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The role of dew and radiation fog inputs in the local water cycling of a temperate grassland in Central Europe

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Abstract. In a warmer climate, non-rainfall water (hereafter NRW) formed from dew and fog potentially plays an increasingly important role in temperate grassland ecosystems under the scarcity of precipitation over prolonged periods. Dew and radiation fog occur in combination during clear and calm nights, and both use ambient water vapor as a source. Research on the combined mechanisms involved in NRW inputs to ecosystems are rare, and the condensation of soil-diffusing vapor, as one of the NRW input pathways for dew formation, has hardly been studied at all. The aim of this paper is thus to investigate the different NRW input pathways into a temperate Swiss grassland at Chamau during prolonged dry periods in summer 2018. We measured the isotopic compositions (δ18O, δ2H, and d = δH − 8δ18O) of both ambient water vapor and the NRW droplets on leaf surfaces combined with eddy covariance and meteorological measurements during one dew-only and two combined dew and radiation fog events. We employed a simple two end-member mixing model using δ18O and δ2H to split the dew input pathways from different sources. Our results showed a decrease of 0.8–5.5 mmol mol⁻¹ in volumetric water vapor mixing ratio and a decrease of 4.8–16.7‰ in ambient water vapor δ2H due to dew formation and radiation fog droplet deposition. A nighttime maximum in ambient water vapor δ18O (−15.5‰ to −14.3‰) and a 3.4–3.7‰ decrease in ambient water vapor d were observed for dew formation in unsaturated conditions. In conditions of slight super-saturation, a stronger decrease of ambient water vapor δ18O (0.3–1.5‰) and a minimum of ambient water vapor d (−6.0‰ to −4.7‰) were observed. The combined foliage NRW and ambient water vapor δ18O and δ2H suggested two different input pathways: (1) condensation of ambient water vapor and (2) of soil-diffusing vapor. The latter contributed 9–42 % to the total foliage NRW. The dew and radiation fog potentially produced 0.06–0.39 mmol night⁻¹ NRW gain on foliage, which was comparable with 2.8 mm day⁻¹ daytime evapotranspiration. The ambient water vapor d was correlated and anti-correlated with ambient temperature and ambient relative humidity respectively, suggesting an only minor influence of large-scale advection and highlighted the dominant role of local moisture as a source for ambient water vapor. Our results thus highlight the importance of NRW inputs to temperate grasslands during prolonged dry periods and reveal the complexity of the local water cycle in such conditions including different pathways of water deposition.

1 Introduction

During extended periods without rainfall, non-rainfall water (hereafter NRW) inputs, namely dew and fog, are an essential water source for plants in arid and semi-arid regions (Agam and Berliner, 2006; del Prado and Sancho, 2007; He and Richards, 2015; Jacobs et al., 2002; Kidron et al., 2002; Malek et al., 1999; McHugh et al., 2015; Tomaszkiewicz et al., 2017; Ucles et al., 2013), Mediterranean coastal regions (Beysens et al., 2007), temperate ecosystems (Jacobs et al., 2006), and tropical climates (Clus et al., 2008). In clear calm nights when dew and radiation fog occur, the atmospheric boundary layer becomes stably stratified, leading to a shallow stable nocturnal boundary layer (hereafter NBL) with a depth on the order of no more than 50–100 m (Garratt, 1992). Dew and radiation fog occur at the bottom of the NBL (Garratt, 1992; Monteith and Unsworth, 2013; Oke, 2002; Stull, 1988). Both dew and radiation fog are formed due to the cooling of the Earth’s surface after sunset by
long-wave radiation losses in clear nights (Oke, 2002). This radiative cooling is a process due to which a body loses heat by long-wave thermal radiation, whereby its surface cools down below the dew point of the adjacent air. Under such conditions, dew can form on plant surfaces while fog forms on activated aerosol particles in the near-surface atmosphere.

NRW inputs contribute to the water budget across many ecosystems including croplands (Atzema et al., 1990; Meng and Wen, 2016; Tomaszkiewicz et al., 2017; Wen et al., 2012), grasslands (He and Richards, 2015; Jacobs et al., 2006; Wen et al., 2012), and forests (Berkelhammer et al., 2013; Dawson, 1998; Fritschen and Doraisswamy, 1973; Hiatt et al., 2012; Lai and Ehleringer, 2011). As compared to forests, grasslands present favorable conditions for dew and radiation fog formations:

1) Grassland surfaces are cooler than forest surfaces due to a higher albedo and thus lower net solar radiation input (Moore, 1976), and higher evapotranspiration (Kelliher et al., 1993; Williams et al., 2012).
2) Canopy resistance of grasslands is lower which reduces the warming effect by ground thermal emission via evaporative cooling (Garratt, 1992).
3) Aerosol particle deposition is weaker over grasslands due to shorter roughness length of grasslands (Gallagher et al., 2002), and thus more aerosol particles remain in the near-surface atmosphere, which consequently results in better conditions for radiation fog formation over grasslands. From the perspective of ecological functions, small amounts of NRW inputs have a more important influence on grasslands than forests because of the lower water use efficiency (hereafter WUE) and lower soil moisture availability in grasslands. At the beginning of drought stress in ecosystems, forests increase their WUE by closing their stomata, which increases stomatal resistance and thus reduces evapotranspiration, while grasslands maintain their evapotranspiration as long as the soil moisture is available to supply evaporative demand (e.g., Wolf et al. (2013)). Therefore, grasslands are more prone to suffer from soil water scarcity. In addition, as opposed to the deep-rooted systems for forest plants, grassland plants take up water from the top soil, where scarcity of soil moisture occurs more frequently during the absence of precipitation, therefore grasslands tend to anticipate lower soil moisture availability compared to forests.

Ambient water vapor is the main vapor source for both dew and radiation fog, therefore dew and radiation fog usually occur in combination. Because of the variability of temperature and humidity conditions, a single NRW night might transit from dew only to dew and radiation fog in combination. Before the atmospheric humidity reaches saturation, dew can only form if the surface temperature drops below air temperature. When the ambient water vapor reaches saturation or even supersaturation, dew and radiation fog can form in combination. Kaseke et al. (2017) used hydrogen and oxygen stable isotope regression to separate the different types of dew and fog, but they focused on dew and fog events separately. Research that focusses on relevant phase change processes during dew and radiation fog in combination is thus rare.

The moisture movement in the soil–plant–atmosphere continuum has been well understood by eddy covariance (hereafter EC) techniques, but the reliability of the method suffers during nighttime with weak turbulence (Berkelhammer et al., 2013). Instead, hydrogen and oxygen stable isotopes are a useful research tool to investigate different fractionation processes in the water cycle (Aemisegger et al., 2014; Delattre et al., 2015; Huang and Wen, 2014; Parkes et al., 2017), and can therefore be used to trace dew formation and radiation fog deposition into ecosystems (Delattre et al., 2015; He and Richards, 2015; Parkes et al., 2017; Spiegel et al., 2012; Wen et al., 2012). The isotopic composition of hydrogen ($^2$H and $^1$H) or oxygen ($^{18}$O and $^{16}$O) is expressed in the delta notation (hereafter $\delta$) as $\delta = (R_{\text{sample}}/R_{\text{standard}} - 1) \cdot 1000$ ‰, where $R_{\text{sample}}$ and $R_{\text{standard}}$ are the molar ratios of either $^2$H/$^1$H or $^{18}$O/$^{16}$O for the sample and standard, respectively. The standard is the Vienna Standard Mean Ocean Water (V-SMOW) controlled and distributed by the International Atomic Energy Agency (IAEA, 2009). With this definition, $\delta^{18}$O and $\delta^2$H are expressed as per mil (‰) discriminations from the standard. Whenever a phase change occurs, water molecules with different isotopes (hereafter isotopologue) as constituting atoms partition into the two phases in a specific way depending on ambient temperature and humidity gradients. Equilibrium fractionation always occurs at the interface between the two phases and results in a $\Delta \delta^2$H/$\Delta \delta^{18}$O ratio of approximately 8:1 in both phases, where $\Delta$ denotes the variabilities of $\delta^2$H and $\delta^{18}$O. When the ambient air is unsaturated, a deviation from the 8:1 ratio becomes measurable due to non-equilibrium fractionation (Dansgaard, 1964). The second order parameter deuterium excess (hereafter $d$), defined as $d = \delta^2$H – 8$\times$ $\delta^{18}$O after Dansgaard (1964), is a useful measure of non-equilibrium fractionation and provides information.
complementary to $\delta^2$H and $\delta^{18}$O. The $d$ is often used as a tracer for the water vapor source of a given water pool in the water cycle (Aemisegger et al., 2014; Galewsky et al., 2016; Gat, 1996; Welp et al., 2012; Yakir and Sternberg, 2000; Yepez et al., 2003). For example, at the local scale, local evaporation is a vapor source with lower $d$, while the entrainment from free troposphere is a vapor source with higher $d$ (Delattre et al., 2015; Parkes et al., 2017). The diurnal cycle of deuterium excess in a well-mixed convective boundary layer has been studied previously (e.g., Lai and Ehleringer (2011)), whereas relevant processes affecting $d$ in the NBL are much less well known, in particular for grasslands.

Monteith (1957) identified two input pathways for dew formation: 1) the downward pathway through the condensation of ambient water vapor onto foliage, and 2) the upward pathway through the condensation of soil-diffusing vapor onto foliage. Soil vapor diffusion is driven by the temperature gradient between the soil and the atmosphere and between different depths of the soil (Monteith, 1957; Oke, 1970). The temperature gradient generally reaches a maximum at the soil–atmosphere interface ($2$–$4$ °C warmer than the adjacent air at 1–2.5 cm in height for short grass or foliage surface (Monteith, 1957; Oke, 1970). The diffusing soil vapor can therefore condense onto cooler foliage. After Monteith (1957) had quantified the downward and upward components of dew formation by absorbing the NRW on foliage with filter paper, research has rarely been focusing on distinguishing these two pathways of dew formation. Furthermore, Monteith (1957) distinguished the two pathways by collecting the NRW in separate nights when only one or the other of the two pathways was assumed to occur. In Monteith (1957), the NRW condensing from soil-diffusing vapor was quantified in very calm nights with a 2 m wind speed (hereafter $u_{2m}$) of less than 0.5 m s$^{-1}$, whereas the maximum NRW condensing from ambient water was assumed to occur in slightly windy nights with $u_{2m}$ in the range of 2–3 m s$^{-1}$. However, for clear calm nights with $u_{2m}$ between 0.5 and 2 m s$^{-1}$, condensation of ambient water vapor and soil-diffusing vapor can occur in combination, with NRW on the foliage being a mix from these two pathways. Stable water isotopes and the “Keeling-plot” approach (Dawson, 1998; Keeling, 1958; Phillips et al., 2005) was used in this study to quantify the individual contributions of these two sources.

Our aim was to (1) investigate the isotopic fractionations during dew-only and dew–fog combined events; (2) estimate the potential gain of NRW from atmospheric vapor and from soil-diffusing vapor; and (3) assess the potential ecological relevance of NRW inputs to a temperate grassland ecosystem. We carried out three 24 h observation campaigns during summer 2018 using stable isotopes combined with EC and meteorological measurements to clarify the meteorological conditions and isotope fractionations for dew and radiation fog formations, to split the dew components from ambient water vapor and soil-diffusing vapor, and to explore the potential role of dew and radiation fog in temperate grasslands.

2 Materials and Methods

2.1 Study site and observation campaigns

The Chamau site (hereafter CH-CHA; 47°12’36.8” N, 8°24’37.6” E) is an intensively managed temperate grassland (4–6 cuts per year) at 393 m a.s.l., located in a valley bottom in Switzerland. The EC and meteorological measurement station (Fig. A1 in Appendix A) have been operational since 2005. The precipitation at the CH-CHA site was 870 mm in 2018, which was 297 mm (about 25%) less than the multiyear average over 2006–2017. From April to September in 2018, with respect to the corresponding monthly climatological values in the period 2006–2017, the monthly precipitation was on average 81 mm, which was averagely 49 mm (38%) less (Fig. 1a), and the monthly average temperature was on average 17.3 °C, which was 1.8 °C higher (Fig. 1b).

Three 24 h observation campaigns were carried out during expected dew/fog events on 25–26 July (event 1), 20–21 August (event 2), and 9–10 September (event 3) 2018. The time series were all recorded in CET (UTC+1). The corresponding consecutive no-rain periods were 23–27 July, 18–21 August, and 8–12 September 2018 respectively. Because of the extreme summer drought in 2018, no harvest of grassland was carried out during the three campaigns, but two harvests were carried out 46 d before event 1 on 9 June 2018, and one day after event 3 on 10 September 2018 respectively. The leaf area index
(hereafter LAI) was 2.5 and 1.5 m² m⁻² during events 1 and 2 respectively (measured 7 d before events 1 and 2 with LAI-2000, LI-COR Biosciences, Lincoln, NE, USA), and was 3.2 m² m⁻² after harvest (measured 1 d after event 3). The mean vegetation height (z) was 0.2 – 0.3 m during the three campaigns. The volumetric soil water content at 10 cm was 18%, 18%, and 21% respectively (ML2x sensors, Delta-T Devices Ltd., Cambridge, UK). The permanent wilting point for the top soil at the site was 16% (calculated from soil texture at 0–20 cm: sand, 35.8%, clay, 19.0% following Roth (2006); the wilting point calculation equation followed Briggs and Shantz (1912)).

2.2 Experiment setup

2.2.1 Eddy covariance and meteorological data and calculations

The EC measurements at 20 Hz were setup at 2.4 m a.g.l. (see Zeeman et al. (2010) for more details), based on measurements with a 3-D sonic anemometer (Gill R3, Gill Instruments Ltd., Lymington, UK), and an open path Infrared Gas Analyzer (IRGA, Li-7500, Li-Cor, Lincoln, NE, USA). The EC measurements were processed to 30 min averages for evapotranspiration rate (mm h⁻¹), horizontal wind speed (hereafter u₂m, in m s⁻¹), H₂O flux (hereafter F₁H₂Os, in mmol m⁻² s⁻¹; minus value means downward flux, whilst positive value means upward flux), atmospheric specific humidity (hereafter qₑ, in g kg⁻¹), and dew point temperature (hereafter Tₑ, in °C) (Buck, 1981; Campbell and Norman, 1998). The meteorological measurements at 0.1 Hz for air temperature (hereafter T₀, in °C), relative humidity (hereafter RH, in %), and long-wave outgoing radiation (hereafter LWₒ rê, in W m⁻²) were setup at 2.0 m a.g.l. (see Zeeman et al. (2010) and Fuchs et al. (2018) for more details). The horizontal visibility (in km) was measured every 10 s by a fog sensor (MiniOFS, Optical Sensors Inc., Göteborg, Sweden) and a present weather detector (PWD10, Vaisala Oyj, Helsinki, Finland). The meteorological measurements were processed to 30 min averages for T₀, RH, and LWₒ rê, and to 1 min averages for visibility.

The vegetation surface temperature (Tˢ, in °C) was determined after Stefan–Boltzmann’s law (Stull, 1988) as:

\[ T^{*}_{s} = \frac{\sqrt{\frac{LW_{o rê}}{\varepsilon - \sigma}} - 273.15}{\varepsilon} \tag{1} \]

where an emissivity (hereafter ε) of 0.98 was used to calculate temperatures for wet leaf surfaces (hereafter index w; T₀ = T₀w), and a value of 0.96 was used for dry leaf surfaces (hereafter index d T₀ = T₀d) after López et al. (2012); σ is Stefan–Boltzmann constant at 5.67 \times 10⁻⁸ W m⁻² K⁻¹.

The saturation specific humidity (qₛ, in g kg⁻¹) and the relative humidity (h₀) with respect to surface temperature T₀ for wet and dry vegetation surfaces was calculated using Tetens formula ((Buck, 1981; Campbell and Norman, 1998), see the equations in Appendix B).

2.2.2 Sampling of the NRW on foliage & isotope ratio mass spectrometer measurements

To analyze the isotopic compositions of the NRW on foliage, the NRW droplets were taken during dew and radiation fog formations. The sampling of the NRW on foliage (hereafter fNRW) was carried out on a grassland area of 100×130 m² around the “EC & meteo” measurements (Fig. A1 in Appendix A). Nine replicated fNRW samples were absorbed from leaf surfaces with cotton balls at the end of the nights of events 1 and 3 (once sampling per event), but bihourly during the night of event 2 (four times of sampling per event). After collection, the samples were immediately transferred into gas tight 12 ml extainers (Labco Exetainer® vial, High Wycombe, UK) and stored in a portable cooling box filled with ice blocks. Before extracting the water in a cryogenic vacuum distillation system (Prechsl et al., 2015), the samples were stored at −19°C. The measurements of the isotopic compositions for fNRW (hereafter δ₁₈O_fNRW) and in soil moisture (hereafter δₒ) of extracted water samples were performed using an isotope ratio mass spectrometer (IRMS, DELTApplusXP, Finnigan MAT, Bremen, Germany). The measured uncertainties of δ₁₈O and δ²H are ±0.1‰ and better than ±1.0‰, respectively (Gehre et al., 2004; Werner and Brand, 2001).
2.2.3 Isotopic compositions and mixing ratio measurements for ambient water vapor

The isotopic compositions and the volumetric mixing ratio for ambient water vapor were measured at 0.5–1 Hz using a cavity ring-down laser spectrometer (L2130-i, Picarro Inc., Santa Clara, CA, USA). The L2130-i was placed in a house 200 m away from the EC & meteo measurements (Fig. A1 in Appendix A). Ambient air was pulled into the L2130-i cavity through a PTFE intake hose, with an inner diameter of 1/4 inch, and a PTFE-filter inlet (FS-15-100 and TF50, Solberg International Ltd., Itasca, IL, USA) fixed at 6 m a.g.l.. The intake hose was thermally isolated, heated using a resistive heating wire (Raychem SBTV2-CT, Von Rotz, Kerns, Switzerland) wrapped around the entire length of the intake tube to prevent condensation, and flushed with an external membrane pump (N022, KNF Neuberger GmbH, Munzingen, Freiburg, Germany) at a rate of 9 L min\(^{-1}\) to minimize memory effects within the inlet system. The isotopic compositions of ambient water vapor (hereafter \(\delta_d\)) and the volumetric ambient water vapor mixing ratio (hereafter \(\omega\)) were measured with an instrumental flow rate of 300 mL min\(^{-1}\).

The instrument’s response time in this setup was found to be on the order of 10 s (Aemisegger et al., 2012). To correct for instrument drifts and to normalize the data to the international VSMOW-SLAP scale, the raw data were calibrated using a Standard Delivery Module (SDM; A0101, Picarro Inc., Santa Clara, CA, USA) by performing two-point calibrations every 12 h (Aemisegger et al., 2012) using two liquid standards (standard 1: \(\delta^{18}O_{a} = -11.43\%o, \delta^{2}H = -81.84\%o, d_1 = 9.64\%oo\); standard 2: \(\delta^{18}O_{a} = -40.66\%o, \delta^{2}H = -325.67\%o, d_2 = -0.37\%o\) measured by an IRMS). The \(\delta^{18}O\) and \(\delta^{2}H\) of the standards thus bracket the range of the measured \(\delta^{18}O\) and \(\delta^{2}H\). Laser spectrometric measurements are known to be affected by a water vapor mixing ratio dependent bias due to spectroscopic effects (absorption peak fitting, and baseline effects). In our study, all measurements were performed at \(w > 12\ 000\ \mu\text{mol}\ \text{mol}^{-1}\), therefore no mixing ratio dependent isotope bias correction was necessary (see more details in Aemisegger et al. (2012)). The L2130-i was calibrated using a dew point generator (LI-COR LI 610, Li-Cor Inc., Lincoln, NE, USA) following the procedure by Thurnherr et al. (2020).

The second-order parameter \(d\) of ambient water vapor (hereafter \(d_0\)) was calculated with the calibrated \(\delta^{18}O_{a}\) and \(\delta^{2}H_{a}\). The overall random uncertainties of \(\delta^{18}O\) and \(\delta^{2}H\) measurements were 0.2%o and 0.8%o respectively (for more details about the uncertainty quantification, see Aemisegger et al. (2012)). Calibrated \(\delta^{18}O_{a}\) and \(\delta^{2}H_{a}\) were then averaged over 30 min intervals.

To compare the ambient water vapor measurements with the fNRW, the NRW equilibrium liquid (aNRW) from this vapor was calculated. Under the assumption of equilibrium fractionation, the isotopic compositions of aNRW (hereafter \(\delta_{aNRW}\)) formed from ambient water vapor (\(\delta_d\)) were calculated using the temperature-dependent equilibrium fractionation factors following Horita and Wesolowski (1994) (see details in Appendix C).

2.3 Determination of atmospheric layer heights

The isotopic fractionation during phase change at the Earth surface is linked to the micrometeorological layers near the surface (Fig. 2). The inclusion of a zero-plane displacement (hereafter \(z_d\), Fig. 2) in wind profiles allows us to separate the downward flux from ambient vapor and the upward flux from soil-diffusing vapor. The height of this z-plane (hereafter \(z_d\)) is typically two-thirds of mean vegetation height (hereafter \(z_e\); Stull (1988)). The roughness length (hereafter \(z_0\)) is a measure of the aerodynamic roughness of the surface, and is around one-tenth of \(z_e\) (Fig. 2; Stull (1988)). The wind speed is zero at \(z_0\) above \(z_d\) (\(z_d + z_0\), Fig. 2; Stull (1988)). Therefore, we consider three pathways of NRW inputs onto the foliage of grasslands for dew and radiation fog: 1) the downward component of dew formation condensing from ambient water vapor (hereafter “dDew”), 2) the upward component of dew formation condensing from soil-diffusing vapor (hereafter “dDew”), and 3) radiation fog deposition (hereafter “aFog”).

We determined the top of the NBL as the lowest height where the vertical stratification of the atmosphere becomes isothermal (\(\partial T/\partial z = 0\), Stull (1988); Tombrou et al. (1998)). During the three events in this study, the NBL top at 1:00 CET of the events was at 730 m, 700 m, and 680 m a.g.l., respectively (Fig. 3); the NBL height was obtained from air pressure after Campbell and Norman (1998); the vertical temperature and pressure profiles were extracted from the hourly European Centre for Medium Range Weather Forecast (ECMWF) reanalysis product ERA5 reanalysis dataset within the Copernicus Climate
Change Service (Hersbach et al., 2020; Horanyi, 2017). Canopy height $z_c$ was 0.2–0.3 m during this time of season (Sect. 2.1), hence the aerodynamic displacement height $z_d$ was roughly 0.13–0.20 m ($\approx 2/3 z_c$), with a roughness length $z_0$ of 0.02–0.03 m. With these assumptions, negligible wind speeds ($u_{2m}$ around zero) could be assumed at heights below $z_d + z_c \approx 0.15–0.23$ m a.g.l. (Fig. 2).

2.4 Partitioning of NRW inputs using a two end-member mixing model

We split the contribution of NRW input pathways into the two main processes described in Sect. 2.3: (1) the downward component of dew formation (aDew) and fog droplet deposition (aFog), and (2) the distillation (dDew) of soil-diffusing vapor on plant leaves. In unsaturated conditions, the NRW on foliage (fNRW) was a mix of aDew and dDew, while in saturated conditions, fNRW was a mix of aDew and aFog. “Unsaturated conditions” in this context refers to the standard 2 m height of meteorological measurements. Both aDew and aFog were condensed from ambient water vapor, thus we used the term “aNRW” if either dew or fog input, or the combination of both, was meant. Dew formed in unsaturated conditions is a mixture of aNRW and dDew but lacks contribution from fog deposition, thus the isotopic signature of the NRW resulting from the isotopic compositions of dDew (hereafter $\delta^{18}O_{dDew}$ and $\delta^2H_{dDew}$) and the proportion of dDew (hereafter $f_{dDew}$) in fNRW was expressed as:

$$\delta^{18}O_{fNRW} = f_{dDew} \cdot \delta^{18}O_{dDew} + f_{aNRW} \cdot \delta^{18}O_{aNRW} \hspace{1cm}(2)$$

$$\delta^2H_{fNRW} = f_{dDew} \cdot \delta^2H_{dDew} + f_{aNRW} \cdot \delta^2H_{aNRW} \hspace{1cm}(3)$$

$$1 = f_{dDew} + f_{aNRW} \hspace{1cm}(4)$$

where $f_{aNRW}$ is the proportion of aNRW in fNRW; $\delta^{18}O_{dDew}$, $\delta^2H_{dDew}$, $f_{dDew}$, and $f_{aNRW}$ are unknown. Therefore, four unknowns with only three equations (Eq. 2–4) required two time points, at 23:00 CET and 1:00 CET in event 2, to obtain empirical estimates for the four unknowns. By doing so, we implicitly assumed that $\delta^{18}O_{dDew}$ and $\delta^2H_{dDew}$ were constant within this 2 h period, and only $f_{dDew}$ and $f_{aNRW}$ were allowed to change between these two sampling times. For $\delta_{fNRW}$, the median value for each sampling was taken, and for $\delta_{aNRW}$ the 2 h average was computed from 30 min data.

2.5 Statistics and imaging

In unspecified explicit, we reported means ± standard deviation. For $\delta_{fNRW}$ and $\delta_{aNRW}$, we reported median considering the heterogeneous distribution for the sampling of NRW on foliage. The calculating and imaging were processed in R version 3.6.3 (R Core Team, 2020). For linear regression between $\delta^2H$ and $\delta^{18}O$ the orthogonal regression was used (total least square, Gat (1981)), whereas the ordinary least-squares method was used for the $d_c$-RH and linear regression.

3 Results

3.1 Atmospheric and surface conditions during dew and radiation fog events

3.1.1 Weak turbulence and high relative humidity

Dew and radiation fog generally form during clear-sky nights with a weak large-scale pressure gradient, low wind speeds and weak turbulence. During the three field campaigns presented in this study, $u_{2m}$ and $F_{H2O}$ showed an abrupt weakening from around 17:00 CET onwards (Fig. 4a, b). With nighttime, $u_{2m}$ remained below 0.7 m s$^{-1}$ (Fig. 4a), and $F_{H2O}$ was at very low (−0.4 to 0.3 mmol m$^{-2}$ s$^{-1}$, minus value means downward flux, and positive value means upward flux; Fig. 4b), indicating a vanishing of turbulent fluxes. These are favorable conditions for dew and radiation fog formations.

The three events with dew or radiation fog were characterized by high relative humidity with respect to air temperature (RH) measured at 2 m above ground level. From around 17:00 CET, RH increased rapidly, and reached 100% around 03:00 CET during event 2, and around 20:30 CET during event 3 (Fig. 4c). These saturated conditions led to the formation of fog...
characterized by a horizontal visibility < 1 km (Fig. 4d). Fog appeared around 05:00 CET during event 2, lasting for less than an hour until sunrise, whilst the onset of fog was much earlier during event 3 (around 23:00 CET), lasting for a longer period until dissipation around sunrise. The visibility was always > 1 km in event 1, indicating that fog was absent during event 1. Therefore, event 1 can be considered as a dew-only event, whilst events 2 and 3 were characterized by a combination of dew and partial influence of radiation fog.

3.1.2 Surface cooling and the sign of condensation

Both grassland surfaces and ambient air started to cool down from around 17:00 CET onwards, due to substantial net long-wave radiation loss, which was not compensated by the low remaining incoming short-wave radiation levels. The leaf surfaces of the grassland cooled more rapidly than the near-surface atmosphere, thus with nightfall, T0 remained cooler than Ta, although both of them gradually decreased (Fig. 5a). The first sign of condensation occurred when the leaf surfaces cooled down below dew point temperature (Fig. 5a, Td < Ta). The level of T0 (Ta,0) became lower than Td at around 0:30 CET in event 1, 21:30 CET in event 2, and 19:00 CET in event 3 (Fig. 5a), determining when the first signs of condensation can be expected. During event 3, the surface already cooled down below the dew point rapidly after sunset (Td < Ta, Fig. 5a), indicating that condensation already started with nightfall.

The specific humidity of the air, qg, steeply increased by 2.0–3.2 g kg⁻¹ from around 17:00 CET until sunset (Fig. 5b), suggesting the inversion of moisture from local evaporation into a shallow inversion layer. The increase of qg over time is enhanced by cold-air drainage down the slopes and along the valley bottom where the CH-CHA site is located as compared to conditions without advection. With nightfall, qg reached a nighttime maximum of 9.6–12.5 g kg⁻¹ (Fig. 5b). Especially, in events 1 and 2, before starting to decrease, qg fluctuated for a short period from sunset until the first sign of condensation (Fig. 5b). When condensation started (Td < Ta, Fig. 5a), qg gradually decreased (Fig. 5b). With qg falling to values below qa (Fig. 5b), super-saturation with respect to the leaf surfaces occurred, thus computed theoretical h0 exceeded 100% (Fig. 4c). The decrease of qg was much faster in event 3 (0.4 g kg⁻¹ h⁻¹, Fig. 5b) than that in events 1 and 2 (0.2 and 0.3 g kg⁻¹ h⁻¹, Fig. 5b), indicating stronger condensation of ambient water vapor.

3.1.3 Characteristics of precondensation and condensation periods

According to the temperature and humidity conditions, the periods from 17:00 CET until sunrise were defined as: 1) precondensation period (hereafter P1) with the weakening of turbulence and with T0 > Ta; and 2) condensation period (hereafter P2) with T0 < Ta. The precondensation period (P1) was further separated into: P1a) starting around 17:00 CET until sunset with the weakening of turbulence and the increase of qa; P1b) from sunset until the first sign of condensation with short-term fluctuations of qa. The condensation period (P2) was further split into: P2a) with dew only under RH < 100%; P2b) with dew and radiation fog occurring in combination under RH = 100%.

3.2 Isotope dynamics of ambient water vapor during dew and fog events

The periods P1 and P2 are reflected in the temporal evolution of w, δw, and δd (Fig. 6). During P1a as from 17:00 CET until sunset, w, δ18Ow, and δ13H2O showed a steep increase by 0.3–0.4 mmol mol⁻¹, 2.0–3.2‰, and 7.4–12.5‰ respectively (Fig. 6a, b, c), whilst δd showed a steep decrease by 11.6–16.9‰ (Fig. 6d). With nightfall when turning into period P1b, w and δw reached a plateau with 15.5 to 17.8 mmol mol⁻¹ in w (Fig. 6a), –15.5 to –14.3‰ in δ18Ow (Fig. 6b), and –128.0‰ to –113.2‰ in δ13H2O (Fig. 6c). During P1b for events 1 and 2, w and δw experienced short-term fluctuations around their nighttime maximum before condensation set in, while δd decreased by 5.1–9.4‰ (Fig. 6d). During P2, w steeply decreased by 0.8–5.5 mmol mol⁻¹ (Fig. 6a), δ13H2O decreased by 3.3–16.7‰ (Fig. 6c), and δd reached its minimum at –11.8‰ to –4.7‰ (Fig. 6d). During P2, the decreasing rate of δ13H2O in event 3 (1.6‰ δ13H2O h⁻¹) was almost double that in events 1 and 2 (0.8 and 1.0 δ13H2O h⁻¹ respectively, Fig. 6c), suggesting stronger condensation in event 3.
Note that the changes of $\delta^{18}O$ and $d_1$ (Fig. 6b, d) depended on the humidity dynamics and the occurrence of dew and fog (Fig. 4c, d). During the dew-only P2a in events 1 and 2 (Fig. 4c, d), $\delta^{18}H$ decreased by 3.3–5.7‰ (Fig. 6c), and $d_1$ slightly decreased by 3.4–3.7‰ (Fig. 6d), while $\delta^{18}O$ showed slight fluctuations around the maximum reached 4 h and 2 h after nightfall of events 1 and 2 respectively (−15.5‰ to −14.3‰, Fig. 6b). During P2b in events 2 and 3 with dew and fog in combination, both $\delta^{18}O$ and $\delta^{18}H$ gradually decreased (by 0.3–1.5‰, and 2.1–12.8‰ respectively) with a ratio of around 8:1 (Fig. 6b, c), hence $d_1$ was approximately constant at the nighttime minimum (−6.0‰ to −4.7‰, Fig. 6d) although with slight fluctuation.

### 3.3 The isotopic signals of non-rainfall water

The NRW on foliage (fNRW) was comparable with the equilibrium liquids from ambient water vapor (aNRW). The isotopic compositions of aNRW was −5.0‰ to −4.3‰ in $\delta^{18}O_{\text{aNRW}}$, −47.4‰ to −38.6‰ in $\delta^{2}H_{\text{aNRW}}$, and −12.1‰ to −2.4‰ in $d_{\text{aNRW}}$ (Fig. 7a, b, c). As a comparison, the NRW on foliage (fNRW) was −6.1‰ to −1.5‰ in $\delta^{18}O_{\text{fNRW}}$, −64.3‰ to −35.6‰ in $\delta^{2}H_{\text{fNRW}}$, and −33.8‰ to 8.0‰ in $d_{\text{fNRW}}$ (Fig. 7a, b, c). The isotopic compositions of fNRW varied in time with gradually decreasing $\delta^{18}O_{\text{fNRW}}$ (Fig. 7a), but gradually increasing $\delta^{2}H_{\text{fNRW}}$ (Fig. 7b) and $d_{\text{fNRW}}$ (Fig. 7c).

The relationships between the isotopic compositions of fNRW and aNRW were related to RH. Under unsaturated conditions during P2a when dew formation occurred, $\delta^{18}O_{\text{aNRW}}$ (−4.4±0.1‰) was more depleted than $\delta^{18}O_{\text{fNRW}}$ (−3.8‰) (Fig. 7a), while $\delta^{2}H_{\text{fNRW}}$ (−42.3±3.8‰) was more enriched than $\delta^{2}H_{\text{aNRW}}$ (−47.7‰) (Fig. 7b), and $d_{\text{fNRW}}$ (−7.1±3.6‰) was higher than $d_{\text{aNRW}}$ (−20.5‰) (Fig. 7c). Under saturated conditions during P2b, the isotopic compositions of aNRW (−4.7±0.2‰ in $\delta^{18}O_{\text{aNRW}}$, −43.0±1.2‰ in $\delta^{2}H_{\text{aNRW}}$, and −5.4±0.3‰ in $d_{\text{aNRW}}$) was identical to the isotopic compositions of fNRW (−4.6‰ in $\delta^{18}O_{\text{fNRW}}$, −41.6‰ in $\delta^{2}H_{\text{fNRW}}$, and −4.7‰ in $d_{\text{fNRW}}$) (Fig. 7a, b, c). Especially, for the sampling at 5:00 CET in event 3 when radiation fog occurred, $\delta^{18}O_{\text{fNRW}}$ and $\delta^{2}H_{\text{fNRW}}$ were depleted by 0.7‰ and 1.4‰ with respect to $\delta^{18}O_{\text{aNRW}}$ and $\delta^{2}H_{\text{aNRW}}$ respectively (Fig. 7a, b), and $d_{\text{fNRW}}$ was 5.5‰ higher than $d_{\text{aNRW}}$ (Fig. 7c).

The relationships of $\delta^{2}H_{\text{aNRW}}$–$\delta^{18}O_{\text{aNRW}}$ and $\delta^{2}H_{\text{fNRW}}$–$\delta^{18}O_{\text{fNRW}}$ with respect to the local meteorological water line (LMWL: $\delta^{2}H = 7.68 \times \delta^{18}O + 6.97$, Prechl et al. (2014)) were shown in Fig. 8. Both $\delta^{2}H_{\text{aNRW}}$–$\delta^{18}O_{\text{aNRW}}$ and $\delta^{2}H_{\text{fNRW}}$–$\delta^{18}O_{\text{fNRW}}$ fell to the right side of LMWL, suggesting lower $d$ from NRW inputs as compared to local precipitation. When we only considered the condensation of ambient water vapor under equilibrium fractionation, $\delta^{2}H_{\text{aNRW}}$ and $\delta^{18}O_{\text{aNRW}}$ distributed on the equilibrium line (orthogonal regression) from $\delta^{18}H_{\text{aNRW}}$–$\delta^{18}O_{\text{aNRW}}$ (for the sampling at 3:00 and 5:00 CET in event 2, Fig. 8). However, with the mix of the component condensing from soil-diffusing vapor (dDew) under RH < 100%, the $\delta^{2}H_{\text{aNRW}}$–$\delta^{18}O_{\text{aNRW}}$ pairs fell to the right-hand side of the equilibrium line (for the sampling at 3:00 CET in event 1, and the samplings at 23:00 and 1:00 in event 2, Fig. 8a), suggesting lower $d_{\text{aNRW}}$ than $d_{\text{aNRW}}$. Whereas, with the mix of the component from radiation fog deposition, $\delta^{2}H_{\text{fNRW}}$–$\delta^{18}O_{\text{fNRW}}$ relation fell to the left-hand sides of the equilibrium line (for the sampling at 5:00 CET in event 3), indicating higher $d_{\text{fNRW}}$ than $d_{\text{aNRW}}$.

### 3.4 Splitting the components of dew using a two end-member mixing model

Under unsaturated conditions, with respect to aNRW, $\delta^{18}O_{\text{aNRW}}$ and $\delta^{2}H_{\text{aNRW}}$ deviated to the enriched and depleted sides of $\delta^{18}O_{\text{dNRW}}$ and $\delta^{2}H_{\text{dNRW}}$, respectively (Fig. 7a, b), suggesting a mix of NRW on foliage (fNRW) from the condensation of soil-diffusing vapor (dDew) and the condensation of ambient water vapor (aNRW). Based on the measurements from 23:00 to 1:00 CET in event 2, the averages of $\delta^{18}O_{\text{dDew}}$, $\delta^{2}H_{\text{dDew}}$, and $d_{\text{dDew}}$ during this 2 h period were estimated as −1.0‰, −71.8‰, and −63.4‰ respectively (Fig. 7a, b, c), and the corresponding contributions of dDew in fNRW were 28% and 9% respectively (Fig. 7d). A linear extrapolation from the two hours between 23:00 CET and 1:00 CET to the beginning of dew formation at 21:30 CET of event 2 increased the contribution of dDew to 42% (Fig. 7d). Similarly, when using the values of $\delta^{18}O_{\text{dDew}}$ from event 2 for estimating the contribution of dDew during event 1, the proportion of dDew was around 18–31% for our sampling at 3:00 CET of event 1 (vertical whiskers in Fig. 7d).
4 Discussion

4.1 Fractionation during condensation of ambient water vapor

We only considered equilibrium fractionation (shown as $\delta_{\text{aNRW}}$ in Fig. 9) when simulating the isotopic compositions of the NRW component condensing from ambient water vapor. An alternative approach would be to consider both equilibrium and non-equilibrium fractionation factors (see Appendix D; see also Lee et al. (2009) and Wen et al. (2012)) because of the laminar sublayer in the leaf boundary layer. To compare these two methods, we applied the method of Wen et al. (2012) on our data to simulate the isotopic composition of the NRW component condensing from ambient water vapor (shown as $\delta_{\text{aNRW}}$ in Fig. 9; see also Appendix D). We found that $\delta_{\text{aNRW}}$ was more depleted as compared to the NRW on foliage (shown as $\delta_{\text{NRW}}$ in Fig. 9), and the depletion of $\delta_{\text{aNRW}}$ was more severe with the increase of $h_0$ (Fig. 4c), which is in agreement with Wen et al. (2012).

The depletion of $\delta_{\text{aNRW}}$ with respect to $\delta_{\text{NRW}}$ was most likely due to the overestimation of the non-equilibrium fractionation factor when $h_0$ exceeded 100% (going up to 132% in our study, see Fig. 4c), because Jouzel et al. (1987) pointed out that non-equilibrium fractionation is negligible above $-10^\circ\text{C}$ in the process of vapor condensing to liquid. However, non-equilibrium fractionation driven by molecular diffusion might have played an important role in a laminar fog boundary layer (hereafter FBL; (Castillo and Rosner, 1989; Epstein et al., 1992)), which led to more depleted $\delta_{\text{NRW}}$ than $\delta_{\text{aNRW}}$ at 5:00 CET in event 3 (Fig. 7a, b) when radiation fog occurred. Heavier isotopologues move more slowly than their lighter counterpart in air (molecular diffusivity: $D[\text{H}_2\text{O}] < D[\text{H}_2^1\text{H}] < D[\text{H}_2^2\text{H}]$, Merlivat (1978)), hence the rate at which heavy isotopologues ($\text{H}_2\text{O}$ and $\text{H}_2^2\text{H}$) in ambient air pass through the laminar FBL to be condensed at the liquid–vapor interface is smaller than the rate of condensation of their lighter counterpart. Therefore, $\delta_{\text{NRW}}$ can become more depleted than $\delta_{\text{aNRW}}$. Fog lasted as from 23:00 CET until sunrise of event 3, and appeared around 5:00 CET within half an hour before sunrise in event 2 (Fig. 4d). However, we only observed a lower $\delta_{\text{aNRW}}$ than $\delta_{\text{NRW}}$ in event 3 (Fig. 7a, b), suggesting that the depletion of $\delta_{\text{aNRW}}$ might also be related to the duration of radiation fog. The condensation of ambient water vapor for dew formation can be approximated as an equilibrium fractionation process accordingly, as was also observed by Wen et al. (2012) and Delattre et al. (2015). The condensation of ambient water vapor to form radiation fog can cause slight depletion of the NRW compared to the equilibrium liquid obtained from ambient water vapor.

4.2 Potential NRW gain from the condensation of soil-diffusing vapor

Splitting the input pathways of dew formation using stable isotopes will allow future studies to quantify dDew gain with the combination of lysimeter or filter paper absorption measurements. In our study, as shown in Sect. 3.4, we estimated that dDew contributed 9–42% of total NRW (Fig. 7d) during our observation periods. Monteith (1957) estimated that the condensation rate of soil-diffusing vapor was 0.01–0.02 mm h$^{-1}$ (with $u_{2m} < 0.5$ m s$^{-1}$) using filter paper absorption measurements. In Monteith (1957), the condensation rate of ambient water vapor varied from 0.004 to 0.035 mm h$^{-1}$ depending on the wind speed (with $u_{2m} < 3$ m s$^{-1}$) and humidity conditions. Thus, the contribution of dDew in the total NRW was potentially 22–83% according to the condensation rate of Monteith (1957). Following the condensation rate of Monteith (1957), the potential total NRW gain was 0.06 – 0.39 mm night$^{-1}$ (see details in Table 2). This amount of NRW gain was comparable with the average evapotranspiration rate of 2.8 mm day$^{-1}$ (daytime) during the continuous no-rain periods of the three events (see details in Table 3).

In future research, combining isotopic compositions measurements with lysimeter measurements to quantify dDew gain would provide useful benchmark data to better evaluate the isotope-based estimates of NRW input. The NRW gain can be measured directly by a lysimeter as the net water gain of the soil and plants (Kaseke et al., 2012; Riedl et al., 2020; Ucles et al., 2013), while dDew is an indirect estimate based on stable water isotope data of the transfer of moisture from one part of the surface (soil surface) to another (foliage) within grassland ecosystems.
4.3 Diurnal patterns of isotopic compositions in ambient water vapor

The diurnal patterns of $d_4$ reflected the main drivers of ambient moisture variability. During the daytime 13:00–17:00 CET, $d_4$ was at a plateau (12.2‰ to 18.0‰, Fig. 6d) compared to condensation periods in the night (P2), when $d_4$ reached its daily minimum (~11.8‰ to ~4.7‰, Fig. 6d). The transition from higher daytime $d_4$ to lower nighttime $d_4$ occurred from 17:00 CET until sunset (P1a, Fig. 6d). Entrainment from the free troposphere played a dominant role in daytime atmospheric moisture during 13:00–17:00 CET, and caused a higher $d_4$ than in the night (Fig. 6d), and a decrease in $\delta$H (Fig. 6b, c) (Delattre et al., 2015; Lai et al., 2006; Lee et al., 2006; Parkes et al., 2017; Welp et al., 2012). On the contrary, during P1a, under reduced entrainment from the free troposphere (weakened $u_{200}$ and reduced $F_{\text{H2O}}$ in Fig. 4a, b) compared to mid-day values, local ET caused a steep decrease of $d_4$ (Fig. 6d) and increases of $\delta$H (Fig. 6b, c), which was in accordance with the previous research by Lai et al. (2006), Huang and Wen (2014), and Parkes et al. (2017). During P1b, the fluctuation of $\delta$H (Fig. 6b, c) was due to short-term variability of the isotopic compositions of soil evaporation (within 1 h before sunset, 0–5 cm soil moisture with $\delta^{18}$O varying from 5.5‰ to ~8.5‰, with $\delta^2$H varying from ~8.5‰ to ~72.8‰, and $d$ varying from ~5.0‰ to ~52.4‰), which was in accordance with the reports by Welp et al. (2012). The decrease of $d_4$ during P2b suggested radiation fog with local moisture as a source for ambient water vapor, which was in contrast with Spiegel et al. (2012) in Greenland that found the increase of $\delta$H with fog during the passage of a cold front. The correlated $d_4 - T_a$ (Fig. 10a) and anti-correlated $d_4 - \text{RH}$ (Fig. 10b) in our study suggested an only minor influence of large-scale air advection and highlighted the dominant role of local moisture as a source for ambient water vapor (Aemisegger et al., 2014).

During dew and radiation fog (P2), the condensation of ambient water vapor could essentially be described by an equilibrium fractionation process, with $d_4$ remained constant at a low nighttime minimum level (Fig. 6d) (Delattre et al., 2015; Huang and Wen, 2014). However, soil evaporation occurred synchronously with condensation. Soil evaporation in saturated ambient air (RH = 100 % at 2 m a.g.l.) is essentially an equilibrium fractionation process (Eichinger et al., 1996; Priestley and Taylor, 1972), which did not affect the variability of $d_4$ during P2b (Fig. 6d). Whereas, non-equilibrium fractionation is intrinsically dominant in the processes of soil evaporation in unsaturated ambient air (RH < 100 % at 2 m a.g.l.), which induced a slight $d_4$ decrease during P2a (Fig. 6d). In addition, cold air drainage along the valley to the bottom where the CH-CHA site is located (Eugster and Merbold, 2015), might have enhanced the effect of local soil evaporation on $\delta$H variability.

4.4 Ecological functions of non-rainfall water

From the perspective of ecological functions, dDew might be more important than previously thought, although it has no large-scale hydrological significance of moisture transfer from one part of the surface to another (Monteith, 1957). This can be expected if the transfer of moisture is from a hydrological pool that is inaccessible to plants (e.g., soil-diffusing vapor) to another that is accessible to plants (e.g., droplets forming or depositing on leaf surfaces or on the surface soil where it can be accessed by the fine roots). The condensation of soil-diffusing vapor was comparable with the condensation of ambient water vapor in our study (contributing 9–42% during our observation periods, Fig. 7d), and was the dominant pathway of NRW inputs during very calm night ($u < 0.7 \text{ m s}^{-1}$; see also Monteith (1957)) when the flux from ambient air to the grassland surface was very small. Soil vapor diffusion occurs as long as a temperature gradient exists (see the soil temperature at different depths in Fig. E1 of Appendix E), which results in vapor pressure differences along that gradient. Therefore, soil vapor diffusion transfers the deeper soil vapor to the surface, from where it moistens the air in contact with the soil surface. Subsequently, this moisture condenses onto foliage and becomes available to the plants. Wang et al. (2017) observed that 0.0092 mm of water were transferred from deeper soil layers to the surface by vapor diffusion in a grassland plot, although it was doubtful whether the water went onto foliage or was absorbed by the top soil. The soil diffusion rate increases with the decrease of soil water content (under volumetric soil moisture content higher than 10% (Barnes and Turner, 1998; Philip and De Vries, 1957)), which makes soil vapor diffusion more important in conditions of low soil moisture. The NRW inputs for dew and radiation fog are expected to be taken up by plants through foliar water uptake or be dripping down to wet the soil surface, thereby potentially
preventing permanent damage of the plants by drought stress (e.g., Schreel and Steppe (2020)). The ecological functions of NRW was also reflected in its thermal effect on the plants. The leaf wetting by NRW which potentially cooled the leaf surfaces by 1.5 °C in comparison to dry leaf surfaces (differences between \( T_{\text{lw}} \) and \( T_{\text{la}} \), Fig. 5a), thereby alleviating potential plant heat stress during the early morning hours when solar radiation quickly increases after sunrise.

Further research should thus focus on the plant water status in response to NRW inputs from dew and radiation fog. In addition, future research focusing on the continuous measurement of the isotopic compositions (\( \delta^{18} \text{O} \) and \( \delta^2 \text{H} \)) of soil vapor would give more quantitative insights on vapor transfer in soils during dew and radiation fog nights. The condensation of soil-diffusing vapor is expected to play a more important role in temperate grasslands than in arid grasslands, if soil salinity and canopy resistance are also taken into consideration: Soil salinity reduces the rate of soil vapor diffusion (Gran et al., 2011). In a laminar boundary layer during dew and radiation fog events, dense canopies in temperate grasslands (LAI was 1.5–3.2 m\(^2\) m\(^{-2}\) for summer 2018 at the CH-CHA site, Sect. 2.1) potentially shield the uppermost soil vapor from being exported into the near-surface atmosphere, while sparse canopies in arid grasslands (LAI around 0.5 m\(^2\) m\(^{-2}\) as e.g. in Wen et al. (2012)) should result in most of the soil-diffusing vapor being emitted into the atmosphere.

5 Conclusion

Our results reveal different input pathways for dew and radiation fog in a temperate grassland during three dry intensive observation periods in summer 2018 in Switzerland. Dew and radiation fog occurred in clear calm nights with very low wind speed (\( u < 0.7 \text{ m s}^{-1} \)) and weak turbulence with near-zero net water vapor flux at the vegetation surface (\( F_{\text{H2O}} \) at –0.4 to 0.3 mmol m\(^{-2}\) s\(^{-1}\)). Condensation of ambient water vapor during dew and radiation fog was found to be predominantly an equilibrium fractionation process, which was deduced from the rather constant \( \delta^2 \text{H} \), h\(^{-1}\) in ambient water vapor during dew and radiation fog. In unsaturated conditions (determined at the meteorological 2 m reference height), condensation occurred from ambient air above the canopy as well as soil-diffusing vapor below the canopy, as was indicated by a 3.4–3.7‰ decrease of \( d_4 \). Local evaporation at high relative humidity from 17:00 CET until sunset caused the lowering of \( d_4 \) to values in the range of 2.4‰ to 4.8‰ as compared to the higher daytime \( d_4 \) (12.2‰ to 18.0‰). A further decrease to \( d_4 \) values in the range of –11.8‰ to –4.7‰ was observed during the occurrence of dew and radiation fog at night. Dew only formed under unsaturated conditions with a mixed NRW condensing from ambient water vapor and soil-diffusing vapor. The comparison between the foliage NRW \( \delta_{\text{NRW}} \) and the equilibrium NRW \( \delta_{\text{NRW}} \) of ambient water vapor allowed us to trace the source of the NRW input pathways during dew formation. The NRW condensing from soil-diffusing vapor contributed 9–42% of the foliage NRW. The correlated \( d_4 – T_v \) and anti-correlated \( d_4 – \text{RH} \) suggested an only minor influence of large-scale air advection and highlighted the dominant role of local moisture as a source for ambient water vapor.

In future studies, continuous isotope measurements of foliage NRW, ambient water vapor and soil vapor should be combined with direct lysimetric and filter paper absorption measurements, as well as physiological measurements to more precisely quantify the NRW input pathways, and the mechanisms of plant water status responding to NRW input on foliage. Confirmation of dew and radiation fog inputs into temperate ecosystems during summer drought by the isotopic compositions of NRW and ambient water vapor would then allow assessing the potential response of these ecosystems to warming and increased frequency of summer droughts under the global climate changes.

The dew and radiation fog potentially produced 0.06–0.39 mm night\(^{-1}\) NRW gain on foliage, which was comparable with 2.8 mm day\(^{-1}\) daytime evapotranspiration. With increasing relative humidity, the share of vapor originating from soil vapor diffusion decreased, whereas the relevance of atmospheric water vapor for dew formation increased. This atmospheric water vapor had a rather local isotopic signature, which suggests that large-scale moisture advection only has a minor influence in the nocturnal NRW gains during dew and radiation fog events. Our results thus highlight the importance of NRW inputs to
temperate grasslands during prolonged dry periods and reveal the complexity of the local water cycle in such conditions including different pathways of water deposition.

6 Data availability

Data will be deposited at the ETH Zurich research collection at https://doi.org/10.3929/ethz-b-000445289.

Author contributions. YL, AR, WE, and FA designed the project. YL and AR performed the field experiment. YL carried out the laboratory work. YL performed all statistical analyses. FA wrote the code of data calibration for the ambient water vapor isotopes. FA and WE commented on the results of the data analysis. YL wrote and revised the manuscript, with contributions and feedbacks by FA, WE, AR, and NB.

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Figure 1. (a) Monthly precipitation and (b) monthly average temperature \( (T_a) \) from April to October in 2018 as compared to the corresponding months over 2006–2017.

Figure 2. Simplified schematics of non-rainfall water (NRW) inputs adapted from Monteith and Unsworth (2013), and Oke (2002): “aDew” means dew formed from ambient water vapor, “aFog” means fog formed from ambient water vapor; “aDew” and “aFog” are both condensed from ambient water vapor, thus “aNRW” represents the condensation of ambient water vapor if either dew or fog input, or the combination of both was meant; “dDew” means dew formed from soil-diffusing vapor. The horizontal wind speed \( (u) \) is zero at \( z_d + z_0 \).

Figure 3. Nocturnal vertical profiles of air temperature \( (T) \) vs. height \( (z, \text{ in m a.g.l.}) \) at 01:00 CET for the three events interpolated to the location of the measurement site based on ERA-5 reanalysis data (Hersbach et al., 2020; Horanyi, 2017).
Figure 4. The meteorological and eddy-covariance (EC) measurements at the CH-CHA site. The 30 min averages of (a) horizontal mean wind speed at 2 m a.g.l. \( u_{2m} \), (b) \( H_2O \) flux at 2.4 m a.g.l. \( F_{H2O} \), (c) relative humidity at 2 m a.g.l \( RH \), relative humidity with respect to the surface temperature \( h_{0w} \) for wet surface, and \( h_{0d} \) for dry surface; (d) 1 min averages of visibility (< 1 km with fog, and > 1 km with the absence of fog).
Figure 5. The meteorological and eddy covariance (EC) measurements at the CH-CHA site. The 30 min averages of (a) air temperature at 680 m a.g.l. \( T_a \), dew-point temperature of the ambient air \( T_d \), surface temperature for wet \( T_{0w} \) and dry \( T_{0d} \) assumptions, and (b) atmospheric specific humidity at 2.4 m a.g.l. \( q_a \), the saturation specific humidity respect to the surface temperature under wet \( q_{0w} \) and dry \( q_{0d} \) assumptions. P1a was from 17:00 CET until sunset with the weakened turbulence and increased specific humidity; P1b was a short-term variability of specific humidity; P2a was dew formation in unsaturated ambient air; P2b was dew and radiation fog in combination in saturated ambient air.
Figure 6. The volumetric mixing ratio and isotopic compositions for ambient water vapor. The 30 min averages and standard deviations of (a) volumetric ambient water vapor mixing ratio ($w$), and (b-d) the isotopic compositions of ambient water vapor ($\delta^{18}O_a$, $\delta^2H_a$, and $d_a$). P1a was from 17:00 CET until sunset with the weakened turbulence and increased specific humidity; P1b was a short-term variability of specific humidity; P2a was dew formation in unsaturated ambient air; P2b was dew and radiation fog in combination in saturated ambient air.
Figure 7. (a–c) Isotopic compositions of different non-rainfall water (fNRW, aNRW, and dDew), and (d) the proportions of dDew (fDew). “fNRW”, the NRW on foliage; “aNRW”, the NRW condensed from ambient water vapor; “dDew”, the dew component condensed from soil-diffusing vapor. P2a was dew formation in unsaturated ambient air; P2b was dew and radiation fog in combination in saturated ambient air.

Figure 8. The relationship of $\delta^{2}H$-fNRW vs. $\delta^{18}O$-fNRW with respect to the orthogonal regression of $\delta^{2}H$-aNRW vs. $\delta^{18}O$-aNRW and local meteorological water line (LMWL: $\delta^{2}H = 7.68 \times \delta^{18}O + 6.97$, Prechsl et al. (2014)). “fNRW” means non-rainfall water (NRW) on foliage, and “aNRW” means the NRW equilibrium from ambient water vapor. “RH” is relative humidity at 2 m a.g.l.
Figure 9. Comparing the isotopic compositions of the simulated NRW to the isotopic compositions of the NRW on foliage ($\delta_{\text{fNRW}}$). The $\delta_{\text{aNRW}}$ was calculated from ambient water vapor $\delta_a$ considering equilibrium fractionation, and $\delta_{\text{naNRW}}$ was calculated from ambient water vapor $\delta_a$ considering both equilibrium and non-equilibrium fractionation. P2a was dew formation in unsaturated ambient air; P2b was dew and radiation fog in combination in saturated ambient air.

Figure 10. The relationships of (a) $d_a$–$T_a$ and (b) $d_a$–RH for the 24 h measurements during the three events. The $d_a$ is the deuterium excess of ambient water vapor; $T_a$ is the ambient temperature at 2 m a.g.l.; RH is relative humidity at 2 m a.g.l.
Table 1. Partitioning the contribution of dDew from a mix of dDew and aDew. The fNRW means the non-rainfall water (NRW) on foliage; aNRW represents either dew or radiation fog, or dew and radiation fog in combination condensed from ambient water vapor; dDew means dew condensed from soil-diffusing vapor; f_dDew means the proportion of dDew in total foliage NRW.

| Event | Time       | Isotope       | fNRW   | aNRW  | dDew | f_dDew |
|-------|------------|---------------|--------|-------|------|--------|
| Event 1 | 3:00 CET   | δ^{18}O (%)  | −3.8   | −4.4±0.2 | −1.0 | 18–31  |
|        |            | δ^{2}H (%)    | −55.1  | −47.4±1.7 | −71.8 |
|        |            | d (%)         | −25.6  | −12.1±1.3 | −63.4 |

Event 2 21:30 CET

| Event 2 21:30 CET | No sampling, but extrapolating from 23:00 and 1:00 CET |
|-------------------|-------------------------------------------------------|
| 23:00 CET         | δ^{18}O (%)  | −3.4 | −4.3±0.2 | −1.0 | 28   |
|                   | δ^{2}H (%)    | −47.7 | −38.6±0.7 | −71.8 |
|                   | d (%)         | −20.7 | −4.4±1.3 | −63.4 |
| 1:00 CET          | δ^{18}O (%)  | −4.2 | −4.5±0.2 | −1.0 | 9    |
|                   | δ^{2}H (%)    | −43.5 | −40.8±1.0 | −71.8 |
|                   | d (%)         | −9.4 | −4.7±1.1 | −63.4 |

Table 2. Estimating the potential non-rainfall water (NRW) gain of the three events in our study according to the condensation rate of Monteith (1957). The fNRW means the non-rainfall water (NRW) on foliage; aNRW represents either dew or radiation fog, or dew and radiation fog in combination condensed from ambient water vapor; dDew means dew condensed from soil-diffusing vapor.

| Event | Period (h night^{-1}) | Condensation rate following Monteith (1957) (mm h^{-1}) | Potential NRW gain (mm) |
|-------|------------------------|--------------------------------------------------------|------------------------|
|       | dDew | aNRW | dDew | aNRW | dDew | aNRW | total |
| Event 1 | 4.0 | 4.0 | 0.01 – 0.02 | 0.004 – 0.035 | 0.04 – 0.08 | 0.02 – 0.14 | 0.06 – 0.22 |
| Event 2 | 5.5 | 8.0 | 0.01 – 0.02 | 0.004 – 0.035 | 0.06 – 0.11 | 0.03 – 0.28 | 0.09 – 0.39 |
| Event 3 | 0.0 | 10.5 | 0.01 – 0.02 | 0.004 – 0.035 | 0 | 0.04 – 0.37 | 0.04 – 0.37 |

Table 3. Evapotranspiration rate during the corresponding continuous no-rain periods of the three events processed from EC measurements.

| Event | Corresponding continuous no-rain periods (yyyy-mm-dd) | Average daily (daytime) evapotranspiration (mm day^{-1}) |
|-------|--------------------------------------------------------|--------------------------------------------------------|
| Event 1 | 2018-07-23 | 2.7 |
|        | 2018-07-24 | 3.3 |
|        | 2018-07-25 | 3.3 |
|        | 2018-07-26 | 3.4 |
|        | 2018-07-27 | 3.0 |
| Event 2 | 2018-08-18 | 2.6 |
|        | 2018-08-19 | 2.8 |
|        | 2018-08-20 | 3.0 |
|        | 2018-08-21 | 2.9 |
| Event 3 | 2018-09-08 | 3.1 |
|        | 2018-09-09 | 2.7 |
|        | 2018-09-10 | 2.1 |
|        | 2018-09-11 | 2.0 |
|        | 2018-09-12 | 1.7 |
Appendix

Appendix A: Measurements and samplings at the CH-CHA site

Figure A1. Measurements and sampling at the CH-CHA site (Satellite Image: © CNES/Spot Image/swisstopo, NPOC). “L2130-i” represents the isotopic compositions and mixing ratio measurements for ambient water vapor; “EC & meteo” represents the eddy-covariance and meteorological measurements; “Sampling of fNRW” represents the sampling of non-rainfall water droplets on foliage.

Appendix B: Meteorological and eddy covariance calculations

The saturation specific humidity \( q_0 \) g kg\(^{-1}\) for wet \( q_{0w} \) and dry \( q_{0d} \) vegetation surfaces was calculated after Campbell and Norman (1998) as:

\[
q_0 = \frac{622 \cdot e_{0a}}{p - 0.378 \cdot e_{0a}} \cdot \tag{B1}
\]

where \( p \) in hPa is air pressure, and \( e_{0a} \) in hPa is saturation vapor pressure at \( T_0 \) calculated after Tetens formula (Buck, 1981) as:

\[
e_{0a} = 6.11 \cdot \exp\left(\frac{17.502 \cdot T_0}{T_0 + 240.97}\right) \cdot \tag{B2}
\]

The dew point temperature \( T_d \) °C was calculated after Campbell and Norman (1998) as:

\[
T_d = 240.97 \cdot \frac{\ln\left(\frac{e_{sa}}{RH} \cdot \frac{1}{T_0 + 240.97}\right)}{\ln\left(\frac{1}{T_0 + 240.97}\right)} \cdot \tag{B3}
\]

where \( e_{sa} \) in hPa is saturation vapor pressure at \( T_a \) calculated after Tetens formula (Buck, 1981) as:

\[
e_{sa} = 6.11 \cdot \exp\left(\frac{17.502 \cdot T_a}{T_a + 240.97}\right) \cdot \tag{B4}
\]
Appendix C: Calculating the isotopic compositions of the NRW from the isotopic compositions of ambient water vapor under equilibrium fractionation

After Horita and Wesolowski (1994), the isotopic compositions of the NRW ($\delta^{18}$O$_{NRW}$ and $\delta^2$H$_{NRW}$) equilibrium from the isotopic compositions of ambient water vapor ($\delta^{18}$O and $\delta^2$H) at surface temperature $T_{sw}$ were calculated as:

$$
\delta^{18}\text{O}_{NRW} = \left[10^3 + \delta^{18}\text{O}\right] \cdot \exp\left(0.35041 \cdot \frac{10^6}{(T_{sw}+273.15)}\right) - 1.6664 \cdot \frac{10^6}{(T_{sw}+273.15)} + \frac{6.7123}{T_{sw}+273.15} - \frac{7.685}{10^3} \cdot 10^3, \quad (C1)
$$

and

$$
\delta^2\text{H}_{NRW} = \left[10^3 + \delta^2\text{H}\right] \cdot \exp \left(1.1588 \cdot \frac{(T_{sw}+273.15)^3}{10^3} - 1.6201 \cdot \frac{(T_{sw}+273.15)^2}{10^3} + 0.79484 \cdot \frac{(T_{sw}+273.15)}{10^3} - 0.16104 + 2.9992 \cdot \frac{10^6}{(T_{sw}+273.15)} \right) = 10^3. \quad (C2)
$$

Appendix D: Simulating the isotopic compositions of the condensate from the isotopic compositions of ambient water vapor considering both equilibrium and non-equilibrium fractionation factors

In our results shown in Fig. 7, we assumed equilibrium fractionation was dominant during the condensation of ambient water vapor, hence we calculated $\delta^{18}$O$_{NRW}$ and $\delta^2$H$_{NRW}$ equilibrium from ambient water vapor (Figs. 7 and 9) using (Eqs. C1 – C2). Whereas, Wen et al. (2012) adopted the method to simulate the isotopic compositions of the NRW from ambient water vapor considering both equilibrium and non-equilibrium fractionation factors ($\delta^{18}$O$_{naNRW}$ and $\delta^2$H$_{naNRW}$ in Fig. 9). To compare these two methods, we applied the method by Wen et al. (2012) on our data, and the equations for calculating $\delta^{18}$O$_{naNRW}$ and $\delta^2$H$_{naNRW}$ was:

$$
\delta^{18}$O$_{naNRW} = \frac{\delta^{18}$O$_{a} + \epsilon_{eq} \cdot \delta^{18}$O$_{a} - \delta^{18}$O$_{a} \cdot \lambda_{h} / (1 - \lambda_{h})}{1 - \lambda_{h}} \cdot \frac{\delta^{18}$O$_{a} - \delta^{18}$O$_{a} \cdot \lambda_{h} / (1 - \lambda_{h})}{1000} \quad (D1)
$$

where $\epsilon_{eq}$ is the non-equilibrium fractionation factor in permil, calculated from $\epsilon_{eq} = m \cdot (1 - D/D_l) \times 1000$ % following Lee et al. (2009), given $D/D_l$ ($^4$H) = 0.9723, $D/D_l$ ($^2$H) = 0.9755 following Merlivat (1978), and $m = 0.67$ for laminar flow following Dongmann et al. (1974); $\epsilon_{eq}$ is equilibrium fractionation factor in permil calculated from $(1 - 1/\alpha) \times 1000$ % with $\alpha$ calculated as:

$$
\alpha \left(^{18}$O\right) = \exp \left(0.35041 \cdot \frac{10^6}{(T_{sw}+273.15)}\right) - 1.6664 \cdot \frac{10^6}{(T_{sw}+273.15)} + \frac{6.7123}{T_{sw}+273.15} - \frac{7.685}{10^3} \cdot 10^3 \quad (D2)
$$

$$
\alpha \left(^{2}$H\right) = \exp \left(0.35041 \cdot \frac{10^6}{(T_{sw}+273.15)}\right) = 1.6664 \cdot \frac{10^6}{(T_{sw}+273.15)} + \frac{6.7123}{T_{sw}+273.15} - \frac{7.685}{10^3} \cdot 10^3 \quad (D3)
$$

Appendix E. Soil vapor diffusion occurred as long as the temperature gradient generated

The dDew was condensed from soil-diffusing vapor, which occurred as long as the temperate gradient generated. The temperature gradient was largest at the land – atmosphere interface (Fig. 5a), but within the soil profile, temperate gradient also generated. The soil temperature at 5 cm, 10 cm, 15 cm, 20 cm, 30 cm, 40 cm, and 50 cm in depth were shown in Fig. E1 (ML2x, Delta-T Devices Ltd., Cambridge, United Kingdom).
Figure E1. Soil temperature at different depths.