Journal of Hydrology 500 (2013) 75–83

Contents lists available at SciVerse ScienceDirect

Journal of Hydrology

journal homepage: www.elsevier.com/locate/jhydrol

Climate and plant cover co-determine the elevational reduction in evapotranspiration in the Swiss Alps

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ARTICLE INFO

Article history:
Received 26 April 2013
Received in revised form 2 July 2013
Accepted 6 July 2013
Available online 19 July 2013
This manuscript was handled by Konstantine P. Georgakakos, Editor-in-Chief.

Keywords:
Atmometer
Grassland
Hydrology
Lysimeter
Mountain
Penman-Monteith equation

SUMMARY

Atmospheric conditions and plant cover are the two main determinants of evapotranspiration (ET), and both strongly depend on elevation. The objective of this work was to separate the two effects across a 1 km elevation gradient and compare theoretical and experimental approaches in order to improve hydrological models for mountainous regions. We used the Penman–Monteith model (FAO-56 procedure) and atmometers (evaporimeters) in combination with grassland weighable lysimeters (WLs) and deep seepage collectors (DSCs) positioned at 1490, 1960 and 2440 m of elevation in the Swiss Alps to quantify climatic and vegetation effects on ET. Atmometers and the Penman–Monteith model were used to quantify atmospheric evaporative demand and WLs were employed to measure ET during clear days (WLET). DSCs were used to calculate ET rates from the water balance equation over longer time periods (DSCET) during summer in 2010 and 2011. During clear days, Penman–Monteith evapotranspiration (PMET) was reduced by 11% at the high site in comparison to the low site, atmometer evaporation (AE) was reduced by 23%, and daytime WLET rates decreased by 37%. Over longer time spans (close to three months), PMET decreased by an average of 10% (14% and 6% for 2010 and 2011) along the entire elevational gradient, and DSCET dropped by a mean 35% (45% and 25% during 2010 and 2011). The steeper elevational gradient in DSCET in 2010 was probably caused by the early onset of the growing season at low elevation during that year. Reductions in WLET and DSCET with elevation were thus consistently larger than the decline in both AE and PMET. Smaller vascular plant biomass at high elevation causes a reduction in ET that is approximately as large as the reduction in ET by reduced atmospheric forcing. These findings call for a revision of the view that evapotranspiration during the growing season is explained by season length only (assuming no effect of elevation on a daily basis). Solving the water balance by hydrological modelling in high elevation catchments needs to account for elevational changes in plant cover (amount of transpiring biomass, LAI), even within the same type of land cover as shown here for grassland.

1. Introduction

One classical question that has long occupied meteorologists and hydrologists concerns the way evapotranspiration (ET) is related to elevation. Meteorologists have focused on effects of elevational shifts in atmospheric conditions on evaporation rates (e.g. the early works by H.B. de Saussure, see Barry, 1978), while hydrologists employed the water balance equation to calculate ET for catchments from different elevations from precipitation and runoff (e.g. Menzel and Lang, 1998). Yet, ET rates are inherently tied to both climatic conditions and land surface properties, both of which are affected by elevation in different ways. Hydrological models for mountain regions often describe vegetation induced variability in ET by using coarseland cover categories such as forest, grassland, permanent snow, etc., (e.g. Gurtz et al., 1999). Grassland is a major vegetation type in montane and alpine catchments (Körner, 2009; Tappeiner et al., 2008) and within this vegetation type, transpiring leaf mass and vegetation height vary considerably with elevation. This study aims to quantify the separate contributions of vegetation and climatic forcing to growing season ET. In the following, we use the term elevation for land surface properties and the surface boundary layer, and the term altitude for conditions in the free atmosphere (McVicar and Körner, 2012).

Climatic conditions change markedly with elevation. Some of those changes are common altitude-related phenomena such as the decline in atmospheric pressure and air temperature. Other climatic factors such as cloudiness, wind speed, and precipitation show strong regional variation and are not related to altitude in a systematic manner (Barry, 1981; Körner, 2003, 2007). In the European Alps, precipitation increases with increasing altitude (Frei and Schär, 1998). Worldwide however, the effect of altitude

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0022-1694/$ - see front matter © 2013 Published by Elsevier B.V.
http://dx.doi.org/10.1016/j.jhydrol.2013.07.013

on precipitation has been shown to depend on latitude (see Lauscher's review, 1976; Körner, 2007). Atmospheric pressure decreases by about 10 kPa km^{-1} (for altitudes up to 4 km; Burman et al., 1987), which causes air temperatures to decline. Generally, for mid latitude mountain ranges, annual air temperature lapse rates are smaller than summer time temperature lapse rates which average about 6.0 K km^{-1} in the central Alps (Dodson and Marks, 1997; McVicar et al., 2007; Rolland, 2003; Tabony, 1985).

From the Chapman–Enskog equation it can be deduced that the diffusion coefficient of water in air is inversely proportional to atmospheric pressure, while at the same time, it shows a positive power relationship with air temperature (Reid et al., 1987). This means that a reduction in air pressure leads to an increase in the diffusivity of water vapour in air (and thus enhances transpiration), while a reduction in air temperature alone has the opposite (but smaller) effect. As a result of declining air temperature, the vapour pressure deficit of the atmosphere decreases with altitude (McVicar et al., 2007). This lower vapour pressure deficit (VPD) should reduce evaporation rates. Under clear sky conditions, net solar radiation (the balance between incoming and outgoing solar radiation), has been reported to be independent of elevation during summer (Marty et al., 2002).

Grassland is affected by elevation in a number of ways. While leaf area per unit of land area (LAI) and leaf biomass decrease with elevation, standing necromass often increases (Cernusca and Seeber, 1981). Although these increasing amounts of dead plant material do not transpire, they increase interception of rainfall and add to the aerodynamic resistance of the canopy (Cernusca and Seeber, 1981). Further, leaf to air temperature differences are larger at high elevations compared to low elevations because low stature vegetation is less coupled to the atmosphere (Cernusca and Seeber, 1981; Körner and Cochrane, 1983). Higher surface to atmosphere temperature differences result in steeper leaf to air water vapour concentration gradients and can be expected to enhance transpiration (Smith and Geller, 1979). Stomatal density and leaf diffusive conductance increase with elevation (Körner and Mayr, 1981; Körner, 2003) and can potentially compensate for the effect of reduced LAI at high elevation. However, the net effect of higher stomatal conductance at high elevation cannot be fully resolved since stomatal, aerodynamic and topography induced resistances to vapour diffusion all act in series. Owing to these complex interactions between atmospheric conditions and vegetation, it is not *a priori* clear how evapotranspiration rates are affected by elevation.

A decrease in atmospheric pressure has been shown to lead to higher rates of pan evaporation (Rohwer, 1931). Decreases in air temperature and VPD however, more than offset this effect and cause evaporation rates from a free water surface to decline with elevation (Blaney, 1958; Longacre and Blaney, 1962 and references therein; Shaw, 1909; Young, 1947). This is supported by theoretical studies (see Gale, 1972; Smith and Geller, 1979). Evaporation may however, increase with elevation due to temperature inversion (e.g. Nullet and Juvik, 1994). Finally, exposure through effects on intercepted solar radiation and on wind speed exerts additional effects (Peck and Pfankuch, 1963; Peck, 1967). At annual time scales, ET rates decrease with elevation (Lang, 1981) because the length of the snow free period decreases with elevation in high latitude mountains (Körner et al., 1989). Daily ET rates on rainless days have been reported to be independent of elevation (Körner et al., 1989; Körner, 2003; Wieser et al., 2008). Three reasons have been put forward to explain this: (1) steeper water vapour pressure gradients because of relatively higher surface temperatures, (2) enhanced contributions of evaporation from the soil surface at high elevation because of larger fractions of open ground (Kelliher et al., 1995), and (3) stomatal down-regulation of transpiration during periods of reduced moisture availability related to overall lower precipitation to ET ratios at low elevation, have been considered (Körner and Mayr, 1981; Körner, 2003).

Climatic effects on elevational changes in ET are well represented in models (e.g. McVicar et al., 2007). However, the influence of elevational changes in plant canopy characteristics on ET has not received the same attention. Here, we aim at quantifying the separate influences of atmospheric conditions and vegetation on ET rates along an elevational gradient. For this purpose, we employed atmometers (often referred to as evaporimeters), a physics based probe and two types of grassland lysimeters (defined containers with vegetated soil monoliths) positioned along a 1 km elevational gradient. We present ET rates for rainless days from weighable lysimeters (WLs) and for longer time periods (identical periods at all sites) from deep seepage collectors (DSCs). At both daily time scales and across longer time spans (close to 3 months during the growing season), we expect atmospheric evaporative demand to decrease with elevation as air temperature and VPD decline. However, for the reasons discussed above, we expect grassland evapotranspiration rates to be unaffected by elevation when equal sampling periods are compared (days or months). To test these hypotheses, we selected grassland types that are typically found along elevational gradients (from the montane to the alpine belt) in the Swiss Alps.

2. Material and methods

In order to determine the relative influences of vegetation structure and atmospheric conditions to evapotranspiration rates in different grassland types along an elevational gradient, we set up an experiment at three sites, all approximately 500 m of elevation apart. The three sites are all located in the Ursern valley in the Swiss central Alps on more or less level terrain. The low site (L) at 1490 m a.s.l. (46°36′N, 8°32′E) had a loamy sand soil and a mean plant canopy height between 15 and 20 cm (always disregarding the height of inflorescences). The mid elevation site (M) at 1960 m a.s.l. (46°35′N, 8°29′E), had a partly podsolized brown earth soil and a mean canopy height of approximately 15–20 cm as well. The high elevation site (H) at 2440 m a.s.l. (46°34′N, 8°25′E), had a shallower, partly podsolized alpine brown earth soil and mean vegetation height varied between 5 and 10 cm. Currently, the potential climatic treeline in the Ursern valley lies between 2000 and 2100 m of elevation.

Each site was equipped with a weather station (Vantage Pro 2, Davis Instruments Corporation, Hayward, CA, USA) including a tipping bucket rain gauge, a cup anemometer, shielded air temperature and humidity sensors and a solar radiation sensor. Sensors were mounted at 1.5–1.8 m above the ground and data were recorded at 10-min intervals. For all water balance calculations (DSCs), we used data from an additional rain gauge (RAINEW, Rainwise Inc., Barharbor, ME, USA) installed at 0.4 m above ground in the vicinity of DSCs, because rainfall measurements at greater height might underestimate true precipitation, especially under windy conditions (Duchon and Essenberg, 2001). We calculated Penman–Monteith evapotranspiration (PMET) assuming a similar short stature 'crop' for all sites. The PMET calculations were parameterized with daily mean climatic values for the comparison with long-term DSC readings, and with hourly means for daytime-only, when we compare PMET with ET rates of weighable lysimeters. Air pressure was calculated from elevation using Eq. (1) (Allen et al., 1998). Daytime-only and daily (24 h) PMET rates were calculated following Allen et al. (1998):

$$P = 101.3\left(\frac{293 - 0.0065z}{293}\right)^{5.26} \tag{1}$$