422.
- 733 Gascoin, S., *et al.*, 2015. A snow cover climatology for the pyrenees from modis snow products.  
734 *Hydrology and Earth System Sciences*.
- 735 Habets, F., *et al.*, 1999. The {ISBA} surface scheme in a macroscale hydrological  
736 model applied to the hapex-mobilhy area: Part ii: Simulation of streamflows and  
737 annual water budget. *Journal of Hydrology*, 217 (12), 97 – 118. Available from:  
738 <http://www.sciencedirect.com/science/article/pii/S0022169499000207>.
- 739 Hall, D.K., *et al.*, 2002. Modis snow-cover products. *Remote Sensing of Environ-*  
740 *ment*, 83 (12), 181 – 194. The Moderate Resolution Imaging Spectroradiome-  
741 ter (MODIS): a new generation of Land Surface Monitoring, Available from:  
742 <http://www.sciencedirect.com/science/article/pii/S0034425702000950>.
- 743 Hargreaves, G.H. and Samani, Z.A., 1982. Estimating potential evapotranspiration. *Journal*  
744 *of the Irrigation and Drainage Division*, 108 (3), 225–230.
- 745 Heynen, M., *et al.*, 2016. Air temperature variability in a high-elevation himalayan catchment.  
746 *Annals of Glaciology*, 57 (71), 212–222.

- 747 Horton, R.E., 1933. The role of infiltration in the hydrologic cycle. *Eos, Transactions Amer-*  
748 *ican Geophysical Union*, 14 (1), 446–460.
- 749 Immerzeel, W.W., Van Beek, L.P., and Bierkens, M.F., 2010. Climate change will affect the  
750 asian water towers. *Science*, 328 (5984), 1382–1385.
- 751 Jain, S.K., Goswami, A., and Saraf, A., 2008. Accuracy assessment of modis, noaa and irs  
752 data in snow cover mapping under himalayan conditions. *International Journal of Remote*  
753 *Sensing*, 29 (20), 5863–5878.
- 754 Knauf, D., 1980. *Die Berechnung des Abflusses aus einer Schneedecke*. Analyse und Berech-  
755 nung oberirdischer Abflüsse DVWK- Schriften, Bonn, Heft 46.
- 756 Kralisch, S. and Krause, P., 2006. JAMS A Framework for Natural Resource Model De-  
757 velopment and Application. In: *Proceedings of the International Environmental Software*  
758 *Society (IEMSS), Vermont, USA*.
- 759 Kralisch, S., *et al.*, 2007. Component based environmental modelling using the JAMS frame-  
760 work. In: *MODSIM 2007 International Congress on Modelling and Simulation*. 812–818.  
761 Peer reviewed.
- 762 Krause, P., 2001. *Das hydrologische Modellsystem J2000: Beschreibung und An-*  
763 *wendung in groen Flueinzugsgebieten, Schriften des Forschungszentrum Jlich*. Reihe  
764 Umwelt/Environment; Band 29.
- 765 Krause, P., 2002. Quantifying the Impact of Land Use Changes on the Water Balance of Large  
766 Catchments using the J2000 Model. *Physics and Chemistry of the Earth*, 27, 663–673.
- 767 Lang, T.J. and Barros, A.P., 2004. Winter storms in the central himalayas. *Journal of Me-*  
768 *teorological Society of Japan*, 82 (3), 829–844.
- 769 Li, H., Haugen, J.E., and Xu, C., 2017. Precipitation Pattern in the Western Himalayas  
770 revealed by Four Datasets. *Hydrology and Earth System Sciences Discussions*, 2017, 1–19.  
771 Available from: <http://www.hydro-earth-syst-sci-discuss.net/hess-2017-296/>.
- 772 Lutz, A., *et al.*, 2014. Consistent increase in high asia’s runoff due to increasing glacier melt  
773 and precipitation. *Nature Climate Change*, 4 (7), 587–592.

- 774 Masson, V., *et al.*, 2013. The SURFEXv7.2 land and ocean surface platform for coupled  
775 or offline simulation of earth surface variables and fluxes. *GEOSCIENTIFIC MODEL*  
776 *DEVELOPMENT*, 6 (4), 929–960.
- 777 Masson, V., *et al.*, 2003. A global database of land surface parameters at 1-km resolution in  
778 meteorological and climate models. *Journal of climate*, 16 (9), 1261–1282.
- 779 Nepal, S., *et al.*, 2014. Understanding the hydrological system dynamics of a glaciated alpine  
780 catchment in the Himalayan region using the J2000 hydrological model. *Hydrological Pro-*  
781 *cesses*, 28 (3), 1329–1344.
- 782 Nepal, S., *et al.*, 2011. Understanding the impact of climate change in the glaciated alpine  
783 catchment of the Himalaya Region using the J2000 hydrological model. *In: Proceedings*  
784 *of the Second International Symposium on Building Knowledge Bridges for a Sustainable*  
785 *Water Future, Panama, 2011.* 55–60.
- 786 Nepal, S., 2012. *Evaluating upstream-downstream linkages of hydrological dynamics in the*  
787 *himalayan region.* Thesis (PhD). PhD Thesis. Friedrich Schiller University, Germany.
- 788 Nepal, S., *et al.*, 2017a. Spatial gr4j conceptualization of the tamor glaciated alpine catch-  
789 ment in eastern nepal: evaluation of gr4jsg against streamflow and modis snow extent.  
790 *Hydrological Processes*, 31 (1), 51–68.
- 791 Nepal, S., *et al.*, 2017b. Assessment of spatial transferability of process-based hydrological  
792 model parameters in two neighbouring catchments in the himalayan region. *Hydrological*  
793 *Processes*, 31 (16), 2812–2826.
- 794 Nepal, S., Flügel, W.A., and Shrestha, A.B., 2014. Upstream-downstream linkages of hydro-  
795 logical processes in the himalayan region. *Ecological Processes*, 3 (1), 1.
- 796 Noilhan, J. and Mahfouf, J.F., 1996. The isba land surface parameterisation scheme. *Global*  
797 *and planetary Change*, 13 (1), 145–159.
- 798 Noilhan, J. and Planton, S., 1989. A Simple Parameterization of Land Surface Processes  
799 for Meteorological Models. *Monthly Weather Review*, 117 (3), 536–549. Available from:  
800 [http://dx.doi.org/10.1175/1520-0493\(1989\)117;0536:ASPOLS;2.0.CO;2](http://dx.doi.org/10.1175/1520-0493(1989)117;0536:ASPOLS;2.0.CO;2).

- 801 Panday, P.K., *et al.*, 2014. Application and evaluation of a snowmelt runoff model in the tamor  
802 river basin, eastern himalaya using a markov chain monte carlo (mcmc) data assimilation  
803 approach. *Hydrological Processes*, 28 (21), 5337–5353.
- 804 Pellicciotti, F., *et al.*, 2012. Challenges and uncertainties in hydrological modeling of re-  
805 mote hindu kush-karakoram-himalayan (hkh) basins: suggestions for calibration strategies.  
806 *Mountain Research and Development*, 32 (1), 39–50.
- 807 Pokhrel, B.K., *et al.*, 2014. Comparison of two snowmelt modelling approaches in the dudh  
808 koshi basin (eastern himalayas, nepal). *Hydrological Sciences Journal*, 59 (8), 1507–1518.
- 809 Racoviteanu, A.E., Armstrong, R., and Williams, M.W., 2013. Evaluation of an ice ablation  
810 model to estimate the contribution of melting glacier ice to annual discharge in the nepal  
811 himalaya. *Water Resources Research*, 49 (9), 5117–5133.
- 812 Salerno, F., *et al.*, 2015. Weak precipitation, warm winters and springs impact glaciers of  
813 south slopes of mt. everest (central himalaya) in the last 2 decades (1994–2013). *The*  
814 *Cryosphere*, 9 (3), 1229–1247.
- 815 Savéan, M., 2014. *Modélisation hydrologique distribuée et perception de la variabilité hydro-*  
816 *climatique par la population du bassin versant de la dudh koshi (népal)*. Thesis (PhD).  
817 Université de Montpellier 2.
- 818 Savéan, M., *et al.*, 2015. Water budget on the Dudh Koshi River (Nepal):  
819 Uncertainties on precipitation. *Journal of Hydrology*. Available from:  
820 <http://linkinghub.elsevier.com/retrieve/pii/S0022169415008082>.
- 821 Sevruk, B., Ondrás, M., and Chvíla, B., 2009. The wmo precipitation measurement inter-  
822 comparisons. *Atmospheric Research*, 92 (3), 376–380.
- 823 Shrestha, M., *et al.*, 2011. Modeling the Spatial Distribution of Snow Cover in the Dudhkoshi  
824 Region of the Nepal Himalayas. *Journal of Hydrometeorology*, 13 (1), 204–222. Available  
825 from: <https://doi.org/10.1175/JHM-D-10-05027.1>.
- 826 Valery, A., Andreassian, V., and Perrin, C., 2010. Regionalization of precip-  
827 itation and air temperature over high-altitude catchments learning from

828 outliers. *Hydrological Sciences Journal*, 55 (6), 928–940. Available from:  
829 <http://dx.doi.org/10.1080/02626667.2010.504676>.

830 Zhang, Y., Liu, S., and Ding, Y., 2007. Glacier meltwater and runoff modelling, keqicar baqi  
831 glacier, southwestern tien shan, china. *Journal of Glaciology*, 53 (180), 91–98.

Table 2: Summary of ISBA surface scheme and J2000 model structures, for precipitation phase distribution, interception, evapotranspiration, snow accumulation and melt, soil water, runoff components, groundwater and flow routing treatments.

ISBA	J2000
<b>Precipitation</b>	
<i>For both models:</i> Precipitation is distributed between rain and snow according to the same threshold temperatures for both models.	
<b>Interception</b>	
<i>For both models:</i> Simple interception storage approach (Dickinson, 1984). The interception storage is computed according to the vegetation type defined by its Leaf Area Index (LAI) for rain and snow.	
<b>Evapotranspiration (ET)</b>	
ET results from the water and energy balance applied on bare soil, vegetation and snow-cover (Noilhan and Planton, 1989).	The potential ET is calculated by Hargreaves and Samani (1982) and is then checked against actual water storage in different landscape compartments (such as interception, soil water etc) to calculate actual ET.
<b>Snow accumulation and melt</b>	
The ISBA-ES implementation (Boone and Etchevers, 2001; Decharme <i>et al.</i> , 2016) provides a twelve-layer discretization of the snow pack. Mass and energy balances are computed for each layer, considering snow-melt and sublimation.	Potential melt from snow pack is estimated with energy input from temperature, rain and ground surface. Accumulation and melting can occur within a time step, controlled by separate accumulation or melt temperatures (Knauf, 1980).
<b>Soil water</b>	
The diffusive approach (ISBA-DIF), (Boone <i>et al.</i> , 2000; Decharme <i>et al.</i> , 2011) uses a 14 layer discretization of the mixed-form richard's equation with vertical soil water fluxes represented by Darcy's law.	Middle/large pore storage (MPS/LPS) partition. MPS refers to the field capacity, whereas LPS refers to the flowing water in the soil that generates subsurface runoff and percolation to groundwater reservoirs.
<b>Runoff components</b>	
<i>For both models:</i> The notions of Dunne's flow (saturation excess runoff) and Horton's flow (infiltration excess runoff) are considered in the computation of surface runoff.	
Dunne's and Horton's runoffs are controlled according to (Dümenil and Todini, 1992). The Dunne runoff for each grid cell depends on the fraction of the cell that is saturated.	Saturation excess runoff and infiltration excess runoff together provide overland flow (RD1) (Krause, 2001, 2002). When LPS is filled, the excess water is divided into sub-surface flow (RD2) and percolation to the groundwater reservoir.
<b>Groundwater</b>	
Groundwater storage is treated by an additional conceptual module. Drainage at the bottom of the soil column is stored in a linear reservoir ( $R_d$ ), controlled by a calibrated residence time ( $t_d$ ).	The percolated water is distributed into two groundwater compartments, which produce interflow 2 (RG1) from shallow aquifers and baseflow (RG2) from deep aquifers.
<b>Routing</b>	
Flow routing is treated by an additional conceptual module. The outflow is computed for each grid cell according to the average slope of the cell, weighted by a calibrated velocity coefficient.	The four different runoff components (RD1, RD2, RG1 and RG2) from each HRU are routed to the next connected HRU until it reaches a river network, using a simplified kinematic wave approach (Krause, 2001).

Table 3: Summary of the spatial discretization methods used in ISBA and in J2000, for the Kharikhola and Tauche catchments.

	Kharikhola catchment		Tauche catchment		
	ISBA	J2000	ISBA	J2000	
Number of units	69 cells	346 HRUs	28 cells	132 HRUs	
Minimum size of units	0.16	0.008	0.16	0.008	$km^2$
Minimum altitude	2050	1997	4070	4021	m.a.s.l.
Maximum altitude	4326	4459	5600	5457	m.a.s.l.

Table 4: Soil and vegetation characteristics of the nine classes defined in Kharikhola and Tauche catchments, respectively. % KK and % Tauche are the fraction of each class on Kharikhola and Tauche catchments. Sand and clay fractions (% Sand and % Clay, respectively), soil depth (SD), root depth (RD) and tree height (TH) are defined based on in situ measurements. The dynamic variables (e.g. the fraction of vegetation and Leaf Area Index) were found in the ECOCLIMAP1 classification (Masson *et al.*, 2003) for representative ecosystems.

ID	Class	% KK	% Tauche	% Sand	% Clay	TH m	SD m	RD m	ECOCLIMAP1 Cover
1	Snow and ice	-	0.7%	0.00	0.00	0.0	0.00	0.00	6
2	Screes	3.1%	31.2%	0.00	0.00	0.0	0.00	0.00	5
3	Steppe	0.6%	33.7%	81.41	1.70	0.0	0.10	0.10	123
4	Shrubs	7.4%	34.4%	70.60	1.55	0.0	0.35	0.27	86
5	Dry Forest	9.7%	-	72.86	1.00	12.0	0.20	0.20	27
6	Intermediary Forest	45.7%	-	84.97	1.01	27.5	0.42	0.40	27
7	Wet Forest	20.6%	-	70.12	1.00	6.8	1.04	0.50	27
8	Slope terraces	11.2%	-	70.89	1.38	5.6	0.56	0.26	171
9	Flat terraces	1.4%	-	67.01	1.69	2.5	1.267	0.20	171

Table 5: Annual volumes for input variables (in millimetres per year): total precipitation, solid precipitation and for variables simulated by ISBA and J2000 models: actual evapotranspiration, discharge at the outlet, snow-melt contribution, snow pack storage variation and soil storage variation, for the 2014–2015 and 2015–2016 hydrological years, for the Kharikhola and Tauche catchments. Performance criteria ( Nash-Sutcliffe Efficiency  $NSE$ , relative bias  $Bias_r$ , determination of coefficient  $r^2$ ,  $NSE$  for the square root of discharges  $NSE_{sqr}$  and  $NSE$  computed for the high-flow period  $NSE_{high}$ ), computed at the daily time scale are also provided.

Observed discharges	Kharikhola catchment				Tauche catchment			
	2014-2015		2015-2016		2014-2015		2015-2016	
	-		1800		440		477	
Model	ISBA	J2000	ISBA	J2000	ISBA	J2000	ISBA	J2000
Total precipitation	3034	3064	2256	2254	837	824	581	607
Solid precipitation	42	36	27	26	403	281	245	148
Actual evapotranspiration	579	548	622	555	292	372	285	363
Discharge at the outlet	2346	2523	1631	1803	373	413	385	303
Snow-melt contribution	53	50	27	21	336	276	309	199
Snow pack storage variation	0	0	0	0	-66	-26	66	24
Soil storage variation	-41	-16	33	15	-8	-7	7	6
$NSE$	0.5018	0.60453	0.9010	0.9158	0.8958	0.9194	0.6760	0.5172
$Bias_r$	-45.7	-37.1	-9.7	0.04	-11.7	-2.8	-19.5	-39.0
$r^2$	0.8613	0.9049	0.9120	0.9327	0.9352	0.9453	0.7203	0.7944
$NSE_{sqr}$	0.6645	0.6985	0.8733	0.9395	0.8553	0.8219	0.6888	0.6956
$NSE_{high}$	0.0742	0.1512	0.7629	0.6640	0.7329	0.8400	0.0193	-0.7239

Table 6: Parametrisation of soil water content introduced in ISBA and in J2000.  $SAND$  and  $CLAY$  are respectively the average sand and clay fractions of the soil for each catchment.  $w_{sat}$  is the water content of the soil column at saturation computed in ISBA.  $maxMPS$  and  $maxLPS$  are the maximal storage capacity in MPS and LPS reservoirs in J2000.

Catchment	SAND	CLAY	$w_{sat}$ mm	$maxMPS$ mm	$maxLPS$ mm
Kharikhola	79.9%	1.1%	207	98	100
Tauche	80.9%	1.7%	52	35	32

Table A.7: Results of the calibration of the HDSM routing module for the Kharikhola and the Tauche catchments, as well as the calibrated values for the Dudh Koshi basin by Savéan *et al.* (2015).

	$NSE$	$Bias_r$	$NSE_{sqr}$	$T_d$ days	$T_s$ hours	$cvel$ m/s
Kharikhola	0.6906	0.2906	0.8051	39.5	0.75	0.0948
Tauche	0.742	0.0244	0.7775	48.1	1.77	0.0162
Dudh Koshi, calibration from (Savéan <i>et al.</i> , 2015)	0.73	-0.5	-	12	1.5	1.7