

Growing season water balance of an inner alpine Scots pine (*Pinus sylvestris* L.) forest

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We estimated components of the water cycle of a 150-year-old *Pinus sylvestris* forest in an inner Alpine dry valley of the Tyrol, Austria throughout five growing seasons. Forest canopy transpiration (T_c) was measured by sap flow measurements scaled to the stand canopy level. Estimates of understory transpiration and forest floor evaporation (ET_u) were derived from the soil water budget method, while interception (I) was modelled. Growing season cumulative evapotranspiration ($ET = T_c + ET_u + I$) varied between 256 and 322 mm or 51 to 79% of the growing season precipitation. The contribution of T_c , ET_u , and I to ET were 33, 40 and 27% respectively. Although these values of each layer (evapo)-transpiration are in good agreement with studies carried out in other European Scots pine forests, our estimated growing season total forest water use ($T_{tot} = T_c + ET_u$) of 200-244 mm is at the lower end of values reported for coniferous forest ecosystems, and thus reflects an adaptation to the low shallow soil water availability. We conclude that Scots pine forests in inner alpine dry valleys are able to cope with high evaporative demand, even when shallow soil water availability is limited.

Keywords: Forest Water Balance, Scots Pine, Dry Inner Alpine Valley, Evapotranspiration, Interception, Runoff

Introduction

Scots pine (*Pinus sylvestris* L.) is a widespread conifer species across Eurasia. In Europe, *P. sylvestris* can be found from the boreal north (70 °N) in Scandinavia to the Mediterranean basin (37 °N) in southern Spain (Prus-Glowack et al. 2012) and it forms open stands in inner alpine dry valleys. As a consequence of climate change, changes of the ecosystem water balance are expected to occur especially in open *Pinus sylvestris* forests in inner-alpine dry valleys in Italy, Switzerland and Austria, which are characterized by high summer temperatures, low precipitation and limited soil water availability (Zweifel et al. 2009, Oberhuber & Gruber 2010, Wieser et al. 2014), and thus are considered to be sensitive to climate change (Rigling et al. 2013, Schlesinger & Bernhardt 2013).

The water balance of a forest ecosystem is determined by the water input through precipitation (P). Some of the P adheres on the above-ground vegetation and evaporates before reaching the ground – the so-called interception (I). The remaining P reaching the soil surface runs off (R_s , surface runoff), infiltrates into the soil, and percolates through the water table (R_D , deep seepage), finally leaving the system as base flow. The amount of water which is held against gravitational forces within the soil matrix can be taken up by the plant roots and is finally transpired by the canopy (T_c) and evapotranspires from the forest floor (ET_u , soil evaporation and understory transpiration) into the atmosphere. Thus, over short periods there are

also changes in the soil water content (ΔW). All the components involved are given in mm (= kg m⁻²), and the water balance equation of a forest ecosystem reads (eqn. 1):

$$P = T_c + ET_u + I + R_s + R_D + \Delta W \quad (1)$$

in which all terms, except ΔW , are flux densities.

As it is frequently very difficult to measure all the components of eqn. 1, the water balance equation (eqn. 1) can be simplified as (eqn. 2):

$$P = ET + R + \Delta W \quad (2)$$

where ET is the sum of all water fluxes from the forest to the atmosphere ($T_c + ET_u + I$) and R (total runoff) is the sum of deep seepage and surface runoff.

Although data on the water balance of Scots pine forests have been reported by several authors (Vincke & Thiry 2008, Giesen et al. 2010, Ilvensiemi et al. 2010), we are not aware of any study focusing on the water balance of Scots pine stands in inner Alpine dry valleys. Therefore, it was the aim of this paper to quantify and describe the main terms of the hydrological cycle in an inner alpine dry Scots pine forest and the parameters that affect total forest water use ($T_{tot} = T_c + ET_u$). One particular task was to quantify the contribution of T_c , ET_u , and I on ET , by applying scaled sap flow rates, estimates of ET_u using the soil water budget method and estimates of I derived from precipitation measurements and literature data. To achieve this goal, we monitored environmental data (air temperature, T ; relative humidity, RH ; global radiation, GR ; precipitation, P ; soil moisture, θ ;

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and sap flow density, Q_s) during the growing seasons (April 1 – October 31) of 2007, 2008, 2009, 2011, and 2012. We, however, were not able to take measurements during the 2010 growing season due to a lack of funding.

Material and methods

Study site

The study was carried out in an open Spring heath-Scots pine forest (*Erico-Pinetum typicum* – Ellenberg & Leuschner 2010) growing on a postglacial rock-slide area, situated in the montane belt (750 m a.s.l.) within the inner Alpine dry Inn valley (Tyrol, Austria – 47° 14' 00" N, 10° 50' 20" E). The study site is south exposed (40° inclination – Wieser et al. 2014) and pioneer vegetation (mainly *Erica carnea* and *Sesleria varia*) prevails in the ground flora. At the time of the study (2007-2012), the trees were 150 ± 30 years old and their height ranged between 4 and 5 m. The stand density was 600 trees ha⁻¹, with a canopy coverage of 30%, a leaf area index (LAI) of 0.55 (Oberhuber & Gruber 2010), and a basal area of 21.8 m² ha⁻¹. The tree diameter at breast height (DBH) averaged 24.5 ± 5.1 cm and varied between 18 and 32 cm (Fig. 1).

The site is characterized by a continental climate with a mean annual temperature of 7.3 °C and a mean annual precipitation of 718 mm (long-term mean 1911-2012 at Ötz, 812 m a.s.l., 5 km south of the study site), and spring drought, occurring every third year at average (Fliri 1975). Furthermore, the water balance is strained by a low water holding capacity of the protorendzina, i.e., a rendzic leptosol (O-A-C profile) according to the World Base for Soil Resources classification system (FAO 2006). The A horizon reaches a depth of 10 cm and consists of coarse-textured materials. The texture is dominated by the sand (54%) and the silt fractions (44%) with almost no clay (2%). According to the water retention curve obtained for the study site (Wieser et al. 2014), the water holding capacity is 42 mm at field saturation. The corresponding values for field capacity (-0.033 MPa, *sensu*

Blume et al. 2010) is 23 mm, and the wilting point (-1.5 MPa) is reached at 7 mm.

Environmental measurements

Air temperature (T), relative humidity (RH), global radiation (GR), and precipitation (P) were monitored 2 m above the ground in an open, non-vegetated ridge in order to minimize interception loss. During the study periods 2007-2009 all these environmental data were recorded with an automatic weather station (ONSET®, Pocasset, MA, USA). In 2011 and 2012 T, RH (HMP45C®, Campbell Scientific, Shepshed, UK), GR (SP-Lite®), and P (ARG100®) were recorded with a Campbell CR10X® data logger (Campbell, Scientific, Shepshed, UK). Both data loggers were programmed to record 30-min averages of all the environmental data sampled every minute during the growing seasons (April 1 – October 31) of 2007, 2008, 2009, 2011, and 2012. Throughout the same periods, soil moisture (θ , m³ m⁻³) was continuously monitored in 5-10 cm soil depth at three sites with capacitive soil moisture sensors (Cyclobis, proprietary development, University Innsbruck, Austria). The measuring interval of these sensors was 30 min, and due to small-scale variability of the soil structure (Oberhuber & Gruber 2010) values of the three soil moisture sensors were averaged. Finally, T, RH, GR, and θ were expressed as daily means, while P was summed up to daily totals (48 values per day each).

Radiation and temperature data obtained 2 m above the ground were used to calculate daily potential evapotranspiration (PET, mm day⁻¹) using the following formula (Turc 1961, see also Kucerova et al. 2010 – eqn. 3):

$$PET = \frac{[(R_g/0.041868)+50] \cdot 0.013 \cdot T_a}{T_a+15} \quad (3)$$

where R_g is the daily sum of global radiation (MJ m⁻² day⁻¹) and T_a is the daily mean air temperature.

Shallow soil water deficit was quantified as relative extractable shallow soil water (REW, dimensionless) and was calculated

as follows (Granier et al. 1999, Vincke & Thiry 2008) – eqn. 4):

$$REW = \frac{W_d - W_{min}}{W_{fc} - W_{min}} \quad (4)$$

where W_d is the actual soil water content (mm) on day d, W_{min} is the minimum soil water content at the permanent wilting point, and W_{fc} is soil water content at field capacity. For *P. sylvestris* trees in nearby more mesic *Erico-Pinetum typicum* forest, an $REW \leq 0.4$ has been shown to limit crown conductance (Leo et al. 2013).

Determination of water fluxes

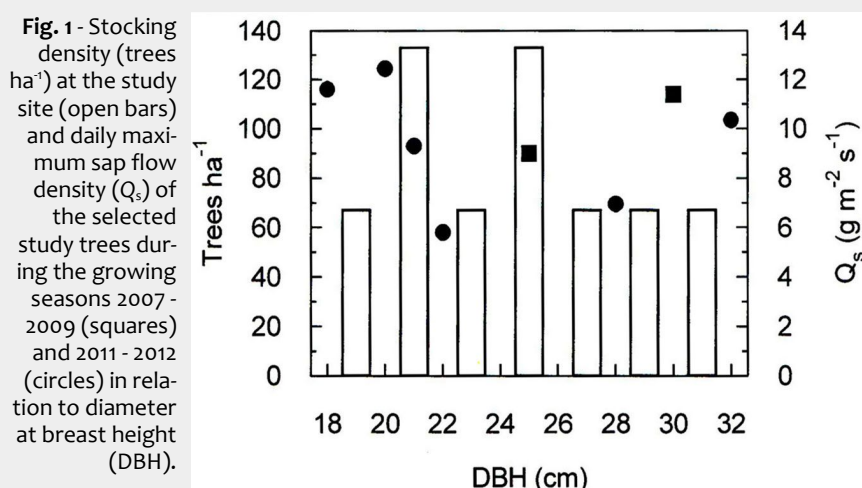
Sap flow density and estimation of canopy transpiration

Sap flow density (Q_s) was monitored with thermal dissipation sensors according to Granier (1985) by battery-operated sap flow systems (M1Sapflow System®, PROSALOG UP, Umweltanalytische Produkte GmbH, Cottbus, Germany). In each study tree (two in 2007-2009 and six in 2011-2012) one 20 mm long sensor was installed into the outer xylem (0-20 mm from the cambium) at breast height. Accounting for variations of Q_s across the cross section of the trunk (Cermak et al. 2004), sensors were also installed in the next 20 mm thick xylem band (20-40 mm sapwood depth, termed “inner xylem”) in two of the selected study trees in 2011 and 2012. The two probes of each sensor were installed vertically 15 cm apart on the north facing side of the 1.3 m above the ground. The upper probe was heated continuously, while the lower one was unheated, remaining at trunk temperature for reference. The temperature difference between the upper heated and the lower reference probe was recorded every 30 min. The sensors were shielded with a thick aluminium-faced foam cover to prevent exposure to rain, and to avoid physical damage and thermal influences from radiation.

For each sensor, Q_s (g m⁻² s⁻¹) was calculated using the original calibration coefficients of Granier (1985). The sensors in the outer and the inner xylem covered most of the active sapwood which in our study trees had an average thickness of 41.5 ± 5.4 mm. In the study trees, Q_s did not vary considerably across sapwood depth ($P > 0.5$). In 2011 Q_s averaged 3.9 ± 2.6 and 4.0 ± 2.8 g m⁻² s⁻¹ in the outer (0-20 mm) and the inner xylem (20-40 mm), respectively. The corresponding values for 2012 were 4.8 ± 1.7 and 4.8 ± 3.01 g m⁻² s⁻¹, respectively. Thus, we assumed uniform Q_s across sapwood depth and our installation sampled 100% of the whole water flow in this conifer. Since Q_s was not correlated with DBH at our study site ($R^2 = 0.001$, $P = 0.45$ – Fig. 1), canopy transpiration (T_c) was determined as follows (Zimmermann et al. 2000 – eqn. 5):

$$T_c = Q_{s-mean} \cdot A_{s-stand} \quad (5)$$

where Q_{s-mean} is the average sapwood den-



sity if the sample trees and $A_{s-stand}$ is the cumulative sapwood area per ground surface area ($m^2 m^{-2}$). A stand specific sapwood area of $10.5 cm^2 m^{-2}$ (Wieser et al. 2014) was used for up-scaling. Scaling up on this method may be applied in uniform stands with a limited range in tree size and a high individual variation of tree specific Q_s (Cermak et al. 2004) due to environmental factors (e.g., drought) as it was the case at our study site. Moreover, there is also evidence that when using 2 or 6 sample trees, the scaling error for estimates of Q_{s-mean} is usually less than 12 and 7%, respectively (Cermak et al. 2015).

Estimation of forest floor evapotranspiration

Forest floor evapotranspiration (ET_U) could not be neglected at our study plot because of the low canopy coverage (30%). Daily ET_U was estimated from shallow soil water content measurements, according to the soil water budget approach (Kucerova et al. 2010). In our experimental plot an important amount of P leaches to deeper soil layers via the macropores (gravitational water) after refilling the shallow soil (personal unpublished observations). Therefore, we assumed that ET_U solely absorbed shallow soil water (Waring & Running 1998) and that ET_U is mainly dependent on evaporative demand reaching the forest floor (Granier et al. 1999). Because of P inputs, water movement within the soil profile, and the ET_U influence on θ , we eliminated all days with rain and also up to two days following heavy rain events ($> 25 mm$) for estimating ET_U from ΔW measurements. Thus, ET_U was calculated as follows (eqn. 6):

$$ET_U = -\Delta W \quad (6)$$

where ΔW (mm) is the change in shallow

soil water content between two consecutive days. This allowed to use approximately 65% of growing season days for estimating ET_U from the soil water budget approach. Missing ET_U values were calculated using a regression between ET_U and PET ($R^2 = 0.92$; $P < 0.001$), as there is evidence that the forest floor vegetation is poorly coupled to the atmosphere (Jarvis & McNaughton 1986, Berbigier et al. 1991).

Interception and runoff

Canopy interception (I) was determined according to measurements of Brechtel (1965) carried out in a 120-year-old *P. sylvestris* forest, with a canopy coverage of 30%. Basically, we assumed I to be 23% of bulk precipitation for rainfall events $< 7 mm$ and 10% of bulk precipitation for rainfall events $> 7 mm$. Finally, runoff (R) was calculated as the residual of eqn. 2.

Data analysis

In this study, we also emphasized the comparison of total forest water use (T_{tot}) as affected by shallow soil water availability (non-limiting vs. limiting shallow soil water availability). Single variable analyses were used to examine the response of T_{tot} to GR, VPD and W . While correlations of T_{tot} with GR and W were obtained by linear regression analysis, the relationship of T_{tot} with VPD was analysed using the following exponential saturation function (eqn. 7):

$$T_{tot} = [1 - (\exp(-a \cdot VPD))] \cdot T_{totMAX} \quad (7)$$

where a is a fitting parameter and T_{totMAX} is the maximum daily mean total forest water use. In addition, we used a multiple linear regression model including GR, VPD, and W , as explanatory variables for T_{tot} . All the analyses were performed using the software package SPSS® ver. 16.0 for Windows (SPSS Inc., Chicago, USA) and nonlinear

curve fits were performed using Fig.P® for Windows (FigP Software Corporation, Hamilton, ON, Canada).

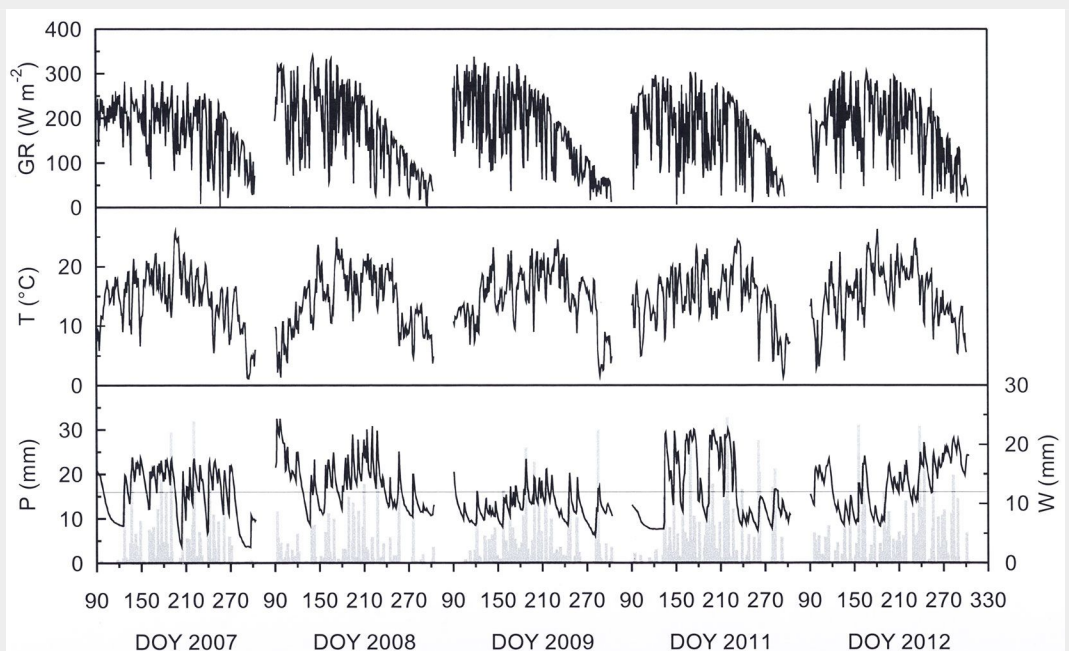
Results

Environmental conditions

Seasonal patterns of environmental conditions obtained during the growing seasons (April 1 – October 31) of 2007, 2008, 2009, 2011, and 2012 were representative for the climatic conditions at the study site, which is characterized by a dry and cool spring and a wet and warm summer. Due to differences in cloud cover, global radiation (GR) varied between $6.0 W m^{-2}$ (8 June 2011) and $339.6 W m^{-2}$ (22 May 2008 – Fig. 2), averaging $177 W m^{-2}$ in 2007, $173 W m^{-2}$ in 2008, $171 W m^{-2}$ in 2009, $168 W m^{-2}$ in 2011, and $174 W m^{-2}$ in 2012. Daily mean air temperature (T) was $14.6^\circ C$ in 2007, $14.0^\circ C$ in 2008, $14.9^\circ C$ in 2009, $14.6^\circ C$ in 2011 and $15.0^\circ C$ in 2012 and varied between $1.1^\circ C$ on October 21, 2007 and $26.3^\circ C$ on June 1, 2012 (Fig. 2). Daily mean vapour pressure deficit (VPD) was $0.53 kPa$ in 2007, $0.50 kPa$ in 2008 and in 2009, $0.62 kPa$ in 2011 and $0.65 kPa$ in 2012. Daily mean VPD approached zero on rainy days and reaching values up to $2.0 kPa$ on warm and sunny days during late spring and summer (data not shown).

Precipitation (P) varied considerably during the course of this investigation (Fig. 2). Of the five years examined, 2008 was the driest growing season ($325 mm$) and 2012 was the wettest growing season with a total of $589 mm$ (Tab. 1). Shallow soil water content (W) varied between a maximum of $24 mm$ on April 3, 2008 and a minimum of $3 mm$ on October 23, 2007 (Fig. 2). Average W over the growing seasons 2007, 2008, 2009, 2011, and 2012 was 12, 13, 10, 12 and 13 mm, respectively. Based on an REW of 0.4 ($W = 12 mm$), soil water deficit characterized 24, 14, 77, 50, and 32% of the growing

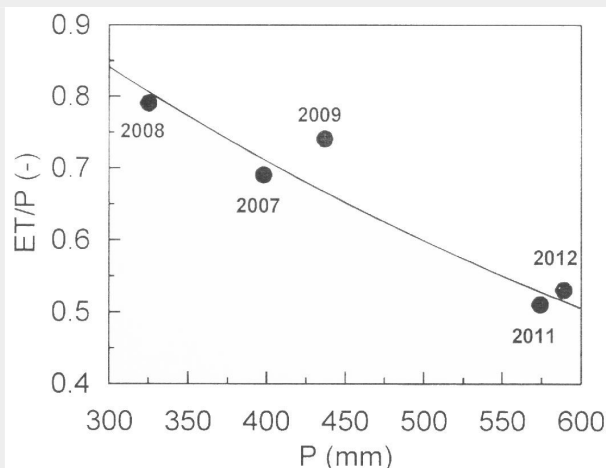
Fig. 2 - Seasonal course of daily mean global radiation (GR), air temperature (T), daily sum of precipitation (P) and shallow soil water content (W) in 0-10 cm soil depth during the growing seasons (April 1 – October 31) 2007, 2008, 2009, 2011, and 2012. The fine solid line indicates shallow soil water deficit.



Tab. 1 - Growing season (April 1 – October 31) totals of precipitation (P , mm), canopy transpiration (T_c , mm), forest floor evapotranspiration (ET_u , mm), total forest water use ($T_{tot} = T_c + ET_u$, mm), interception (I , mm), total actual evapotranspiration ($ET = T_c + ET_u + I$, mm), change in the soil water content (ΔW , mm), runoff (R , mm), the ratio of each layer (evapo)-transpiration to ET , as well as evapotranspiration (ET/P) and runoff (R/P) normalized to P . (CV): coefficient of variation.

Year	P	T_c	ET_u	T_{tot}	I	ET	ΔW	R	T_c/ET	ET_u/ET	T_{tot}/ET	I/ET	ET/P	R/P
2007	398	90	118	208	65	273	-8	133	0.33	0.43	0.76	0.24	0.69	0.33
2008	325	116	86	202	55	256	-6	75	0.45	0.34	0.79	0.21	0.79	0.23
2009	437	109	135	244	78	322	-7	123	0.34	0.42	0.76	0.24	0.74	0.28
2011	574	78	122	200	96	295	-2	281	0.26	0.41	0.68	0.32	0.51	0.49
2012	589	85	126	211	100	310	7	273	0.27	0.41	0.68	0.32	0.53	0.46
Average	465	95	117	213	79	291	-3	177	0.33	0.40	0.73	0.27	0.65	0.36
CV (%)	25	16	15	8	25	9	-224	53	20	11	6	17	19	31

Fig. 3 - Growing season evapotranspiration normalized with precipitation (ET/P) against precipitation (P). Points were fit by the exponential regression: $y = 1.40 \cdot \exp(-0.0017 \cdot x)$; $R^2 = 0.88$; $p < 0.001$.



seasons in 2007, 2008, 2009, 2011, and 2012, respectively (Fig. 2).

Seasonal water fluxes and influencing factors

Growing season cumulative evapotranspiration (ET) ranged from 256 mm (2008) to 322 mm (2009) (Tab. 1), or 51% (2011) to 79% (2008) of growing season precipitation with coefficients of variation of 9 and 19%, respectively (Tab. 1). Growing season cumulative evapotranspiration was composed of the following fluxes: 78 (2011) to 116 mm (2009) from T_c , 86 mm (2008) to 135 mm (2008 and 2009) from ET_u , and 55 mm (2008) to 100 mm (2012) from I (Tab. 1). Consequently, growing season total forest water use ($T_{tot} = T_c + ET_u$) varied between 200 mm in 2011 and 244 mm in 2009 (Tab. 1). Accordingly, for the stand T_{tot}/ET varied between 68 and 79% and the contribution of I to ET was 21 to 32% (Tab. 1).

The potential driving forces considered (global radiation, air temperature, vapour pressure deficit, soil water availability, and precipitation) explained little of the variability in growing season ET (all $p > 0.67$ – data not shown). The best predictor was the growing season mean air temperature, which in a linear regression explained 23% of the variability in growing season ET ($p = 0.41$). Restricting the analysis to T_{tot} ($= T_c + ET_u$), where most of the ET occurs (Tab. 1), improved regression statistics only slightly (all $p > 0.40$ – data not shown).

When ET was normalized with P , we found a statistically significant non-linear relationship ($R^2 = 0.88$; $p < 0.001$) with P (Fig. 3), indicating that the fraction of P evaporated to the atmosphere increased with decreasing P , while the runoff normalized to precipitation increased during years with ample P (Tab. 1).

Seasonal trends in daily total forest water use (T_{tot}) per unit ground area in general followed those of GR and T (Fig. 4, cf. Fig. 2). T_{tot} varied between 2.07 mm d⁻¹ during cloudless days in summer (July 21, 2007) and 0.14 mm d⁻¹ (October 8, 2011 and October 15, 2012) during rainy days in fall (Fig. 4). Annual growing season T_{tot} varied between 200 mm in 2011 and 244 mm in 2009 (Tab. 1), increased almost linearly from April throughout August and in general

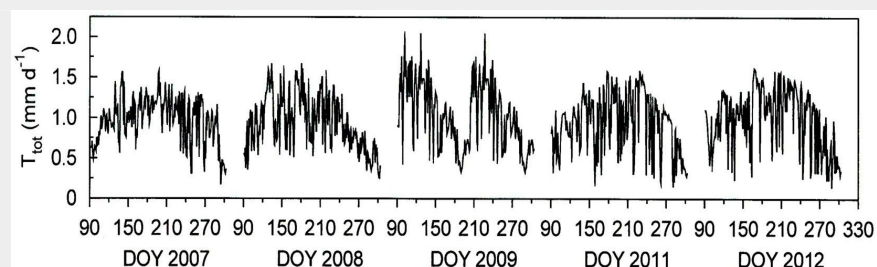


Fig. 4 - Seasonal course of daily total forest water use (T_{tot}) per unit ground surface area during the growing seasons (April 1 – October 31) 2007, 2008, 2009, 2011, and 2012.

Fig. 5 - Inter-annual variation of cumulative total forest water use (T_{tot}) per unit ground surface area during the growing seasons (April 1 – October 31) 2007, 2008, 2009, 2011, and 2012.

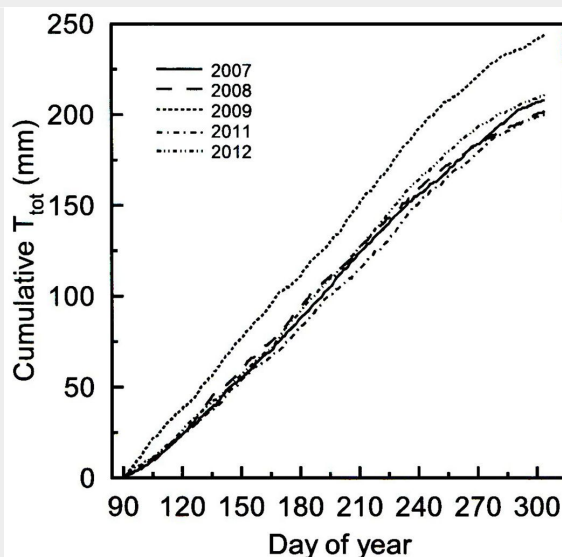
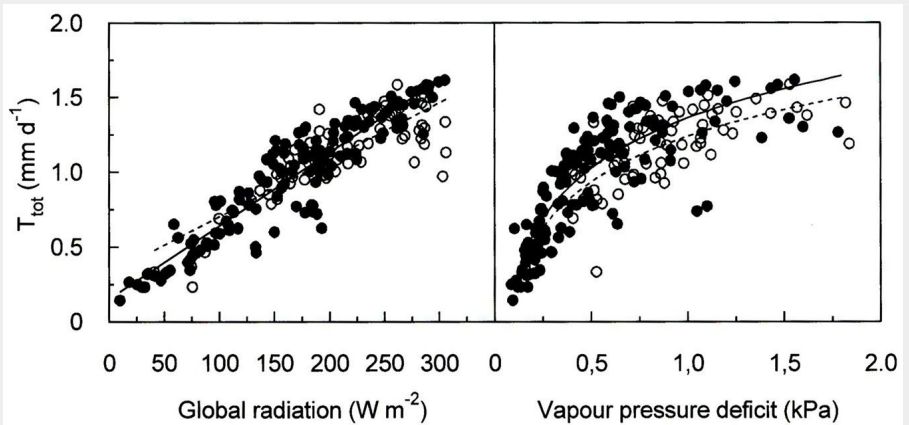


Fig. 6 – Daily total forest water use (T_{tot}) as a function of global radiation (GR, left) and vapour pressure deficit (VPD, right) under non-limiting (solid symbols and solid line) and limiting (open symbols and dotted line) shallow soil water content. Points were fit by linear and exponential saturation functions for GR and VPD, respectively. Global radiation: non-limiting $y = 0.005 \cdot \text{GR} + 0.15$, $R^2 = 0.87$; limiting W: $y = 0.004 \cdot \text{GR} + 0.32$, $R^2 = 0.69$. Vapour pressure deficit: non-limiting W: $y = [1 - (\exp(2.03 \cdot \text{VPD}))] \cdot 1.63$, $R^2 = 0.74$; limiting W: $y = (1 - (\exp(1.59 \cdot \text{VPD})) \cdot 1.63$, $R^2 = 0.63$. All $p < 0.001$.



tended to decline gradually towards the end of the growing season throughout fall (Fig. 5).

Topsoil water availability did not considerably modify the response of T_{tot} to GR and VPD. When examined at a daily timescale, these results generally reflected positive correlations between T_{tot} and both environmental factors, as shown for the growing season 2012 in Fig. 5. We obtained linear correlations between T_{tot} and GR at R^2 levels of 0.87 and 0.69 under conditions of non-limiting and limiting soil water availability (both p -values < 0.001 – Fig. 6), respectively. With respect to VPD, T_{tot} increased sharply at low VPD and tended to saturate at mean daily VPD values > 1.2 kPa under conditions of non-limiting shallow soil water availability and VPD values > 1.5 kPa when shallow soil water availability was limiting (Fig. 6). Vapour pressure deficit explained 74 and 63% (both p -values > 0.001) of the variation in T_{tot} under conditions of non-limiting and limiting shallow soil water availability, respectively. We, however, did not obtain an effect of shallow soil water availability on T_{tot} ($p = 0.24$, data not shown). A multiple linear regression model including GR, VPD and W explained 82% of the observed variability in T_{tot} ($p < 0.001$). In addition, the multiple linear regression analysis also clearly indicated that GR (β -coefficient = 0.79) had a greater effect on T_{tot} than VPD (β -coefficient = 0.20), while the effect of W on T_{tot} was nil (β -coefficient = 0.04).

Discussion

In the present study we focused on the growing season water balance of a 150-year-old *P. sylvestris* forest in an inner Alpine dry valley in Tyrol, Austria, where soil water availability is limited by a low water holding capacity of the shallow soil, and where trees (Leo et al. 2013) and understory vegetation do not have access to groundwater. Shallow soil water reserves demonstrated clear seasonal variations (Fig. 3). Based on an RWE of 0.4 ($W < 12$ mm) limited shallow soil water availability prevailed for 14% (36 days in 2008) to 77% (165 days in 2009) during an entire growing season. In contrast to findings of Sturm et

al. (1996) who showed that water loss of a Scots pine forest in Germany was significantly reduced when θ dropped below $0.16 \text{ m}^3 \text{ m}^{-3}$, we did not detect any considerable effect of shallow soil drought on daily and total forest water use ($T_{\text{tot}} = T_c + ET_u$). Even during a 90-day period in spring 2009 (DOY 95–185) when W was continuously below 12 mm ($\approx \theta$ of $0.12 \text{ m}^3 \text{ m}^{-3}$ – Fig. 2), we hardly detected any significant effect on daily and total T_{tot} (Fig. 4 and Fig. 5).

Our observed similar linear increase of T_{tot} with increasing GR (Fig. 6) under conditions of limiting and non-limiting shallow soil water availability reflects the energy demand for the evaporation of water (Campbell & Norman 1998) and indicates that limitations of transpiration by stomatal closure; due to low soil water availability and/or dry air played a minor role at this site (Wieser et al. 2014, Schuster et al. 2016). This is corroborated by the dependency of daily T_{tot} on the vapour pressure deficit whose slope tended to decrease only at very high VPD values (Fig. 6), as also observed by eddy covariance measurements of water vapour exchange over forests at other study sites associated in the FLUXNET project (Law et al. 2002).

Our estimated growing season ET averaged 291 ± 27 mm (Tab. 1) and was composed of the following fluxes: $33 \pm 7\%$ from T_c , $40 \pm 4\%$ from ET_u , $27 \pm 5\%$ from I (Tab. 1). These values obtained for each layer (evapo)-transpiration (T_c , ET_u and I – Tab. 1) are in good agreement with T_c , ET_u and I values reported for other *P. sylvestris* forest ecosystems in Europe (Wedler et al. 1996, Lüttenschwager et al. 1999, Llores et al. 2008, Vincke & Thiry 2008, Gielen et al. 2010, Ilvensiemi et al. 2010), and Central Siberia (Kelliher et al. 1998, Zimmermann et al. 2000).

Our estimated growing season T_{tot} of 200 to 244 mm was within the lower end of values reported for other coniferous forest ecosystems (e.g., 90–700 mm – Law et al. 2002, Meiresone et al. 2003, Delzon & Loustau 2005, Vincke & Thiry 2008, Gielen et al. 2010, Ilvensiemi et al. 2010, Schlesinger & Bernhardt 2013). Obviously, such low total growing season T_{tot} values as obtained at our study site are restricted to extreme cli-

matic and/or soil conditions (Moore et al. 2000, Poyatos et al. 2008), and a sparse understory. In addition, the low LAI of 0.55 and the low canopy coverage of 33% (Oberhuber & Gruber 2010), may have also contributed to the low T_c (Tab. 1) and hence also to the low T_{tot} , although maximum whole tree water use rates obtained for *P. sylvestris* at our study site ($25.4\text{--}8.1 \text{ kg day}^{-1}$ – Wieser et al. 2014) were in good agreement with data reported for other conifer tree species comparable in DBH (18–32 cm – Wullschlegel et al. 1998).

Conclusions

Precipitation measurements, scaled sap flow rates, estimates of ET_u and I by the soil water budget method (Kucerova et al. 2010), and derived from the literature (Brechtel 1965), respectively made it possible to investigate the water balance of an inner Alpine *P. sylvestris* forest which had no access to groundwater. Our data suggest that *P. sylvestris* forests in inner Alpine dry valleys are able to cope with high evaporative demand even under limited soil water availability (SWA) in the topsoil. In all the growing seasons investigated the amount of precipitation exceeded the total actual evapotranspiration ($= T_c + ET_u + I$). Although shallow soil drought did not play a major role at our study site, further research has to include estimates of $\delta^{18}\text{O}$ isotope ratios of precipitation, plant tissues (xylem and foliage), and soil water in different soil depths (Sarris et al. 2013), as well as the quantification of rooting depths (Waring & Running 1998). All these assessments should be combined with an evaluation of absorptive root areas in different soil layers by earth impedance methods (Cermak et al. 2013), in order to assess the accessibility of different soil water horizons as sources for transpiration (Leo et al. 2013) of trees and understory vegetation in inner Alpine dry valleys and other drought-prone areas indicated by low LAI, a low canopy coverage and a sparse understory.

List of abbreviations

The following abbreviations have been used throughout the paper:

• $A_{\text{S-stand}}$: cumulative sapwood area per

- ground surface area;
- DBH: diameter at breast height;
- ET: growing season cumulative evapotranspiration;
- ET_U : soil evaporation and understory transpiration;
- I: rainfall interception;
- P: precipitation;
- PET: potential evapotranspiration (Turc equation);
- Q_s : sap flow density;
- GR: global radiation;
- R: total run off = $R_s + R_D$;
- R_D : deep seepage;
- REW: relative extractable water;
- RH: relative humidity;
- R_s : surface run off;
- SWA: soil water availability;
- T: air temperature;
- T_c : canopy transpiration;
- T_{tot} : total forest water use ($T_c + ET_U$);
- VPD: vapour pressure deficit;
- W: shallow soil water content;
- ΔW : change in the soil water content;
- θ : soil moisture.

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