Authors' response - Manuscript "Why has catchment evaporation increased in the past 40 years? A data-based study in Austria" by D. Duethmann and G. Blöschl

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Replies to the comments by Axel Thomas

We would like to thank Axel Thomas for his interest and for his comments on our manuscript. His main comment addresses the potential effect of changes in wind speeds on evaporation. Two further comments address a statement in the introduction that relates increases in evaporation to global warming, and the use of the term evaporation.

Potential effect of changes in wind speed

There is a vast body of evidence (see in particular the review paper of McVicar et al. (2012, cited) that both radiation and aerodynamic terms are determining potential evapotranspiration (Ep). Depending on region wind may account for the major part of Ep variance. Even though the authors argue with a lack of spatial homogeneity of wind speeds in their study area their averaged wind speeds (Supplement Figure S2b) show quite well the general decrease of wind speeds that has been observed world-wide. The large variability is to be expected in largely mountainous Austria and disregarding this variability does introduce a major error into the analysis. In this respect it is most unfortunate to see that both calculation AND attribution analysis of Penman-Monteith Ep are based on spatially and temporally averaged wind data. Even using averaged wind data as additional variable in the attribution analysis would already show that both radiative and aerodynamic forcing largely explain most of the variance. In the present form – without a realistic inclusion of wind data – the results are misleading. I would propose to recalculate Ep with wind speed data that contains as much spatial and temporal variance as possible. In addition attribution itself is variable both in spatial and temporal terms (Fan and Thomas, https://doi.org/10.1016/j.jhydrol.2018.02.080) so an extended analysis taking into account attribution variability would offer the reader a considerably improved analysis of Ep dynamics.

Response: We have now analyzed the effect of changes in wind speed. In a first analysis, we applied average monthly trends derived from station observations of wind speed to the wind speeds used in the original analysis. In a second analysis, we aimed at also including spatial heterogeneities in wind speed and its trends. For this purpose, we derived spatially smoothed patterns of average monthly trends in wind speed from station observations. These were applied to spatial patterns of wind speeds derived from high-resolution downscaled reanalysis data. The results suggest that wind speeds have indeed decreased in Austria (by about 3% per decade) but the effect on trends in reference evaporation is small. According to the first analysis, the trend in E_0 averaged over all catchments is 2.4 ± 0.7 % per decade when allowing for decreasing wind speeds, as compared to 2.8 ± 0.7 % per decade when assuming no trends in wind speed. In the second analysis, E_0 estimates and trends in wind speed were lower due to lower wind speeds in the reanalysis data compared to the averages of the station data. This led to a smaller effect of the trends in wind speed on E_0 than in the first analysis (average E_0 trend of 2.9 ± 0.6 % per decade when allowing for trends in wind speed as compared to 3.1 ± 0.6 % per decade when assuming no trends in wind speed). The low impact of the changes in wind speed on reference evaporation can be explained by the generally humid climate in Austria, where wind speed has a much lower impact on reference evaporation than in an arid climate (Irmak et al., 2006). We have added the analyses to the supplement and we refer to it in sections 2.1.3, 3.2.1 and 4.2 of the main text.

The temporal variability of the drivers of reference evaporation (E_0) is already shown in the manuscript (Fig. 3). The spatial variability of the trends in the drivers of E_0 is not large and not shown in order to keep the number of figures low.

Statement in the introduction

There are two smaller points I would also like to mention: the authors rightly point out in their paper that temperature is not an important driver of Ep but at the same time begin their introduction with the much-too–often-heard statement that global warming (hence temperature) has increased regional evapotranspiration. Even the IPCC still voices this scientifically incorrect statement. I would propose to rephrase this sentence to clarify that in the context of global CLIMATIC change Ep also has seen changes.

Response: We agree and this has been changed as suggested: "In the context of global climate change, regional *E* has increased in many parts of the world in the last decades (Huntington, 2006)."

Use of the term evaporation

Another point to clarify is the sometimes misleading way 'evapotranspiration' is used in this paper. Evapotranspiration is an umbrella term that has many definitions and can be estimated in different ways. In the Introduction most of the papers cited deal with actual evapotranspiration as do the authors when they use the abbreviation 'E' in their data analysis and results. On p 2/l 15 however they appear to mean potential evapotranspiration (at least most of the cited papers deal with Penman-Monteith potential evapotranspiration). On p 3/l 27 it is PET (again perhaps potential evapotranspiration) while reference evapotranspiration (E0, p 6/l 22) is actually crop reference evapotranspiration as the method of Allen et al. 1998 is cited. 'potential evapotranspiration' is used twice in section headlines 2.3 and 2.3.2 but is not defined elsewhere; on p 7/l 23ff 'potential evapotranspiration' and 'reference evapotranspiration' are used almost synonymously. Perhaps the authors might consider adding a short section pointing out the differences between different measures and methods of evapotranspiration cited or used in their paper and then use the appropriate terms consistently throughout their paper.

Response: Thank you very much for pointing this out. Indeed, the different terms for evaporation should be used more carefully and we have changed this in the revised version of the manuscript. We now also consistently use the term evaporation and explain that it includes evaporation from plants via transpiration. Reference evaporation is defined in section 2.3.1. With these two explanations, we found an additional section on the different terms of evaporation not necessary.

In detail, we have included the following changes:

P2/L15: Here we discuss potential drivers of changes in actual evaporation, changes in available energy and atmospheric evaporative demand being one of them. Since this may be estimated by pan evaporation, the cited papers in this section deal with pan evaporation.

P3/L27: Reference evaporation has been used for this analysis and this has been changed accordingly.

P6/L22: To our knowledge, both terms, reference evaporation and reference crop evaporation (as well as reference evapotranspiration and reference crop evapotranspiration), may be used for the method of Allen et al. (1998).

Headlines 2.3 and 2.3.2: potential evapotranspiration has been replaced by reference evaporation.

P7/L23 ff: All uses of potential evaporation have been changed to reference evaporation.

Reference:

Irmak, S., J.O. Payero, D.L. Martin, A. Irmak, and T.A. Howell (2006): Sensitivity analyses and sensitivity coefficients of standardized daily ASCE-Penman-Monteith equation, *Journal of Irrigation and Drainage Engineering*, *132*(6), 564-578.

Replies to the comments by Ryan Teuling

We thank Ryan Teuling for his interest in our study and for his review of our manuscript. The comments of Ryan Teuling concern the potential effect of variations in wind speed, the strong control of P on E trends, and the interrelation between trends in vegetation and trends in soil moisture.

Potential effect of changes in wind speed

Concerning the effect of wind: the potential impact should be discussed in more detail. Wind speed is known to have seen significant trends in many regions (Vautard et al., Nature Geosci. 3, 756-761), and this could have impacted ET trends also in this study. According to the offline PM-equation used by the authors, the impact of wind is direct. It should be noted, however, that when coupled to an atmospheric model, the sensitivity of PM ET for wind becomes much smaller (see e.g. Van Heerwaarden et al., Geophys. Res. Lett. 37, L21401). So I consider it unlikely that wind is a strong driver of ET trends locally, but I agree with the other referee that this warrants an in-depth discussion.

Response: We have now analyzed the effect of changes in wind speed. In a first analysis, we applied average monthly trends derived from station observations of wind speed to the wind speeds used in the original analysis. In a second analysis, we aimed at also including spatial heterogeneities in wind speed and its trends. For this purpose, we derived spatially smoothed patterns of average monthly trends in wind speed from station observations. These were applied to spatial patterns of wind speeds derived from high-resolution downscaled reanalysis data. The results suggest that wind speeds have indeed decreased in Austria (by about 3% per decade) but the effect on trends in reference evaporation is small. According to the first analysis, the trend in E_0 averaged over all catchments is 2.4 ± 0.7 % per decade when allowing for decreasing wind speeds, as compared to 2.8 ± 0.7 % per decade when assuming no trends in wind speed. In the second analysis, E_0 estimates and trends in wind speed were lower due to lower wind speeds in the reanalysis data compared to the averages of the station data. This led to a smaller effect of the trends in wind speed on E_0 than in the first analysis (average E_0 trend of 2.9 ± 0.6 % per decade when allowing for trends in wind speed as compared to 3.1 ± 0.6 % per decade when assuming no trends in wind speed). We have added the analyses to the supplement and we refer to it in sections 2.1.3, 3.2.1 and 4.2 of the main text.

Strong control of *P* on *E* trends, interrelation between trends in vegetation and trends in soil moisture

My main concern related to the interpretation of Figure 8c. This figure shows the relation between inferred trends in P and ET. It suggests a very strong control of P on ET trends, which seems somewhat suspicious given the general humid climate conditions in Austria. In my view, two possible explanations exist.

Response: We thank the reviewer for raising this point. We agree that the estimated sensitivity of trends in E to trends in annual P as derived from Fig. 8c is relatively high for a generally humid region.

One reason why we expect a relatively high sensitivity of changes in E to changes in P in our study is the seasonality of the observed changes in P. P increased not in a uniform way over the entire year but the increases in P were concentrated in the summer season (Supplementary Figure S7 and S8). Thus, the estimated sensitivity is approximately an estimate of the sensitivity of changes in E to changes in summer P, which can be expected higher than the sensitivity to changes in annual P. While changes in summer P are expected to contribute more strongly to changes in E, changes in winter P more likely result in changes in discharge. We discuss this in section 4.2 of the revised manuscript.

However, we carefully rethought the analysis and became aware that the sensitivity derived from the regression in Fig. 8c might be overestimated since water balanced derived E (calculated as P - Q) and P are not independent variables. We have performed Monte Carlo simulations with correlated annual P and Q series to estimate the magnitude of this effect. This analysis aimed at investigating the strength of the relationship between trends in P and trends in Q resulting from the dependency of the two variables when assuming that trends in E are independent of trends in P. Means and standard deviations of P and Q, the covariance between P and Q, and the spatial variability of the trends in P have been derived from the data. This results in regression relationships with a slope of 0.08 ± 0.03 (i.e. 1 mm y^{-2} increase in P is related to an increase of E by 0.08 ± 0.03 mm y^{-2}) and a correlation coefficient of 0.06 ± 0.04. Based on these results, we estimate that the slope derived from Fig. 8c overestimates the sensitivity of changes in E to changes in P by 0.08 ± 0.03. In the revised paper, we consider this by subtracting the value derived by the Monte Carlo analysis from the regression slope derived from Fig. 8c. This reduces the sensitivity of changes in E to changes in P from 0.30 ± 0.04 to 0.22 ± 0.05. Consequently, we have revised the attribution. The revised estimates suggest that changes in atmospheric conditions, vegetation activity, and precipitation have contributed 43 ± 15 %, 34 ± 14 %, and 24 ± 5 %, respectively, to the average increase in E_{wb} in the study catchments.

It might be that in general, soil moisture constraints on ET have weakened because of increased P. In this case, one would expect inferred actual ET to be significantly lower than potential ET. This relation between actual and potential ET, however, is not explored in the manuscript. I believe such an analysis should be included in a revised version, as it provides important insight into the possible background of the drivers of ET trends.

Response: An analysis of the ratio between E_{wb} to E_0 shows ratios close to or even above unity, indicating generally humid conditions. However, estimates of the AET/PET ratio based on the ratio of E_{wb} to E_0 likely overestimate the AET/PET ratio. E_{max} (the maximum possible evaporation under the actual vegetation when soils are saturated) is likely much higher than E_0. The land cover in the study catchments is dominated by forest and grassland, with average fractions over all study catchments of 0.52 and 0.25. Analyses from non-weighable lysimeters indicate that E_{max} for sites with non-deciduous trees (pine forests) was about 20-30% higher than E_0 (ATV-DVWK, 2001). We estimated E_{max} for each catchment as $E_{max} = E_0 \cdot \sum(l_i \cdot f_i)$, where l_i is the fraction of land cover *i* and f_i is the ratio of E_{max}/E_0 for land cover type *i*, which was approximated as 1.2 for forests and 1 for all other land cover types. This results in median (upper/lower quantile) values for E_{wb}/E_{max} of 0.84 (0.77/0.91), suggesting that in most catchments the AET/PET ratio is <1, even though the study area is classified as humid.

It should be noted that uncertainties in the estimated E_{wb} contribute to uncertainties in the estimated AET/PET ratio. These uncertainties arise to a large part from uncertainties in *P*, e.g. due to undercatch errors and the uncertainties in their correction. While the effect of the undercatch correction on the estimated trends in E_{wb} is small, it has a relatively high influence on uncertainties of the absolute E_{wb} estimates (see section 3.1; table 3).

The estimates of the E_{wb}/E_{max} ratios have been added to the end of section 3.1 of the revised manuscript.

It should be noted that trends in soil moisture and vegetation might possibly be related. This needs to be discussed, along with the implications for the results shown in Figure 9 which assumes soil moisture/P and vegetation effects to be independent.

Response: Regarding the interrelation between trends in vegetation and trends in soil moisture, an analysis of the covariance between trends in *P* and trends in NDVI showed no significant relationship (r = -0.01). This suggests that, in the study area, increases in *P* were not an important driver for the changes in vegetation activity, and that increases in NDVI are rather driven by increases in air temperature and a longer growing season, increases in atmospheric CO_2 and land cover changes (such as the increase in forest at the expense of grassland). We have added this point to section 3.2.4 and to the discussion section of the revised manuscript.

A second explanation for the relation in Fig. 8c could be that trends in ET are induced by overestimation of trends in P, for instance due to a too strong correction for undercatch. This possibility needs to be explored and discussed.

Response: An overestimation of trends in *P* by a too strong correction for undercatch has only a small influence on the estimated relationship between trends in *P* and trends in E_{wb}. As shown in Table 3, the effect of the applied undercatch correction on average trends in *P* and E_{wb} is small (in contrast to the effect on absolute values). This is also reflected by a small effect on the estimated regression slope between trends in *P* and trends in E_{wb} and the attribution result. When considering no undercatch correction of precipitation 5.9 ± 1.8 mm y^{-1} $decade^{-1}$ of the E_{wb} trend is estimated to be due to increases in precipitation, as compared to 6.9 ± 1.6 mm y^{-1} $decade^{-1}$ when correcting precipitation for undercatch using parameters for moderately sheltered locations.

So in summary, if the correlation is physical/causal, the authors should provide additional evidence for the underlying process, for instance by showing increasing ET/PET ratios. In addition, the dependency of trends in vegetation and soil moisture needs to be explored. Fig 9 is interesting, but these results are currently not sufficiently robust to be published.

Response: We revised the analysis taking into account the dependency between trends in *P* and trends in E_{wb}, analyzed the relationship between trends in vegetation and trends in *P*, and provided additional explanation for a high sensitivity of changes in *E* to changes in *P* in the study region. In the revised paper, we formulate the attribution more cautious and more clearly mention the uncertainties.

Despite the remaining uncertainties, we believe that presenting the results in Fig. 9 is a useful contribution.

Reference:

ATV-DVWK (2001): Verdunstung in Bezug zu Landnutzung, Bewuchs und Boden, GFA-Ges. zur Förderung d. Abwassertechnik e.V.

Replies to the comments by Referee #2

We would like to thank Referee #2 for his/her interest and the useful comments on our manuscript. These were related to the attribution to soil moisture change, changes in seasonality and feedback effects between the drivers.

Attribution to soil moisture change

The authors explain the dependence of the ET change (partly) on change of precipitation is as a result of increased soil moisture arising from the precipitation change. Here is where I have some concern - I have trouble grasping the attribution to soil moisture change. Why would there be increased soil moisture when P and ET are changing in the same direction? I have trouble understanding it. Furthermore, if there is increased soil moisture why is Q not increasing? I would not be so quick to jump to this conclusion: perhaps they can take it more slowly, and in steps. Precipitation and atmospheric demand are both increasing, but at the same time for these same reasons (and other reasons, e.g., temperature) I can understand vegetation activity being increased, which probably removes the increased soil moisture in the root zone due to the increased precipitation (leading to increased ET) but without increased recharge, which keeps Q the same. Not sure if this logic is right.

Response: Thank you for this comment. It made us aware that using soil moisture is not suited well as one of the drivers for the attribution because of the feedbacks between changes in E and changes in soil moisture and because it neglects changes in interception. The feedbacks between changes in E and changes in soil moisture lead to some ambiguity. For example, under conditions of increasing E due to increasing atmospheric demand, E increases more strongly at a location with increasing precipitation compared to a location with stationary precipitation but it is unclear whether this effect should be ascribed to changes in soil moisture (as done in the original manuscript). In the revised manuscript, we use precipitation instead of soil moisture as one of the drivers for the attribution analysis.

Changes in seasonality

In any case I am afraid the observed phenomenon may not be fully explained without invoking changes to seasonal variability (of everything, especially NDVI). There must be some kind of nonlinearity caused by changes to the seasonality, which may contribute to the phenomenon. In other words, changes in precipitation and radiation (and wind) propagate through the system in more complex ways than the authors have concluded in the paper. For the present, the paper requires some moderate revisions to address these issues. I suggest that the authors try to refine their attribution exercise to account for this complex system perspective, and to allow for seasonality changes to play a role in contributing to the phenomenon.

Response: Seasonality effects are partly already considered in the analysis. E_0 was calculated on a daily basis considering daily inputs of global radiation, air temperature, and vapor pressure deficit. Thus, variable rates of increase in these inputs during different seasons and their variable effects on E_0 during different seasons are considered in the analysis. We investigated changes in NDVI over the year and these were considered in the analysis of changes in E_{0v}. With respect to the analysis of the water balance components, the manuscript already showed seasonal changes of precipitation (for the summer and winter half year; Supplementary Figure S7 and S8).

We have now also analyzed changes in *P-Q* and discharge on a seasonal basis (Fig. R 1). *P-Q* shows increases during the summer half year (May-Oct) and decreases over the winter half year (Nov-Apr). Precipitation increases during the summer half year but shows no trends or decreases over the winter half year. Discharge does not show trends over the summer or the winter half year.

Due to intraannual storage variations *P-Q* for the winter or summer half year cannot be interpreted as *E* in the winter or summer half year. Changes in *P-Q* represent a combination of changes in *E* and changes in storage. The negative trend in *P-Q* during the winter half year suggests an increase in evaporation during the winter half year and/or a lower transfer of stored water from the winter to the summer half year. One possible explanation for a lower transfer of stored water might be the decrease in snow, i.e. a greater proportion of the precipitation that falls during winter contributes to discharge during the winter instead of being stored as snow and contributing to discharge or *E* during the summer half year.

Consideration of feedbacks

A further suggestion, anticipating future studies, is to present a conceptual model (in the form of a causal loop) that possibly accounts for some kinds of feedbacks that may need to be invoked to fully explain the phenomenon. The current paper looks like a stepping stone towards a more comprehensive model of the system in the future.

Response: We agree with the referee about the importance of considering changes in E_{wb} within a systems approach since the changes in *E*, vegetation, soil moisture, etc. are related through multiple feedbacks (Fig. R 2). We now include a causal loop diagram that visualizes these feedbacks and supports the discussion section of our paper. We explicitly discuss which feedbacks are included or excluded with the different drivers in the attribution.

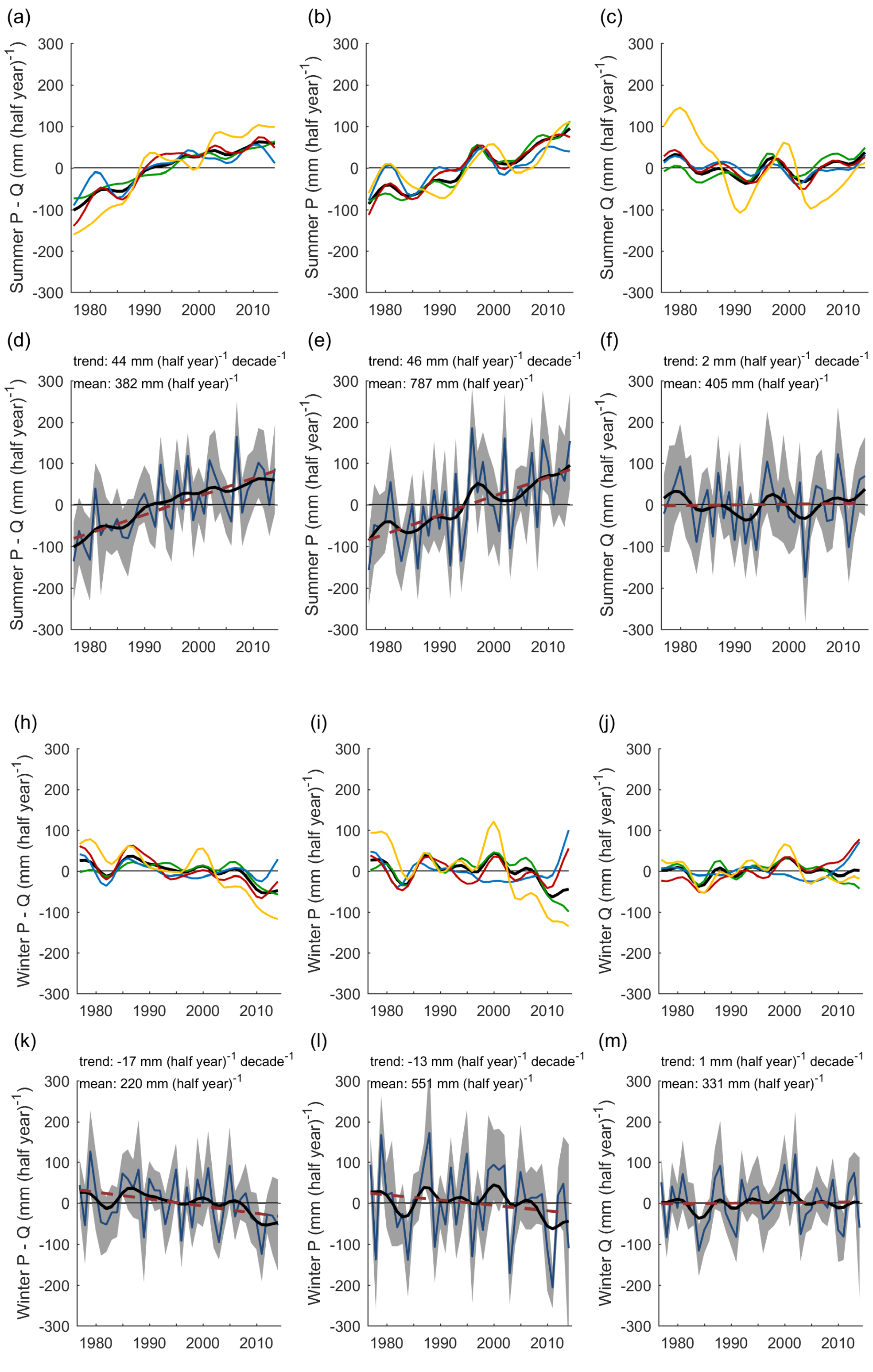

Fig. R 1: Anomalies of (a, d, h, k) *P-Q*, (b, e, i, l) precipitation and (c, f, j, m) discharge for (a-g) summer half years and (h-m) winter half years over 1977–2014. (a–c and h–j) show anomalies by region. Data smoothed using a Gaussian filter with a standard deviation of 2 years. (d–f and k–m) show anomalies over all catchments. The thin blue line shows the mean over all catchments, the grey shaded area the variability between catchments (± 1 standard deviation), the bold black line the smoothed mean, and the red dashed line the trend.

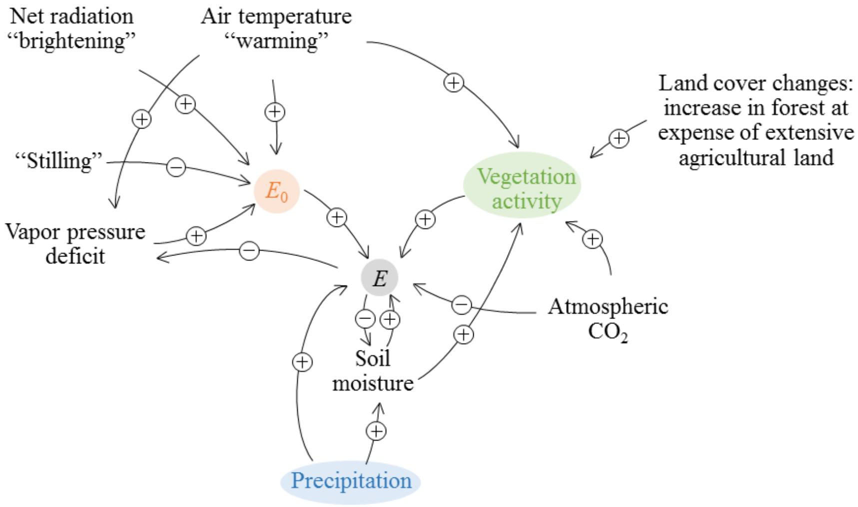

Fig. R 2: Drivers of changes in evaporation, *E*, including feedback effects.

Replies to the comments by Referee #3

We would like to thank Referee #3 for his/her interest and the comments on our manuscript. These relate to the relationship between the driving variables and the effect on the attribution estimates, the possible effect of variations in storage, the calculation of the trends, and the effect of variations in wind speed.

Relationship between the driving variables and the effect on the attribution estimates

[1] The authors clearly acknowledge that drivers of ETwb are tightly interlinked in the discussion. Does the attribution methodology adequately take this into account though? Have you explored the covariance among attributing variables? For example, if increases in P lead to increases in NDVI, are increases in P being overstated in the current attribution?

Response: We agree that the effect of increases in P would be overestimated if trends in P and trends in NDVI were correlated and thus Fig. 8c included indirect effects of P on vegetation activity. We checked partial correlation coefficients when trends in NDVI and E_0 were accounted for, and this did practically not affect the correlation between P and E_{wb}. In the study region, increases in P are not related to increases in NDVI (r = -0.01) or increases in E_0 (r = -0.12) (Fig. R 3). The attribution analysis is therefore not influenced by covariances between trends in P and E_0 or P and NDVI. We added the analysis of correlations between trends in P and E_0 or P and NDVI to section 3.2.4 of the manuscript.

However, the effect of increases in P on increases in E_{wb} was overestimated due to dependencies between the P and E_{wb} series. This is accounted for in the revised manuscript (see the reply to the comments of Ryan Teuling).

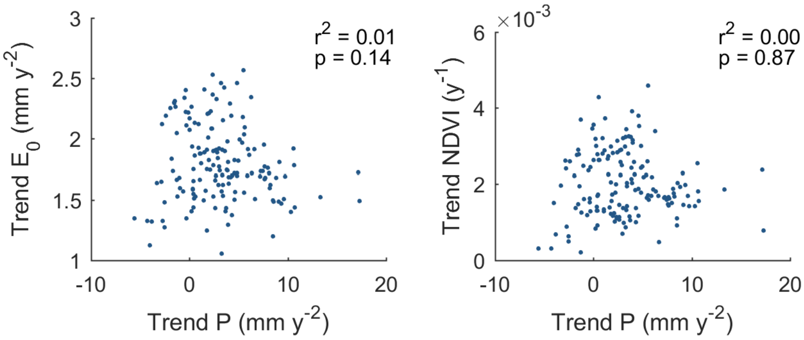

Fig. R 3: Scatter plots of the relationships between trends in E_0 and NDVI against the trend in *P*.

Possible effect of variations in storage

[2] Section 2.1.3. I am a little confused about the time scale of ETwb data being smoothed and plotted in Fig. 2. Are you estimating <annual ETwb> = <annual P> - <annual Q> and smoothing these annual estimates with the Gaussian filter? If so, could increases in precipitation add to storage and not necessarily ET?

Response: Yes, Fig. 2a and d show variations in anomalies of E_{wb} estimated as <annual E_{wb} > = <annual *P*> - <annual *Q*>. Values for individual years cannot be interpreted as variation in E_{wb} since variations in *P* may have added to storage. The data were therefore smoothed by a Gaussian filter. Changes in surface water, soil, and snow storage can be assumed small over periods of several years. Studies on groundwater level changes in Austria do not show large-scale groundwater changes over the study period (Blaschke et al., 2011; Neunteufel et al., 2017). We therefore assume that changes in groundwater storage (and changes in any groundwater fluxes) are small. Since changes in glaciers can result in significant storage changes, catchments including glaciers were excluded. This suggests that changes in storage are likely small and that the trend in *P*-*Q* can be interpreted as trend in *E*.

Calculation of trends

Also, are the trends in ETwb consistent with changes in ETwb inferred by actually dividing the time series into larger time intervals where changes in storage can assumed to be much smaller (e.g. <ETwb> estimated from average data 1977 to 1995, and <ETwb> estimated from average data 1996 to 2014)?

Response: Trends calculated from the annual series are consistent with changes in E_{wb} derived from dividing the series into two parts from 1977 to 1995 and from 1996 to 2014, as in this equation:

$$t = \frac{\overline{E_{wb96-14}} - \overline{E_{wb77-95}}}{\overline{y_{96-14}} - \overline{y_{77-95}} + 1} * 10$$

where *t* is the trend (mm y^{-1} decade^{-1}), $\overline{E_{wb96-14}}$ ($\overline{E_{wb77-95}}$) is the average E_{wb} over 1996-2014 (over 1977-1995), and $\overline{y_{96-14}}$ ($\overline{y_{77-95}}$) is the average year of the second (first) half of the study period (i.e. 2005 and 1986). Calculating the trend in E_{wb} this way results on average over all catchments in a trend of 30.4 mm y^{-1} decade^{-1}, compared to an average trend of 29.3 mm y^{-1} decade^{-1} when calculated from the annual series and using Sen's slope as in the manuscript.

Effect of variations in wind speed

[3] Lastly, I agree with the three previous comments that the attribution would benefit from a more thorough discussion of the impacts of wind speed and water availability. I think these suggestions were well elaborated in the previous comments. With regards to wind speed though, it could be useful to perform a sensitivity analysis of ETo to possible wind speed trends (e.g. based on the magnitude of the trend inferred from ERA Interim data) rather than using uniform monthly wind speed.

Response: We have now analyzed the effect of changes in wind speed. In a first analysis, we applied average monthly trends derived from station observations of wind speed to the wind speeds used in the original analysis. In a second analysis, we aimed at also including spatial heterogeneities in wind speed and its trends. For this purpose, we derived spatially smoothed patterns of average monthly trends in wind speed from station observations. These were applied to spatial patterns of wind speeds derived from high-resolution downscaled reanalysis data. This approach was chosen since suggested drivers for the trends in wind speed are changes in the atmospheric circulation and an increase in surface roughness, which however is not captured by reanalysis data (Vautard et al., 2010). The results suggest that wind speeds have indeed decreased in Austria (by about 3% per decade) but the effect on trends in reference evaporation is small. According to the first analysis, the trend in E_0 averaged over all catchments is 2.4 ± 0.7 % per decade when allowing for decreasing wind speeds, as compared to 2.8 ± 0.7 % per decade when assuming no trends in wind speed. In the second analysis, E_0 estimates and trends in wind speed were lower due to lower wind speeds in the reanalysis data compared to the averages of the station data. This led to a smaller effect of the trends in wind speed on E_0 than in the first analysis (average E_0 trend of 2.9 ± 0.6 % per decade when allowing for trends in wind speed as compared to 3.1 ± 0.6 % per decade when assuming no trends in wind speed). We have added the analyses to the supplement and we refer to it in sections 2.1.3, 3.2.1 and 4.2 of the main text.

Minor comments:

[1] Page 2 (L26): In addition to CO2, stomata also respond to changes in atmospheric demand, temperature, and soil moisture.

Response: We have reformulated this sentence to "Rising atmospheric CO_2 concentrations increase plant growth and influence stomata closure (Gedney et al., 2006)." Since the focus in this sentence is on the changes in atmospheric CO_2 as a further driver of changes in *E*, other factors that influence stomata closure are not mentioned here.

[2] Page 3 (L27): PET, "Potential evapotranspiration".

Response: Has been changed to reference evaporation.

[3] Page 5 (L14): It may be helpful to explain this in more detail: "wind data were regarded as not representative with respect to evaporation trends", i.e. "not representative" is vague.

Response: The sentence has been changed to "Trends in wind data were not included in the analysis since station observations of wind speeds are known to be prone to inhomogeneities (Böhm, 2008), annual anomalies of wind speed data from 85 stations in Austria appear unrelated to each other

(Supplementary Figure S1a), and temporal trends over 1977–2014 do not show any spatial pattern (Supplementary Figure S2a) (see Supplement S1).”

[4] Since the phrase “vegetation activity” is used frequently, it might be useful to add a sentence in the 2.3.2 explaining the potential physical mechanisms driving “vegetation activity” (as represented to NDVI), e.g. vegetation fraction, vegetation type, LAI, phenology, etc.

Response: Good idea. We have added the following sentence: “Changes in vegetation activity as observed by the NDVI represent an integrated signal of changes in the phenology, the leaf area index, the vegetation fraction, the vegetation type and the land cover.”

[5] Page 12 (L20): with higher values during the early 1990s, right?

Response: Yes, thank you.

[5] Clarify titles in Figures S7-S8.

Response: Has been changed to “…for biweekly averages over the course of the year (the plot titles indicate the starting day of the two-week period).”

References:

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Why has catchment evaporation increased in the past 40 years? A data-based study in Austria

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Abstract. Regional evaporation has increased in many parts of the world in the last decades~~Global warming has increased regional evapotranspiration in many parts of the world in the last decades~~, but the drivers of these increases are widely debated. Part of the difficulty lies in the scarcity of high-quality long term data on evapo~~transpi~~ration. In this paper, we analyze changes in catchment evapo~~transpi~~ration estimated from the water balances of 156 catchments in Austria over the period 1977–2014 and attribute them to changes in atmospheric demand and available energy, vegetation, and ~~soil moisture~~precipitation as possible drivers. Trend analyses suggest that evapo~~transpi~~ration has significantly increased in 60 % of the catchments ($p \leq 0.05$) with an average increase of 29 ± 14 mm y^{-1} decade^{-1} (± standard deviation) or 4.9 ± 2.3 % decade^{-1}. ~~A pooled~~P~~ p~~an evaporation ~~series~~ based on 22 stations has, on average, increased by 29 ± 5 mm y^{-1} decade^{-1} or 6.0 ± 1.0 % decade^{-1}. Reference evaporation over the 156 catchments estimated by the Penman-Monteith equation has increased by 18 ± 5 mm y^{-1} decade^{-1} or 2.8 ± 0.7 % decade^{-1}. Of these, 2.1 % are due to increased global radiation and 0.5 % due to increased air temperature according to the Penman-Monteith equation. A satellite-based vegetation index (NDVI) has increased by 0.02 ± 0.01 decade^{-1} or 3.1 ± 1.1 % decade^{-1}. Estimates of reference evaporation accounting for changes in stomata resistance due to changes in NDVI indicate that the increase in vegetation activity has led to a similar increase in reference evaporation as changes in the climate parameters. A regression between trends in evapo~~transpi~~ration and precipitation~~, as a proxy of soil moisture,~~ yields a sensitivity of 0.22~~30~~ ± 0.05~~04~~ mm y^{-2} increase in evapo~~transpi~~ration to 1 mm y^{-2} increase in precipitation. A synthesis of the data analyses suggests that 43 ± 15 ~~38 ± 13~~ % of the observed increase in catchment evapo~~transpi~~ration ~~can~~ may be directly attributed to increased atmospheric demand and available energy, 34 ± 14 ~~30 ± 12~~ % to increased vegetation activity, and 24 ± 5 ~~32 ± 5~~ % to ~~increased soil moisture due to~~ increases in precipitation.

1 Introduction

Evapo~~transpi~~ration (E), which includes transpiration through plants, is an important process in the water, energy, and carbon cycles and directly controls agricultural productivity and water availability for human purposes. In the context of global climate change, ~~Global warming has increased~~ regional E has increased in many parts of the world in the last decades (Huntington, 2006). However, due to the difficulty of measuring E, especially at large spatial scales, the drivers of changing E are still debated.

Decadal changes in catchment E may be inferred from the catchment water balance, as storage changes are usually small over decadal scales. Surprisingly few studies have investigated trends in water balance based evapo~~transpi~~ration (E_{wb}). Those studies that exist generally found increases in E_{wb} in the 20th century. Examples include large river basins in the US (Milly and Dunne, 2001; Walter et al., 2004; Kramer et al., 2015), the Tibetan Plateau (Zhang et al., 2007), and catchments in Switzerland (Spreafico et al., 2007). A study of 109 basins around the world for the period 1961–1999 found only few significant trends but a tendency towards positive trends in North and South America and Europe and a tendency towards negative trends in Africa and Siberia (Ukkola and Prentice, 2013).

Observed trends in catchment E_{wb} can be complemented by point measurements, although it is invariably difficult to link point and catchments scales. Lysimeter data from Rietholzbach, Switzerland over 1976–2007 showed a decreasing trend in E in the first half of the period and an increasing trend in the second half (Teuling et al., 2009). Observations based on eddy covariance are usually too short for trend analyses (Wang and Dickinson, 2012), but they have been used to train models that use satellite and climate data at longer time scales and larger space scales ~~(Jung et al., 2010; Wang et al., 2010; Zhang et al., 2010; Miralles et al., 2014)~~ (Jung et al., 2010; Wang et al., 2010; Zhang et al., 2010). While these studies generally agree on positive trends since the 1980s, they differ in the magnitude of the estimated trends (Dong and Dai, 2017). It has been noted, however, that the results of such models need to be treated with care as they are sometimes inconsistent with trends from the water balance, particularly for wet basins (Zhang et al., 2012; Liu et al., 2016).

Potential drivers for changes in E are changes in available energy and atmospheric evaporative demand, which is driven by variations in wind, vapor deficit, and air temperature. Available energy and evaporative demand of the atmosphere can be estimated based on climatic drivers or measured with evaporation pans. In many parts of the world (including North America, China, India), annual pan evaporation (E_{pan}) has decreased in the second half of the 20th century with rates of 10–40 mm y^{-1} decade^{-1}, despite increases in air temperature (Peterson et al., 1995; Roderick et al., 2009; McVicar et al., 2012). In some instances, this decrease in E_{pan} has been explained by decreases in net radiation and/or wind speed (Roderick and Farquhar, 2002; Roderick et al., 2007). In other instances, decreasing E_{pan} has been interpreted as a consequence of increasing actual evaporation (Brutsaert and Parlange, 1998; Brutsaert, 2013). In Europe, most studies found increasing trends of E_{pan} (e.g. Ireland (1963–2005) (Stanhill and Möller, 2008), England (1957–2004 and 1986–2010) (Stanhill and Möller, 2008; Clark, 2013), Greece (1983–1999) (Papaioannou et al., 2011), and the Czech Republic (1968–2010) (Trnka et al., 2015)). In this paper, we use the term atmospheric conditions to summarize the drivers available energy and atmospheric demand.

~~Another~~ Other potential drivers are changes in land cover and vegetation (Piao et al., 2007). Rising atmospheric CO_2 concentrations increase plant growth (Piao et al., 2007) and~~,~~ ~~including~~ influence stomata closure ~~caused by increasing atmospheric CO_2 concentrations~~ (Gedney et al., 2006). Finally, changes in ~~terrestrial~~ water availability resulting from changes in precipitation may contribute to changing E. For example, Jung et al. (2010) attributed the hiatus of the increasing trend in global terrestrial E during 1998–2008 to the limited moisture supply in the southern hemisphere.

Several global scale studies attributed modelled changes in E to their drivers. The land surface models of Douville et al. (2013) could only explain variations in E over 1950–2005 if natural forcings, enhanced greenhouse gas concentrations and aerosols were considered. Based on an ensemble of land surface models that considered variations in climate, land cover, atmospheric CO_2 concentration and nitrogen deposition, Mao et al. (2015) found that E trends over 1982–2013 were dominantly driven by variations in climate, in particular precipitation. Miralles et al. (2014) showed that variations in E in the tropics were strongly influenced by variations in precipitation driven by El Niño/Southern Oscillations. In contrast, using a modified Penman-Monteith equation with detrended input variables, Zhang et al. (2015) concluded that the increase in global terrestrial E over 1982–2013 was largely driven by vegetation greening while changes in global radiation, wind speed, air vapor pressure, air temperature, and atmospheric CO_2 concentration had only minor effects. There does not seem to exist a full consensus regarding the drivers of increasing E that has been observed. The above cited studies looked at changes in modeled E at the global scale, based on globally available meteorological data. Complementary to these global studies, there is a need for data-based studies focusing at smaller regions with high quality data.

The aim of this study is to (a) identify changes in catchment evaporation in the past 40 years, and (b) identify the drivers of these changes. We use high-quality data sets of discharge, precipitation, and other climate variables from 156 catchments in Austria during the period 1977–2014. We analyze regional averages over these catchments in order to increase the robustness of the analysis. The potential drivers of changes in catchment evaporation examined are the atmospheric conditions, quantified by reference evaporation (E_0) and E_{pan}; vegetation, quantified by a satellite-based vegetation index; and ~~soil moisture, quantified by~~ precipitation as a proxy for the available water.

2 Data and methods

2.1 Water balance data and water balance estimates

2.1.1 Discharge data and catchment attributes

For the analysis of changes in the water balance based evapo~~transpi~~ration (E_{wb}), we identified all catchments in Austria where daily discharge data in the period 1977–2014 (hydrological years, November to October) with a maximum of two years missing were available. The beginning of the analysis period was set to 1977 because most discharge series in Austria start in the mid-1970s. Catchments with substantial anthropogenic influences from dams or water withdrawals (Viglione et al., 2013), catchments containing glaciers, and a few high-mountain catchments where observed discharge exceeded observed precipitation were excluded. This selection resulted in a total of 156 catchments (Fig. 1) ranging in size from 23 to 6214 km² (average 316 km²). Land cover was derived from the Corine 2000 data (European Environment Agency, 2016). The land cover is largely dominated by forest and grassland (Table 1). Median catchment elevations were calculated from the SRTM digital elevation model (Jarvis et al., 2008). They range from 287 to 1920 m (average 910 m). All catchments have a ~~PET/P~~ ratio of reference evaporation to precipitation smaller than one and are thus classified as energy-limited or humid according to Budyko

(1974). The catchments were assigned to regions with homogeneous variations in climate (Fig. 1), derived from a multi-variable (temperature, precipitation, sunshine duration, air pressure) Principal Component Analysis (Matulla et al., 2003; Matulla, 2005; Auer et al., 2007).

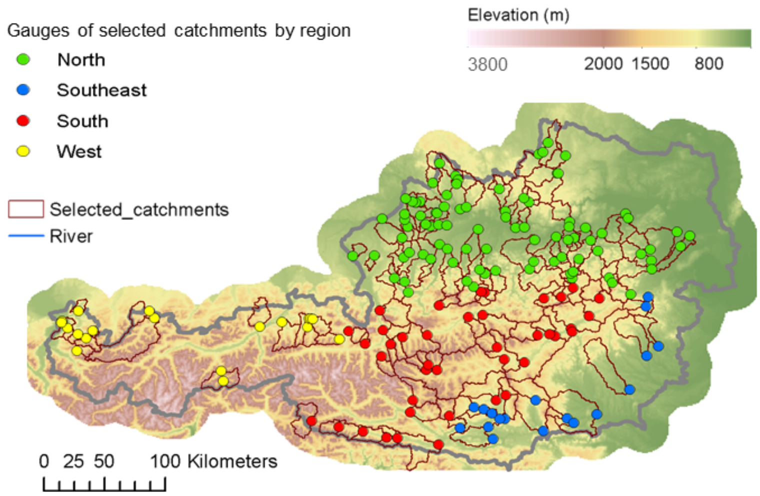


Fig. 1 Distribution of the study catchments in Austria and their association to one of four regions.

Table 1 Characteristics of the 156 study catchments in Austria.

	Median (lower quartile/ upper quartile)
Area (km²)	198 (95/368)
Median elevation (m)	825 (571/1218)
Coniferous forest (%)	31 (13/47)
Mixed and broadleaf forest (%)	12 (3/34)
Natural grassland (%)	2 (0/16)
Pasture grassland (%)	12 (6/19)
Arable (incl. heterogeneous agricul-tural areas) (%)	5 (0/29)
~~PET/P~~Ratio of reference evaporation to precipitation	0.50 (0.38/0.65)

2.1.2 Catchment average meteorological data

Air temperature and precipitation were obtained from the gridded SPARTACUS data set (Hiebl and Frei, 2016; Hiebl and Frei, 2017). This data set has ~~a~~ temporal and spatial resolutions of 24 h and 1 km, respectively, and was designed to be suitable for trend analyses. The interpolation method of minimum and maximum air temperature accounts for nonlinearities in the thermal profile and uses a constant station network of 150 stations. Precipitation is based on a two-step interpolation scheme, in which 1249 stations (including 119 totalizer precipitation gauges) were used for obtaining a daily background climatology for 1977–2006 and a constant number of 523 stations was used for interpolating ratios between the daily precipitation and the background climatology. To account for systematic underestimation from gauge undercatch we corrected the gridded precipitation data set for gauge undercatch using the following equation (Richter, 1995)

$$P_{\mathrm{corr}} = P_{\mathrm{orig}} + b \cdot P_{\mathrm{orig}}{}^{e} \tag{1}$$

where P_{corr} is undercatch corrected precipitation, P_{orig} uncorrected precipitation, and b, e are coefficients that depend on precipitation type and wind exposure. We estimated the precipitation type as snow for mean air temperatures below −1°C, as rain for mean air temperatures above 3°C, and as mixed precipitation between −1°C and 3°C (ATV-DVWK, 2001). The coefficients of Richter (1995) for moderately sheltered locations were applied to all grid points.

Measurements of relative humidity at 7:00 and 14:00 and global radiation were provided by the Austrian Central Institute for Meteorology and Geodynamics (ZAMG). Stations with more than 5 % (15 % for global radiation) missing data during 1977–2014 (hydrological years, November to October) were excluded, which resulted in 125 and 6 stations for relative humidity and global radiation, respectively. Data gaps were filled using linear regression to the station with the highest correlation. The data were interpolated onto a 1 km^2 grid using local ordinary least squares regression with elevation. The local neighborhood was set to a default radius of 100 km for relative humidity and 200 km for global radiation. This was adjusted to include a minimum of 10 (global radiation 4) and a maximum of 40 stations. The grid values were aggregated to catchment average series.

Trends in wind data were not included in the analysis since station observations of wind speeds are known to be prone to inhomogeneities (Böhm, 2008), annual anomalies of wind speed data from 85 stations in Austria appear unrelated to each other (Supplementary Figure S1a) and temporal trends over 1977–2014 do not show any spatial pattern (Supplementary Figure S2a) (see Supplement S1). Furthermore, interpolating wind data in space results in high uncertainties. We therefore used ~~Since wind data were regarded as not representative with respect to evaporation trends (see supplement),~~ uniform monthly wind speeds averaged over all years and over all stations in Austria ~~were used~~. The potential effect of changes in wind speed on evaporation is analyzed in Supplement S2.

2.1.3 Estimating evapo~~transpi~~ration from the water balance (E_{wb})

The catchment water balance can be written as

$$\frac{dS_{\mathrm{snow}}}{dt} + \frac{dS_{\mathrm{ice}}}{dt} + \frac{dS_{\mathrm{sw}}}{dt} + \frac{dS_{\mathrm{soil}}}{dt} + \frac{dS_{\mathrm{gw}}}{dt} = P - Q - E \tag{2}$$

where S_{snow} is snow, S_{ice} ice, S_{sw} surface water, S_{soil} soil water, and S_{gw} ground water storage; P precipitation; and Q discharge. In order to be able to estimate E from the water balance some assumptions on storage changes need to be made.

Over periods of several years, we may assume that changes in surface water, soil, and snow storage are small. Studies on groundwater level changes in Austria do not show large-scale groundwater changes over the study period (Blaschke et al., 2011; Neunteufel et al., 2017). Trends in annual mean groundwater levels at 2114 sites in Austria over 1976–2006 showed a heterogeneous picture, with decreasing trends ($p \leq 0.05$) at 18 % of the sites, increasing trends at 12 % of the sites and insignificant trends at 70 % of the sites (Blaschke et al., 2011). We therefore assume that changes in groundwater storage (and changes in any groundwater fluxes) are small. Estimates of absolute values of E_{wb} (trends in E_{wb} are not affected) furthermore depend on the assumption that groundwater fluxes across the catchment boundaries can be largely neglected, which is supported by prior rainfall-runoff studies in the catchments that suggest that the water balance can be closed (Parajka et al., 2005). Catchments with glaciers had been excluded from the analysis. For the time scale of decades, catchment evaporation can therefore simply be estimated as $E_{wb} = P - Q$. Since annual data of E_{wb} ~~should be considered with caution~~might be influenced by storage effects, we applied a Gaussian filter with a standard deviation of two years for the graphical presentation of variations of E_{wb} over the study period.

2.2 Pan evaporation data

Daily pan evaporation (E_{pan}) data using the GGI-3000 evaporimeter were provided by the Central Hydrographical Bureau (HZB), Vienna, and from ZAMG, Vienna. Further monthly E_{pan} data were obtained from the Meteorological Yearbook (ZAMG, 1977–1990).The coordinates, elevation, mean, and standard deviation of warm-season E_{pan} are listed in the Appendix (Supplementary Table S1). Missing values in the ZAMG data were replaced by estimates from an empirical Dalton-type formula with locally derived coefficients if wind speed and saturation deficit were available (Neuwirth, 1978). Missing values in the HZB data were replaced by estimates from the climate factor method (Hydrographischen Zentralbüro, 1996) with locally derived coefficients if wind speed, air temperature, and relative humidity were available. Remaining negative values and values larger than 15 mm d^{-1} were considered erroneous and flagged as missing data. Gaps with a maximum gap size of 4 days were linearly interpolated. Totals over the summer half year (May to October) were calculated if more than 90 % of the daily values or all monthly values were available. Due to the uneven record lengths~~,~~ we analyzed trends for three periods, 1979–2005 (13 stations), 1983–2014 (8 stations), and 1993–2014 (16 stations), which includes only series with a length of at least 20 years and five years or less missing from the record. Since trend analyses at individual stations did not show regional differences, normalized data of all series with a length of at least 20 years during 1977–2014 (24 stations) were pooled to a common series. The data were normalized by subtracting the mean over the overlapping period 1993–2005, excluding 1995 and 1998, which had many missing values at several stations. Summer E_{pan} was upscaled to the full year by the average ratio of annual and summer reference evapo~~transpi~~ration (1.33) for comparison.

2.3 Estimation of reference evapo~~transpi~~ration~~ and potential evapotranspiration~~

2.3.1 Reference evapo~~transpi~~ration

In order to examine effects of changes in atmospheric conditions we estimated reference evapo~~transpi~~ration (E_0) by the Penman-Monteith equation for well-watered short grass vegetation (Allen et al., 1998):

$$E_0 = 0.408 \cdot \frac{\Delta \cdot (R_\mathrm{n} - G) + \gamma \cdot \frac{185400}{(T + 273) \cdot r_\mathrm{a}} \cdot (e_\mathrm{s} - e_\mathrm{a})}{\Delta + \gamma \cdot (1 + \frac{r_\mathrm{s}}{r_\mathrm{a}})} \tag{3}$$

where R_n is the net radiation at the crop surface (MJ m^{-2} d^{-1}), G is the soil heat flux density (MJ m^{-2} d^{-1}), T is the mean air temperature at 2 m height (°C), r_a is the aerodynamic resistance (s m^{-1}), r_s is the surface resistance (s m^{-1}), e_s is the saturation vapor pressure (kPa), e_a is the actual vapor pressure (kPa), Δ is the slope of the vapor pressure curve (kPa °C^{-1}), and γ is the psychrometric consstant (kPa °C^{-1}). According to the reference conditions of a vegetated surface with a height of 0.12 m, r_s = 70 s m^{-1} and $r_\mathrm{a} = 208/u_2$ where u_2 is the wind speed at 2 m height (m s^{-1}), which was derived from the wind speed at 10 m height based on a logarithmic wind speed profile (Allen et al., 1998). The ground heat flux was neglected. The vapor pressure deficit $e_\mathrm{s} - e_\mathrm{a}$ was calculated as the average of the vapor pressure deficit at the minimum air temperature (using relative humidity at 7:00) and at the maximum air temperature (using relative humidity at 14:00). R_n was calculated from global radiation (R_s; MJ m^{-2} d^{-1}), albedo (α; set to 0.23) and net longwave radiation (R_nl; MJ m^{-2} d^{-1})

$$R_\mathrm{n} = \alpha \cdot R_\mathrm{s} + R_\mathrm{nl} \tag{4}$$

where R_nl was estimated according to Allen et al. (1998) based on minimum and maximum air temperature, clear-sky solar radiation, measured R_s~~global radiation~~, and the mean daily vapor pressure. E_0 was calculated on a daily basis on a 1 km^2 grid and aggregated to catchment average annual values (hydrological years, November to October).

The average contributions of the input variables net radiation, air temperature, and vapor pressure deficit to the trend of E_0 were evaluated using estimates of E_0 with one or several of the input variables held fixed to a particular year. The year 1994 was selected since annual mean values of E_0 and its input variables were close to the mean value over the study period. For this analysis, daily series of catchment evaporation were estimated from input data aggregated to catchment average daily values in order to reduce the computing time.

The contribution $\varphi_{\mathrm{i,E0}}$ of variable i to the trend of E_0 in catchment k was calculated as (see e.g., Galbraith et al. (2010)):

$$\varphi_{\mathrm{i,E0}}(k) = \frac{\tau_\mathrm{i}(k) - \tau_\mathrm{c}(k)}{\tau_{\mathrm{E0}}(k)} \tag{5}$$

where $\tau_{\mathrm{c}}(k)$ is the trend of the control (where all input variables are kept to those of 1994), $\tau_{\mathrm{i}}(k)$ is the trend of E_0 calculated with only variable i varying over the study period (and all other inputs as for the control), and $\tau_{E0}(k)$ is the trend of E_0 with all input variables varying. The two-way interaction effect $\varphi_{\mathrm{i\times j,E0}}$ of the variables i and j in catchment k was calculated as:

$$\varphi_{\mathrm{i\times j,E0}}(k) = \frac{\tau_{\mathrm{i\times j}}(k) - \tau_{\mathrm{c}}(k)}{\tau_{\mathrm{E0}}(k)} - \varphi_{\mathrm{i,E0}}(k) - \varphi_{\mathrm{j,E0}}(k) \tag{6}$$

where $\tau_{\mathrm{i\times j}}(k)$ is the trend of E_0 calculated with variable i and j varying over the study period (and all other inputs as for the control). For average effects and their variability, we calculated averages and spatial standard deviations of $\varphi_{\mathrm{i,E0}}(k)$ and $\varphi_{\mathrm{i\times j,E0}}\,(k)$ over all catchments.

2.3.2 Effect of changes in vegetation activity on ~~potential~~ reference evapo~~transpi~~ration

In order to examine changes in vegetation we used the Normalized Difference Vegetation Index (NDVI) based on satellite data. Changes in vegetation activity as observed by the NDVI represent an integrated signal of changes in the phenology, the leaf area index, the vegetation fraction, the vegetation type and the land cover. Observed 15-day maximum value composite NDVI data at a resolution of 8 km from the Advanced Very High Resolution Radiometer (AVHRR) for 1982–2014 were obtained from Tucker et al. (2005). The NDVI data were aggregated to catchment averages and linearly interpolated to daily catchment average series.

In the Penman-Monteith equation, an increase of the vegetation activity reduces r_{s}, which increases the potential evapo~~transpi~~ration. We calculated the reference evapo~~transpi~~ration considering a variable r_{s} ($E_{0\mathrm{v}}$) using Eq. (3), applying a variable r_{s} derived from the satellite data instead of a constant r_{s} of 70 s m^{-1}. The vegetation effect was estimated by calculating (i) $E_{0\mathrm{v}}$ from the original NDVI series and (ii) $E_{0\mathrm{c}}$ from a detrended NDVI series.

We applied two approaches for estimating r_s from NDVI to consider the uncertainty of these estimates. In the first approach, r_{s} was estimated from the leaf area index (LAI) and the fraction of photosynthetically active radiation (FPAR), which was estimated from (Sellers et al., 1996):

$$\mathrm{FPAR} = \frac{(S - S_{\mathrm{min}})}{(S_{\mathrm{max}} - S_{\mathrm{min}})} \cdot (\mathrm{FPAR_{max}} - \mathrm{FPAR_{min}}) + \mathrm{FPAR_{min}} \tag{7}$$

w~~W~~here S is a transformed NDVI value $(1 + \mathrm{NDVI})/(1 - \mathrm{NDVI})$, and S_{min} and S_{max} are the 5 % and 98 % quantiles of S for a given land cover class. LAI was estimated from FPAR (Sellers et al., 1996):

$$\mathrm{LAI} = \mathrm{LAI_{max}} \cdot \frac{\log(1 - \mathrm{FPAR})}{\log(1 - \mathrm{FPAR_{max}})} \tag{8}$$

where $\mathrm{LAI_{max}}$ is the maximum LAI of a land cover class. In Eq. (7) and Eq. (8), we applied the following coefficients for grassland: $\mathrm{NDVI_{min}} = 0.039$, $\mathrm{NDVI_{max}} = 0.674$, $\mathrm{FPAR_{max}} = 0.95$, $\mathrm{FPAR_{min}} = 0.001$, and $\mathrm{LAI_{max}} = 5$ (Sellers et al., 1996).

r_s was estimated as $r_s = r_l \cdot (LAI \cdot 0.5)^{-1}$ assuming a leaf stomata resistance r_l of 100 s m^{-1} for well-watered grass (Allen et al., 1998).

In the second approach, we used the relationship between r_s and NDVI of Zhang et al. (2010):

$$r_s(\mathrm{NDVI}) = \left(\frac{1}{b_1 + b_2 \cdot \exp(-b_3 \cdot \mathrm{NDVI})} + b_4\right)^{-1} \tag{9}$$

where b_1 =175 s m^{-1}, b_2=2000 s m^{-1}, b_3= 6, and $b_4 = -1/(b_1 + b_2)$ (coefficients for grass).

The average contributions of changes in atmospheric conditions and of changes in vegetation to the trend in E_{0v} ($\overline{\varphi_{\mathrm{atm,E0v}}}$ and $\overline{\varphi_{\mathrm{veg,E0v}}}$) averaged over all catchments and the two approaches for estimating r_s from NDVI were estimated as follows:

$$\begin{aligned} \overline{\varphi_{\mathrm{atm,E0v}}} &= \frac{1}{2n}\sum_{k=1}^{n}\sum_{l=1}^{2}\frac{\tau_{\mathrm{E0c}}(k,l)}{\tau_{\mathrm{E0v}}(k,l)} \\ \overline{\varphi_{\mathrm{veg,E0v}}} &= \frac{1}{2n}\sum_{k=1}^{n}\sum_{l=1}^{2}\frac{\tau_{\mathrm{E0v}}(k,l) - \tau_{\mathrm{E0c}}(k,l)}{\tau_{\mathrm{E0v}}(k,l)} \end{aligned} \tag{10}$$

where k is the catchment index, n the total number of catchments, l refers to one of the two approaches for estimating r_s from NDVI, $\tau_{\mathrm{E0c}}(k,l)$ is the trend in E_{0c} and $\tau_{\mathrm{E0v}}(k,l)$ is the trend in E_{0v} of catchment k when using approach l.

2.4 Trend analyses, regression analyses, and attribution of the trend in E_{wb}

Trends were estimated by the Sen's slope estimator (Sen, 1968). Trend significance was assessed by the nonparametric Mann-Kendall test (Mann, 1945; Kendall, 1975). The trend free pre-whitening technique was applied to remove lag-one serial correlation (Yue et al., 2002). Uncertainties in the trend magnitude were estimated using a bootstrapping approach. For this purpose, 1000 samples of size N were drawn, with replacement, from the record of length N years. The Sen's slope was calculated from each of the 1000 samples and the standard deviation was determined. Trends and the standard deviations were first calculated for each catchment and then averaged over the catchments to derive average trends and their uncertainties over a number of catchments.

The trends in E_{wb} in the individual catchments were related to the respective trends in E_0, NDVI, and mean annual precipitation by regression analysis in order to unravel the relation between changes in E_{wb} to changes in atmospheric conditions, changes in vegetation, and changes in precipitation~~soil moisture~~.

The contributions of the different drivers to the increase in E_{wb} were estimated as follows. The sensitivity of the trend in E_{wb} to trends in precipitation was estimated based on the slope of the linear regression between trends in E_{wb} and trends in precipitation. Since E_{wb} is estimated from precipitation and discharge, trends in E_{wb} are not independent from trends in precipitation and the regression relationship may overestimate the effect of trends in precipitation on trends in E_{wb}. The magnitude of this overestimation was estimated by Monte Carlo simulations with correlated annual precipitation and discharge series generated according to the statistics of the data (see Supplement S3). The sensitivity of the trend in E_{wb} to trends in precipitation ($s_{\mathrm{prec,Ewb}}$) was estimated as the slope of the linear regression between trends in E_{wb} and trends in precipitation corrected by the overestimation effect estimated by the Monte Carlo simulations. The average contribution of changes in ~~soil moisture~~precipitation to the trend in E_{wb} ($\overline{\varphi_{\mathrm{prec,Ewb}}}$) was estimated ~~based on the relation between the average trend in E_{wb} and the average trend in annual precipitation~~as:

$$\overline{\varphi_{\mathrm{prec,Ewb}}} = \frac{s_{\mathrm{prec,Ewb}} \cdot \overline{\tau_{\mathrm{prec}}}}{\overline{\tau_{\mathrm{Ewb}}}} \tag{11}$$

where ~~$s_{\mathrm{prec,Ewb}}$ is the slope of the linear regression of the trend in E_{wb} against the trend in precipitation, and~~ $\overline{\tau_{\mathrm{prec}}}$ and $\overline{\tau_{\mathrm{Ewb}}}$ are average trends over all catchments in precipitation and E_{wb}.

The contributions of changes in atmospheric conditions and vegetation could not be estimated in a similar way since the trends in E_0 or in NDVI were not related to trends in E_{wb} (see Sect. 3.2.1 and 3.2.3~~0~~). This is probably due to a relatively low spatial variability in changes in E_0 and changes in NDVI. While the spatial variability of changes in precipitation is relatively high, the spatial variability of changes in available energy, that is an important driver for changes in E_0 and in NDVI, is low.

A different approach was therefore used for estimating the average contributions of changes in vegetation and atmospheric conditions to the trend in E_{wb} ($\overline{\varphi_{\mathrm{atm,Ewb}}}$ and $\overline{\varphi_{\mathrm{veg,Ewb}}}$). Assuming that the remainder of the trend in E_{wb} is caused by changes in atmospheric conditions and vegetation, their contributions were estimated according to the ratio of their effects on E_{0v}:

$$\begin{aligned} \overline{\varphi_{\mathrm{atm,Ewb}}} &= \left(1 - \overline{\varphi_{\mathrm{prec,Ewb}}}\right) \cdot \overline{\varphi_{\mathrm{atm,E0v}}} \\ \overline{\varphi_{\mathrm{veg,Ewb}}} &= \left(1 - \overline{\varphi_{\mathrm{prec,Ewb}}}\right) \cdot \overline{\varphi_{\mathrm{veg,E0v}}} \end{aligned} \tag{12}$$

Where $\overline{\varphi_{\mathrm{atm,E0v}}}$ and $\overline{\varphi_{\mathrm{veg,E0v}}}$ are the contributions of changes in atmospheric conditions and in vegetation to the trend in E_{0v} (see Eq. (10)) averaged over all catchments and both parameterizations for r_s.

Uncertainties in the attribution estimate are based on the standard deviation of the regression slope of the trend in precipitation against the trend in E_{wb} and the standard deviations of $\varphi_{\mathrm{atm,E0v}}$ and $\varphi_{\mathrm{veg,E0v}}$ over all catchments and the two parameterizations for r_s.

3 Results

3.1 Changes in evapo~~transpi~~ration estimated from the water balance (trend detection)

Catchment E_{wb} trends increased significantly ($p \leq 0.05$) in 93 out of the 156 catchments (60 %) during 1977–2014. One catchment shows a significant decreasing trend. On average over all catchments, the annual E_{wb} increased with a rate of 29 ± 14 mm y^{-1} or 4.9 ± 2.3 % per decade (± standard deviation of the trend; the standard deviation refers to the average uncertainties of the trend estimates). The increase was largest during 1980–1995 and flattened out later (Fig. 2a,d). The increase in E_{wb} is more consistent over space and time than the changes in precipitation and discharge (Table 2, Fig. 2b–c, ~~Supplementary Figure S 3~~Supplementary Figure S6), which would be expected and adds credence to the estimates.

Annual precipitation trends increased significantly ($p \leq 0.05$) in 64 out of the 156 catchments (41 %), with an average increase of 32 ± 23 mm y^{-1} or 2.4 ± 1.7 % per decade. Two catchments show significant decreasing trends. Increases were particularly large in the eastern Alpine region of the study domain and generally occurred in summer (Supplementary Figure S7~~Supplementary Figure S 4~~ and S~~-5~~8).

Table 2 Means and trends of catchment evaporation estimated from the water balance (E_{wb}), precipitation, and discharge. Numbers given are the spatial averages over regions (and the entire study area) of the mean and the standard deviation.

	E_{wb}		Precipitation		Discharge	
	Mean (mm y^{-1})	Trend (mm y^{-1} $decade^{-1}$)	Mean (mm y^{-1})	Trend (mm y^{-1} $decade^{-1}$)	Mean (mm y^{-1})	Trend (mm y^{-1} $decade^{-1}$)
North	625	31 ± 13	1235	38 ± 22	610	7 ± 21
Southeast	668	27 ± 13	1060	17 ± 20	392	-8 ± 16
South	556	27 ± 15	1411	45 ± 25	855	16 ± 24
West	546	27 ± 19	1931	-13 ± 28	1385	-43 ± 31
All	604	29 ± 14	1339	32 ± 23	735	2 ± 23

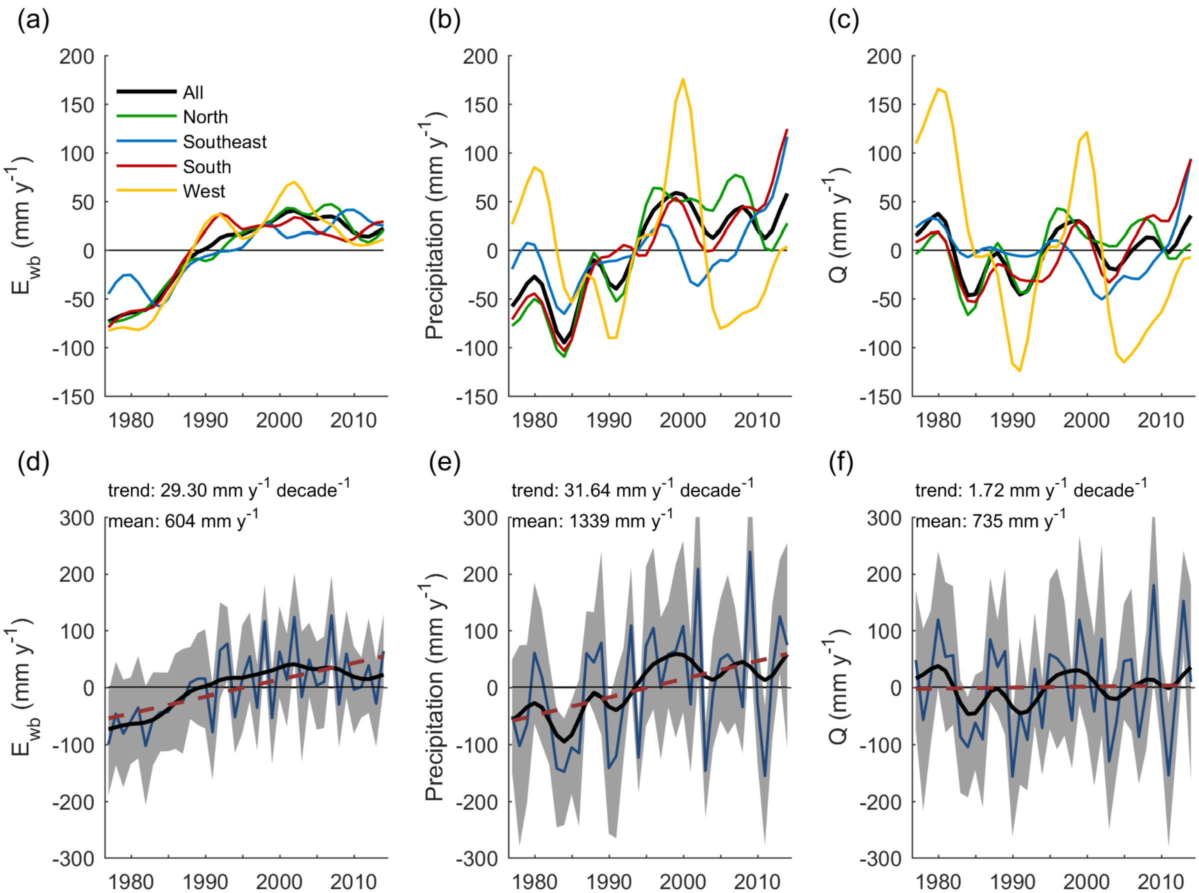


Fig. 2 Anomalies of (a, d) catchment evaporation estimated from the water balance (E_{wb}), (b, e) precipitation and (c, f) discharge over 1977–2014. (a)–(c) mean anomalies by region. Data smoothed using a Gaussian filter with a standard deviation of 2 years. (d)–(f) mean anomalies over all catchments. The thin blue line shows the mean over all catchments, the grey shaded area the variability between catchments (± 1 standard deviation), the bold black line the smoothed mean, and the red dashed line the trend.

Discharge trends increased significantly ($p \leq 0.05$) in 16 and out of the 156 catchments (10 %) and decreased significantly in 9 catchments (6 %), resulting in an average trend of 2 ± 23 mm y^{-1} or 0.2 ± 3.1 % per decade. Catchments with increasing discharge are located in the eastern Alpine region where precipitation increased most. Catchments with significant decreasing trends are located in the West of Austria where precipitation did not change much (Supplementary Figure S6~~Supplementary Figure S 3~~c). Interestingly, the decadal fluctuations of the discharge series within the study period are very similar to those of the precipitation series (Fig. 2b,c).

The analyses in this study are based on undercatch corrected precipitation using coefficients for moderately protected locations. In order to analyze the sensitivity of the correction assumption on the estimated trends in E_{wb}, we estimated trends in E_{wb} using uncorrected precipitation and undercatch corrected precipitation with correction parameters for unprotected locations. Without undercatch correction, estimates of average catchment precipitation would be 9 % lower and the resulting estimates of E_{wb} would be 20 % lower than with undercatch correction for moderately protected locations (Table 3). The percentage of catchments with significant increasing trends in E_{wb} would increase from 60 to 65 %, and the average trend in E_{wb} would increase from 29 to 31 mm y^{-1} or from 4.9 to 6.4 % per decade. Using undercatch correction for wind exposed stations has an effect of similar magnitude but of opposite direction. These results show that, while undercatch correction of precipitation has a strong effect on average E_{wb}, it only moderately affects its trends. Please note that in all figures and tables of this paper, with the exception of Table 3, precipitation undercatch has been corrected.

The E_{wb} estimates were used to estimate ratios of actual E to E_{max}, the maximum possible evaporation under the actual vegetation when soils are wet. A ratio close to unity would suggest precipitation not to be a likely driver of increases in E_{wb}. Since the land cover in the study catchments is dominated by forest (average fraction over all study catchments of 0.52), E_{max} is likely much higher than E_0 (e.g., Teuling, 2018). Analyses from non-weighable lysimeters suggest E_{max} to be 20–30% higher than E_0 for sites with pine forests at typical stand ages of 80–100 years compared to sites with grass (ATV-DVWK, 2001). We estimated E_{max} for each catchment as $E_{max} = E_0 \cdot \sum(l_i \cdot f_i)$, where l_i is the fraction of land cover i and f_i is the ratio of E_{max}/E_0 for land cover i, which was approximated as 1.2 for forests and 1 for all other land cover types. This results in median (upper/lower quantile) values for E_{wb}/E_{max} of 0.84 (0.77/0.91), suggesting that, in most catchments, E_{wb} is significantly lower than E_{max}. It has to be noted that these estimates are uncertain due to uncertainties of the factor f and the absolute estimates of E_{wb}.

Table 3 Effect of undercatch correction on estimates of average precipitation (P), E_{wb}, and their trends (averages over all study catchments and 1977–2014, significant trends for $p \leq 0.05$). *Parameters for moderately protected locations are used for all other analyses in this paper.

Undercatch correction	Average P (mm y^{-1})	Percentage of catchments with sign. increases in P (%)	Average increase in P (mm y^{-1} decade^{-1})	Average E_{wb} (mm y^{-1})	Percentage of catchments with sign. increases in E_{wb} (%)	Average increase in E_{wb} (mm y^{-1} decade^{-1})
Parameters for wind exposed locations	1414	35	27.1	679	54	25.8
Parameters for moderately protected locations*	1339	41	31.6	604	60	29.3
No undercatch correction	1221	45	33.7	486	65	31.0

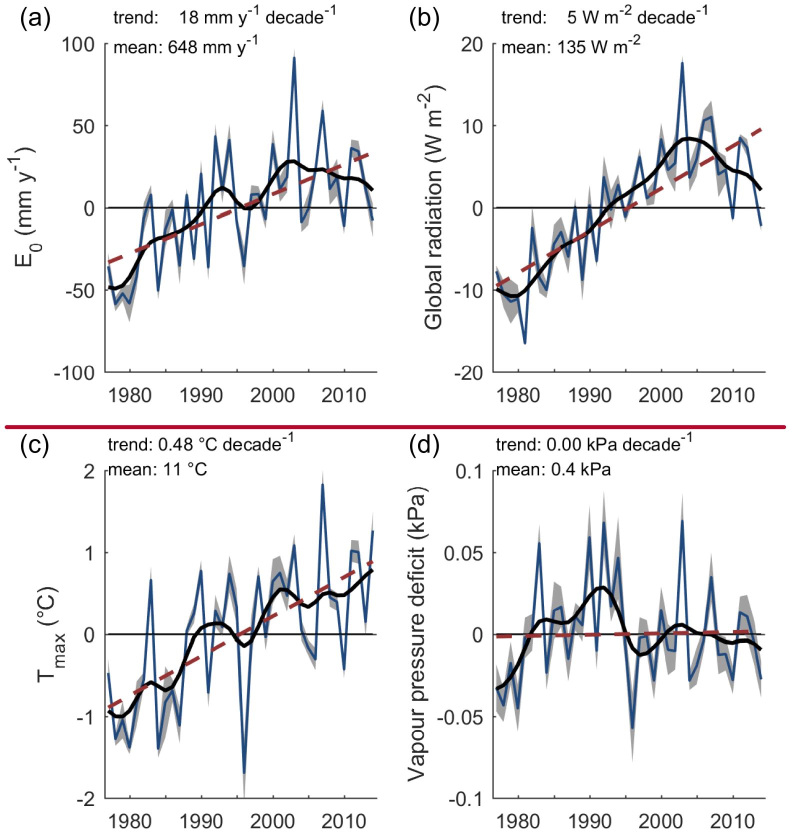
3.2 Drivers of the increases in evapo~~transpi~~ration (attribution)

3.2.1 Changes in atmospheric conditions – reference evaporation

Changes in the atmospheric conditions were examined by analyzing E_0 and E_{pan}. Averaged over all catchments, annual E_0 increased by 18 ± 5 mm y^{-1} or 2.8 ± 0.7 % per decade during 1977–2014 (Fig. 3a). Spatial variations in the increase in E_0 are small and there is no significant correlation between trends in E_{wb} and trends in E_0 (r^2=0.02, p=0.09) (Fig. 8a). Partial correlations between trends in E_{wb} and E_0, when trends in annual precipitation and NDVI were accounted for, are not significant either.

Over our study period, global radiation on average increased by 5.1 ± 0.9 W m^{-2} or 3.8 ± 0.7 % per decade (Fig. 3b). ~~Maximum~~ Mean air temperature (calculated as the average of minimum and maximum air temperatures) increased by 0.4~~8~~5 ± 0.09~~11~~ °C $decade^{-1}$ (Fig. 3c), while vapor pressure deficit showed variations, with higher values during the ~~late~~ early 1990s, but no trend over 1977–2014 (Fig. 3d). The analysis of the E_0 estimates with one or several of the input variables held fixed to a particular year showed that the increase in E_0 was largely driven by increasing net radiation, which contributed 76 ± 8 % to the trend in E_0, and increases in air temperature, which contributed 19 ± 6 % to the trend in E_0 (Fig. 4).

A trend analysis of wind speed observations shows large scatter between the stations. Averaged over all stations, wind speeds decreased by -3.0 ± 2.7 % per decade (see Supplement S3). According to a simple analysis, the average trend in E_0 reduces to 2.4 ± 0.7 % per decade when allowing for decreasing wind speeds, as compared to 2.8 ± 0.7 % per decade when assuming no trends in wind speed. Thus, negative trends in wind speed probably slightly reduced the effects of the positive trends in global radiation and air temperature on E_0.


(a)
trend: 18 mm y-1 decade-1
mean: 648 mm y-1
E0 (mm y-1)
(b)
trend: 5 W m-2 decade-1
mean: 135 W m-2
Global radiation (W m-2)
(c)
trend: 0.48 °C decade-1
mean: 11 °C
Tmax (°C)
(d)
trend: 0.00 kPa decade-1
mean: 0.4 kPa
Vapour pressure deficit (kPa)

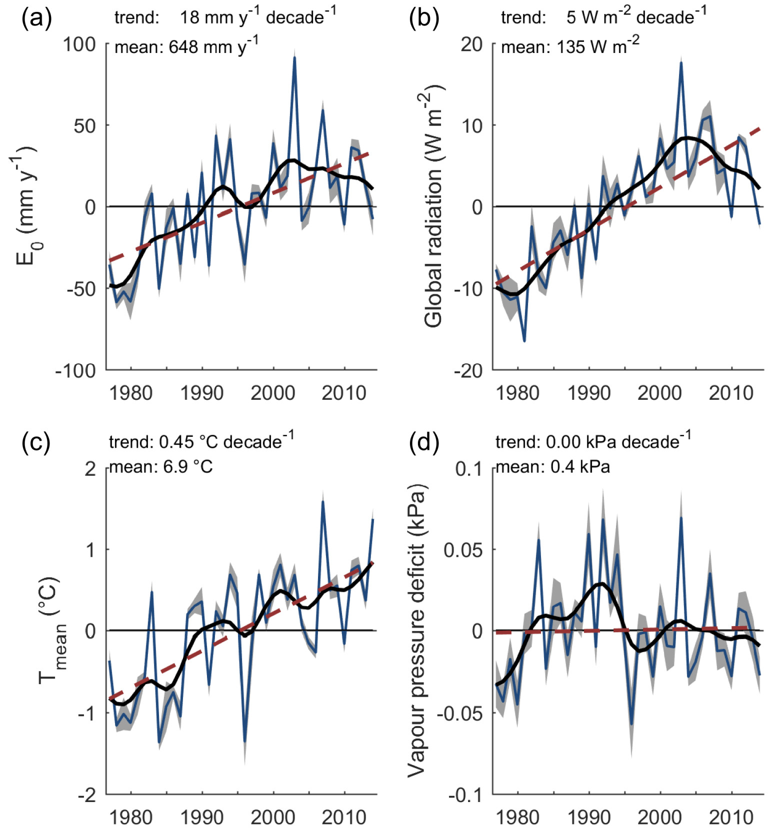


Fig. 3 Anomalies of (a) E_0, (b) global radiation, (c) ~~maximum~~ mean air temperature, (d) vapor pressure deficit over 1977–2014 and all study catchments. The thin blue line shows the mean over all catchments, the grey shaded area shows the variability between catchments (± 1 standard deviation), the bold black line shows the filtered mean (10-year Gauss filter with a standard deviation of 2 years), and the dashed red line the linear trend.

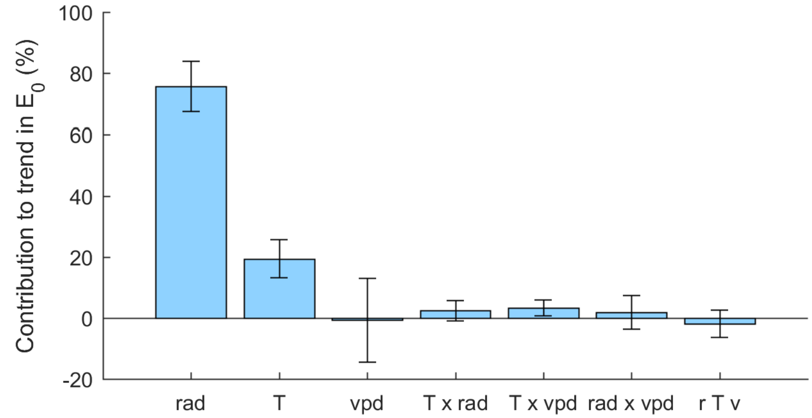


Fig. 4 Mean contributions of variations in net radiation (rad), air temperature (T) and vapor pressure deficit (vpd) and their two-way and three-way interaction effects to the trend in E_0. Bars show means over all catchments, error bars show the standard deviation of the variation between catchments. Percent are relative to trends in E_0.

3.2.2 Changes in atmospheric conditions – pan evaporation

Pan evaporation significantly ($p \leq 0.05$) increased at 5 of 13 stations over 1979–2005, at 4 of 8 stations over 1983–2014, and at 5 of 16 stations over 1993–2014; there are no significant negative trends (Supplementary Figure S9~~Supplementary Figure S 6~~, Supplementary Table S1) and the regional differences in the trends are small. The average normalized E_{pan} series shows a highly significant ($p \leq 0.01$) increase of 29 ± 5 mm y^{-1} or 6.0 ± 1.0 % decade^{-1} over 1977–2014 (Fig. 5).

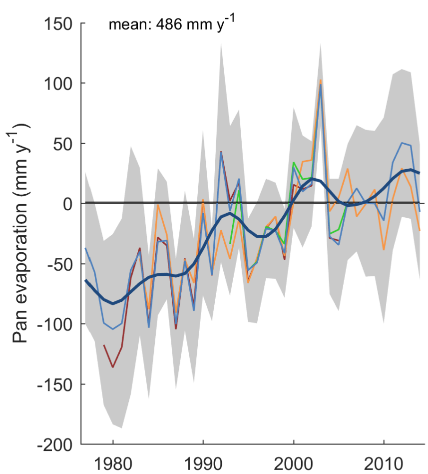


Fig. 5 Pan evaporation anomalies of 25 stations with a minimum of 20 years available data during 1977–2014. Thin blue line: mean, grey area: ± 1 standard deviation, thick blue line: filtered mean (Gauss filter with a standard deviation of 2 years). The red (orange, green) lines show means for subsets of stations for 1979–2005 (1983–2014, 1993–2014; maximum of 5 years missing during the period).

3.2.3 Changes in vegetation activity

Catchment average NDVI shows a clear seasonal cycle with high values in summer and low values in winter (Fig. 6a, Supplementary Figure S10~~Supplementary Figure S 7~~). Mean trends in NDVI over all catchments for 15-day composites are positive nearly over the entire year and particularly strong during March/April and November/December (Fig. 6b). Timing of the trends is correlated to median catchment elevation. During October–March, positive trends are mostly observed in catchments with low median elevation, while during May–June, this reverses and stronger positive trends are observed in high elevation catchments (Supplementary Figure S11). The average and standard deviation over all catchments of the trend in the average annual NDVI is 0.02 ± 0.01 decade^{-1} or 3.1 ± 1.1 % decade^{-1}.

To estimate the effect of these vegetation changes on E we calculated E_{0v} with r_s estimated based on the original observed NDVI data and E_{0c} with r_s based on detrended NDVI data. E_{0v}, which reflects changes in vegetation activity and atmospheric conditions, showed a stronger increase over the study period than E_{0c}, which reflects changes in atmospheric conditions only (Fig. 7a, b). Estimated as average and standard deviation over all catchments and over both approaches for estimating r_s, the contribution of changes in atmospheric conditions to the trend in E_{0v} was 56 ± 15 % and the contribution of changes in vegetation to the trend in E_{0v} was 44 ± 15 % (Fig. 7c).

Changes in annual cumulative NDVI are not correlated to changes in E_{wb} (r^2=0.01, p=0.23) (Fig. 8b). This was also the case for partial correlations between trends in E_{wb} and NDVI, when trends in annual precipitation or trends in E_0 were taken into account.

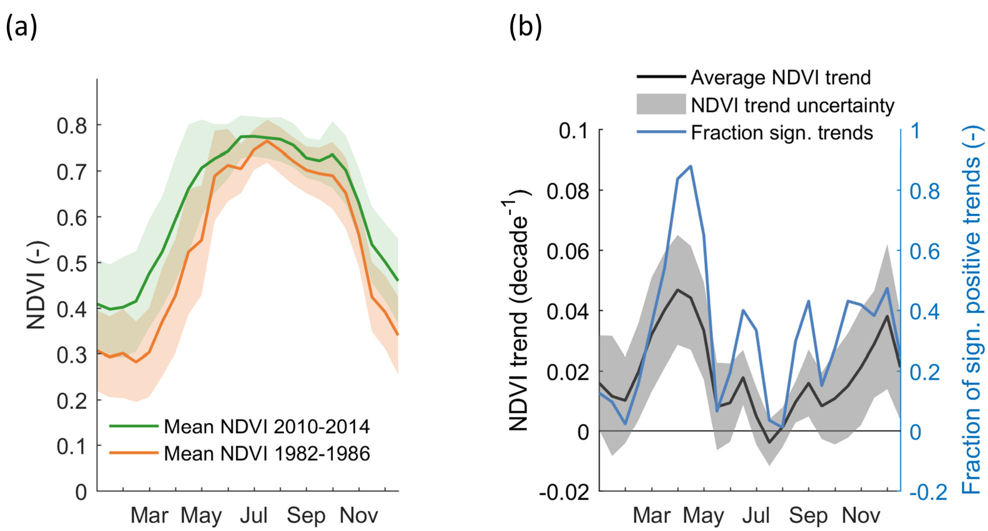


Fig. 6 Changes in NDVI. (a) Seasonal cycle of NDVI averaged over 1982–1986 and 2010–2014. Solid lines show averages over all catchments and shaded areas show ± 1 standard deviation. (b) Seasonal cycle of trends in catchment average NDVI values over 1982–2014. The solid line shows the mean NDVI trend over all catchments, the grey shaded area the spatial variability of the trends between catchments (± 1 standard deviation), and the blue line the fraction of significant positive trends (p≤0.05).

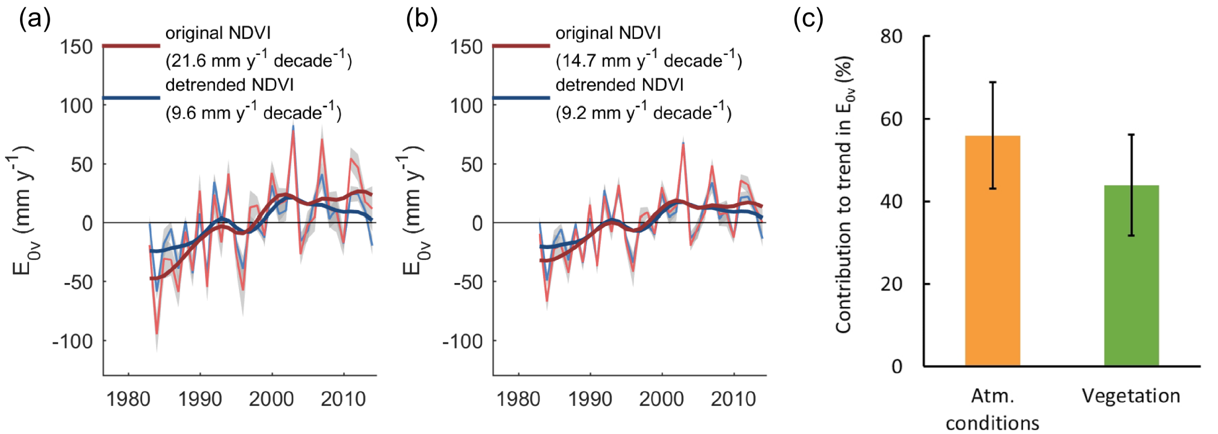


Fig. 7 Effect of variations in atmospheric conditions and vegetation on E_{0v} (a, b) Anomalies of E_{0v} with r_s estimated with original and detrended NDVI data, (a) r_s estimated based on Sellers et al. (1996), (b) r_s estimated based on Zhang *et al* (2010). The thin line shows the mean over all catchments, the grey shaded area shows the variability between catchments (± 1 standard deviation), and the thick blue line shows the filtered mean (Gauss filter with a standard deviation of 2 years). The number in brackets gives the trend estimate over 1982–2014. (c) Contributions of variations in atmospheric conditions and variations in vegetation to the trend in E_{0v}. Bars show means over all catchments and over both approaches for estimating r_s, error bars show the variability over all catchments and over both approaches for estimating r_s (± 1 standard deviation). Percent are relative to trends in E_{0v}.

3.2.4 Changes in ~~soil moisture~~precipitation

We used annual precipitation as a proxy for the water available for E. Trends in annual precipitation are described in Sect. 3.1. Higher increases in E_{wb} are observed in catchments with higher increases in annual precipitation (r^2=0.24, p<0.0001; Fig. 8c). The trends in precipitation are not related to trends in NDVI (r=-0.01, p=0.87), which suggests that the relationship between trends in E_{wb} and trends in precipitation does not include indirect effects of changes in precipitation on trends in E_{wb} through changes in vegetation activity. Trends in precipitation are also not related to trends in E_0 (r=-0.12, p=0.14). A linear regression of the trend in E_{wb} against the trend in annual precipitation results in a slope of 0.30 ± 0.04. This relationship may however overestimate the influence of changes in precipitation on changes in E_{wb} since E_{wb} is derived from the water balance and trends in E_{wb} are thus not independent from trends in precipitation. The magnitude of this overestimation effect was estimated using Monte Carlo simulations as 0.08 ± 0.03 (see Supplement S3) and the estimate for the sensitivity of trends in E_{wb} to trends in precipitation was therefore corrected to 0.22 ± 0.05. This suggests that 1 mm y^{-2} increase in precipitation is associated with 0.~~30~~22 ± 0.~~04~~05 mm y^{-2} increase in E_{wb}. Thus, with an average precipitation trend of 32 mm y^{-1} decade^{-1}, on average ~~9.4~~6.9 ± ~~1.4~~1.6 mm y^{-1} decade^{-1} (uncertainty ~~relates~~ derived from the~~to~~ standard deviation of the trend slope and of the correction value) of the E_{wb} trend may be related to the increase in precipitation.

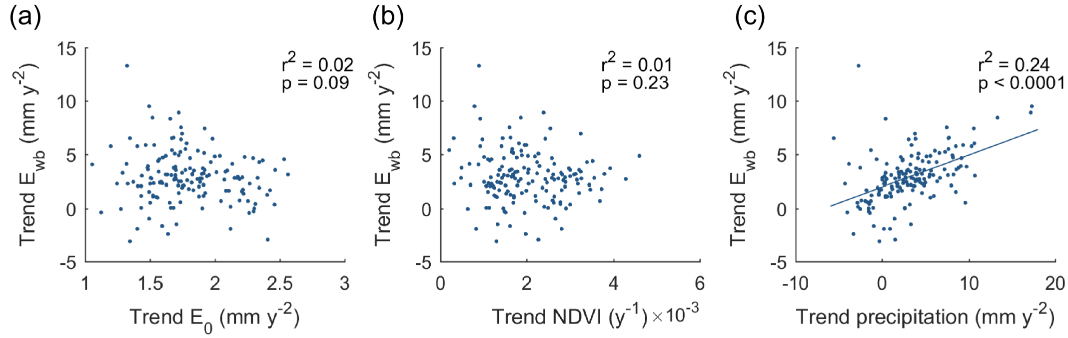
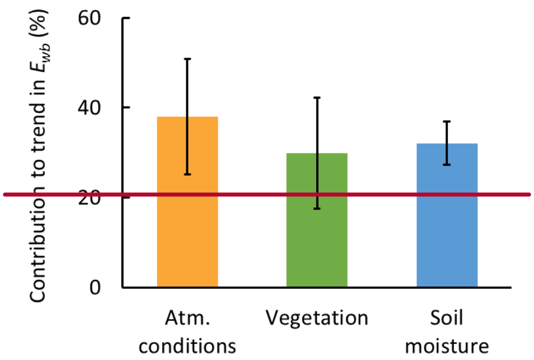


Fig. 8 Scatter plots of trends in (a) E_0, (b) NDVI and (c) annual precipitation against the trend in E_{wb}.

3.3 Synthesis of attribution

We may now estimate the contributions of the different drivers to the increase in E_{wb}. ~~From~~ T~~t~~he regression of the trend in E_{wb} against the trend in annual precipitation (Fig. 8c) ~~we found~~suggests that, on average, 6.9 ± 1.6~~9.4 ± 1.4~~ mm y^{-1} decade^{-1} of the E_{wb} trend of 29.3 mm y^{-1} decade^{-1} ~~are~~ may be related to the increase in precipitation (Sect. 3.2.4). The relative contributions of atmospheric conditions and vegetation were assumed to conform to their relative effects on E_{0v} (Sect. 3.2.3, Fig. 7c). Thus the remaining 22.4 ± 1.6~~19.9 ± 1.4~~ mm y^{-1} decade^{-1} are split at a ratio of 0.56 ± 0.15 to 0.44 ± 0.15 into being due to atmospheric conditions and vegetation, respectively. This results in a contribution of changes in atmospheric conditions of 12.5 ± 4.2~~11.1 ± 3.8~~ mm y^{-1} decade^{-1} and in a contribution of changes in vegetation of 9.8 ± 4.0 mm ~~8.8 ± 3.6~~ mm y^{-1} decade^{-1}. In summary, the data suggest that changes in atmospheric conditions, vegetation activity, and ~~soil moisture~~ precipitation have contributed 43 ± 15 ~~38 ± 13~~ %, 34 ± 14 ~~30 ± 12~~ %, and 24 ± 5 ~~32 ± 5~~ %, respectively, to the average increase in E_{wb} in the study catchments (Fig. 9).



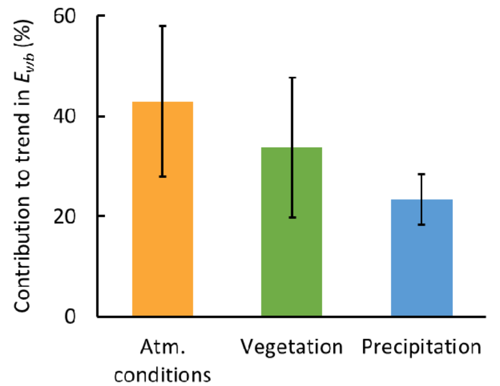


Fig. 9 Average contributions to the average trend of catchment evaporation estimated from the water balance (E_{wb}) from changes in atmospheric conditions, vegetation activity, and ~~soil moisture~~precipitation. Error bars relate to the standard deviation of the estimate. Percent relate to the total average trend of E_{wb} of 29 mm y^{-1} decade^{-1}.

4 Discussion

4.1 Changes in catchment evapo~~transpi~~ration in Austria over the past four decades

Based on the analysis of the water balances, we found an average E_{wb} increase of 29 ± 14 mm y^{-1} decade^{-1}, i.e. a total increase of 108 ± 52 mm y^{-1} or 18.0~~7.9~~ ± 8.5 % over 37 years. This increase is consistent between the different regions, which points towards the importance of drivers with spatially consistent changes across the study region, i.e. global radiation or air temperature rather than changes in precipitation. The increase is strongest in the beginning of the study period and seems to have stopped around 2000. A similar pattern is observed for E_0 and E_{0v} (Fig. 3a and Fig. 7a,b) where the decreasing tendency in global radiation (Fig. 3b) appears to be the main cause for the decreasing tendency after 2000.

The E_{wb} estimates could potentially be influenced by changes in groundwater storage. Changes in groundwater storage were assumed to be small over time scales of decades. This assumption is supported by the absence of general trends in discharge, together with the incoherent picture of trends in groundwater levels (Blaschke et al., 2011; Neunteufel et al., 2017), which makes it unlikely that groundwater storage changes have a big influence on the E_{wb} estimates.

Increases in E_{wb} of a similar magnitude have also been observed for Switzerland, where E_{wb} increased by ~20 mm y^{-1} decade^{-1} over 1977–2007 (Spreafico et al., 2007, from Figure 1 on plate 6.6). Lower rates of increase in E_{wb} were observed in other regions and other periods. Estimates based on the water balance were 10 ± 5 mm y^{-1} decade^{-1} in several large catchments across the conterminous US during 1950–2000 (Walter et al., 2004), 6 ± 4 mm y^{-1} decade^{-1} in catchments in the eastern US during 1901–2009 (Kramer et al., 2015), and 7 mm y^{-1} decade^{-1} in 16 catchments on the Tibetan Plateau during 1966–2000 (Zhang et al., 2007). The larger increases of E in our study may be related to the large increases in air temperature and global radiation in the study region over the period considered.

4.2 Drivers of the observed changes in catchment evapo~~transpi~~ration

All three drivers investigated – changes in atmospheric conditions, changes in vegetation activity and changes in ~~soil moisture~~precipitation – are found to be important for the observed increase in catchment evapo~~transpi~~ration~~and the direct effects of these drivers on E are in a similar order of magnitude~~. The drivers are closely interlinked (Fig. 10). For example, increases in air temperature ~~and precipitation~~ not only contribute directly to changes in E_0 and thus E, but also indirectly through increases in the vegetation activity. Attributing changes in E to the drivers can therefore be done in different ways. While, in this study, vegetation effects on E include land use changes and indirect effects via climate and atmospheric CO_2 on vegetation, other studies have treated the indirect effects separately. Using a global biosphere model, Piao et al. (2007) attributed a global change in E of +0.3 mm y^{-1} decade^{-1} during 1901–1999 to climate (+0.7 mm y^{-1} decade^{-1}; including indirect effects of climate on vegetation, atmospheric conditions and precipitation) and land use changes (−0.8 mm y^{-1} decade^{-1}; mainly deforestation). Changes in atmospheric CO_2 had a further positive effect of +0.4 mm y^{-1} decade^{-1} through increasing LAI and stomata resistance (Piao et al., 2007). Based on global land surface model simulations, Mao et al. (2015) found that climate effects (including indirect effects of climate on vegetation, atmospheric conditions and precipitation) on E were larger than the effects of land use, atmospheric CO_2 and nitrogen deposition during 1982–2013, with precipitation being the most important climate variable. The latter finding is in line with several other studies (Jung et al., 2010; Miralles et al., 2014) although Zhang et al. (2015) found vegetation change (including indirect effects of climate on vegetation) to be more important.

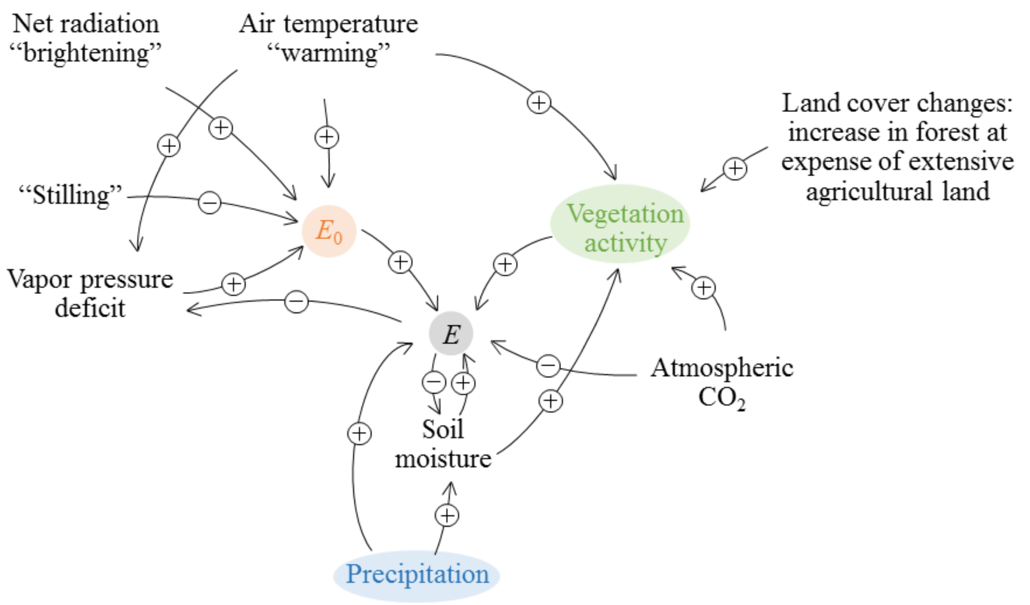


Fig. 10 Drivers of changes in evaporation, *E*, including feedback effects between them.

In this study, increases in E driven by changes in atmospheric conditions have been identified by increases in E_0 and E_{pan}. The main driver for the increase in E_0 in our study is an increase in global radiation of on average 5.1 ± 0.9 W m^{-2} decade^{-1} with a further contribution from an increase in mean air temperature of 0.45~~8~~ ± 0.09~~11~~ °C decade^{-1}. Even though the scatter is large, the data suggest that the predominantly decreasing wind speeds have likely reduced the increase in E_0. While attribution studies summarized in McVicar et al. (2012; Table 7) found trends in wind speed to be the most frequent dominant driver for changes in E_0 and E_{pan} (followed by air temperature), trends in wind speed play a minor role in our study region. Teuling et al. (2009) and Wang et al. (2010) reported a strong influence of global radiation on changes in E in Europe. Global radiation in Europe generally decreased during the 1960s–1980s ('global dimming') and increased from the 1980s ('global brightening') (Norris and Wild, 2007; Wild, 2009; Sanchez-Lorenzo et al., 2015). The magnitude of the increase in the present study is slightly higher than those reported in other studies (5.1 ± 0.9 W m^{-2} decade^{-1} in this study; and 2.0 ± 1.2 W m^{-2} decade^{-1} in Central Europe during 1971–2012 found by Sanchez-Lorenzo et al. (2015) and 3.7 W m^{-2} decade^{-1} in Austria/Switzerland during 1985–2005 found by Wild et al. (2009)). The increases in global radiation over Europe since the mid 1980s have mainly been attributed to the reduction in aerosols (Norris and Wild, 2007). The strong sensitivity of E_0 to global radiation suggests that projected air temperature increases do not necessarily imply increased evaporation in the future.

The observed increase in E_{pan} in our study is consistent with other European studies. In Greece, non-significant increases in E_{pan} over 1983–1999 were observed for a pooled series from 14 stations (Papaioannou et al., 2011). Three out of eight sites in

Ireland showed a significant increase in E_{pan} over 1963–2005, and one site showed a significant decrease (Stanhill and Möller, 2008). In England, significant increases in E_{pan} were observed at two stations over 1957–2004 and 1986–2010, respectively (Stanhill and Möller, 2008; Clark, 2013), and in the Czech Republic April–June E_{pan} significantly increased at three out of five stations during 1968–~~-~~2010 (Trnka et al., 2015).

The NDVI data shows a marked increase in vegetation activity in Austria, similar to many other studies in the Northern Hemisphere (Myneni et al., 1997; Slayback et al., 2003; Liu et al., 2015). Generally, i~~I~~ncreases in air temperature and precipitation and CO_2 fertilization have been identified as drivers (Piao et al., 2006; Los, 2013). In the study area, trends in precipitation and trends in NDVI show no significant relationship (*r*=-0.01), suggesting that increases in precipitation were not important for the changes in vegetation activity. We found ~~the~~ strong~~est~~ increases in NDVI in spring and autumn, indicating a lengthening of the active growing season which has, e.g., been noted by Myneni et al. (1997). Increases in NDVI may further be enhanced by land cover changes. In Austria, the forest area increased from 43 % to 47 % over 1977–2010 at the expense of cropland and extensive grassland (Krausmann et al., 2003; Gingrich et al., 2015) as agricultural land in remote areas with low productivity was abandoned due to economic pressure (Tasser and Tappeiner, 2002; Rutherford et al., 2008).

According to ~~this~~ our study, the effect of increased vegetation activity on E_{0v} is of a similar magnitude as the effect of changes in the atmospheric conditions. A strong influence of changes in vegetation activity on E was also suggested for the eastern US based on correlations between NDVI and E_{wb} (Kramer et al., 2015). In two forested catchments in the southern Appalachians, indirect effects of climate on vegetation dynamics were found to be a much more important driver for long-term increases in E (plus potential storage changes) than direct climate impacts (Taehee et al.). Using simulations from a global, coupled biosphere-atmosphere model, Bounoua et al. (2000) found an E increase of 43 mm y^{-1} for an NDVI increase of 0.08, which is slightly higher than the average NDVI increase observed in this study of 0.06. ~~This~~ Our study, however, does not account for the effect of increasing atmospheric CO_2 concentrations on increasing stomata resistance, which means that the effect of vegetation might be overestimated. Stomata closure due to increased atmospheric CO_2 may have reduced global E since 1960 by 1.6 to 2.0 mm y^{-1} decade^{-1} (Gedney et al., 2006; Piao et al., 2007).

Even though the study area can generally be classified as humid, we found a strong influence of changes in precipitation on changes in E. ~~While~~ E is generally energy-limited in the study region (Sect. 2.1.1). However, as indicated by E_{wb}/E_{max}~~,~~ ratios smaller than unity, E may frequently be limited by available moisture, particularly interception and E from the soil and non-vegetated areas. Thus, an ~~so that an~~ increase in ~~soil moisture~~ precipitation can lead to an increase in E ~~(Parajka et al., 2007; Parajka et al., 2009)~~. One reason for the strong sensitivity of changes in E to the increases in annual precipitation is the seasonality of the observed changes in precipitation. Increases in precipitation were concentrated in the summer season and changes in summer precipitation are expected to contribute more strongly to changes in E as this is the period when E is highest, whereas changes in winter precipitation more likely result in changes in discharge. It should be noted that the strength of the relationship between changes in E and changes in precipitation would be overestimated if there were artifacts in the trends of the precipitation data, for example due to inconsistent record lengths. In this study, consistent record lengths were used. Several studies around the world and in particular in the tropics identified precipitation as an ~~At the global scale and in particular in the tropics, several studies found that changes in precipitation were the most~~ important factor for changes in E (Jung et al., 2010; Miralles et al., 2014; Mao et al., 2015).

5 Conclusions and implications

Over the past four decades (1977–2014), catchment evapo~~transpi~~ration increased on average over 156 study catchments in Austria by 29 ± 14 mm y^{-1} decade^{-1}. This increase was attributed ~~with similar orders of magnitude~~ to changes in atmospheric demand and available energy (~~that~~ suggested to account~~ed~~ for 12.5 ± 4.2~~11.1 ± 3.8~~ mm y^{-1} decade^{-1}), changes in vegetation (9.8 ± 4.0 ~~8.8 ± 3.6~~ mm y^{-1} decade^{-1}), and changes in ~~soil moisture~~precipitation (6.9 ± 1.6~~9.4 ± 1.4~~ mm y^{-1} decade^{-1}).

E_0 increased on average over all study catchments by 18 ± 5 mm y^{-1} decade^{-1}. The increase in E_0 was largely driven by the increase in global radiation with further contributions from increasing air temperature. Rising atmospheric demand and energy available for E was also revealed by increases in the available E_{pan} data. Satellite-derived NDVI data for 1982–2014 indicate an increase in vegetation activity. This increase may have led to a similar increase of E_{0v} as the climate variables. A positive correlation between increases in E and increases in precipitation furthermore points to increases in ~~soil~~ water availability as a third driver for the increases in E_{wb}.

Over the study period, trends in annual discharge were close to zero and increases in E were balanced by increases in precipitation. If the increase in precipitation had been lower and the increase in E had been similar as in this study, notable reductions in discharge would have been likely. A lower increase in precipitation would likely have reduced the increase in E as 24 ± 5 %~~about a third~~ of the E increase was directly attributed to the increase in precipitation. ~~Furthermore, the increase in E due to the increase in vegetation activity might have been slightly lower without the increase in precipitation.~~

Estimates of future changes in E of climate impact assessments are often based on predicted air temperature and precipitation changes. This study clearly shows that, despite the large air temperature increase in the recent decades, global radiation was much more important for changes in E_0 than air temperature. Potential future changes in global radiation, due to e.g. changes in cloud cover or air pollution, should therefore be explicitly be accounted for in climate impact studies on E. ~~Climate impact studies on E should therefore explicitly account for possible changes in global radiation.~~ Furthermore, hydrologic models used in such studies should consider the effects of possible changes in vegetation on E that for example result from a longer growing period.

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Supplement

Why has catchment evaporation increased in the past 40 years? A data-based study in Austria

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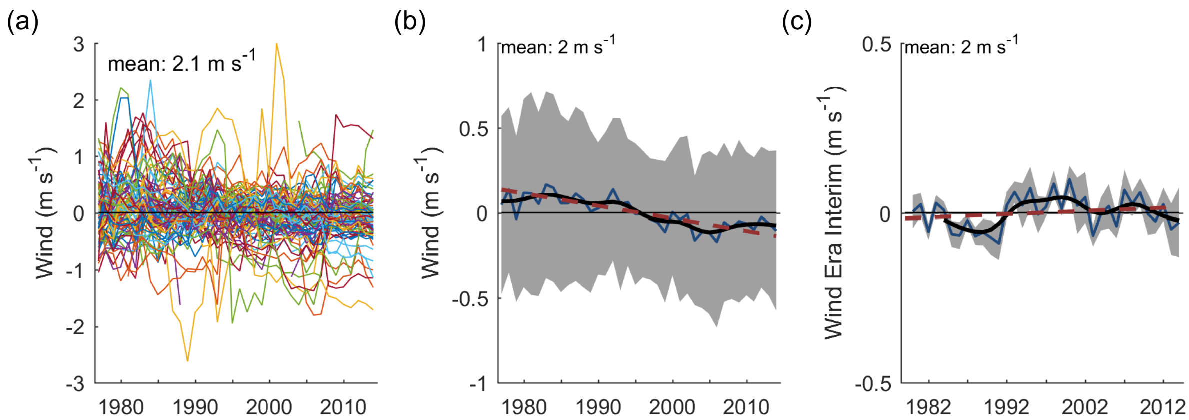
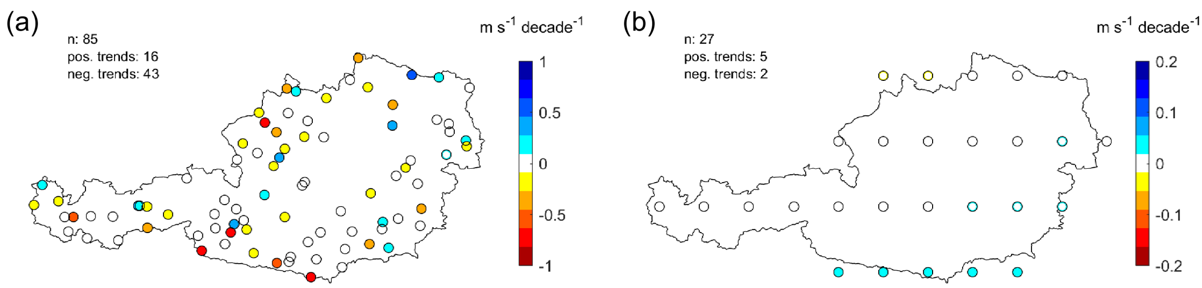
Supplement S1

Wind data

Wind data were regarded as not representative with respect to trends. The reasons for this are: i) annual anomalies of wind speed data from 85 stations in Austria appear unrelated to each other (Supplementary Figure S 1a) and temporal trends over 1977–2014 do not show any spatial pattern (Supplementary Figure S 2a); ii) averaged anomalies of annual wind speeds from station data and ERA Interim data (Dee et al., 2011) show for most part of the series opposing patterns (Supplementary Figure S 2); and iii) wind data are known to be prone to inhomogeneities (Böhm, 2008). We therefore used uniform monthly wind speeds averaged over all years and over all stations in Austria in this study. The potential effect of changes in wind speed was analyzed in Supplement S2.



Supplementary Figure S 1 Anomalies of wind speeds(a, b) from station data (85 stations) over 1977–2014 (c) from ERA Interim data over Austria (27 grid points) over 1980–2014. (a) Each line refers to one station. (b, c) The thin blue line shows the mean over all catchments, the grey shaded area the variability between catchments (± 1 standard deviation), the bold black line the smoothed mean, and the red dashed line a trend line.



Supplementary Figure S 2: Spatial pattern of trends in wind speed (a) from 85 stations over 1977–2014 (b) from ERA Interim data over Austria over 1980–2014. Filled circles indicate significant trends at *p*≤0.05.

Supplement S2

Analysis of the potential influence of trends in wind speed on reference evaporation

Calculation of E_0 including trends in wind speed

We performed two analyses on the effect of trends in wind speed on E_0. In the first analysis, we applied average monthly trends derived from station observations of wind speed to the wind speeds used in the original analysis. Heterogeneities in the observations of wind speed (measured at 10 m height) were identified as periods where all annual averages deviate by more than three standard deviations from the rest of the series (Vautard et al., 2010) and removed from the series. Series with three or more missing years were removed from the data set, which resulted in a data set of 58 stations. We then derived relative trends in monthly wind speeds, averaged these over all stations, and applied them to the wind speeds used in the original analysis.

The second analysis aimed at including spatial heterogeneities in wind speed and its trends. Due to the high spatial heterogeneity of wind speeds, spatial fields of wind speed cannot be directly inferred from the interpolation of the station observations. Spatial fields of average monthly wind speeds were therefore derived as monthly averages from the high-resolution reanalysis data set COSMO-REA6 (Bollmeyer et al., 2015; Kaiser-Weiss et al., 2015), which is based on ERA-Interim (Dee et al., 2011), has a horizontal resolution of about 6 km, and is available during 1995–2015.

Trends in wind speed were again derived from station observations since downscaled reanalysis data can only capture trends in wind speeds caused by atmospheric circulation changes. Trends caused by changes in surface roughness, caused for example by changes in land use, cannot be represented (Vautard et al., 2010). The relative trends in monthly wind speeds derived from the station observations were interpolated onto a 1 km grid. In order to focus on the general patterns (i.e. obtain smooth surfaces), a weighted linear least-squares regression (LOWESS method) with a span of 0.75 was used. Gridded fields (1 km resolution) of monthly wind speeds for the period 1977–2014 that represent the general monthly trends were estimated by multiplying absolute values of average monthly wind speeds derived from the reanalysis data with the relative trends derived from the station data:

$$u_{\mathrm{m,t}}(x,y) = u_{\mathrm{m}}(x,y) \cdot \left(1 + (t - t_{\mathrm{ave}}) * \frac{\tau_{\mathrm{m}}(x,y)}{100}\right)$$

where $u_{\mathrm{m,t}}(x,y)$ is the wind speed at point (x,y) in month *m* and year *t*, $u_{\mathrm{m}}(x,y)$ is the average monthly wind speed (m s^{-1}) at point (x,y) derived from the reanalysis data, $\tau_{\mathrm{m}}(x,y)$ is the trend in monthly wind speed derived from the station data (% y^{-1}), t_{ave} is the year represented by the average monthly wind speeds (2004).

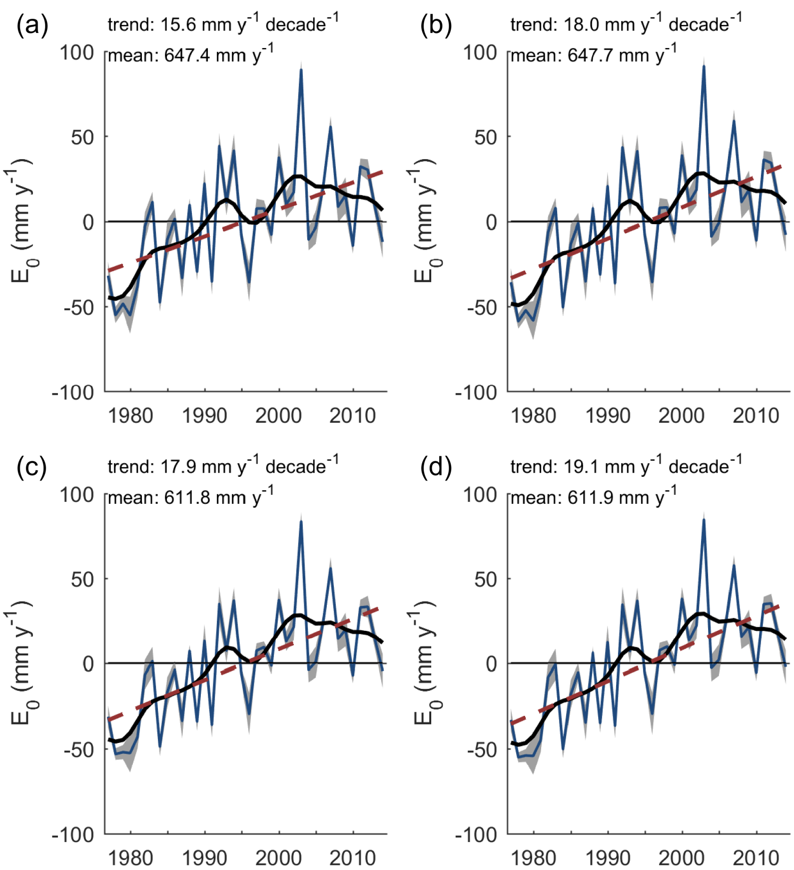
For both analyses, E_0 was calculated i) including trends in wind speed, and ii) with average monthly wind speeds of 1994 on a 1 km grid and summarized to catchment averages.

Trends in wind speed and their effect on reference evaporation

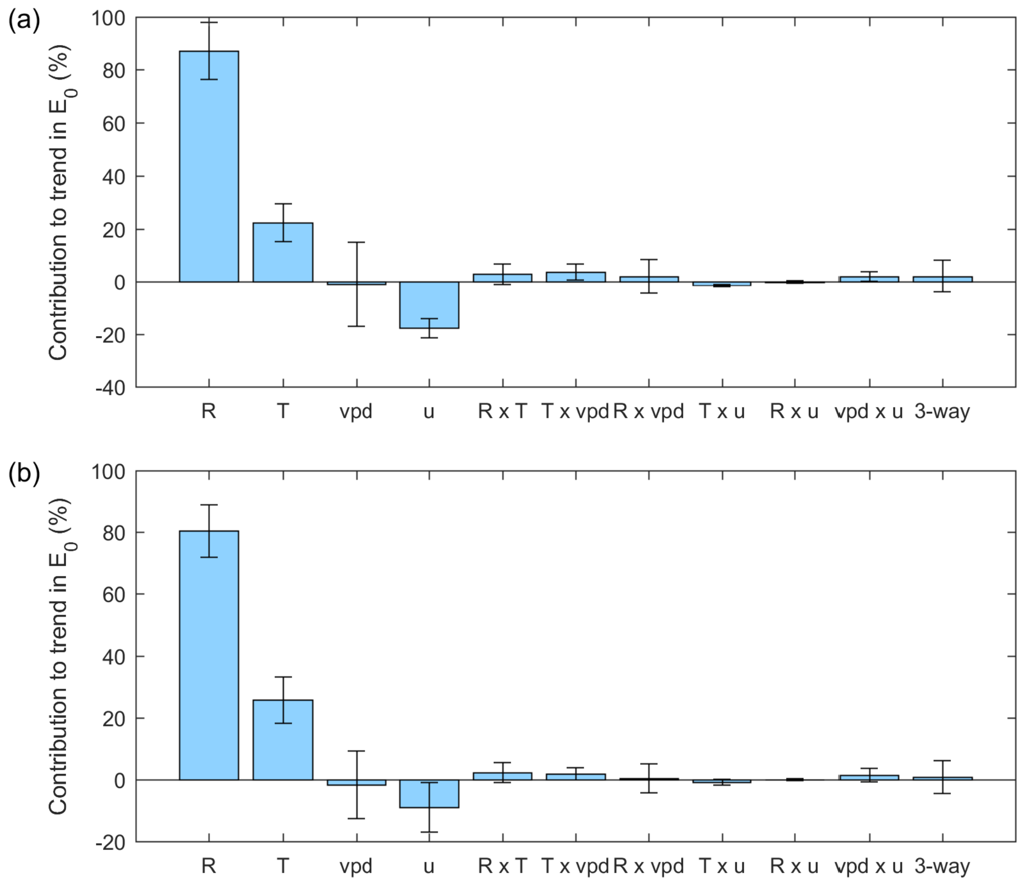
The annual trend in wind speed averaged over all stations is -0.07 ± 0.06 m s^{-1} per decade (or -3.0 ± 2.5 % per decade) during 1977–2014, which is within the range of the values reported by McVicar et al. (2012) for Europe and very similar to the trend over 276 stations in Europe by Vautard et al. (2010) (-0.09 m s^{-1} per decade or -2.9% per decade during 1979–2008). Trends vary strongly from station to station without a clear spatial pattern. However, the smoothed spatial patterns of trends in wind speed indicate negative wind trends particularly in the southwest of Austria and very small negative or no trends in the east.

The effect of the trends in wind speed on E_0 is small. According to the first analysis, the trend in E_0 averaged over all catchments is 2.4 ± 0.7 % per decade when allowing for decreasing wind speeds, as compared to 2.8 ± 0.7 % per decade when assuming no trends in wind speed (Supplementary Figure S 3). The direct effects of the changes in net radiation, air temperature and wind speed contributed 87 ± 11 %, 22 ± 7 %, and -18 ± 4 % to the trend in E_0 (Supplementary Figure S 4a). In the second analysis, E_0 estimates and trends in wind speed were lower due to lower wind speeds in the reanalysis data compared to the averages of the station data, which led to a smaller effect of the trends in wind speed on E_0 than in the first analysis. When allowing for decreasing wind speeds, the average trend in E_0 is 2.9 ± 0.6 % per decade, as compared to 3.1 ± 0.6 % per decade when assuming no trends in wind speed (Supplementary Figure S 3). The direct effects of the changes in net radiation, air temperature and wind speed contributed 80 ± 9 %, 26 ± 7 %, and -9 ± 8 % to the trend in E_0 (Supplementary Figure S 4b).

The low influence of the changes in wind speed on E_0 can be explained by the generally humid climate in Austria, where wind speed has a much lower impact on E_0 than in an arid climate (Irmak et al., 2006). The estimated effect of decreasing wind speed may be even lower when the Penman-Monteith equation is coupled to an atmospheric model (van Heerwaarden et al., 2010).



Supplementary Figure S 3: Anomalies of E_0 (a, c) considering trends in wind speed (b, d) with wind speeds as of 1994 for all years. (a, b) refers to an analysis that applied average monthly relative trends to the wind speeds used in the original study. (c, d) refers to a second analysis that considered spatial heterogeneities in wind speed and its trends. The thin blue line shows the mean over all catchments, the grey shaded area shows the variability between catchments (± 1 standard deviation), the bold black line shows the filtered mean (10-year Gauss filter with a standard deviation of 2 years), and the dashed red line the linear trend.



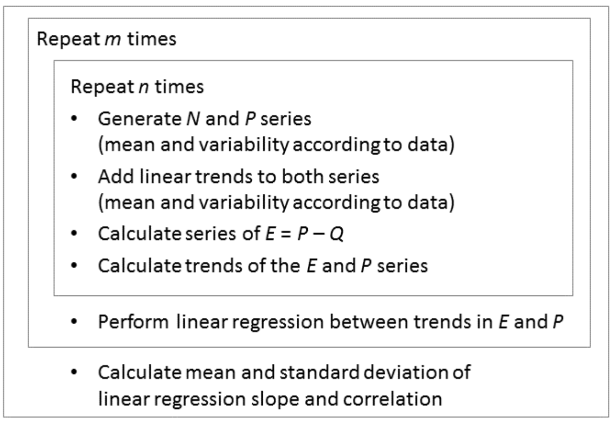
Supplementary Figure S 4: Mean contributions of variations in net radiation (R), air temperature (T), vapor pressure deficit (vpd), wind speed (u), their two-way interaction effects and all three way interaction effects (3-way) to the trend in E_0. Bars show means over all catchments, error bars show the standard deviation of the variation between catchments. Percent are relative to trends in E_0. (a) refers to an analysis that applied average monthly relative trends to the wind speeds used in the original study, (b) refers to a second analysis that considered spatial heterogeneities in wind speed and its trends.

Supplement S3

Monte Carlo simulations for estimating the overestimation of the regression relationship between trends in precipitation and trends in evaporation

Since E_{wb} is estimated from precipitation (P) and discharge (Q), trends in E_{wb} are not independent from trends in P and a regression relationship between these two variables may overestimate the effect of trends in P on trends in E_{wb}. We therefore performed Monte Carlo simulations that aimed at investigating the strength of the relationship between trends in P and trends in Q resulting from the dependency of the two variables when assuming that trends in E are independent of trends in P. The results depend on the assumed statistical properties of the data, amongst others on the spatial variability of the trend in P (the stronger the spatial variability of the P trend, the weaker the relationship when assuming trends in E independent of trends in P). For the Monte Carlo simulations, we generated n correlated, normal distributed series of annual P and annual Q (Supplementary Figure S 5). Means and standard deviations of annual P and Q, and the covariance between annual P and Q were set according to the data set of this study, and n was set to the number of study catchments. Trends in P were considered by adding linear trends to the P series. The variability of the trends between catchments was assumed normal distributed and the mean and variability were derived from the study data. In accordance with the assumption that trends in E are independent of trends in Q, the trend added to the P series was also added to the associated Q series. Annual E series were calculated as P minus Q. Trends in the E and P series were estimated using Sen's slope. We performed a linear regression between trends in E and trends in P over the n data points and calculated the slope and the coefficient of determination. This procedure was repeated m=1000 times and the mean and standard deviation of the slope and the coefficient of determination over these m repetitions were calculated.

The resulting regression relationships have a slope of 0.08 ± 0.03 (mean ± standard deviation) and a correlation coefficient of 0.06 ± 0.04, suggesting that the slope derived from the regression of E_{wb} against P overestimates the sensitivity of changes in E to changes in P by 0.08 ± 0.03. The sensitivity of the trend in E_{wb} to trends in P was therefore estimated as the slope of the linear regression between trends in E_{wb} and trends in P corrected by this value.



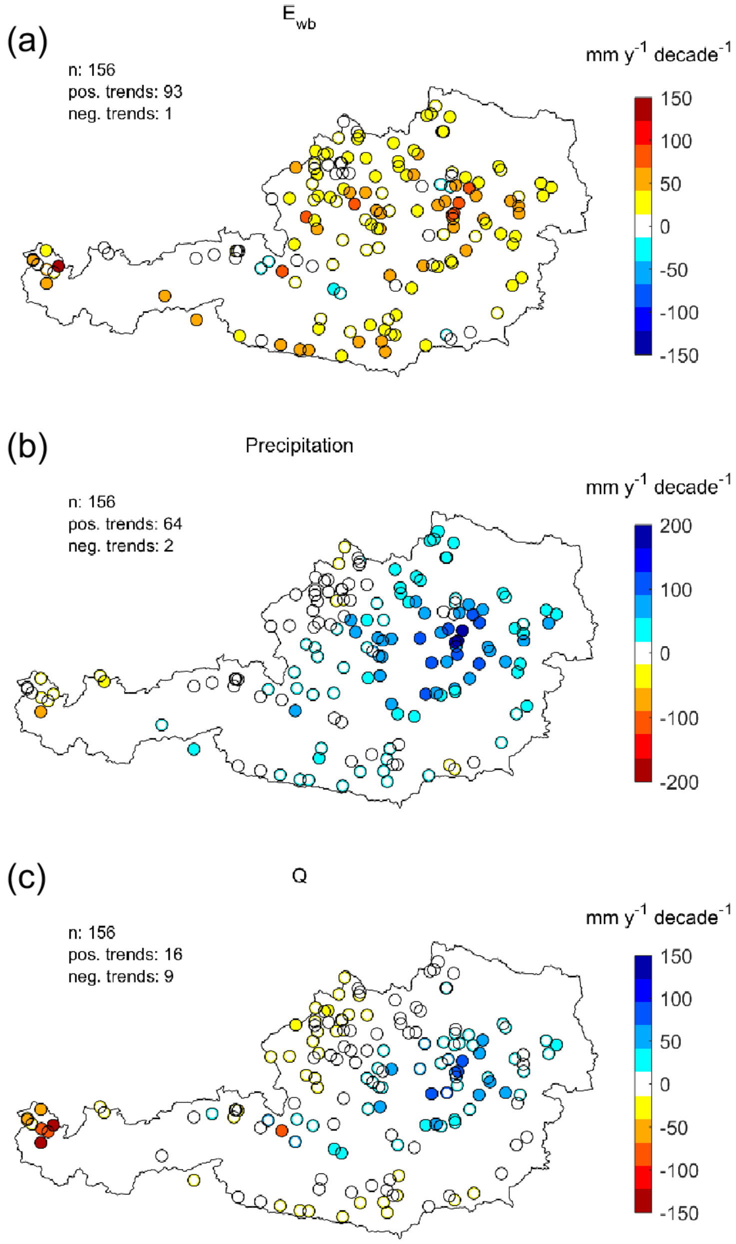
Supplementary Figure S 5: Estimating the overestimation of the regression relationship between trends in precipitation and trends in evaporation, caused by the dependency of the precipitation and evaporation series, by Monte Carlo simulations.

Supplement S4

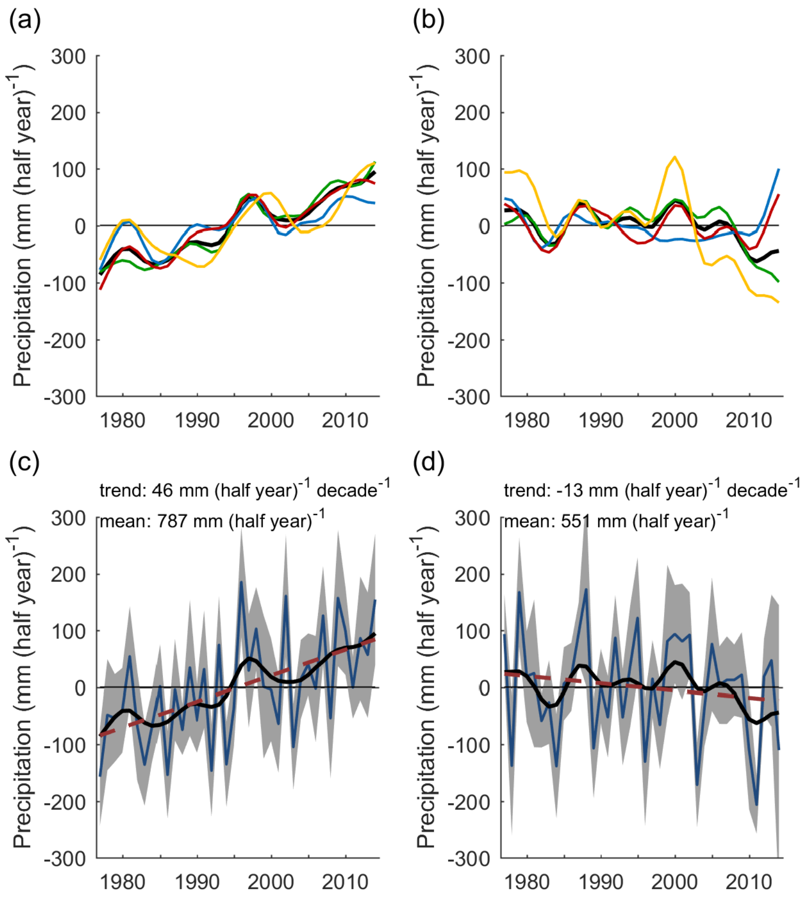
Further supplementary tables and figures

Supplementary Table S 1 Trends in summer (May-Oct) pan evaporation of individual stations for three periods. The table includes the mean, standard deviation (Std), and coefficient of variation (CV) of summer pan evaporation over the available period.
5 **Stars indicate trend significance: *** $p \leq 0.01$, ** $p \leq 0.05$, * $p \leq 0.1$**

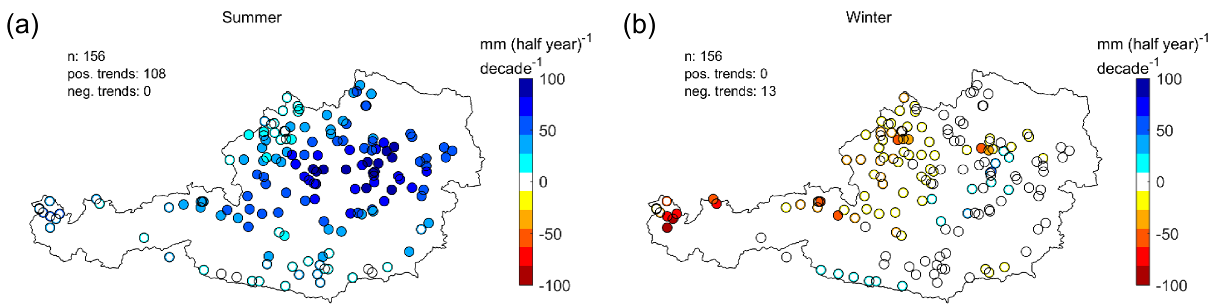
Name	Source	Elev. (m)	East (°)	North (°)	Mean (mm)	Std (mm)	CV (-)	Year start	Year end	Missing data	Trend (% decade^{-1}) 1979-2005	1983-2015	1993-2015
Elmen-Martinau	HZB Tirol	954	10.54	47.36	348	43	0.12	1982	2014	1983	-0.5	-1	3.1
Leutasch-Kirchplatzl	HZB Tirol	1135	11.14	47.37	373	41	0.11	1982	2014	2007	0.6	5.9**	12.2***
Ladis-Neuegg	HZB Tirol	1350	10.65	47.10	400	39	0.10	1982	2014		5.7	2.6	-4.4
St.Johann	HZB Tirol	667	12.44	47.52	336	51	0.15	1982	2014		12.0***	13.1***	13.5***
Aschau	HZB Tirol	1005	12.31	47.38	290	34	0.12	1982	2014		0.2	-2.6	-3.1*
Stetten	HZB Tirol	179	16.38	48.37	369	63	0.17	1992	2014		-	-	9.3
Franzensdorf	HZB Tirol	152	16.64	48.19	382	52	0.14	1992	2014	2010,2011	-	-	-2.9
Hochberg	HZB Tirol	1672	12.36	46.82	373	44	0.12	1983	2014	2008	13.8***	6.2***	3.4
Prägraten	HZB Tirol	1340	12.38	47.02	346	37	0.11	1983	2014		10.8***	4.4*	-1.2
Matrei	HZB Tirol	1040	12.54	47.00	317	48	0.15	1983	2013	2010	-	12.2***	10.1***
Waidring	HZB NÖ	775	12.55	47.59	303	50	0.17	1993	2014	2011	-	-	14.3**
Lunz	HZB NÖ	611	15.07	47.86	292	48	0.17	1992	2014	2010,2013	-	-	-2.1
Frankenfels	HZB NÖ	468	15.33	47.98	293	61	0.21	1993	2014	2007	-	-	22.5***
Ottenstein	HZB NÖ	554	15.34	48.58	336	35	0.10	1994	2014		-	-	5.9
Pyhra	HZB NÖ	298	15.70	48.15	404	44	0.11	1993	2014	2000,2001	-	-	4.4
Hollenthon	HZB NÖ	685	16.26	47.59	369	53	0.14	1993	2014		-	-	6.5
Retz	ZAMG	242	15.95	48.77	424	63	0.15	1975	2005	1987,1995,1996,1997,1998	6.9	-	-
Schwarzenau	ZAMG	500	15.27	48.75	337	38	0.11	1975	2001	1985,1995,1997,1998,1999	-	-	-
Hörsching	ZAMG	298	14.19	48.24	442	60	0.14	1978	2005	1995,1998,2004	5.2	-	-
Wien	ZAMG	163	16.40	48.25	470	73	0.16	1979	2005	1994,1995,1997,1998	12.6***	-	-
Innsbruck Flugh.	ZAMG	579	11.36	47.26	459	64	0.14	1976	2005	1995,1998,2002	6.4	-	-
Vandans	ZAMG	670	9.86	47.09	317	47	0.15	1979	2005	1983,1995,1998,1999,2000	6.8	-	-
Zeltweg	ZAMG	669	14.78	47.20	389	63	0.16	1978	2004	1981,1982,1984,1995,1998	-	-	-
Klagenfurt	ZAMG	447	14.33	46.65	454	81	0.18	1976	2005	1978,1991,1995,1996,1998,2000	13.8**	-	-



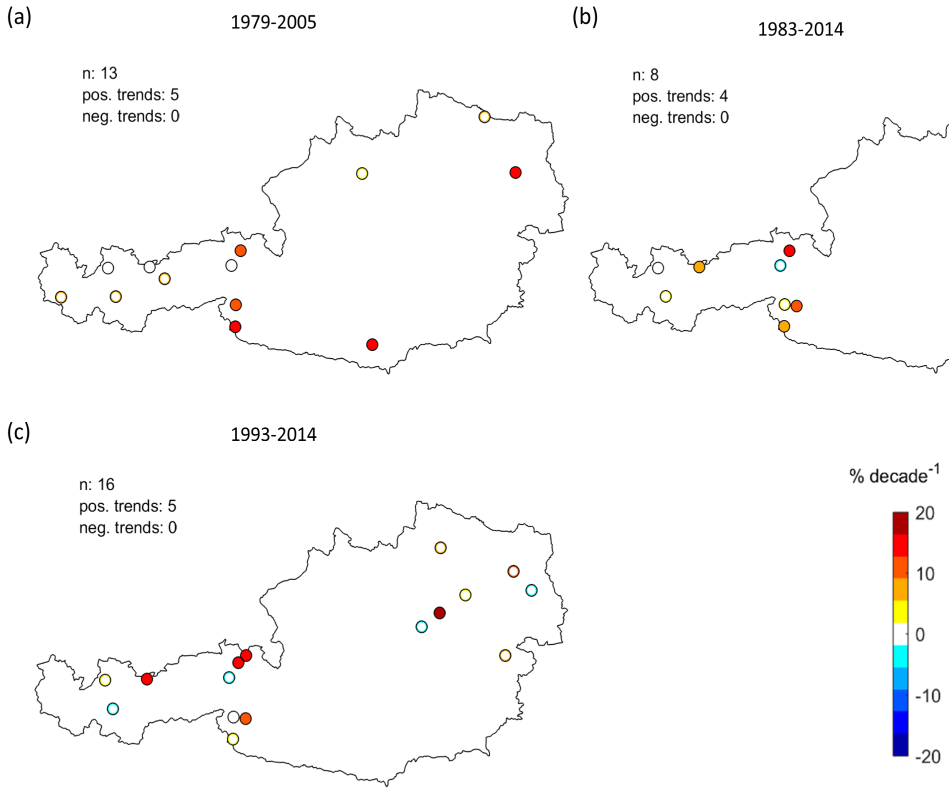
Supplementary Figure S 6 Spatial pattern of trends in (a) E_{wb}, (b) precipitation and (c) discharge over 1977–2014. Each circle indicates the outlet of one catchment. Filled circles indicate significant trends at $p \leq 0.05$.



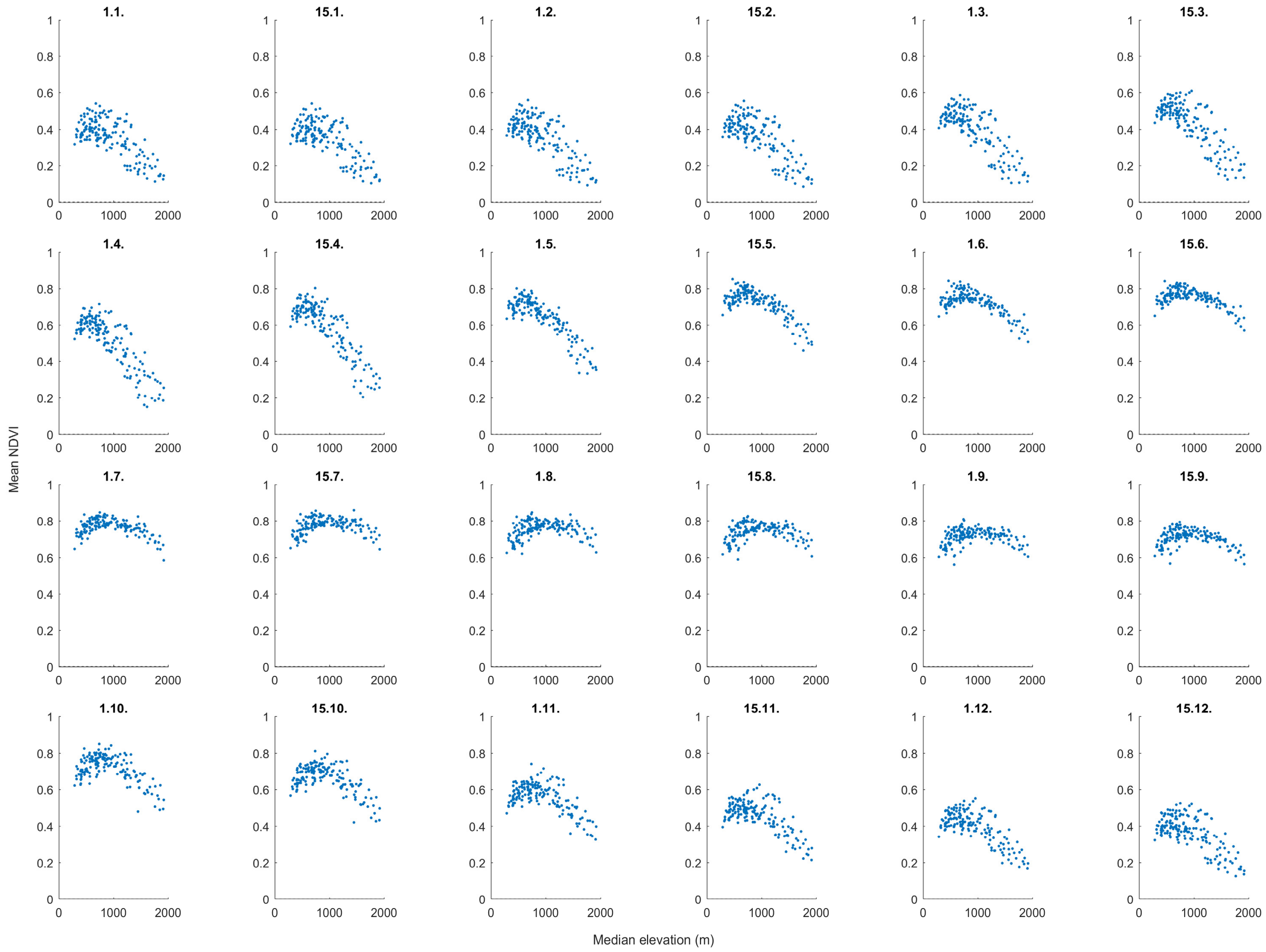
Supplementary Figure S 7 Anomalies of (a, c) summer precipitation (May–Oct), (b, d) winter precipitation (Nov–Apr) over 1977–2014. (a)–(b) mean anomalies by region. Data smoothed using a Gaussian filter with a standard deviation of 2 years. (c)–(d) mean anomalies over all catchments. The thin blue line shows the mean over all catchments, the grey shaded area the variability between catchments (± 1 standard deviation), the bold black line the smoothed mean, and the red dashed line the trend.



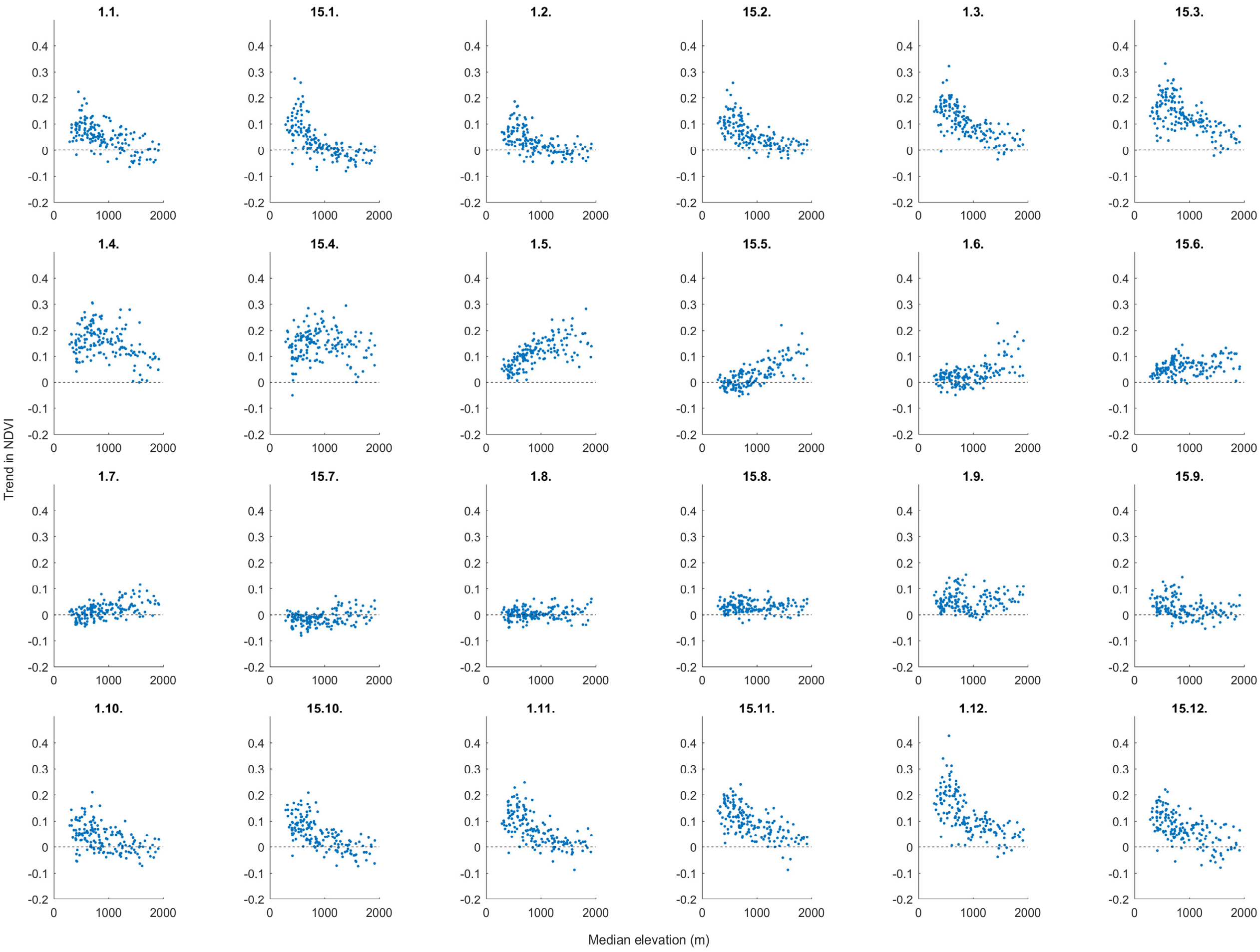
Supplementary Figure S 8 Spatial pattern of trends in (a) summer precipitation (May–Oct), and (b) winter precipitation (Nov–Apr) over 1977–2014. Each circle indicates the outlet of one catchment. Filled circles indicate significant trends at *p*≤0.05.



Supplementary Figure S 9 Trends in summer pan evaporation over three different periods. Filled circles indicate significant trends at *p*≤0.05.



Supplementary Figure S 10 Scatterplots of catchment average NDVI (*y*-axis) versus median catchment elevation (*x*-axis) for biweekly averages over the course of the year (the plot titles indicate the starting day of the two-week period).



Supplementary Figure S 11 Scatterplots of catchment average NDVI trend over 1982–2014 (*y*-axis) versus median catchment elevation (*x*-axis) for biweekly averages over the course of the year (the plot titles indicate the starting day of the two-week period).

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