More green and less blue water in the Alps during warmer summers

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Climate change can reduce surface-water supply by enhancing evapotranspiration in forested mountains, especially during heatwaves. We investigate this "drought paradox" for the European Alps using a 1212-station database and hyper-resolution ecohydrological simulations to quantify blue (runoff) and green (evapotranspiration) water fluxes. During the 2003 heatwave, evapotranspiration in large areas over the Alps was above average despite low precipitation, amplifying the runoff deficit by 32% in the most runoff-productive areas (1300-3000 m above sea level). A 3 °C air temperature increase could enhance annual evapotranspiration by up to 100 mm (45 mm on average), which would reduce annual runoff at a rate similar to a 3% precipitation decrease. This

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suggests that green water feedbacks--often poorly represented in large-scale model simulations--pose an additional threat to water resources, especially in dry summers. Despite uncertainty in the validation of the hyper-resolution ecohydrological modelling with observations, this approach permits more realistic predictions of mountain region water availability.

Although relatively small in size, the European Alps (hereafter "Alps") contribute a disproportionally large amount of water, especially during summer, to four major European rivers[1], in the basins of which reside more than 170 million people[2]. For this reason, they are referred to as "the water towers of Europe"[3]. At the same time, water scarcity and droughts in central Europe are becoming more frequent[4]. The summer droughts of 2003, 2010, 2015 and 2018 have raised concerns about the vulnerability of the European water budget to climate change[2,5] as these events have affected more than 17% of the European population with an annual economic impact exceeding € 6.2 billion between 2001 and 2006[6]. Temperature in the Alps is increasing at a fast pace[7], relative humidity is generally decreasing[8], evapotranspiration (ET) is increasing[9], Alpine glaciers are shrinking and the distribution of snow is shifting to higher elevation[10], while climatic extremes are becoming more frequent[11]. The complex topography, the interactions between water and vegetation and the multiple processes shaping the water cycle in mountainous areas hinder the quantification of the different water budget components in traditional large-scale climate change impact assessment studies[12]. For example, climate change can shift the partitioning of water fluxes in the hydrosphere and biosphere moving blue water (runoff and streamflow) into green water (ET)[13,14]. Quantifying how these fluxes change with elevation, seasonally, and interannually is an important and challenging scientific question.

Large uncertainties are associated with the vegetation response to water stress[15,16]. Studies in different parts of the Alps have found contrasting impacts of droughts on vegetation[17], spanning from increased mortality in dry inner-Alpine valleys[18] to enhanced productivity in wet pre-Alpine hills[19]. These discrepancies emphasize that extrapolating results of specific case studies[20] to the entire Alpine domain or downscaling results of coarse resolution modelling studies[14] is problematic[21]. The largest component of land ET is plant transpiration[22], which is poorly quantified due to the large variability in plant water use strategies and stomata sensitivity to water stress[23,24]. Land-surface models often represent the links between soil moisture and transpiration in a simplistic way, without mechanistic representation of the soil-plant-atmosphere continuum[5,25,26]. Most importantly, they do not resolve land-surface energy and water fluxes at sufficient spatial resolution to capture local topographic and microclimatic effects, and often ignore lateral flows of water[12,27–29].

Here, to overcome these limitations, we combined a new pan-Alpine database with hyper-resolution ecohydrological simulations to test the "drought paradox" hypothesis, i.e. that during droughts ET may increase at high elevations in a large part of the Alps, thus amplifying the runoff deficit. Plot-scale observations provided evidence of the drought paradox occurrence[17,30], but the extent and implications of ET changes during heatwaves at the pan-Alpine scale remain unexplored. We compiled a new dataset comprising meteorological, discharge, and snow depth measurements at 1212 stations and combined it with distributed hydrometeorological products to generate high-resolution forcing (precipitation (P), air temperature, solar radiation, wind speed, and relative humidity; see Methods) to drive and validate the ecohydrological model Tethys-Chloris (T&C)[21]. The model resolves the water, carbon and energy budgets at the hourly time scale in a physically based and spatially explicit manner, accounting for lateral water transfer and topographic effects on radiation. The model has been extensively validated in many regions worldwide[21,22,31], including Alpine ecosystems[32]. To account for the high spatial heterogeneity of the region, we performed massively parallel simulations ($6.1*10^5$ CPU hours) at an unprecedented high resolution (250 m grid) for the entire Alpine arch (257,045 km^2, 4.12 million pixels, Supplementary Fig. 1).

The simulation period consists of three hydrological years (2001-2003, Fig. 1 and Supplementary Tables 1-4), including a very wet and a very dry year (2001 and 2003, respectively, Supplementary Fig. 2). We used these simulations for partitioning the pan-Alpine water budget into blue and green water fluxes and we quantified the sensitivity of each component to changes in precipitation and temperature. We validated the model output with daily discharge and snow depth measurements, as well as with CO_2, water, and energy fluxes observed by eddy covariance flux towers located in the Alps or nearby (see Methods and Supplementary Table 5). This validation yielded very satisfactory results, considering that model parameterization was only based on literature values and calibration with previous plot-scale model applications in the Alpine region (see Methods, Extended Data Figure 1 and 2, and Supplementary Figs 3, 4, and 5). We further compared the simulated distributions of ET and gross primary production (GPP) over the entire domain with other products (TRENDY[33,34], FLUXCOM[35] and ERA Interim[36]). Averaged in space, these products provide estimates comparable to the T&C simulations (Extended Data Figure 3, Supplementary Figs 6, see Methods). However, due to their coarse resolution compared to our analysis, they largely underestimate the spatial variability of ET and GPP and they provide spatial patterns that are poorly resolved. This reinforces the need for hyper-resolution simulations in complex terrain.

Figure 1 | Simulation results highlight the spatial heterogeneity in latent heat (evapotranspiration in energy units). **a**, The spatial extent of the European Alps. **b**, November 2000-October 2003 average latent heat flux (W m^{-2}) for the entire 257,045 km^2 domain simulated with Tethys-Chloris. **c-d** zoom on the Bernese highlands, Switzerland, and illustration of the small-scale spatial heterogeneity captured with the hyper-resolution simulation (250 m x 250 m pixels).

Dissecting the water towers of Europe

Simulation results indicate that latent heat, i.e. total ET expressed in units of equivalent energy, reaches its maximum values in wetter areas (e.g., in the north in Fig. 1c, where annual precipitation exceeds 2000 mm), especially on south-facing slopes (Fig. 1d) confirming that energy is the dominant driver of ET in this area. In drier regions, such as in the SE valley in Fig. 1c (upper Rhone valley), latent heat is overall lower because P (~500 mm yr^{-1}) becomes the critical constraint for annual ET. High-elevation areas are clearly distinguishable because rocks, snow, and ice emit low latent heat (Fig. 1b and c).

Analyzing the three-year average water fluxes across the entire elevation range, we found that the areas between 200 and 300 m a.s.l. contribute proportionally the largest fraction to total ET. At higher elevations the contribution to ET declines faster than the fractional area, despite the slightly increasing contribution of precipitation (Fig. 2a). The elevational distribution of ET varies considerably between different catchments, due to climatic and vegetation heterogeneity and the interplay between water and energy limitations (Supplementary Fig. 7)[37]. We used three-year average P-ET as a proxy for runoff (Extended Data Figure 2)[38], since changes in soil and snow water storage over three hydrological years could be considered small, and ice melt only marginally contributed to the total water budget (less than 3%) at the annual scale at a rate of roughly 4 km^3 yr^{-1} (Extended Data Figure 4). Runoff production, P-ET, peaks at around 800 m a.s.l. (Fig. 2a). However, more than 50% of the blue water originates from the areas between 1300 and 3000 m a.s.l., which correspond to only 35% of the total Alpine domain. This can be explained by the sharp decrease of ET with elevation due to temperature constraints (Fig. 2a). The runoff production shows a great spatial variability, and even neighboring catchments, such as the upper Rhone and the Aare catchment in Switzerland, may exhibit distinct P-ET patterns (Fig. 2b).

Blue vs. green water during an exceptionally dry and warm summer

The Alpine water budget also displays high temporal variability (Fig. 2c); P-ET in 2001 was 53% higher than in 2003 (Extended Data Figure 4), which can be explained by both higher precipitation and lower ET. More specifically, the Alps received 225 mm more P in 2001 compared to 2003 (1363 and 1138 mm, respectively, averaged over the entire domain) while ET was 30 mm lower on average (Fig. 2c).

We used discharge measurements from 381 catchments across the entire domain to validate the simulations and assess the severity of the 2003 drought. We focused on analyzing runoff deficits, thus we excluded 47 locations in which observed 2003 runoff was higher than the long-term average and/or catchment P in 2003 was higher than the 2001-2003 mean. For the remaining 334 catchments, we computed how much ET contributed to amplifying the effect of precipitation deficit on runoff during the 2003 growing season (here defined as the May-September period to isolate the period with active vegetation, when green water feedback can be pronounced, Supplementary Fig. 8). We found that in 75% of the catchments, ET amplified the drought impact on runoff. The remaining 25% of the catchments - mostly located in the SW and NE of the pan-Alpine domain - experienced dry conditions with water-stressed vegetation and reduced ET. Considering the entire domain, ET increased during the drought in an area covering more than 144,000 km^2 (Supplementary Fig. 9). Overall, the increase in green water flux amplified the precipitation-driven deficit by roughly 22% (Fig. 2c and Supplementary Fig. 8). In the areas between 1300 and 3000 m a.s.l., enhanced ET created an additional water loss of almost 4 km^3 during the 2003 growing season compared with the 2001-2003 growing season average, amplifying the runoff decrease due to precipitation by 32% (mean weighted by the area of each catchment).

Figure 2 | Relationship between elevation and blue and green water fluxes. a, Percent contribution of each elevation class (grouped in 100 m elevation bins) to precipitation (P), evapotranspiration (ET) and P-ET (the fluxes are averaged over the entire simulation period, i.e. 2001-2003), including the fractional area of each class for the entire domain. For the latter, the areas between 1300 and 3000 m a.s.l. (from which more than 50% of the total pan-Alpine runoff originates) are highlighted. **b**, P-ET in mm yr^{-1}. The dashed blue line and the shaded area represent the median P-ET (averaged over the entire simulation period) over the entire domain and the interquartile range, respectively. Colored lines show the median P-ET elevational distribution for selected catchments to illustrate the spatial variability. A locator map is included. **c**, ET anomalies during the 2003 growing season and for the +3 °C scenario (solid lines; left y-axis) and ET contribution to the runoff deficits compared with the precipitation deficit during the 2003 growing season (solid points; right y-axis). The anomalies were computed based on the 2001-2003 mean and the fluxes were averaged in space based on the 100 m elevation bins. Each point represents one of the 334 catchments (out of the 381 in total) for which catchment-averaged precipitation in 2003 was lower than in the period 2001-2003 and runoff in 2003 was below the long-term average (the latter was computed using the entire available record for each station, as summarized in the Supplementary Table 3). For ET anomalies, the interquartile ranges are also shown (shaded cyan and magenta areas).

To quantify the sensitivity of ET to warming only and remove the effect of reduced precipitation (2003 was not only the warmest summer but also exceptionally dry, see Supplementary Fig. 2), we performed a scenario analysis exploiting the correlation between the spatial distribution of simulated mean annual ET and mean annual temperature (space-for-time substitution, see Methods). This procedure was necessary because multiple simulations with T&C at a so high spatial resolution are computationally too expensive. We found that with a 3 °C increase in air temperature, annual ET would increase on average by 6% (evaporation will increase by 9% and transpiration will increase by 5%) while P-ET would decrease by roughly 5% under the assumption of constant precipitation (Fig. 2c). A similar effect on runoff is expected if annual precipitation is reduced by only 3% (change in P-ET for unchanged ET), which is likely to happen by the end of the century[39]. The sensitivity of annual ET to temperature shifts from positive to negative below 700 m a.s.l., which implies that in a warmer climate ET will likely decrease at low elevations (Fig. 2c and Extended Data Figure 4), because of higher water stress and earlier senescence of grass and leaf-shedding in deciduous forests. A recent study found similar patterns with a general positive ET sensitivity to temperature increases in the Sierra Nevada mountains in California[40].

During the growing season, precipitation is still the main source of blue water (81% on average), snowmelt comes second (16%), and ice melt accounts for the remaining 3% (Extended Data Figure 4). Below 500 m a.s.l., the contribution of soil water stored during winter and early spring and depleted during the growing season is substantial (99 mm, compared with 464 mm of ET, Extended Data Figure 4). The simulated growing season ET for the +3 °C scenario is generally higher than the simulated ET during 2003 (when the growing season temperature anomaly was roughly +3 °C) because precipitation during 2003 was very low (Supplementary Fig. 2), but overall the two patterns agree well (Extended Data Figure 4). This reinforces our confidence on the space-for-time approach employed.

Blue and green water fluxes averaged over the entire domain show that ET sporadically exceeded precipitation during the growing season of 2001 and 2002. For 2003, however, ET was higher than precipitation (which was 32% lower than the long-term average) already before the beginning of the growing season (Supplementary Fig. 10). The earlier snowmelt, which peaked before the start of the growing season and plummeted afterwards, amplified this precipitation deficit. Early snowmelt is becoming more frequent with rising temperature[10] and in 2003 it was only partly compensated by increased ice melt. The simulated ice melt during August 2003 was 38% of the total Alpine runoff in August, corresponding to about 2 km^3 of water.

Figure 3 | Analysis of anomalies in blue and green water fluxes during the 2003 drought. a, Histogram of observed May-September 2003 total runoff anomalies in mm for 381 locations. **b**, Histogram of observed May-September 2003 runoff anomalies in percentage for the same locations colored according to the magnitude of the anomaly (<-75%: yellow, -50 to -75%: cyan, -25 to -50%: red and >-25%: black), growing season 2003 is compared with the mean of each station for the period 2001-2003. **c**, Spatial distribution of the simulated ET anomaly in mm during the 2003 growing season (May-September, the reference period for ET is also 2001-2003). The dots represent the 381 locations with hydrological measurements and are colored as described in (b). The three insets in the lower right part show the boxplots of simulated ET anomaly in May-September 2003 for three different vegetation types (evergreen, deciduous, and grassland) in three elevation classes (<1000 m a.s.l., 1000-2000 m a.s.l. and >2000 m a.s.l.). The box length provides the interquartile range (I_{QR}), the bottom of the box the 25^{th} percentile (first quartile, q_1), the top of the box the 75^{th} percentile (third quartile, q_3), and the horizontal line within the box the median value. The lower whisker corresponds to q_1 - $1.5I_{QR}$, and the upper whisker corresponds to $q_3 + 1.5I_{QR}$.

Considering all 381 discharge stations, the observed runoff in May-September 2003 was on average 50% lower than the long-term mean at each station in the period May-September (Fig. 3 and Supplementary Fig. 10). Higher-than-average runoff occurred at a few locations, mostly in highly-glacierized catchments (Fig. 3)[41]. During this period, precipitation over the Alps was the lowest recorded between 1992-2008 and mean temperature was record-breaking high (Supplementary Fig. 2). The detailed vegetation scheme in T&C allows an analysis of different vegetation responses to the 2003 drought (Fig. 3). Most evergreen forests strongly benefited from the increased radiation and temperature, and did more so as the drought intensified during the summer months, mostly at high elevations, where ET increased. Grasslands and deciduous forests were water-stressed below 1000 m a.s.l., but benefited from the drought above 2000 m a.s.l., especially at the beginning of the growing season, when monthly ET increased by up to 60 mm (Fig. 3). In most areas of the Northern Alps the simulated ET anomaly was positive throughout the summer, in agreement with local measurements[19]. In all dry inner Alpine valleys (such as Valle d'Aosta and Val d'Adige in Italy, Valais in Switzerland and Murtal in Austria), lateral subsurface water redistribution leads to higher soil moisture and ET only in proximity of the streams (Extended Data Figure 5), but it cannot outweigh the low precipitation over the entire valley. These conditions render irrigation necessary to sustain vegetation productivity during summer in these areas[42]. This becomes even clearer during the 2003 drought, when very low soil moisture (Extended Data Figure 5) led to increased water stress and reduced ET for all vegetation types in the valleys[43]. However, the fractional area occupied by valleys is disproportionally smaller than the area with increased ET at high elevation (Fig 3), which implies that the role of these valleys for the overall Alpine water budget is minimal (Fig. 4). We remark that due to poor knowledge of subsurface conditions, we assumed 1 m uniform soil depth over the entire domain (except rocky areas and lakes), which leads to a simplified simulation of subsurface unsaturated and saturated dynamics and likely de-emphasizes the role of deep-storage and groundwater flow[44,45]. This could explain the excessive drying during summer in certain locations. This simplification together with the use of invariable rooting depths for the various vegetation types across space are potentially important limitations, but the relatively good performance of T&C for discharge, even during the most extreme low-flow conditions of August 2003 (Extended Data Figure 6), suggests that our inferences about ET are also robust, despite some discrepancy in the ET anomaly with other distributed products (Extended Data Figure 7).

Discussion

Our results –derived from a single mechanistic model– indicate that ET considerably contributed to reduce water yield during the 2003 growing season because vegetation benefited from the unusually warm and sunny conditions in a large part of the Alpine region at higher elevations. At the annual timescale, however, the temperature-driven ET feedbacks on runoff are less important than the direct effect of changes in precipitation; a 3% reduction in annual precipitation would affect runoff production over the entire pan-Alpine domain similarly to a hypothetical increase in mean annual air temperature by 3 °C. Note, however, that the scenario of +3 °C change in air temperature is simply based on a space-for-time analysis. Thus, it could overestimate the response because other meteorological variables covary with elevation. It further does not account for non-climatic factors (e.g., plant acclimation) that might lead to different vegetation responses. Similarly to our results, previous research in the California's Sierra Nevada showed that the 2012-2015 drought in that region caused severe runoff deficits, while ET was less influenced[46]. However, for unchanged precipitation, increases in temperature are projected to mostly enhance ET[40], which implies a negative feedback on runoff. Annual precipitation in the Alps has shown no long-term trends so far[47], but summer precipitation in central Europe has decreased and could further do so in the future due to changes in atmospheric circulation patterns leading to more intense summer droughts[48]. Combined with the expected decrease in ice melt and earlier snowmelt[10], our results demonstrate that blue water could be considerably reduced in the European Alps during warmer summers, but green water will continue to increase[9], leading to the oxymoron "lush vegetation-drier rivers" (Fig. 4). In the near future events like the 2003 drought may be no longer classified as extreme[11], but the combined effects of spatially variable changes in the timing and magnitude of precipitation will likely be more complicated, which could lead to different ET changes and potentially affect plant water stress patterns. On the other hand, the expected increase in plant water use efficiency with higher levels of CO_2 concentration than present[49] as well as large-scale disturbances (e.g. forest mortality[50]), species changes, and plant acclimation, which are not considered here, may partially offset this ET feedback during warmer summers in the long-term but they will unlikely have a major role in the near future. Furthermore, in certain regions of the Alps, vegetation management is intense and past disturbances such as wildfires or forest logging may have influenced vegetation composition and function in ways that are not accounted for in the model initialization.

While the presented concepts are general, the extension of the results to other mountain regions strongly depends on the relative magnitude of P and ET at the annual scale and during summer (Fig. 4). Important factors are also the elevation at which P-ET shifts from positive to negative during warm and dry summers and the areal extent covered by different elevation bands and vegetation types (Fig. 4). Nevertheless, results from Sierra Nevada[40,46] largely agree with our findings.

Figure 4 | Conceptual representation of the drought paradox in the Alps. a, Elevation profile and P-ET fluxes for average (in light blue) and dry and warm (in yellow) growing seasons across contrasting subregions of the Alps (1, plains and hills; 2, dry valleys; 3, forested mountains; 4, rocks and ice). **b**, Blue (P and P-ET) and green (ET) water fluxes in average (filled bars) and warm and dry (empty bars) growing seasons in contrasting Alpine subregions (as explained in (a) and shown in the map in (c), in mm). **c**, Map of the four distinct areas and the contribution of each area to runoff production, the circle size is proportional to the km^3. Contribution to runoff production was computed as P-ET + snowmelt + ice melt - ΔS, where ΔS denotes the difference in soil water storage from the beginning to the end of the growing season. The filled circles show the water volumes for average growing seasons and the empty circles show the water volumes for warm and dry growing seasons for the four subregions.

Understanding the partitioning of green and blue water fluxes and their spatial distribution from few square kilometers to the entire Alps is essential to manage the European water resources under current and future climatic conditions[2,14]. This partition has implications on ecosystem functioning, energy production, and water supply. We showed that ecohydrological simulations driven by high-resolution hydrometeorological forcing improve the quantification and understanding of the water budget in mountainous areas and its vulnerability to climate, providing insights into processes that coarser-scale approaches fail to reproduce[12,27,51]. The need for more realistic, high-resolution quantifications of water availability is urgent[52]. Our study demonstrates that recent advances in ecohydrological modeling, combined with large scale datasets and new computational capabilities, offer the possibility to address this urgent need, thus helping to define strategies to counteract or adapt to climate change impacts on water resources.

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Author contributions

S.F. and T.M. designed the study, J.P., R.R., B.S. and M.B. contributed with data. P.H. contributed to the optimization of the parallel simulations. T.M. performed the simulations and the analyses. T.M. designed the figures with contributions from G.M and S.F. The results were synthesized by T.M., S.F., C.P. and P.M. T.M. and S. F. wrote the manuscript with contributions from all other authors.

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Competing financial interests

The authors declare no competing financial interests

Methods

The Tethys-Chloris model (T&C). We performed simulations with the distributed ecohydrological model Tethys-Chloris (T&C). T&C simulates the coupled dynamics of energy, water, and vegetation and has been successfully applied to a very large spectrum of ecosystems and environmental conditions as summarized elsewhere[31,32,49,53–55].

The model simulates the energy, water, and carbon exchanges between the land surface and the atmospheric surface layer accounting for aerodynamic, undercanopy, and leaf boundary layer resistances, as well as for stomatal and soil resistances[56]. In each simulated grid cell, vegetation can occupy two vertical layers to mimic the coexistence of trees and bushes/grasses. Horizontal composition of vegetation is also possible since each element can account for multiple species or plant functional types[54,57]. Dynamics of water content in the soil profile are solved using a quasi 3-D approach: the one-dimensional (1-D) Richards equation is used for vertical flow and the kinematic wave equation is used for lateral subsurface flow. Saturated and unsaturated parts of the soil column are explicitly identified. Surface overland flow and channel flow are also solved by the kinematic equation. Snowpack dynamics are computed using the energy balance: snow can be intercepted by the vegetation or fall to the ground, where it accumulates and successively melts. Runoff generation occurs via saturation excess and infiltration excess mechanisms and depends on lateral moisture fluxes in the unsaturated and saturated zones as well as on overland flow [58]. Soil water content changes, infiltration, and runoff production are estimated with an adaptive time step based on the stability of Richards equation solution with the method of lines[59]. Overland and channel flow routing are computed with an adaptive time step that satisfies the Courant-Friedrichs-Lewy condition (seconds to 5 min). The version of T&C used in this study does not include soil freezing and thawing processes and the water present in soil pores is always considered to be in a liquid state. The soil biogeochemistry module is not activated, thus there is no modeling of soil carbon and ground evaporation does not occur from litter, but only from soil.

Photosynthesis is simulated using the Farquhar biochemical model[60] adapted with subsequent modifications and temperature dependence of biochemical parameters. The model follows the two big leaves scheme, where sunlit and shaded leaves are treated separately for estimating net assimilation and stomatal resistance[61]. Leaf maintenance respiration is assumed equal to the leaf dark respiration and acclimation effects are not accounted for. For upscaling photosynthesis from leaf to plant scale, photosynthetic capacity is assumed to decay exponentially with canopy depth[62]. The stomatal conductance parameterization accounts for net assimilation rate, leaf internal CO_2 concentration and vapor pressure deficit following the Leuning model[63]. The dynamics of seven carbon pools are explicitly simulated in the model and include leaves, living sapwood, fine roots, carbohydrate reserve (non-structural carbohydrates), reproductive tissues (fruits and flowers), standing dead leaves and heartwood/dead sapwood. The carbon assimilated through photosynthetic activity is used for maintenance and growth respiration, otherwise it is allocated to one of the first five pools. The different pools are undergoing tissue turnover in function of tissue longevity and environmental stresses, i.e., drought and low temperatures. Carbon allocation is a dynamic process that accounts for resource availability (light and water) and allometric constraints[64], e.g., a minimum ratio of fine root to foliage carbon; and an upper limit for the storage of carbohydrate reserves[65]. Carbon allocated to reserves can be subsequently translocated to favor leaf and fine root expansion at the onset of the growing season or after severe disturbances[54]. Phenology for extratropical species is simulated considering four states[66]: dormant, maximum growth, normal growth, and senescence. Patterns of plant allocation are influenced by the phenological phase[67]. Transition between phenological phases are prognostic in the model and controlled by soil temperature, soil moisture and photoperiod. The model assumes that vegetation is in a mature phase and in equilibrium with its nutritional environment. Further details of model computational set-up, structure, and description of process parameterizations are presented elsewhere[54]. Vegetation dynamics are solved at the daily time scale, energy fluxes at the hourly time scale.

Domain setup. The land surface for the simulations was defined based on the 90 m resolution STRM digital elevation model (DEM) following the definition of the Alpine Region of the European Soil Data Center (ESDAC, project name: Eco-pedological Map for the Alpine Territory, https://esdac.jrc.ec.europa.eu/projects/Alpsis/Ecalp_data.html)[68]. This includes the entire alpine arch from the French coast in the southwest to the Austrian lowlands in the north east and from the Swiss plateau on the northwest to the Slovenian coast in the southeast. This DEM was resampled to the final resolution of the simulations (250 m). The flow matrix used by T&C was calculated based on the D-infinity method[69]. To compute the stream network, we set the threshold of upslope area above which a pixel is considered to belong to the channel network to 40 km^2. Therefore, small streams are not explicitly resolved.

Soil texture data were derived from the SoilGrids product at 250 m resolution[70]. We obtained sand, silt, clay, and organic content for seven different depths (at the ground surface and at 0.05, 0.10, 0.30, and 1 m below the ground surface). Average soil properties were then computed for each grid cell considering the vertical discretization of the soil and assuming linearity. A uniform soil depth equal to 1 m was used for the entire area. This is a simplification of the real system, since variability in soil depth can affect runoff generation mechanisms, but our choice is considered reasonable for large parts of the pan-Alpine domain, in the absence of consistent information about soil depth. We used land cover data from Corine[71], using seven different classes to summarize the information provided in the dataset. The classification we used is shown in the Supplementary Table 1. Glaciers were initialized with a 50 m depth. Lake depth was not explicitly simulated. In other words, we assured that there was always water available for evaporation for every lake-pixel (by starting with a very large lake depth), but the rivers flowing into the lakes did not interact with this "lake" water.

For reasons of computational efficiency, the domain was divided into eight subdomains, roughly equal in size, following the divides (Supplementary Fig. 1); these subdomains were run independently of each other. No lateral exchange between the eight domains is anyhow expected because of the kinematic wave assumption (this is indeed a simplification because groundwater rarely follows the surface topography). Further, each domain was run in parallel with 24-36 cores on the Euler cluster at ETH (https://scicomp.ethz.ch/wiki/Euler). For this parallelization, each subdomain was automatically split into smaller subdomains according to the number of cores. The vegetation and hydrology modules of T&C were run independently for each of these small subdomains. Lateral exchanges were then performed serially by a master process (e.g. flow routing, avalanches) at each time step. Each completed job on the cluster automatically saved a report where the CPU time was stated. The total computational demand for the three-year simulation for each of these sub-regions was roughly 76,000 CPU hours (information about the cores available can be found at the following link: https://scicomp.ethz.ch/wiki/Euler# - Euler III and IV). We initialized the carbon pools in the model using values for the different vegetation types corresponding to mid-summer conditions (e.g., leaf biomass for evergreen forests was set to 270 gC m^{-2}) derived from decadal simulations at the plot-scale run for each vegetation type[32]. Furthermore, we started from a relatively dry domain and we run the distributed model twice with observed meteorological forcing for the period September-October 2000. The final values of all state variables (e.g., soil moisture, channel water storage, surface temperature, snow water equivalent, carbon pools) were considered the initial conditions for the numerical experiment presented here, which started on the 1st of November 2000.

Meteorological input. Obtaining hourly meteorological fields at the desired 250 m resolution for an area spanning 257,000 km^2 and seven countries (Austria, France, Germany, Italy, Lichtenstein, Slovenia and Switzerland) required the combination of distributed products and station data as described below.

<u>Precipitation</u>

We used the daily Alpine precipitation grid dataset[72], which has a grid spacing of 5 km and daily time resolution. To obtain the required model input at hourly temporal resolution and 250 m spatial resolution we used ground observations from 111 stations in Austria, France, Italy, Liechtenstein, and Switzerland (Supplementary Table 2). The computation of hourly precipitation was performed as the simulation was running, because of the impracticability of saving hourly precipitation fields for the entire period. Prior to the simulation, Thiessen polygons were defined based on the 111 stations, so that each pixel of the simulated domain was assigned to a single station. Each pixel was also assigned to the corresponding grid of the

Alpine precipitation. Then, during the simulation, in hour 1:00 of each day, the daily sum of the distributed product was disaggregated to hourly precipitation proportionally to the hourly measurements of the corresponding station but preserving the amount of precipitation based on the gridded daily product. In case that the corresponding station recorded no precipitation in a day in which the gridded product did record some, the daily gridded sum was assigned uniformly to the hours 18:00-22:00. These cases were however rare and mostly occurring at low rainfall intensities.

Air Temperature

For air temperature we used the ECMWF ERA INTERIM product[36] using time 00:00:00 and 12:00:00 and step 3/6/9/12. This corresponds to the forecasts that are issued twice a day for the next 12 hours with 3-hourly time step. Thus, we obtained 3-hourly data. We compared this to the product that includes analyses of four times a day and only 3-hourly forecasts and the difference was minimal. We used the finest resolution available (0.125°, data downloaded from https://apps.ecmwf.int/datasets/data/interim-full-daily/levtype=sfc/ on 23/2/2018). We first assigned the ERA INTERIM temperature to all the T&C cells within the corresponding ERA INTERIM grid and then we readjusted the temperature for each T&C grid cell to account for the elevation variability within each grid cell of the ERA INTERIM. For this readjustment, we used a single, temporally dynamic lapse-rate over each subdomain. The lapse-rate was computed every three hours by using all the ERA INTERIM grid cells within the subdomain. The two hours between the three-hourly ERA INTERIM data were computed using linear interpolation.

Wind Speed

For wind speed, we used the corresponding variable of ERA INTERIM, as described for temperature above. A constant lapse rate was used for downscaling from ERA INTERIM to T&C grid cells throughout the entire simulation based on observations in Switzerland (0.48 m

s^{-1}/ 100 m). A lower limit to wind speed was also imposed to avoid numerical instabilities in the computation of turbulent fluxes at the surface (0.01 m s^{-1}).

Atmospheric Pressure

For surface pressure, we used the corresponding ERA INTERIM variable, as described for temperature and wind speed above. Between the three-hourly data, we performed linear interpolation and applied a specific exponential correction to account for the variations of atmospheric pressure with elevation within each ERA INTERIM grid cell.

Relative Humidity

To compute the relative humidity, which is an additional necessary model input, we used dew temperature data from the ERA INTERIM product, and extrapolated it in space and time as described above for air temperature. Then, we combined hourly air temperature and dew point temperature to compute the hourly values of relative humidity for each cell in the T&C domain.

Solar Radiation

For computing distributed hourly radiation components for each pixel, we used measurements from 90 stations across the entire domain (stations with an asterisk (*) in the Supplementary Table 1). We used the inverse distance weight to compute radiation in each T&C grid cell from the meteorological stations, which means that more than one station was taken into account for each T&C grid cell. Only direct radiation was corrected based on constant lapse rates that varied according to the radiation wavelength (lapse rate 0.0015 W m^{-2} m^{-1} for direct shortwave radiation between 0.29 and 0.70 µm, 0.0027 W m^{-2} m^{-1} for 0.70 - 4.0 µm and 0.0014 W m^{-2} m^{-1} for direct photosynthetically active radiation). Diffuse shortwave radiation was not rescaled with elevation. Subsequently, T&C modified the incoming radiation to account for local and remote topographic effects, (i.e., shading and backscatter from nearby terrain, slope and aspects of the grid cell). Sky view factor and terrain configuration factor scaling between 0 and 1 are used for this purpose[73].

Longwave Radiation

To compute the incoming longwave radiation we followed an empirical formula for the clear sky emissivity[74], based on air temperature, vapor pressure (computed from relative humidity) and cloud cover. The latter was estimated for each of the stations with solar radiation measurements by comparing observed radiation and clear-sky radiation, simulated by a weather generator[75].

Space-for-time substitution. We estimated ET for a +3 °C change in mean annual temperature using a space-for-time substitution[76]. For each vegetation type (i.e. grassland, evergreen, deciduous, and mixed forest) we fit linear (for E) and non-linear (for T) models to describe the current spatial relationship between mean annual evapotranspiration and mean annual temperature. We performed separate regressions for evaporation and transpiration against temperature for each subdomain (Supplementary Fig. 1). Once the new reference values of ET were computed for each cell at the higher temperature, we also added the residuals of the regressions to preserve the spatial variability and the non-temperature-driven effects.

Model validation. We validated the model against remote sensing MODIS snow cover estimates (https://nsidc.org/data/mod10a2)[77]. Using the distributed dataset with temporal resolution of 8 days, we computed the time fraction with snow cover over the entire domain and compared it with the simulation results (Supplementary Fig. 3). The model tends to underestimate snow cover at lower elevations, and especially on the north facing slopes of the inner Alpine valleys. It also overestimates snow cover in the very wet pre-Alpine mountains, which might be partly an artefact of the model spin-up (precipitation in October 2000 was very high). Overall, results are satisfactory, considering that the model was run with parameters derived from previous plot-scale studies[22,32] and MODIS may tend to overestimate snow cover at low elevations, because it assigns to all eight days the largest snow cover recorded in that period. In fact, comparison of simulated snow cover with station observations (Extended Data Figure 1) suggests that the bias in snow cover duration only sporadically exceeds 30 days (Supplementary Fig. 3) and much larger values are likely to be a remote sensing artefact.

We further validated our analysis by comparing T&C output with gross primary production (GPP) and latent heat (LE) observations in 10 FLUXNET sites in the Alps or in the nearby areas (Supplementary Figs 4 and 5, and Supplementary Tables 5 and 6). Since the periods with available data vary between these sites and most often do not cover the 2001-2003 period, we performed an independent plot-scale simulation with T&C with hourly time step for each site. For these simulations, we used local meteorological forcing (downloaded from the FLUXNET database, see Supplementary Tables 5 and 6); these data were gap-filled whenever needed with linear regression when data gaps lasted less than a few hours or using the mean of the corresponding hour and day of the year over the entire period with available data. For all the sites we applied the same parameterization (according to the vegetation type) of the pan-Alpine setup.

We further compared the model with ET and GPP estimates from three distributed products: FLUXCOM[35] for ET (0.0833° spatial resolution) and GPP (0.5° spatial resolution), TRENDY[33,34] for GPP (spatial resolution 1°, simulations S2: CO_2 and climate (time-invariant present-day land use mask, models used: CLM4CN, HYLAND, LPJ_GUESS, LPJ, OCN, ORCHIDEE, SDGVM, TRIFFID VEGAS-2.1), and ERA Interim[36] for ET (derived from monthly estimates of latent heat at 0.125° spatial resolution, Extended Data Figure 3 and Supplementary Figs 6). The comparison shows that all products provide estimates close to the T&C average, but they all underestimate spatial variability due to their coarse spatial resolution. For instance, a single grid cell corresponding to the spatial resolution of TRENDY multimodel ensemble, includes areas with annual precipitation ranging from less than 700 mm to more than

3300 mm, while the entire domain is represented by only 52 grid cells. All of these distributed products cannot represent the complex topography of the Alps, which implies that other methods are needed for a robust quantification of green and blue water fluxes in mountainous regions as shown here.

We validated the model against daily runoff measurements in 381 stations (Supplementary Table 3) and daily snow depth measurements in 720 stations in Switzerland, Austria, and Italy (Supplementary Table 4). The results of the validation against station measurements are summarized in Extended Data Figure 1. Since T&C simulates only natural flows, i.e., it does not account for human regulation, which is quite important in the Alps, we removed heavily regulated catchments. To do so, in the absence of a pan-Alpine database indicating which station is affected by regulation, dams, or water withdrawals, we visually inspected all the observed hydrographs to flag each station as regulated or natural. We found that the model shows a considerably higher performance when excluding the regulated catchments (Extended Data Figure 2). The correlation between simulated and observed runoff time series is higher for the natural catchments, compared with the regulated ones (R^2 equals 0.69 and 0.47, respectively). Mean bias is less than 0.01 mm hr^{-1} for both groups of catchments, but the sign is opposite. As expected, T&C tends to overestimate runoff in heavily regulated catchments due to water abstractions, which were not simulated. Since in our analysis we used P-ET as a proxy for runoff, the inset of Extended Data Figure 2 compares area-averaged P-ET with measured runoff and shows that this assumption is largely valid, especially for the natural and non-glacierized catchments. For glacierized catchments, icemelt should be added to P-ET to obtain a reliable estimate of runoff.

Data availability. Data sources are summarized in the Supplementary Table 6. The DEM, land cover product and MODIS snow cover product used for this study are publicly available (https://esdac.jrc.ec.europa.eu/projects/Alpsis/Ecalp_data.html, https://www.eea.europa.eu/data-and-maps/data/clc-2006-raster-4, and https://doi.org/10.5067/MODIS/MOD10A2.006, respectively). FLUXNET data for running and validating T&C plot-scale simulations were downloaded from https://fluxnet.fluxdata.org/. FLUXCOM GPP and ET estimates were downloaded from http://www.fluxcom.org/, ERA Interim latent heat estimates were downloaded from https://www.ecmwf.int/en/forecasts/datasets/reanalysis-datasets/era-interim and TRENDY estimates of GPP were downloaded from http://www-lscedods.cea.fr/invsat/RECCAP/. Hydrological and meteorological/snow depth data from Switzerland are publicly available (https://www.hydrodaten.admin.ch/, and https://gate.meteoswiss.ch/idaweb/more.do, respectively). Daily data from the Austrian network are provided by the Austrian hydrological service (https://ehyd.gv.at/), and are publicly available. Hydrological and meteorological data for the French stations are publicly available (http://www.hydro.eaufrance.fr/, and https://donneespubliques.meteofrance.fr/, respectively). Hydrological data for the Slovenian stations are publicly available (http://www.arso.gov.si/vode/podatki/). Hydrological, meteorological and snow depth data for the Piemonte region and the Friuli-Venezia-Giulia region (Italy) are publicly available (http://www.arpa.piemonte.it/ and http://www.meteo.fvg.it/archivio.php?ln=&p=dati, respectively). Hydrological, meteorological and snow depth data for the Trentino-Alto Adige region (Italy) are provided by the University of Trento and are available upon request.

Code availability. Tethys-Chloris source code used for these simulations (11/2015 version) is available at https://hyd.ifu.ethz.ch/forschung/models.html, saved simulations results and the analysis scripts are available upon request to the corresponding author (S.F.). The software used to generate all the results is MATLAB 2016b®.

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