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Spatial extrapolation of agrometeorological variables

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Abstract

Agrometeorological variables are especially subject to variations in space – a fact that makes calculations of areal mean values for e.g. evaporation extremely elusive. Wind, temperature, and humidity cannot be measured continuously in space. In this paper we present simple methods to determine values of those variables for non-ideal positions from measurements at a single ideal point and test these spatial functions by measurements at three non-ideal points in a catchment. For wind, we were able to obtain a relatively good fit by taking the sheltering effects of the catchment rims into account. Temperature was not regionalizable with simple approaches, but needs to be regionalized with spatially and temporally differentiated models. Specific humidity can be assumed to be homogeneously distributed even though there were sinks and sources for water vapor in the study area.

Keywords: Temperature; Wind speed; Absolute humidity; Shelter factor; Spatial scale; Regionalization

1. Introduction

The microclimates at most sites of a region are influenced in some way by the landscape elements that form their respective surroundings. These 'non-ideal sites' are not usually assumed to be suitable for meteorological measurements for at least two reasons. Firstly, they are not representative, i.e. it is assumed that comparable data cannot be obtained at different distinct non-ideal sites (Liljequist and Cehak, 1990). Secondly, advective effects at non-ideal sites are strongly dependent on the respective wind direction due to fetch effects, so that the micrometeorological gradient and eddy correlation methods are not conclusively applicable (Bernhofer, 1992).

Due to these difficulties, measurements from meteorologically ideal exposures are often taken as representative for larger areas. If records of more than one meteorological station are available, statistical interpolation routines are used, such as Principal Component Analysis (e.g. Boyer and Feldhake, 1994) or geostatistical methods (Söderström and Magnusson, 1995; Bland and Clayton, 1994). Geiger (1961) reasons that a deterministic approach to regionalize microclimate is to identify 'microclimate laws' to extrapolate meteorological variables at a certain point from records of a nearby meteorological station. The differences between different sites are a result of many complex, interacting processes that are caused by orography, vegetation, buildings, or open water

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surfaces. Geiger's microclimate laws are meant to explicitly account for each of these alterating factors.

The purpose of this study was to determine spatial extrapolation functions in order to describe spatial modifications of micrometeorological variables based only on the orography of the catchment under investigation. The rationale was, that areal modelers would like to have continuous surfaces for meteorological variables rather than single point measurements or discrete fields of e.g. wind vectors, but on the other hand, often have only sparse information about the factors that influence the microclimate. The only information that is always given when an area is modeled, e.g. with an geographic information system (GIS), is the relief, mostly represented by a digital elevation model (DEM). The literature provides some ideas for continuous functions that could serve for a spatial extrapolation of single ideal-point measurements. It is important to emphasize at this point, that these functions in order to be applicable must not contain fitted parameters, or they would require intense meteorological measurements to be transferable to different sites even within the same area. The functions we used are described in Section 3 of this paper. For the purpose of testing these functions, micrometeorological variables at four different locations within a small agricultural catchment were measured. The estimation by the homogeneous assumption, i.e. the microclimate at every point is the same as in a single ideal measurement point, was compared to the estimation by the regionalization function. In Section 4 of this paper, the results for wind speed, air temperature, and air humidity are presented, which are among the driving forces of energy and matter transfer between soil/vegetation and the atmosphere. Section 5 gives an overall evaluation and outlook.

2. Measurements

2.1. Case study area

The Weiherbach area is a small agricultural watershed of 6.3 km^2. It is located in the Southwest of Germany in Kraichgau in the triangle Karlsruhe, Heilbronn and Heidelberg. Predominant crops are grain, corn, sunflower and sugar beet. Due to the loess substrate, loam is the dominant soil texture class. Altitudes range from 125 to 250 m above sea level. Fig. 1 shows a digital elevation model of the catchment. The mean annual rainfall in the region is 750 mm while the mean temperature is 10°C.

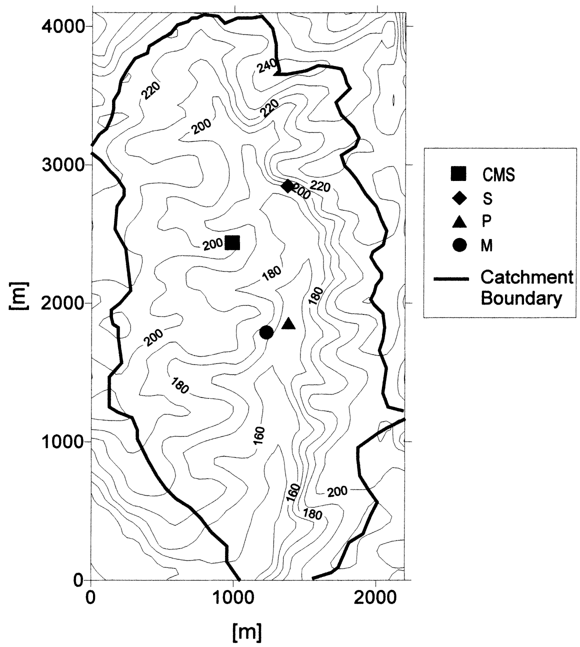

Fig. 1. Digital elevation model of the Weiherbach catchment. CMS: Central Meteorological Station; mobile Stations: S, P, and M.

2.2. Measurement sites

Reference measurements were taken at a central meteorological station (CMS, Fig. 1, altitude 204 m) within the catchment. Additional measurements were carried out at three non-ideal sites with mobile stations (S, M, P, Fig. 1). The station sites were chosen so that small landscape elements were unlikely to exhibit much influence on the microclimate at these. Only the orography of the valley, that can be represented in a digital elevation model (DEM), was to modify the variables in question. One exception to this rule was a small stand of trees that formed the horizon looking south from Station P (180–225°, Fig. 2).

The distribution of the stations within the catchment was to reflect as wide a range of different elevations, slopes and aspects as possible. Station P was close to

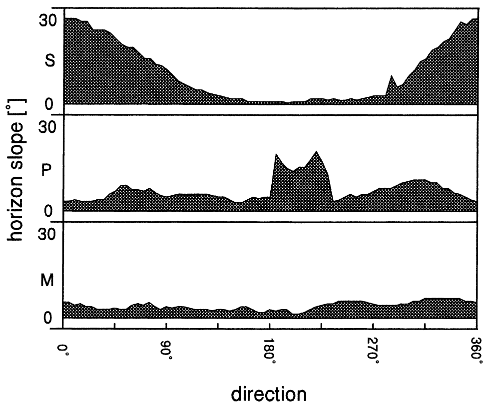

Fig. 2. Measured slope of the horizon upwind at the non-ideal sites. The CMS has a constant horizon line of 0° elevation.

the bottom of the valley and Weiherbach creek at an altitude of 162 m. Station M was located about 8 m higher in altitude on the western catchment slope. Both stations were close to wheat fields but the ground around the stations was covered with grass. At an altitude of 186 m, Station S was positioned on a south-looking slope that was used as a sheep pasture. With an inclinometer the slope and the horizon line were measured at every station (Fig. 2). For the calculation of the horizon slope upwind (see Section 3) moving averages over 25° segments of the horizon line were used.

The number of measurement stations was subject to four constraints: equipment was limited and so was the data handling capacity. There were also only few places in the catchment not evidently influenced by any small-scale disturbances and owned by farmers willing to cooperate. The single most important water source in the catchment is the Weiherbach brook which runs close to Station P in a deep trench with high riparian vegetation.

2.3. Instruments

Wind speeds were measured with cup anemometers, wind directions with wind vanes, at P, M, S, and CMS (all Vector Instruments). Air temperature measurements were taken with shaded thermistors (Fenwal) at the mobile Stations P and M, and shaded Pt100 thermoelements (Thies/Heraeus) at the CMS. Relative air humidity was measured with capacitive sensors (Skye Instruments) at P and M, and psychrometers (assembled at the Technical University of Karlsruhe) at CMS. The specifications of the instruments can be seen in Table 1. The instruments at CMS and the mobile stations were compared by control measurements with a capacitive sensor at CMS which showed that the deviations between the instrument types were not greater than between sensors of the same type. The data were recorded as 10 min averages

Table 1
Specifications of the instruments

Variable	Station	Screen height (m)	Instrument	Stall speeds[a]	Range[a]	Accuracy[a]
Wind speed	CMS, S, M, P	2	Cup anemometer	0.25 m s^{-1}	0–75 m s^{-1}	1% ±0.1 m s^{-1}
Wind direction	CMS	8	Wind vane	0.6 m s^{-1}	0–358.2°	±2° ± 0.2°
	S, M, P	2				
Air temperature	CMS	2	Pt 100			0.05 K
	M, P	2	Thermistor		min −30°C max +70°C	0.2 K (0 to 60°C)
Air humidity	CMS	2	Psychrometer (with Pt 100)			<1%
	M, P	2	SKH 2011 (capacitive)		0–100%	<2% >96% : 4%

[a]As specified by manufacturer.

from 21 March 1996 to 1 August 1996 with data loggers (Delta-T).

3. Preliminary theoretical remarks

The results of this study may be used to regionalize driving variables of a spatially distributed hydrological catchment model. Therefore, the mathematical extrapolation schemes have to meet certain requirements: they must have a form of transfer or correction functions based on the recording of the CMS. The description of the characteristics of the respective non-ideal site may contain only the information of the DEM of the catchment. Vegetation and soil related data are not to be used. The functions should not depend on the specific placement of our measurement or contain fitted parameters, but rather be widely applicable and easily transferable to other sites.

3.1. Wind speed

The wind field in an orographically uneven terrain is the result of the superposition of influences on different scales (Ryan, 1977). The modifications of the free wind speed by orography consists of weather side and lee effects (i.e. sheltering, diverting, creation of turbulence) and the velocity affecting consequences of the diversion. In this study, we assumed the sheltering effect of the Weiherbach catchment rim to be the only relevant modification of free wind speed. Local wind systems, such as sea breeze or cold air drainage, are the result of thermal inhomogeneities and are a separate source of momentum that is subject to the same modifications as the free wind (Stull, 1993; Ryan, 1977). Local wind systems have characteristic directions determined by the geographic conditions (Oke, 1978). Under certain circumstances they can contribute considerable wind speeds. Clements et al. (1989) reported low-level jets of 5–6 m s^{-1} in Brush Creek Valley at 80 m above the ground, depending on the atmospheric stability. However, the geography of the Weiherbach valley does not favor the genesis of strong endogenous wind systems. Since the slopes of the Weiherbach catchment are not long, cold air drainage probably does not result in high flow speeds.

The sheltering effect of elevated horizons upwind was considered to predominantly affect the wind speed. Ryan (1977) developed a trigonometric function for the relationship between wind speed in a non-ideal position, free wind speed and the slope of the horizon upwind by a correction factor:

$$v_n = F_R v_f \tag{1}$$

with:

$$F_R = 1 - 0.01 \arctan(0.17\,Y) \tag{2}$$

where v_n is equal to wind speed of non-ideal sites (m s^{-1}), v_f equal to free wind speed (m s^{-1}), and Y is equal to the slope of the horizon upwind (%). This formula is of general applicability and can be used on different scales. It has been used for the calculation of orographic sheltering on different scales: Ryan (1977) developed it based on van Eimern's (1955) study of the influence of the Harburg Mountains and Geest Ranges of Germany on wind flow, and on Kaiser's (1959) analysis of wind speed reduction by wind breaks. He then has applied them to the sheltering effect of the Rocky Mountains close to San Bernardino. The scale of our catchment is between the features that served Ryan for the original derivation. Bergold (1993) tested the equation for nine pairs of mobile and basis stations for greater distances between ideal and non-ideal stations as we and reports showed considerable improvements compared to the homogeneous assumption in eight out of the nine cases.

3.2. Air temperature

Air temperature varies systematically with the height above sea level following the adiabatic gradient. Other influencing factors are surface temperatures and atmospheric exchange. Surface temperature is a function of exposure to the sun, of the physical properties of the effective surface, the water supply of the vegetation and again the actual surface-atmosphere exchange characteristics. Cloud cover plays a major role on sunshine duration and intensity and nightly outflux of longwave radiation, and thereby cold air drainage. While sunshine duration during the day might be assumed to equally affect every point in the catchment, the nightly cold air drainage leads to lower temperatures in valleys than in higher regions. This can lead to continuing inversion in the early hours of the day when air mass exchange is still inhibited by a stable atmosphere. Thus, atmospheric stability has a major affect on the temperatures within a landscape unit like a small basin (Brockhaus, 1995). These complex interdependencies cannot be simulated only on the basis of a DEM.

For the Weiherbach catchment a well mixed layer of air throughout the catchment is a reasonable assumption for daytimes after the morning inversion has been overcome. Then, one may assume the mean differences in air temperature to be a mere function of the height differences. The correction function is a simple linear equation:

$$T_{\mathrm{X}} - T_{\mathrm{CMS}} = f(z_{\mathrm{CMS}} - z_{\mathrm{X}}) \tag{3}$$

with T_{X} is equal to temperature at the respective point X in the catchment (°C), T_{CMS} is equal to temperature at the CMS (°C), z_{CMS} equal to height (a.s.l.) of the CMS (m), z_{X} = height (a.s.l.) of site X (m) and f a meaningful lapse rate (°C m^{-1}). The lapse rate has to be an interpretable combination of the dry and moist adiabatic lapse rates (0.01 and 0.006°C m^{-1}) and the stable cases (<0.006°C m^{-1}) that can occur on cloudy days without large temperature differences. A mean lapse rate of the size of 0.006°C m^{-1} would result in a mean temperature difference of 0.3°C between our highest and lowest stations.

For the times of the day, when the atmosphere is well mixed, an 'empirical lapse rate' can be determined by fitting a linear regression between the measurements of the two stations of an intermediate temporal resolution, e.g. hourly means. If the slope of the regression equation is equal to 1, i.e. the oscillations of the temperatures at the two sites are the same, dividing the intercept by the height difference gives a mean empirical lapse rate. The value can then be compared to theoretical lapse rates and conclusions may be drawn in order to find rules for the transfer of the lapse rates to other places. If the slope of this regression does not equal 1, or the empirical lapse rate does not give clues about transfer rules, however, no simple regionalization functions on the basis of a lapse rate can be formulated, and the need for more complex models is unavoidable.

3.3. Air humidity

The humidity of the air is measured as relative humidity. Many evaporation equations make use of the relative humidity as an expression of the degree of saturation of the air. But since relative humidity is strongly dependent on air temperature, it is similarly complex in its spatial variation as air temperature, given that there is no systematic spatial variation in the absolute water vapor content of the air. Such systematic variations can be caused by different sources and sinks for water vapor, e.g. open water surfaces like creeks, brooks or ponds, or significantly higher dewfall caused by certain vegetation covers.

If a catchment's atmosphere is well mixed, with no strong vapor sources around, and the vegetation similar and in similar growth stages, one may assume the water vapor to be homogeneously distributed over the whole catchment, i.e. the specific humidity is the same everywhere in the area. We can calculate the specific humidity at the different stations from measurements of relative humidity and air temperature by using the following approximations (Liljequist and Cehak, 1990):

$$q = \frac{0.622\, e_{\mathrm{a}}}{p} \tag{4}$$

with

$$e_{\mathrm{a}} = \frac{h\, e^*(T)}{100} \tag{5}$$

$$e^*(T) = 611 \exp\left(\frac{17.08\, T}{234.18 + T}\right) \tag{6}$$

where h is equal to relative humidity (%), e_{a} equal to actual water vapor pressure (Pa), q is equal to specific humidity, $e^*(T)$ equal to saturated water vapor pressure (Pa), T equal air temperature (°C), and p is equal to barometric pressure at the point in question (Pa). The specific humidity of the ideal station (CMS) is then used as an estimation for the specific humidity at every point in the catchment and compared to the measured/derived specific humidities at the mobile stations P and M.

4. Results and discussion

4.1. Wind speed

4.1.1. Results

At the three non-ideal measurement sites, wind speeds are significantly reduced compared to the

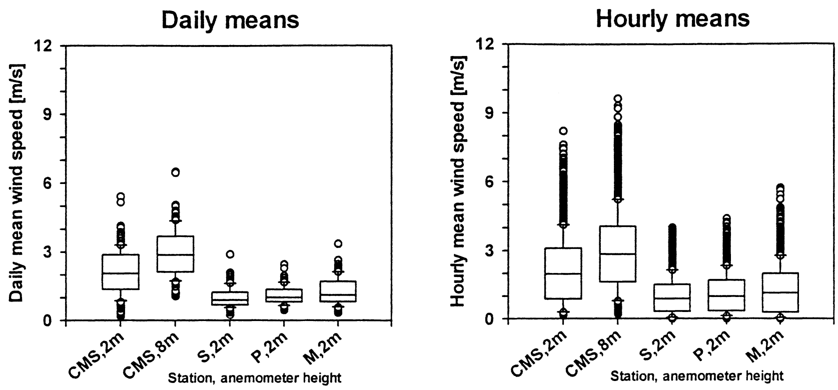

Fig. 3. Box-plots for the mean daily (left) and hourly (right) wind speed data at the CMS (2 m and 8 m screen heights), and the sites S, P, and M. The boxes contain the mean 50% of the values, the lines indicate 10% and 90% percentiles.

CMS for both, hourly as well as daily mean values (Fig. 3). Using the measured values from the CMS as estimates for the wind speed at non-ideal situations will therefore result in systematic overestimation as is evident when comparing the mean deviations with the mean absolute differences in Table 2.

An analysis of the thermally induced wind systems made use of the fact that thermal wind systems have typical wind directions, i.e. the direction of the slope or of the whole valley. Thus, we verified the hypothesis that thermal wind systems do not occur frequently. They also never result in high flow velocities. It could be shown that even at the south-facing slope of Station S thermal circulation was rarely stable over more than 10 min in many cases since the wind directions kept changing.

Table 2
Wind speed estimates for the mobile meteorological stations

E = estimated by		Hourly mean values		Daily mean values	
		Data of CMS	Ryan	Data of CMS	Ryan
Station S					
n		2316	2289	96	96
MD	(m s^{-1})	1.43	−0.049	1.43	−0.056
MSD	(m^2 s^{-2})	3.05	0.36	2.47	0.19
MAD	(m s^{-1})	1.44	0.47	1.43	0.39
Station P					
n		1589	1580	64	63
MD	(m s^{-1})	1.34	−0.29	1.34	-0.28
MSD	(m^2 s^{-2})	2.59	0.33	2.14	0.12
MAD	(m s^{-1})	1.35	0.45	1.34	0.30
Station M					
n		1581	1565	65	65
MD	(m s^{-1})	1.06	−0.17	1.05	−0.16
MSD	(m^2 s^{-2})	1.63	0.29	1.27	0.10
MAD	(m s^{-1})	1.07	0.42	1.05	0.21

x: represents measured values at the mobile stations, e: estimated data using two different approaches (data of CMS and the Ryan approach), n: is the number of available data. The mean deviations (MD) were calculated by $\Sigma(e-x)/n$, the mean squared deviations (MSD) as $\Sigma(e-x)^2/n$, and the mean absolute differences (MAD) as $\Sigma|e-x|/n$.

Wind speeds at the measurement sites can also be estimated by the Ryan approach (Eqs. (1) and (2)) using the observations of the CMS. Table 2 uses three measures for the quality of fit: mean deviations, mean squared deviations and mean absolute differences. The smaller the values are, the better the fit is. A comparison of the mean deviation with the mean absolute deviation assesses as to what degree a neutralization of over- and underestimations occurs, i.e. to what degree the over- or underestimation is a systematic error. A comparison of the mean absolute deviation and the mean squared deviation gives clues about outliers. It is shown in Table 2, that the extrapolation with the Ryan approach results in much lower errors than to use the data of the CMS as an estimate for the wind speeds at sheltered sites. All measures for the quality of fit of the estimates are highly improved by replacing the uncorrected value of CMS by estimates using the Ryan approach (Eqs. (1) and (2)).

4.1.2. Discussion

The significant decrease in wind speed may be caused by the lee effect of the catchment rim. However, turbulence created by the rim (lee eddies) or other landscape elements does not explain the measured decrease of the wind speed, since cup anemometers are not sensitive enough towards turbulence but tend to overestimate the mean speed of the real flow due to their inertia.

The estimation of the wind speed in the sheltered valley by Ryan's equation results in values of the same order of magnitude as the measurements. In Fig. 4, estimates and measurements of mean daily wind speed for Stations P and S are shown for part of April 1996. During this period there is a systematic under-estimate using the Ryan formula, however over the whole comparison period there is a small over-estimation of wind speed. At low wind speeds, measurements had to be excluded occasionally from the vectorial averaging procedure due to the low sensitivity of the wind vanes. Thus, the averaged measurements overestimate the real mean winds, and the underestimation might be smaller.

For P and M, we were able to achieve very good fits with the Ryan approach (cf. Table 2). For S, due to the different characters of the southern and northern horizon line, a more detailed discussion of the sheltering factor's influence is possible. Intuitively, one would assume that the distance from a sheltering landscape feature is more important than the horizon slope that is formed by it. However, slope represents a combination of size and distance of the features. The horizon line looking south from Station S is formed by the northern end of the Black Forest. Even though this mountain range is comparatively high, its shelter factor is very low, since it is far away. The best estimates in our application are obtained for landscape features like this. The steep slope on which S is located, however, results in high sheltering factors for northerly winds. A reduction in measurement and calculation accuracy occurred during the averaging process due to overall low wind speeds in this sheltered position ($\leq$1.6 m s^{-1}). More important however, divergence, channeling, and especially turbulence, have a stronger

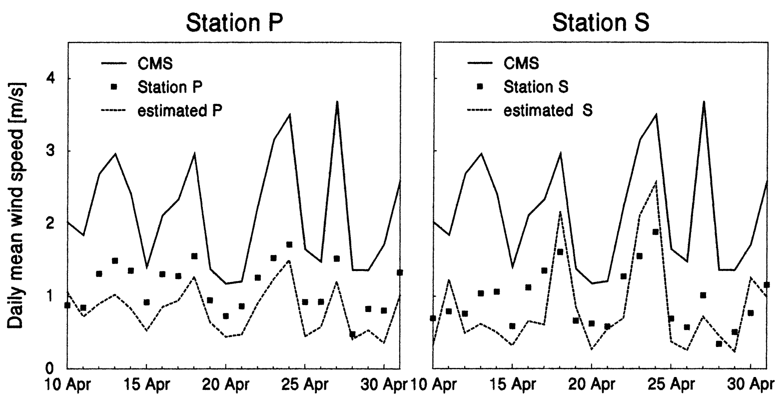

Fig. 4. Mean daily wind speeds and Ryan estimates at the Stations P (left) and S (right) in April 1996. Station M shows similar behavior as Station P.

impact at this site than in less extreme situations which led to a worse fit compared to Stations M and P (Table 2). Nevertheless, Ryan's equation gives a good approximation of the non-ideal wind situation at all sites.

4.2. Air temperature

4.2.1. Results

During cloudy days, temperatures were very similar at all stations. The typical sunny day showed only small or no differences between the stations during daytime, but considerably lower temperatures at the valley bottom during the night. The daily amplitudes of the temperatures were largest at Station P. Mean daily temperatures (24 h) at P and M were closely related with the CMS (R^2 of 0.994 and 0.998), although they are usually highest at the CMS.

Since the atmosphere is expected to be best mixed in the late morning and the afternoon, regressions (Eq. (3)) were calculated for temperatures between 10 and 6 p.m. The results for M were $T_M = 0.289°\text{C} + 0.983\,T_{CMS}$ (714 degrees of freedom, and $R^2 = 0.996$) and for P, $T_P = 0.279°\text{C} + 0.982\,T_{CMS}$ (with 662 degrees of freedom, and $R^2 = 0.996$). Although these regression equations show good fits, the slopes of the two regressions are significantly different from 1. If this was not the case, the empirical lapse rates (Eq. (3)) could be determined from the constants of these regressions as 0.0085 K m^{-1} for M and 0.0066 K m^{-1} for P.

4.2.2. Discussion

The low temperatures in the 24 h means at the valley Stations M and P are due to an accumulation of cold air in the deeper regions of the valley. During the day, similar temperatures are measured at every station, while night cooling is much stronger in all valley situations. Even though the thermally driven wind systems do not play a role for the regionalization of wind speeds due to their low velocity and stability, they seem to be efficient advection mechanisms for cold air, thereby causing complex spatial variations in air temperatures. The influence of cold air drainage on daily mean temperatures is overriding the lapse rate-following a lapse rate, the differences between ideally situated measurement sites should have the opposite signs. Due to this oscillation of inverted and 'normal' situations, the regionalization of mean temperatures requires a model that calculates the mean temperatures at a site in question from different regionalization functions for different times of the day. Some of these will include the complexity of the valley's geometry and memory effects of the atmosphere. For e.g. a model for cold air drainage would have to account for the radiative behavior of the heterogeneous surface as well as for the constellation of slopes and the 'history of the atmosphere', i.e. cloudiness and radiation of the day before the one in question.

The results of the regression for the daytime only temperatures (10 a.m. to 6 p.m.) show that a proper meaningful lapse rate for regionalization purposes could not be found. The calculated values for the empirical lapse rates seem to lie in a physically justifiable range. One may speculate that the lapse rate for the P-CMS couple is lower than the one for the M-CMS couple because of the water vapor source of the Weiherbach brook near P, but it may as well be due to other factors (cf. Section 4.3.). However, since the slopes are significantly different from 1 there are systematic differences in the thermal behavior of the sites. Thus, the estimated values are not reliable enough to support a physical interpretation or serve for the derivation of parameters in a simulation model. Correcting the temperatures by a constant value would instead worsen the temperature estimate compared to the homogeneous assumption since the slopes of the regressions do not equal 1. The coefficients summarize our measurements, but do not allow any further conclusions. So the best 'simple' three estimates that can be given for temperature with our approach is the homogeneous assumption.

4.3. Air humidity

4.3.1. Results

Relative humidity (as measured) was lowest at the CMS during all phases of the measurement period. Considerably, higher values were recorded at the Stations M and P. In Fig. 5, the direct comparison between the hourly means at the CMS and the mobile stations is displayed. The covariation of relative humidity and temperature is very high.

As described above, the relative humidity records of the meteorological stations were converted into specific humidities (Eqs. (4),(5) and (6)) and compared

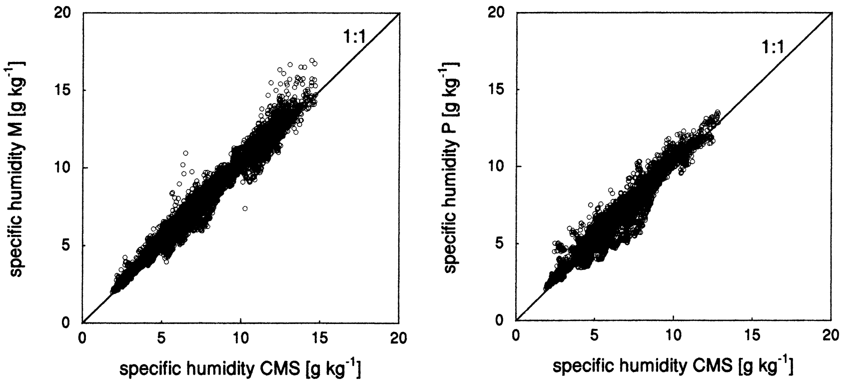

Fig. 5. Scatterplots of the mean specific humidities from hourly measurements measured at the mobile Stations M (left) and P (right) versus measurements at the CMS (left column) and the approximations (right column) for the hourly means of relative humidity at the sites P and M.

between the stations as illustrated in Fig. 5. The highest differences between the stations (-2.59 and $2.20\,\mathrm{g\,kg^{-1}}$ for CMS–P, -1.79 and $2.88\,\mathrm{g\,kg^{-1}}$ for CMS–M) are not large compared with the differences between the two measurement techniques, that are -1.30 and $0.64\,\mathrm{g\,kg^{-1}}$ for the two different sensors at CMS (not shown). The results for the regressions between the specific humidities at CMS and sites P and M, respectively are: $q_{\mathrm{P}} = 0.21\,\mathrm{g\,kg^{-1}} + 1.007 q_{\mathrm{CMS}}$ (with 1681 degrees of freedom, $R^2 = 0.954$) and $q_{\mathrm{M}} = 0.075\,\mathrm{g\,kg^{-1}} + 1.024 q_{\mathrm{CMS}}$ (with 1807 degrees of freedom, and R^2=0.978) where q_{P}, q_{M} and q_{CMS} denote the specific humidities at stations P, M, and CMS, respectively. The slopes are mostly significantly different from 1, but the deviation from 1 is smaller than 3%. The intercepts are very small, even though they are significantly different from 0.

5. Discussion

The regression results for specific humidity show that the homogeneous assumption gives reasonable results with extremely low effort. That the intercepts are bigger for P than for M might be due to the Weiherbach creek's influence as a water vapor source. However, both intercepts are very small, i.e. smaller than the measurement and calculation errors. Since the creek is the strongest source for water vapor in the whole area, it is justified to generalize the approach for the area.

In contrast to the temperature approach (Section 4.2), the method of the homogeneous specific humidity is not dependent on times of the day. A distributed catchment model can simulate small local dynamics of condensation and re-evaporation and their diurnal variation in dependence on the closeness to water vapor source. This would have to be combined with a local temperature model, where the dewfall temperature is playing a crucial role. At certain times of the day, specific humidity would be supposed to be equal in the area. Further research remains to be done in this field of locally distributed agrometeorological modeling (cf. Bergold, 1993) on high temporal resolution. For daily means, however, the presented method is a good approximation.

6. Conclusions

In this study, simple functions for the extrapolation of the agrometeorological variables wind speed, air temperature, and air humidity were tested by using measurements of mobile meteorological stations at different non-ideal sites and data from an ideally exposed, standard central measurement station.

For wind speed, the sheltering function of Ryan (1977) resulted in good estimates for the wind speeds in all our non-ideal sites. The estimations are best on a time scale of about 1 h, but still are good for time scales of up to one day. Even for very large and close obstacles like the slope on which one of the stations was installed, and which may produce additional wind speed affecting phenomena, the magnitude of the wind speeds could be estimated with reasonable quality by only accounting for sheltering. The necessary orographic information that is needed to calculate the estimate is easily obtainable either by on-site measurements or by calculation from a DEM. Thus, the proposed regionalization scheme is a quick and powerful instrument to calculate continuous wind fields close to the ground and can easily be combined with a GIS.

For air temperature, the naïve assumption of a constant lapse rate could not be justified. Cold air advection processes were responsible for an inverted temperature profile of 24 h mean temperatures. In order to capture the fast changing factors, at least an hourly resolution and a combination of different process models is necessary. Many of these processes (e.g. cold air drainage in complex terrain) also require a two or three dimensional, spatially explicit calculation (e.g. Kondo and Okusa, 1990). An analysis of daytime-only temperatures (10–6 p.m.) lead to mean empirical lapse rates of sizes between the moist and dry adiabatic gradients. However, no generalization scheme could be found, i.e. we could not discover any rules for a transfer of our results to other sites in the basin.

The results further suggests that specific humidity can be assumed to be homogeneously distributed in catchments similar to that under investigation having no extraordinarily strong sources or sinks for water vapor. Under this assumption, very good estimates for relative humidity or saturation pressure deficit could be calculated for distant places if reliable temperature measurements were obtainable. Future research is suggested towards a spatiotemporal regionalization approach that accounts for the covariation in variables (e.g. humidity and temperature or wind and heat advection) and memory effects (first- or higher-order integration of the processes).

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