

EVAPORATION IN CONCEPTUAL RAINFALL-RUNOFF MODELS
TESTING MODEL REALISM USING REMOTELY SENSED EVAPORATION

MSC THESIS

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ABSTRACT

The procedure to determine evaporation in hydrological models is considered to be unsatisfactory by some researchers; ‘too’ accurate by others. In this procedure catchment scale evaporation is related to some form of potential evaporation, determined with point scale meteorological data. The main criticism is that the potential evaporation is not representative for the catchment and that spatio-temporal dynamics in vegetation cannot appropriately be expressed with the time-invariant, spatially lumped model parameters in the above mentioned procedure.

Using remotely sensed observations, catchment scale estimates on evaporation and vegetation dynamics can be derived. It is hypothesized that by integrating remotely sensed evaporation estimates and additional information on vegetation dynamics in conceptual rainfall-runoff models, we can get more insight into the realism of the modelled evaporation flux and the role of vegetation dynamics. With a more realistic representation of evaporation, the water partitioning can be modelled more accurately, eventually improving our understanding of the catchment behaviour. The way the evaporation estimates can be used depends on the spatial and temporal resolution and the reliability of the products.

The hypothesis is tested in the well studied Ourthe catchment, located in Belgium. The climate is Atlantic temperate, the streamflow is characterized by a quick response.

Vegetation dynamics in both space and time are investigated in a principal component analysis on multi year MODIS Normalized Difference Vegetation Index (NDVI) data. Areas with similar temporal dynamics are distinguished. The most important temporal dynamics are related to phenology and agricultural growing seasons. Areas with an increasing trend in NDVI are identified as well, but the spatial extent is too small to be relevant for hydrological applications.

In a validation study three remotely sensed evaporation products are examined in terms of their reliability and applicability in conceptual models, namely EARS (daily, 4km x 9km), WACMOS (daily, clear days, 1km x 1km) and MOD16 (8-daily, 1km x 1km). EARS and WACMOS are surface energy balance models, based on land surface temperature observations. MOD16 uses the Penman-Monteith equation, with distributed surface characteristics and a Jarvis-like approach to calculate the surface resistance. The remotely sensed products are validated with ground measurements of evaporation from five eddy covariance towers and - if applicable - with a multi year water balance analysis. Mainly based on the water balance analysis, we concluded that the EARS product gives relatively accurate evaporation estimates and can be used in the last step of the research. WACMOS (SEBS) was shown to have an extremely poor correlation between the remotely sensed evaporative fraction and the evaporative fraction determined from eddy covariance measurements and was not further considered. MOD16, often criticized for not taking into account soil moisture constraints on evaporation, did deviate from EC measurements in the growing season, but this was shown not to be related to limited soil moisture. The main drawback of MOD16 in relatively moist climates such as the Ourthe catchment, seems to be the application of constant canopy conductance and the large scale of the meteorological input data. Catchment scale evaporation appeared to be slightly underestimated, which may be attributed to the fact that interception is neglected in this version of the product.

In the last step of the research, i.e. confronting a water balance model with the ancillary data on vegetation dynamics and remotely sensed evaporation, the hypothesis is further specified. Lead by the spatio-temporal resolution of the EARS product, EARS evaporation ($E_{RS,EARS}$) is used as forcing of a daily lumped conceptual model. Disadvantage of the coarse spatial resolution and the lumped modelling approach, is that the link between vegetation dynamics and evaporation cannot be made. This is left for further research.

In a comparative modelling approach i) the realism of the conventional procedure to determine evaporation is examined, and ii) we investigate whether by imposing $E_{RS,EARS}$ on the model, a more realistic representation of the water partitioning can be simulated. Three models are compared: $FLEX^{Ep}$ with the conventional procedure to determine evaporation, $FLEX^{Ep,RS}$ also forced with potential evaporation, but determined from the catchment average net radiation, and $FLEX^E$, forced with $E_{RS,EARS}$. The hypothesis is that especially the seasonal dynamics in streamflow generation can be better simulated by $FLEX^E$. This was not the case. It was shown that the evaporation modelled with the conventional conceptualization of the evaporation flux ($FLEX^{Ep}$), is fundamentally different from the EARS remotely sensed evaporation estimates ($E_{RS,EARS}$). Yet in terms of streamflow simulation, $FLEX^{Ep}$ outperforms ($FLEX^E$), especially in spring and autumn. Furthermore, parameter identifiability was shown not to be better for $FLEX^E$ than for $FLEX^{Ep}$.

These results indicate that either the EARS product is not as accurate as we expected, or that the lumped conceptual model does not represent the dominant processes occurring in the catchment. Assuming the latter, this can be explained by i) a too complex model with too many degrees of freedom as the limited parameter identifiability indicates, ii) an erroneous representation of the dominant processes or a wrong hypothesis on the occurring dominant processes and iii) the level of aggregation of evaporation and other hydrological processes is too high and the model too simple to be representative for the heterogeneous catchment. To look further into the latter, we suggest the combined use of topography driven semi-distributed models and the patterns in vegetation as derived by the PCA analysis.

Although we assumed the EARS product to reliably estimate the catchment scale evaporation, there are some unexplained issues. Forest evaporation seems to be overestimated by the EARS product, whereas in the water balance analysis the catchment scale evaporation appears to be in the right order of magnitude. If the overprediction indeed is the case, somewhere in the catchment or at some time evaporation is underestimated. Furthermore on clear days in winter, evaporation tends to be zero according to the parametrization of the EARS product, which is not observed by the eddy covariance measurements. It is recommended to further look into the parameterization of the EARS algorithm especially during the changing seasons.

Concluding, the use of remotely sensed evaporation estimates can be valuable for improving our understanding of hydrological processes. The value however strongly depends on the spatio-temporal scale of the product and the possibility to validate the data. In EARS, although not perfect, after a first modelling exercise, we seem to have found a rather reliable product. The relatively coarse spatial resolution does not allow for a direct use of the data in semi-distributed models or to relate evaporation to vegetation dynamics. However, the use of continuous timeseries of catchment scale evaporation in a daily lumped water balance model as direct forcing, provides the possibility to investigate the catchment scale water partitioning (internal model processes), without having to conceptualize evaporation.

CONTENTS

1	INTRODUCTION	1
1.1	On hydrological modelling	1
1.2	On evaporation dynamics	2
1.3	On evaporation modelling	2
1.3.1	Water balance models	2
1.3.2	Remote sensing	3
1.4	Motivation	4
1.5	Research questions and hypotheses	4
1.6	Approach	5
1.7	Outline	6
2	THEORY ON EVAPORATION MODELLING AND MEASUREMENTS	7
2.1	Evaporation modelling	7
2.1.1	Energy balance	7
2.1.2	Penman-Monteith	10
2.1.3	Vegetation control (Jarvis-Stewart)	10
2.2	Phenology	11
2.2.1	GSI	11
2.2.2	NDVI	11
2.3	Eddy covariance measurements	12
2.3.1	Data quality	12
2.4	Remotely sensed evaporation	13
2.4.1	WACMOS evaporation product	13
2.4.2	MOD16 evaporation	15
2.4.3	EARS	18
3	STUDY AREA AND DATA	23
3.1	Ourthe catchment	23
3.1.1	Climate	24
3.1.2	Flow regime	26
3.1.3	Anomalies	26
3.1.4	Landscape and geology	26
3.1.5	Land cover	29
3.2	Eddy covariance measurement sites	32
3.3	Data	35
3.3.1	Meteorological data, precipitation, discharge	35
3.3.2	Eddy covariance data	35
3.3.3	NDVI	35
4	VEGETATION DYNAMICS	37
4.1	Introduction	37
4.2	Method	37
4.2.1	Principle of Principal Component Analysis	38
4.2.2	Interpretation	39
4.2.3	Application to the Ourthe Catchment	39
4.3	Results and discussion	41
4.3.1	NDVI statistics	41
4.3.2	Principal component analysis	44
4.4	Conclusion	56
5	VALIDATION OF THE REMOTELY SENSED PRODUCTS	59
5.1	Introduction	59
5.2	Method	59

5.2.1	Validation against EC measurements	59
5.2.2	Validation at catchment scale	60
5.3	Results and discussion	63
5.3.1	Validation against EC measurements	63
5.3.2	Validation at catchment scale	78
5.4	Conclusions	82
6	WATER BALANCE MODEL	85
6.1	Introduction	85
6.2	Method	85
6.2.1	Forcing data and subcatchment	86
6.2.2	Model structure	87
6.2.3	Model evaluation	89
6.2.4	Calibration	89
6.2.5	Validation	90
6.3	Results and discussion	91
6.3.1	Parameter identifiability	91
6.3.2	Model performance in terms of objective functions	91
6.3.3	Comparison of the evaporation	92
6.3.4	Comparison of the discharge	92
6.3.5	Where could it go wrong?	93
6.4	Conclusion	95
7	CONCLUSIONS AND OUTLOOK	101
7.1	Conclusion	101
7.2	Outlook	102
A	EARS EWBMS ALGORITHM	103
A.1	MFG versus MSG	103
B	CORINE LAND COVER CHARACTERIZATION	105
C	PRINCIPAL COMPONENT ANALYSIS	111
C.1	Distribution principal component scores per landuse class	111
D	VALIDATION REMOTELY SENSED EVAPORATION	117
D.1	Method	117
D.2	Validation against EC measurements 1-day interval	118
D.3	Evaporative fraction	125
D.3.1	Self-preservation of Λ	125
D.3.2	Comparison $\Lambda_{RS,WACMOS}, \Lambda_{EC}$	125
E	MODEL RESULTS	129
E.1	Forcing - decision on catchment	129
E.2	Model calibration	131
E.3	Results Model 0	138
E.4	Results Model 3	139
	BIBLIOGRAPHY	141

LIST OF FIGURES

Figure 3.1	Overview study area	23
Figure 3.2	Catchment characteristics	25
Figure 3.3	Temperature, precipitation and runoff	27
Figure 3.4	Spatial distribution annual precipitation	28
Figure 3.5	Spatial distribution T	28
Figure 3.6	Catchment characteristics	30
Figure 3.7	Land cover and EC towers	34
Figure 4.1	Principle of PCA for time series analysis	38
Figure 4.2	NDVI time series (2000-2006)	42
Figure 4.3	NDVI per vegetated land use class	42
Figure 4.4	NDVI statistics of the Ourthe catchment	43
Figure 4.5	Seasonal mean NDVI	43
Figure 4.6	Characteristics of the Ourthe catchment (repeated)	43
Figure 4.7	Principal component scores	45
Figure 4.8	Coefficients of the principal components	46
Figure 4.9	Comparison of Y_2 and Y_3	48
Figure 4.10	Comparison of the seasonality of Y_4 subsets for agriculture	48
Figure 4.11	Spatial pattern of agriculture subsets	48
Figure 4.12	NDVI of Y_5 subsets	49
Figure 4.13	NDVI of Y_6 subsets	49
Figure 4.14	Characteristics Y_1	50
Figure 4.15	Characteristics Y_2	51
Figure 4.16	Characteristics Y_3	52
Figure 4.17	Characteristics Y_4	53
Figure 4.18	Characteristics Y_5	54
Figure 4.19	Characteristics Y_6	55
Figure 5.1	Point scale PM vs MOD16	64
Figure 5.2	Comparison of the time series of E_{EC} and $E_{RS,MOD16}$ for the 8 day interval	68
Figure 5.3	Comparison of the time series of E_{EC} and $E_{RS,WACMOS}$ for the 8 day interval	69
Figure 5.4	Comparison of the time series of E_{EC} and $E_{RS,EARS}$ for the 8 day interval	70
Figure 5.5	Correlation of E_{EC} and $E_{RS,MOD16}$ for the 8 day interval	71
Figure 5.6	Correlation of E_{EC} and $E_{RS,WACMOS}$ for the 8 day interval	72
Figure 5.7	Correlation of E_{EC} and $E_{RS,EARS}$ for the 8 day interval	73
Figure 5.8	Correlation of E_{EC} and E_{RS} for the ensemble of stations - 8 day interval	74
Figure 5.9	Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for MOD16 and the 8 day interval	75
Figure 5.10	Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for WACMOS and the 8 day interval.	76
Figure 5.11	Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for EARS and the 8 day interval.	77
Figure 5.12	Evaporation in the Ourthe Catchment	78
Figure 5.13	Cumulative water balance	79
Figure 5.14	Annual water balance	81
Figure 6.1	Forcing data for the Ourthe Orientale 2000-2005	86
Figure 6.2	Structure lumped model	87

Figure 6.3	Runoff coefficient	88
Figure 6.4	Model performance with respect to F_Q , F_{logQ} , F_{fdc} and F_{seas}	96
Figure 6.5	Model performance FLEX ^E , FLEX ^{Ep} and FLEX ^{Ep,RS}	97
Figure 6.6	Seasonality in evaporation according to EARS, FLEX ^{Ep,0} and FLEX ^{Ep,RS,0}	98
Figure 6.7	Hydrograph with the mean seasonal flow	98
Figure 6.8	Flow duration curve of modelled and observed discharge	98
Figure 6.9	Storage in catchment	99
Figure 6.10	Hydrograph of the daily discharge	99
Figure 6.11	Hydrograph of the logarithm of the discharge	99
Figure A.1	EARS annual evaporation estimate for the Ourthe catchment in 2005. Left: MVIRI (MFG) images as input data. Right: based on SEVIRI (MSG) images.	104
Figure A.2	EARS annual potential evaporation estimate (calculated as $0.8R_N$) for the Ourthe catchment in 2005. Left: MVIRI (MFG) images as input data. Right: based on SEVIRI (MSG) images.	104
Figure A.3	Difference in EARS annual evaporation estimate based on MVIRI (MFG) and SEVIRI (MSG) images in 2005. Left: Difference in potential evaporation ($E_{PRS,EARS}$). Right: Difference in evaporation ($E_{RS,EARS}$).	104
Figure C.1	Distribution of the normalized score of PC1	111
Figure C.2	Distribution of the normalized score of PC2	111
Figure C.3	Distribution of the PC3 score	112
Figure C.4	Distribution of the normalized score of PC4	112
Figure C.5	Distribution of the normalized score of PC5	112
Figure C.6	Distribution of the normalized score of PC6	112
Figure D.1	Comparison of the time series of E_{EC} and $E_{RS,WACMOS}$ for the daily interval	119
Figure D.2	Time series of E_{EC} and $E_{RS,EARS}$ at daily interval	120
Figure D.3	Correlation of E_{EC} and $E_{RS,WACMOS}$ for the daily interval	121
Figure D.4	Correlation of E_{EC} and $E_{RS,EARS}$ for the daily interval	122
Figure D.5	Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for WACMOS and the daily interval	123
Figure D.6	Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for EARS and the daily interval	124
Figure D.7	Correlation of Λ_{inst} and Λ_{day}	126
Figure D.8	Correlation of $\Lambda_{RS,WACMOS}$ and Λ_{EC}	127
Figure E.1	Forcing data for the Ourthe 2000-2010	129
Figure E.2	Forcing data for the Ourthe Orientale 2000-2010	130
Figure E.3	Forcing data for the Ourthe Occidentale 2000-2010	130
Figure E.4	Parameter identifiability FLEX ^E - F_Q	132
Figure E.5	Parameter identifiability FLEX ^{Ea} - F_{logQ}	133
Figure E.6	Parameter identifiability FLEX ^{Ep}	134
Figure E.7	Parameter identifiability FLEX ^{Ep2}	135
Figure E.8	Performance per calibration step, F_{logQ} and F_Q	136
Figure E.9	Model performance FLEX ^E , FLEX ^{Ep} and FLEX ^{Ep,RS}	137
Figure E.10	Hydrograph during validation for model variant 0	138
Figure E.11	Hydrograph with the mean seasonal flow	139
Figure E.12	Flow duration curve of modelled and observed discharge	139
Figure E.13	Hydrograph during validation for model variant 3	140

LIST OF TABLES

Table 3.1	Crops cultivated under a rotational system in the Ourthe catchment. In the majority of the catchment winter wheat and fodder maize predominate. In the northern part of the catchment (part of the Condroz region) sugar beet and summer barley are cultivated as well.	30
Table 3.2	Summary of the characteristics of the Ourthe Catchment and subcatchments	31
Table 3.3	Site characteristics EC measurements	33
Table 3.4	Land cover per grid cell of the E_{RS} products	33
Table 4.1	Characteristics of principal components 1-6 In bold the principal component number (Y), the explained variance (ψ) and the main characteristic that is distinguished by the principal component is given. In normal text the specific characteristic of the pixels with a high, respectively a low score (y) is given, along with the dominant land cover in these subsets.	57
Table 5.1	Comparability EC sites and RS grid cells	61
Table 5.2	Statistics of the comparison of E_{EC} and E_{RS} for MOD16 with 8 day interval	67
Table 5.3	Statistics of the comparison of E_{EC} and E_{RS} for WACMOS with 8 day interval	67
Table 5.4	Statistics of the comparison of E_{EC} and E_{RS} for EARS with 8 day interval	67
Table 5.5	Annual water balance for the Ourthe catchment	80
Table 5.6	Annual water balance for the Ourthe Orientale subcatchment . . .	80
Table 5.7	Annual water balance for the Ourthe Occidentale subcatchment . .	80
Table 6.1	Model variants with different forcing and D_z and α	89
Table 6.2	Initial parameter ranges for the lumped models	90
Table A.1	EARS based on MFG vs MSG	103
Table B.1	Land cover classes in the CORINE land cover database	106
Table C.1	Characteristics of principal components 1-3	114
Table C.2	Characteristics principal components 4-6	115
Table D.1	Statistics of the comparison of E_{EC} and E_{RS} for WACMOS with daily interval	118
Table D.2	Statistics of the comparison of E_{EC} and E_{RS} for EARS with daily interval.	118
Table D.3	Comparison Λ_{inst} and Λ_{day}	125
Table E.1	Posterior parameter range for models with $D_f > 0$ and $\alpha \neq 1$. . .	131
Table E.2	Posterior parameter range for models with $D_f = 0$ and $\alpha = 1$. . .	131

INTRODUCTION

“Evaporation is one of the least understood components of the hydrological cycle. Yet, in terms of its magnitude alone, it is as important as precipitation, runoff, and groundwater flow. (...) It is generally agreed that [the procedure to determine actual regional evaporation in rainfall-runoff models] is unsatisfactory.”

Brutsaert (1986)

The procedure Brutsaert refers to is exactly the same procedure that - after three decades of studying evaporation at a range of spatial and temporal scales - is still commonly applied in conceptual rainfall-runoff models. Catchment or regional scale evaporation still seems to be poorly understood, which motivated the subject of this thesis. In general terms this is getting more insight into the spatio-temporally variable evaporation from the catchment and in the realism of the modelled evaporation flux in conceptual rainfall-runoff models.

Before elaborating on the above mentioned procedure and why it is unsatisfactory, a short introduction on hydrological modelling is given.

1.1 ON HYDROLOGICAL MODELLING

One of the main topics in hydrology is understanding catchment's behaviour, i.e. understanding spatio-temporal variable catchment responses to climatic inputs (water and energy) (Sivapalan, 2005). To test and improve our understanding of the catchment's behaviour we translate our hypotheses of the real world's functioning to hydrological models. As such the hypotheses can be tested against observations and adjusted. Improving the model thus should improve our understanding of reality (Savenije, 2009).

Commonly applied hydrological models are conceptual rainfall-runoff models, also called water balance models. In these models spatio-temporal variable hydrological processes occurring in the watershed are spatially aggregated into a number of key responses, represented by storage components and their interactions (e.g. Beven, 2007). The effect inherent to the aggregation process, is that at the larger scales (e.g. catchment scale, functional unit scale) the processes can be described by relatively simple hydrological laws (Savenije, 2001). It also means that the model parameters characterizing the processes are scale dependent and have to be determined by calibration (Savenije, 2001). In the calibration process parameters are adjusted until relevant signatures of observed data within some limits of acceptability are reproduced (e.g. Winsemius et al., 2009). Usually data contained in streamflow observations are used for this purpose, as streamflow has the appropriate scale (Wagener, 2007) and data (point measurements) are relatively easily and accurately obtained. If, however, we are actually interested in how the catchment functions, and not just in the ability of our models to predict streamflow (under historical conditions that is), the internal processes should be known to represent the (dominant) hydrological processes in the catchment correctly as well. Or as Kirchner (2006) puts it, what we want is that we get the right answers for the right reasons. Calibration and validation of internal model processes themselves is difficult (Seibert and McDonnell, 2002), because of the level of aggregation of the modelled processes versus our point-scale understanding and observations of the 'real' processes taking place in the catchment.

Using multiple objectives in the calibration process, e.g. based on Pareto-optimality (Gupta et al., 1998) or using a stepwise model improvement approach (Fenicia et al., 2008),

is widely applied, and, although usually different aspects of the hydrograph itself are targeted, it gives some confidence in better internal model performance and consistency.

1.2 ON EVAPORATION DYNAMICS

Evaporation is, depending on the climate, one of the major components in a site's water balance (Van der Ent and Savenije, 2011, Oki and Kanae, 2006) and importantly affects the hydrologic partitioning (e.g. Franks and Beven, 1997, Winsemius et al., 2008, Thompson et al., 2011) and thus the streamflow regime. Evaporation (E) is defined here as all processes in which water is transferred from the liquid to the vapour phase. It includes interception (E_I), transpiration (E_T), soil evaporation (E_S) and open water evaporation (E_O), all in $[\text{L T}^{-1}]$ (Savenije, 2004), see Equation 1.1.

$$E = E_I + E_T + E_S + E_O \quad (1.1)$$

The physical process and time scale of these components differ.

Depending on the scale we are looking at, vegetation has an important role in the evaporation processes. Plants can directly control transpiration, and influence interception and soil evaporation through vegetation cover and leaf area. In turn, vegetation is controlled by its environment, through feedback mechanisms that vary along climates and that are still not fully understood (e.g. Asbjornsen et al. (2011), Thompson et al. (2011), Sivapalan et al. (2011)).

At small spatial (leaf) and temporal (minutes to hours) scales vegetation is known to actively control transpiration by opening and closure of stomata, under the influence of e.g. light (K), water potential gradient ($\delta\psi$), leaf-air vapour pressure deficit (D), leaf temperature (T_l) and CO_2 concentration (Jarvis, 1976). On seasonal time scales, phenology (i.e. events associated with seasonal and interannual variations in climate such as bud burst and senescence) as well as agricultural activity influence transpiration. Over even longer timescales (years or decades) transpiration of trees is found to decrease due to a reduction in the hydraulic conductivity with increasing height, an (associated) reduction in stomatal conductance and decreased leaf area index (Waring and Landsberg, 2011, Delzon and Loustau, 2005). Delzon and Loustau (2005) found that the total evaporation of the stand did not depend on stand age, and that mainly the water partitioning between overstorey and understorey was affected. However, observations at catchment scale have shown (non-linear) changes in stream flow regime after de- and reforestation (see e.g. Asbjornsen et al. (2011)).

1.3 ON EVAPORATION MODELLING

1.3.1 Water balance models

In conceptual rainfall-runoff models the climatic input in practice exists of precipitation (P) as a source of water, and potential evaporation (E_p) as a measure for the amount of energy available for evaporation.

The common way evaporation is modelled, and this is where Brutsaert (1986) referred to, is by scaling some form of potential evaporation to 'actual' evaporation by a reduction factor c_E [-], which may depend on the soil moisture content (Equation 1.2).

$$E = c_E \cdot E_p \quad (1.2)$$

with E $[\text{L T}^{-1}]$ the evaporation at the level of aggregation of the model, but E_p $[\text{L T}^{-1}]$ calculated from point scale meteorological data. The reduction factor is obtained after

some type of model calibration. Currently it typically has the form of Equation 1.3 (Feddes et al., 2001).

$$c_E = \min \left(1, \frac{S_u}{S_{u,max}} \frac{1}{L_p} \right) \quad (1.3)$$

with S_u [L] the storage in the 'unsaturated zone', $S_{u,max}$ [L] the storage capacity of the unsaturated zone and L_p [-] the threshold for moisture constrained evaporation. L_p and $S_{u,max}$ are calibration parameters, S_u is a state variable.

The main criticism against this procedure is the representativeness of the potential evaporation based on point scale meteorological and surface data (Brutseart, 1986). Furthermore, vegetation specific variables, e.g. vegetation cover and composition, rooting depth, sensitivity to water stress, and dynamics associated with the seasons are generally not explicitly taken into account, but expressed in time invariant calibration parameters. This means that the dynamic character of evaporation in these models is only expressed by the dynamics in water availability – i.e. a function of precipitation and the partitioning of water – and a form of available energy for vaporization. It is argued that the interaction between hydrology and vegetation is not sufficiently represented and that e.g. phenology should be included in evaporation modelling (e.g. Thompson et al., 2011, Ye et al. 2012), or that we should focus on a changing environment and the co-evolution of hydrology and nature (e.g. McDonnell et al. (2007), Wagener et al. (2010), Gao et al. (2014), Berghuijs et al. (2014), Harman and Troch (2014)).

The influence of the applied potential evaporation formula on model performance has been subject to several studies (e.g. Federer et al., 1996, Vörösmarty et al., 1998, Oudin et al., 2005a). And the choice in formulae is ample. Some depend on meteorologic conditions only (e.g. Hamon (1961)), sometimes surface characteristics can be taken into account in the form of albedo (Penman (1948), Priestley-Taylor (1972), Makkink (1957)) and the aerodynamic and surface resistances (e.g. Penman-Monteith (Monteith)). Oudin et al. (2005a) show that for a wide range of potential evaporation formulae - simple one parameter formulae to more complex Penman-Monteith like formulae - with differing seasonal cycle, the difference in streamflow modelling performance is relatively small (tested in a wide range of climates, with amongst others the HBV model). Moreover, it was shown that applying long term mean potential evaporation as model forcing resulted in similar performance as time-varying potential evaporation, both based on Penman-Monteith (Oudin et al., 2005b). They see this as an encouragement to apply one of the simplest formulae, only depending on temperature. However, if this 'lack of required accuracy' is true it is more likely that different internal model processes can compensate for erroneous potential evaporation, meaning a lack of model realism.

'There are more than 50 models [to calculate E_p] to choose from' (Fisher et al., 2011)

1.3.2 Remote sensing

Since the 1970s a wide range of remotely sensed evaporation estimates have been developed based on vegetation indices and land surface temperature. Evaporation can thus be observed at the scale of the catchment and is equivalent to evaporation in from a hydrological model (Winsemius et al., 2008). The integration of remotely sensed evaporation estimates and hydrological applications seems to be more common in distributed hydrological modeling. However, remotely sensed actual evaporation has been used successfully as soft data to decide on a model structure (Franks and Beven, 1997 and Winsemius et al. 2008) and for updating parameter probability distributions of parameters that relate transpiration to soil moisture states (Winsemius et al., 2008). Immerzeel and Droogers (2008) used the evaporation signal for parameter calibration in a distributed model. Contrary to this thesis, in the before mentioned studies stream flow data was not available or could not be used, so that the models relied on evaporation data only.

Depending on the temporal and spatial resolution of the remotely sensed evaporation products they can be used to:

- give insight into the spatio-temporal patterns in the catchment and determine functional areas within the catchment with comparable temporal dynamics in evaporation. With that information the switch to a semi-distributed model can be made, providing the possibility to capture more of the spatial and temporal character of evaporation by adjusting the conceptualization and/or parameterization of the evaporation flux per ‘functional area’. This is a possibility if only snapshots of data are available at a sufficiently high spatial resolution.
- update parameter probability distributions of parameters that relate transpiration to soil moisture states (Winsemius et al., 2009)
- calibrate or validate the modelled evaporation flux, and therewith the internal model processes. This is more commonly applied in distributed hydrological models (e.g. Immerzeel and Droogers 2008, Schuurmans et al. 2003, Jhorar et al. 2011)
- force the water balance model with observed evaporation to get insight into the internal water partitioning in the catchment.

1.4 MOTIVATION

What is emphasized with the above, is that, although research towards evaporation in hydrology has not stood still in the past thirty years, our understanding of the catchment scale evaporation hasn’t improved enough to change the procedure that thirty years ago already was described as unsatisfactory. Based on this procedure hydrological theories have been developed. Questions are: To what extent is the modelled evaporation the real evaporation integrated over the entire catchment? What are the consequences of neglecting spatial variability of temporal dynamics in vegetation?

In this thesis the realism of the evaporation flux is examined and the consequences of the - potentially erroneous - evaporation conceptualization for the realism of hydrological models are investigated

1.5 RESEARCH QUESTIONS AND HYPOTHESES

It is hypothesized that by integrating remotely sensed evaporation estimates and additional information on vegetation dynamics in conceptual rainfall-runoff models, we can get a better understanding of catchment behaviour and the role of vegetation dynamics in evaporation. The way the evaporation estimates can be used and which questions can be answered depend on the spatial and temporal resolution of the suitable products.

The main research questions are:

- Being an important cause of evaporation dynamics, how variable is the vegetation in space and time, in a catchment?
- How reliable are the available remotely sensed evaporation products, i.e. WACMOS, MOD16 and EARS? And how can they be embedded in the modelling process?
- How realistic is the modelled evaporation flux in water balance models when we apply the conventional procedure to describe the evaporation flux (Equation 1.2)?
- Can we get a more realistic representation of the internal model processes, (i.e. better water partitioning in the catchment), if we apply more realistic evaporation in our models?

1.6 APPROACH

The selection of a catchment subject to this research was lead by the availability of remotely sensed evaporation estimates (E_{RS}) at the appropriate spatial and temporal scale, together with ground based evaporation observations for validation purposes. Furthermore, hourly to daily discharge measurements, precipitation and meteorologic data had to be available at a as high as possible spatial resolution. The well measured and studied Ourthe catchment, a subbasin of the river Meuse, located in Belgium, was selected. In the vicinity of this catchment five eddy covariance (EC) measurement sites are located, from which direct measurements of evaporation (E_{EC}) are available. Three remotely sensed evaporation products could be used, which are based on different algorithms and data sources, and with different spatial and temporal resolutions. The products are:

- MOD16A2 8 day composite product, version 2010 (MOD16), available through the MODIS Land Products Subset Database. The MOD16 algorithm uses the Penman-Monteith method with 8 day remotely sensed input data and daily weather data. Evaporation is calculated as the 8 day sum, with a spatial resolution of ca. 1 km by 1 km. Data is available for the years 2000-2006.
- the Water Cycle Multimission Observation Strategy Evapotranspiration product, version 1.0 (WACMOS), produced by the Faculty of Geo-Information Science and Earth Observation (ITC) of the University of Twente in the context of the European Space Agency (ESA) WACMOS project of 2009. WACMOS is based on the surface energy balance. Daily evaporation is provided at cloud free days, at ca. 1 km by 1 km spatial resolution. Available for the year 2008.
- The evaporation product of the Energy and Water Balance Monitoring System (EWBMS) from EARS-E2M (EARS), provided for the purpose of this research. This product is also based on the surface energy balance and gives temporally continuous daily evaporation estimates with a spatial resolution of ca. 5 km (lon) by 9 km (lat) in the period 2000-2005 and ca. 3 km (lon) by 7 km (lat) in the period 2005-2010.

As described in the previous sections, the dynamics in vegetation is - potentially - an important source of dynamics in evaporation and water partitioning. Moreover, it can rather easily be observed using remotely sensed data. The first step in this research is the analysis of the vegetation dynamics in the catchment, over multiple years. The main objective is to distinguish areas with comparable seasonality or agricultural activity, and to detect longer term (multiple year) changes in vegetation - if present. The vegetation dynamics we are interested in are spatial and temporal variations in vegetation type, vegetation cover, leaf area, vegetation health, harvest of agricultural crops, grazing, forest clear cutting, etc. For this purpose the Normalized Difference Vegetation Index (NDVI) is used, from the MODIS sensors, at quasi 8-day temporal and 250m spatial resolution.

The second step is the validation of the remotely sensed evaporation estimates. The validation is performed at point scale as well as at catchment scale, if the spatial- and temporal resolution of the RS products allow so. In point scale validation the time series of the ground-based evaporation measurements from the five EC sites are compared with the E_{RS} time series of the grid cells in which the EC towers are located. Critical in this analysis is the spatial representativeness of E_{EC} for E_{RS} (footprint of E_{EC} versus grid cell of E_{RS}). The catchment scale validation is based on the multi-year annual water balance of the catchment, assuming that the storage change in the catchment over multiple years is zero. This validation method is only suitable for temporally continuous products with long time series. A second more qualitative validation at the catchment scale is performed by comparing the patterns of the evaporation estimates with patterns in vegetation and potential evaporation. From this step follows the selection of the most reliable RS product(s) and therewith the direction of the last step of the research.

The direction of this last step depends on the spatio-temporal resolution and reliability of the RS products. In anticipation of the results of the validation study, the EARS product was found to be best suitable. Since EARS provides temporally continuous evaporation estimates at a daily time scale, the data is used to force a water balance model. As such the water partitioning between the catchment and the atmosphere is imposed on the model. Parameters that define the partitioning in the fast and slow runoff components still need to be determined/calibrated. By comparing models with different evaporation forcing the research questions can be answered.

1.7 OUTLINE

The first part of this report contains the information that is relevant for all research questions: in *Chapter 1 Theory on evaporation modelling and measurements*, the relevant formulae to predict or to model evaporation based on meteorological data and surface resistance are given. Furthermore the theoretical background of the evaporation measurements is described, from the ground based measurements as well as the remotely sensed evaporation products.

Chapter 2 describes the *Study area and Data* that were used to test the hypotheses.

Each of the subsequent three chapters of the report concerns one of the research questions and give a more elaborate method description, the results, discussion of the results and the conclusion. *Chapter 3 Vegetation dynamics* concerns the analysis of spatial and temporal dynamics of vegetation in the catchment - associated with seasons and agricultural activity. Areas with similar dynamics are determined, and its use in the conceptualization of water balance models is discussed.

In *Chapter 4 Validation of the remotely sensed evaporation* the suitability of WACMOS, MOD16 and EARS for giving more insight into the evaporation dynamics in the Ourthe catchment is examined. This part concludes with the selection of the most reliable evaporation estimates to be used in the last part of the research.

In *Chapter 5 Water balance model* the realism of the modelled evaporation flux, and therefore with the realism in modelled water partitioning in the catchment is examined.

In *Conclusions and outlook* the main findings of this research are summarized.

THEORY ON EVAPORATION MODELLING AND MEASUREMENTS

In this chapter firstly the surface energy balance, turbulent fluxes and the Penman-Monteith formula (in fact the energy balance with rewritten sensible heat flux), are introduced. These concepts are at the heart of almost all evaporation modelling approaches, from point-scale to landscape scale, from in situ to remotely sensed approaches.

As introduced in the previous chapter, in evaporation modelling - as for any hydrological process in fact - scale is an important issue. Furthermore, the interaction with vegetation is important, although not fully understood. After introducing the Penman-Monteith model, the Jarvis model is introduced. This model describes the temporal variable canopy (stomatal) conductance. As argued by Jarvis and McNaughton (1986) this is predominantly relevant at smaller spatial and temporal scales. At larger temporal scales (seasons), phenology can be an important control on evaporation. A model to predict global phenological changes is introduced. Furthermore the background of the Normalized Difference Vegetation Index (NDVI) is given, which is used to observe phenological events remotely.

Concerning measurements of evaporation, at field scale measurements of evaporation over relatively homogeneous surfaces can be performed by e.g. the Bowen method, lysimeters, and eddy covariance measurements. The latter are used in this study and will be shortly described. At larger scales - field to landscape - remotely sensed evaporation becomes relevant. The chapter is closed with a description of the theoretical background of remotely sensed evaporation from the three RS products that are examined in this thesis.

2.1 EVAPORATION MODELLING

2.1.1 Energy balance

The energy balance of the earth's surface is given by:

$$R_n = G + S_t + F - A_h + H + \rho\lambda E \quad (2.1)$$

where R_n is the net radiation (W m^{-2}), G is the soil heat flux (W m^{-2}), S_t is the physical energy storage per unit horizontal area (W m^{-2}), F the biochemical energy storage per unit horizontal area (W m^{-2}), A_h the net advected energy flux (W m^{-2}), H the turbulent sensible heat flux (W m^{-2}) and $\rho\lambda E$ is the turbulent latent heat flux (W m^{-2}), with ρ the density of water (kg m^{-3}), λ the latent heat of vaporization (J kg^{-1}), and E evaporation (m s^{-1}). In the next sections the components of the energy balance are shortly described. See further e.g. Shuttleworth (2012) and Brutsaert (2005).

Net radiation

Net radiation R_n is the sum of outgoing and incoming short and long wave radiation (Equation 2.2).

$$R_n = (1 - \alpha) R_g + \epsilon_s R_{l,d} - R_{l,u} \quad (2.2)$$

with R_g the global short wave radiation (W m^{-2}), α the surface albedo (-), $\epsilon_s R_{l,d}$ the fraction of the downward radiation that is absorbed by the earth surface (W m^{-2}), ϵ_s the surface emissivity (equal to the absorptivity) (-), $R_{l,d}$ the downward long wave (atmospheric) radiation, and $R_{l,u}$ the upward long wave radiation, emitted by the earth's surface (W m^{-2}).

Global radiation at the earth's surface comprises direct solar radiation and diffuse sky radiation. In absence of measurements, R_g can be calculated from a variety of theoretical models and empirical formulae. It is a function of extraterrestrial radiation R_{se} , optical air mass, atmospheric turbidity, water vapour content of the air and cloud cover. The instantaneous R_{se} on a horizontal surface can be calculated with Equation 2.3.

$$R_{se} = I_S \cos \beta \quad (2.3)$$

where R_{se} is the extraterrestrial radiation, or solar radiation at the top of the atmosphere, on a horizontal surface (W m^{-2}), I_S is the extraterrestrial radiation normal to the solar beam (W m^{-2}) and β is the solar zenith angle (rad).

$$I_S = I_{pS} \left(1 + e \cos \frac{2\pi}{365} (d - 3) \right) \quad (2.4)$$

The zenith angle can be calculated according to Equation 2.5.

$$\cos \beta = \cos \phi \cos \omega \cos \delta + \sin \phi \sin \delta \quad (2.5)$$

where ϕ is the latitude, ω is the hour angle and δ is the solar declination.

The daily extraterrestrial radiation can be obtained by integration of Equation 2.3 over dt between sunrise $\omega = -\omega_s$ and sunset $\omega = \omega_s$, giving:

$$R_{se}^{day} = \frac{2}{2\pi} I_S (\sin \omega_s \cos \phi \cos \delta + \omega_s \sin \phi \sin \delta) \quad (2.6)$$

The upward long wave radiation from the surface is described by the Stephan-Boltzmann law (Equation 2.7).

$$R_{lu} = \epsilon_s \sigma T_s^4 \quad (2.7)$$

where σ is the Stefan-Boltzmann constant ($5.67 \cdot 10^{-8} \text{ W m}^{-2} \text{ K}^{-4}$) and T_s is the surface temperature (K). ϵ_s for natural surfaces ranges between 0.95 for bare soil to 0.99 for fresh snow, vegetation having values in between (see e.g. Brutsaert, 2005).

The downward long wave radiation can be calculated based on measurements of air humidity and temperature, according to a variety of equations usually of the form:

$$R_{ld} = \epsilon_a \sigma T_a^4 \quad (2.8)$$

where ϵ_a is the atmospheric emissivity (-) and T_a is the air temperature near the ground (K). Several (empirical) formula exist to estimate ϵ_a , for clear sky as well as for clouded conditions.

Turbulent fluxes

The ABL is the part of the atmosphere that is directly influenced by the presence of the earth's surface. The response time is less than one hour. It comprises the atmospheric surface layer (ASL), which is the bottom ca. 10% of the ABL, where the variation of the turbulent fluxes is less than 10% of their magnitude, and the roughness sublayer or interfacial layer. In the latter molecular transport dominates over turbulent transport, and its height is approximated by 35 times the surface roughness height or three times the vegetation height.

The main transport mechanism of latent heat $\rho \lambda E$, sensible heat H and momentum τ in the atmosphere is by turbulence. Assuming that the atmosphere nearest the surface can be considered as a steady boundary layer above a quasi-homogeneous surface (the Atmospheric Boundary Layer (ABL)), the turbulent fluxes can be defined in terms of the turbulent components of wind velocity, moisture and heat (Moors, 2012).

$$\rho \lambda E = \lambda \rho_a \overline{w'q'} \quad (2.9)$$

$$H = \rho_a c_p \overline{w'\theta'} \quad (2.10)$$

where ρ_a is the air density (kg m^{-3}), w' is the turbulent fluctuation of the vertical velocity component around the mean vertical velocity $\bar{w}$ (m s^{-1}), q' the turbulent fluctuation of specific humidity around $\bar{q}$ (kg kg^{-1}), c_p the specific heat of the air at constant pressure ($\text{J kg}^{-1} \text{K}^{-1}$) and θ the potential temperature (K). In the lower layers of the surface layer the difference between θ and T_a is often small, and θ can be replaced by T_a (Brutsaert, 2005). $\overline{w'q'}$ and $\overline{w'\theta'}$ statistically speaking are covariances.

The fluxes can also be described, in analogy to electrical circuits, using a resistance to diffusion of water vapour and heat from the surface at z_1 to the reference level z_{ref} :

$$\rho \lambda E = -\rho_a \lambda \frac{q_{z_{ref}} - q_{z_1}}{r_{av}} \quad (2.11)$$

$$H = -\rho_a c_p \frac{\theta_{z_{ref}} - \theta_{z_1}}{r_{ah}} \quad (2.12)$$

with r_{av} and r_{ah} the aerodynamic resistance to vapour transfer, respectively turbulent heat from z_1 to z_{ref} (s m^{-1}).

Equation 2.11 can be rewritten in terms of vapour pressure:

$$\rho \lambda E = -\frac{\rho_a c_p}{\gamma} \frac{e_{z_{ref}} - e_{z_s}}{r_{av}} \quad (2.13)$$

with γ the psychrometric constant ($\text{kPa } ^\circ \text{C}^{-1}$) and e_{z_s} and $e_{z_{ref}}$ the actual vapour pressure at the surface and at the reference height (kPa).

Energy storage and advection

G , S_t , F and A_h are often considered to be negligible compared to the (daily) net radiation and turbulent fluxes. The soil heat flux has a seasonal cycle with predominantly negative values in winter (heat is released) and positive values in summer (heat is stored into the soil), which averages out over the year. Upon this seasonal cycle a diurnal cycle is superimposed. The diurnal cycle more or less averages out over the day. Midday however, G can be as large as 25% of R_n (Shuttleworth, 2012, Ch.6). S_t , the physical energy storage, is attributed to the thermal capacity of the (moist) air and vegetation. The amount of energy stored per unit time per unit area between level z_1 in the soil (m) and z_2 in the atmosphere (m) in the considered volume may be calculated by Equation 2.14. Total stored energy is the sum of energy storage associated with a change in temperature of constituents i in the volume (soil, vegetation, air) (first term) and energy associated with a change of air humidity (second term):

$$S_t = \int_{z_1}^{z_2} \frac{\delta}{\delta t} \left(\sum_i \rho_i c_i T_i \right) dz + \int_{z_1}^{z_2} \frac{\delta}{\delta t} (\rho_a \lambda q) dz \quad (2.14)$$

where t is time (s), z is elevation from the surface (m), ρ_i (kg m^{-3}), c_i ($\text{J kg}^{-1} \text{K}^{-1}$) and T_i (K) respectively the density, specific heat and temperature of constituent i . In tall vegetation (forests) especially after sunrise and sunset S_t can be a significant part of the energy balance (Stewart and Thom, 1973), yet on a daily basis it is usually assumed to be negligible (Brutsaert, 2005, Ch.2).

The biochemical storage F represents the energy change due to photosynthesis and respiration of vegetation. On a day with intense photosynthetic activity F can be up to 5% of the global radiation (Brutsaert, 2005).

Net advected energy A_h is the energy that enters (positive) the volume under consideration due to advection by wind. For homogeneous surfaces it is considered negligible.

2% according to Shuttleworth (2012), 10% according to Rosema et al. (1998).

2.1.2 Penman-Monteith

The Penman-Monteith equation (Monteith, 1965) is used to calculate evaporation. Originally it is applied as a single layer model, in which the canopy is treated as a 'big leaf'. Depending on the surface resistance term that is applied, the evaporation integrates transpiration and soil evaporation. The Penman-Monteith equation is given by:

$$E = \frac{s(R_n - G) + c_p \rho_a (e_s - e_a) / r_a}{\left(s + \gamma \left(1 + \frac{r_s}{r_a}\right)\right) \lambda \rho} \quad (2.15)$$

where E is the (potential) evaporation (m d^{-1}), R_n is the net radiation on the earth's surface ($\text{J d}^{-1} \text{m}^{-2}$), G the ground heat flux ($\text{J d}^{-1} \text{m}^{-2}$), λ the latent heat of vaporization (J kg^{-1}), s is the slope of the saturation vapour pressure-temperature curve ($\text{kPa } ^\circ \text{C}^{-1}$), c_p the specific heat of air at constant pressure ($\text{J kg}^{-1} \text{K}^{-1}$), ρ_a and ρ the density of air and water, respectively (kg m^{-3}), e_a and e_s are the actual and saturation vapour pressure in the air at z m height (kPa), γ is the psychrometric constant ($\text{kPa } ^\circ \text{C}^{-1}$), $r_a = r_{av} = r_{ah}$ the aerodynamic resistance to turbulent heat and vapour transfer from the surface to some height z above the surface (d m^{-1}) and r_s is the bulk surface resistance to flow of water vapour from inside the leaf, vegetation canopy, or soil to outside the surface (d m^{-1}).

2.1.3 Vegetation control (Jarvis-Stewart)

As described in the Introduction, vegetation actively controls transpiration, predominantly through opening and closure of stomata. The stomatal conductance (g_{stomata}), the reciprocal of the stomatal resistance (r_{stomata}), is the conductance to water vapour transport from inside the leaf, through the stomata, to the atmosphere. g_{stomata} can rather accurately be described by (a modified form of) the semi-empirical Jarvis equation (Jarvis, 1976), see Equation 2.16.

$$g_{\text{stomata}} = g_{\text{stomata,max}} \Psi_{\delta\psi} \Psi_{T_l} \Psi_D \Psi_K \Psi_{CO_2} \quad (2.16)$$

in which $g_{\text{stomata,max}}$ (m s^{-1}) is the maximum stomatal conductance per unit leaf area (fully open stomata), and Ψ_i are non-linear stress functions of variable i , having values between 0 and 1, reducing the stomatal conductance under non-optimal conditions. The parametrisation of these stress functions is species and location specific. Scaling to the canopy level is done by assuming that the stomata have equal conductances and act in parallel. Thus, the stomatal conductance can be scaled to the canopy conductance by the leaf area index (LAI, i.e. the leaf area per unit surface area (m^2/m^2)) Stewart (1977):

$$g_c = g_{\text{stomata}} \frac{LAI}{LAI_{\text{max}}} \quad (2.17)$$

in which g_c (m s^{-1}) is the canopy conductance and LAI_{max} is the maximum LAI. To calculate the transpiration of a canopy, the reciprocal of g_c can be applied for $r_s = 1/g_c$ in Equation 2.15.

The term LAI/LAI_{max} is most importantly influenced by phenology or artificial changes (agricultural practice), and over longer time scales by physiological changes related to ageing. To account for natural pholiar changes in Equation 2.17 either observations of vegetation indices combined with a descriptive model or a predictive model can be used. E.g. Thompson et al. (2011) and Ye et al. (2012) use the Growing Season Index (GSI) developed by Jolly et al. (2005) to directly scale the potential evaporation in hydrological models. In the next section phenology modelling and observations are discussed.

2.2 PHENOLOGY

Phenological changes (i.e. changes associated with seasonal and interannual variations in climate such as bud burst and senescence) are studied extensively and are generally related to a combination of minimum soil temperature, day length and vapour pressure deficit (Jolly et al., 2005), and vegetation species and age. With satellite imagery changes in vegetation greenness can be monitored at large spatial scales, using vegetation indices such as the Normalized Difference Vegetation Index (NDVI) (e.g. Deering (1978), Tucker (1979), Reed et al. (1994)) and Enhanced Vegetation Index (EVI), Huete et al. (2002), and derived parameters such as Leaf Area Index (LAI) and Fractional Vegetation Cover (Myneni et al. 2002). The next sections describe the theoretical background of the NDVI and a model to predict changes in vegetation greenness on global scales, the Growing Season Index (GSI), developed by Jolly et al. (2005).

2.2.1 GSI

The Growing Season Index (GSI, Jolly et al. (2005)) is an index bounded by 0 and 1 that was originally developed to quantify the greenness of vegetation (NDVI) throughout the year, without *a priori* knowledge of vegetation and climate. It is calculated as the 21 day moving average of the product (i_{GSI}) of individual daily indicators (i_x) based on minimum temperature ($T_{\min}$), vapour pressure deficit (D) and day length (N):

$$i_{GSI} = i_{T_{\min}} \cdot i_D \cdot i_N \quad (2.18)$$

with daily indicators:

$$i_{T_{\min}} = \begin{cases} 0 & \text{if } T_{\min} \leq T_{Mmin} \\ \frac{T_{\min} - T_{Mmin}}{T_{Mmax} - T_{Mmin}} & \text{if } T_{Mmin} < T_{\min} < T_{Mmax} \\ 1 & \text{if } T_{\min} \geq T_{Mmax} \end{cases} \quad (2.19)$$

$$i_D = \begin{cases} 0 & \text{if } D \geq D_{max} \\ 1 - \frac{D - D_{min}}{D_{max} - D_{min}} & \text{if } D_{min} < D < D_{max} \\ 1 & \text{if } D \leq D_{min} \end{cases} \quad (2.20)$$

$$i_N = \begin{cases} 0 & \text{if } N \geq N_{max} \\ \frac{N - N_{min}}{N_{max} - N_{min}} & \text{if } N_{min} < N < N_{max} \\ 1 & \text{if } N \leq N_{min} \end{cases} \quad (2.21)$$

2.2.2 NDVI

The Normalized Vegetation index (NDVI) initially formulated by Deering (1978) uses differences in the spectral signature of different surfaces to distinguish between them. The NDVI expresses the ratio of the spectral reflectance in the near-infrared (ρ_{NIR}) and the spectral reflectance in the visible (red) band (ρ_{red}):

$$NDVI = \frac{\rho_{NIR} - \rho_{red}}{\rho_{NIR} + \rho_{red}} \quad (2.22)$$

There are however anisotropic effects (NDVI depends on the angular geometry of illumination and observation, thus on the position of the target within the swath of the instrument or the time of passage of the satellite over the site), atmospheric effects, and soil wetness that influence NDVI.

Mathematically, the index can take on values between -1 and 1 and is - opposed to the simple ratio $\rho_{\text{NIR}} / \rho_{\text{red}}$ - relatively insensitive to the solar zenith.

The spectral signature of vegetation is characterized by a relatively low reflectance in the visible wavelength ranges (green higher than red and blue), and a relatively high reflectance for the near infrared wavelengths. Blue and red light (wavelengths from 0.4 to 0.7 μm) is absorbed by chlorophyll and is converted into metabolic energy (up to 5% of the incoming radiation) in the process of photosynthesis (ref:RS of vegetation). Radiation in the near-infrared region (wavelengths from 0.7 to 1.1 μm) would only heat the tissues, and is scattered by the structure of the leaf tissues. The exact spectral signature of a canopy depends on, among others, the leaf structure and chemical composition (thickness, tissue density, air spaces, presence and distribution of waxes or hairs covering the leaves, pigment composition, leaf age, leaf water content), vegetation cover, and secondary and tertiary interactions between leaves at different levels in the canopy and the soil. As such, different plant species and conditions can be distinguished. Dry bare soil has a very different spectral signature with uniformly rising reflectance from the visible to the near-infrared wavelengths and with the highest reflectance in the middle infrared range. The NDVI of dry bare soils, having NIR slightly higher than VIS, is small positive (0 to 0.2). Areas covered by green vegetation will have an NDVI between 0.3 and 0.8 or 0.9. (Jackson and Huete, 1991)

NDVI is closely related to transpiration, as it reflects the chlorophyll content of vegetation (Glenn et al., 2007). Furthermore, it is related to leaf area index and vegetation cover (Carlson and Ripley, 1997), two important parameters in many evaporation formulae.

2.3 EDDY COVARIANCE MEASUREMENTS

Eddy covariance measurements provide direct observations of, among others, turbulent fluxes of sensible heat (H) and latent heat ($\rho\lambda E$). In the eddy covariance method the formulation of the turbulent fluxes according to Equation 2.9 and Equation 2.10 is used. What is measured is the covariance between the vertical wind speed and specific humidity (for $\rho\lambda E$), respectively vertical wind speed and air temperature (for H). This requires high frequency, simultaneous, co-located measurements of the variables (Shuttleworth, 2012).

The latent heat flux per unit energy required for the vaporization of one cubic meter of water ($\rho \cdot \lambda$) is equivalent to the evaporation representative for the area within the footprint of the EC measurement. This evaporation, E_{EC} , is the total evaporation, comprising soil evaporation, transpiration, interception and open water evaporation - if present within the footprint.

The EC footprint is in the range of 0.1-1km², depending on the height of the EC instrument, the surface roughness and atmospheric stability conditions (Burba and Anderson, 2010).

2.3.1 Data quality

Data uncertainty of eddy covariance measurements originates from 1) random errors, comprising errors due to sensor noise, sampling errors associated with the random nature of turbulence (e.g. incomplete sampling of large eddies) and changes in the footprint over which the measurements integrate, and 2) systematic errors such as misapplication of the eddy covariance technique for certain measurement conditions, errors resulting from instrument calibration and design and errors in the data processing, (Aubinet et al., Ch.7, 2012 and Baldocchi et al., 2001). Random errors tend to be quite large at the half-hourly time scale (Aubinet et al., 2012, Ch.7), but diminish greatly as the number of samples increase (Baldocchi et al., 2001). Some of the systematic errors can be corrected (e.g. by u^* filtering for advection effects), so that the uncertainty is reduced. Studies comparing integrated annual measurements of evaporation with lysimeters and watersheds show good agreement (Barr et al. 2000, Wilson and Baldocchi 2000, cited in Baldocchi et al. (2001)), implying that the

accuracy of $\rho\lambda E$ is relatively high. Energy balance closure checks at several sites however, show that turbulent fluxes are systematically 10% to 30% too small (Baldocchi et al., 2001). Underestimation of the night time CO_2 flux is acknowledged, but the night time latent heat flux was generally assumed to be close to zero. However, *e.g.* Fisher et al. (2007) showed by comparing sapflow measurements and EC measurements that night time transpiration is present in different tree species. Furthermore, night time interception cannot be ignored either.

2.4 REMOTELY SENSED EVAPORATION

Remotely sensed (RS) evaporation estimates are especially relevant to estimate evaporation at field and regional scales. A wide range of methods has been developed since the 1970s. The methods can be divided in roughly two classes: i) upscaling point scale estimates or measurements using spatially distributed vegetation indices, and ii) methods using the surface energy balance based on land surface temperature. The latter mainly differ in the way the sensible heat flux (Equation 2.10) is parametrised.

Three RS evaporation products have been examined in this research, namely the MODIS MOD16A2 8 day composite product (MOD16); the WACMOS ET product (WACMOS) produced by the Faculty of Geo-Information Science and Earth Observation (ITC) of the University of Twente in the context of the European Space Agency (ESA) WACMOS project of 2009; and the evaporation product of the Energy and Water Balance Monitoring System (EWBMS) from EARS-E2M (EARS). The products differ in the algorithm that is applied to determine the evaporation, in the satellite from which the observations of the land surface and atmosphere are gained and (thus) in the spatial and temporal resolution of the evaporation product. EARS and WACMOS are based on the surface energy balance, in which the sensible heat is determined based on remotely sensed land surface temperature. MOD16 is based on the Penman-Monteith equation with distributed remotely sensed input of surface characteristics. In fact this is a surface energy balance as well. However, the temperature gradient driving the sensible heat flux is not derived from surface temperature estimates, but from the saturation vapour pressure - temperature relation and modelled atmospheric vapour pressure estimates.

In the following sections the characteristics and theoretical basis of the products are described.

2.4.1 WACMOS evaporation product

The version 1.0 WACMOS evaporation product is available from <http://wacmos.itc.nl/> for the year 2008. It provides evaporation estimates and additional parameters (*e.g.* instantaneous latent heat at time of satellite overpass and the evaporative fraction) at 1 km² (0.0083°) spatial and daily temporal resolution, given cloud free satellite overpass times. The evaporation estimates are based on SEBS, the Surface Energy Balance System (Su, 2002), with optical and thermal remotely sensed images from the MODIS, MERIS and AATSR instruments of the Terra/Aqua and Envisat satellites, respectively, and meteorological data from the European Centre for Medium-Range Weather Forecasts (ECMWF). Validation results of the WACMOS product are not yet available, from SEBS there are. The principle of SEBS is the determination of the evaporative fraction on the basis of the surface energy balance at limiting cases (wet and dry pixels). The determination of the turbulent sensible heat flux is based on atmospheric similarity theories and a dynamic model for the aerodynamic roughness length for heat kB^{-1} .

Algorithm

The surface energy balance applied in the SEBS algorithm is:

$$R_n = G + H + \rho \lambda E \quad (2.23)$$

F , S_t and A_h thus are neglected (see Equation 2.1). R_n is calculated with Equation 2.2. R_g and R_{ld} are obtained from the atmospheric model. R_{lu} is calculated according to Equation 2.7, with ε_s and T_s from VIS and TIR images. G is calculated based on an empirical equation:

$$G = R_n (\Gamma_c + (1 - f_c) (\Gamma_b - \Gamma_c)) \quad (2.24)$$

where Γ_i is the ratio ground heat flux to net radiation (-) for full canopy coverage ($i = c$) and bare soil ($i = s$), and f_c is the vegetation cover fraction (-). Γ thus is linearly scaled with f_c between $\Gamma_c = 0.05$ and $\Gamma_s = 0.315$.

f_c is derived from the NDVI:

$$f_c = \frac{NDVI - NDVI_{min}}{NDVI_{max} - NDVI_{min}} \quad (2.25)$$

The sensible heat flux is determined using similarity theory, relating surface fluxes to surface variables and mixed layer atmospheric variables, i.e. describing the profiles of mean wind speed (u , ms^{-1}) and the mean potential temperature θ ($^{\circ}\text{C}$) over the atmospheric boundary layer (ABL). Integration from the surface to height z yields for the mean wind speed and mean potential temperature at height z :

$$u = \frac{u_*}{k} \left[\ln \left(\frac{z - d_0}{z_{0m}} \right) - \Psi_m \left(\frac{z - d_0}{L} \right) + \Psi_m \left(\frac{z_{0m}}{L} \right) \right] \quad (2.26)$$

$$\theta_s - \theta_a = \frac{H}{ku_* \rho c_p} \left[\ln \left(\frac{z - d_0}{z_{0m}} \right) - \Psi_h \left(\frac{z - d_0}{L} \right) + \Psi_h \left(\frac{z_{0m}}{L} \right) \right] \quad (2.27)$$

where u_* is the friction velocity (-), $k=0.4$ is the von Karman constant (-), d_0 is the displacement height (m), z_{0m} is the roughness length for momentum transfer (m), θ_s is the potential surface temperature (K), θ_a is the potential air temperature at height z (K), z_{0h} is the roughness length for heat (m), Ψ_m and Ψ_h are the stability correction functions for momentum and sensible heat respectively (-) and L is the Obukhov stability length (m). L is defined by:

$$L = -\frac{\rho c_p u_*^3 \theta_v}{kgH} \quad (2.28)$$

where θ_v is the potential virtual temperature near the surface (K). u_* is given by:

$$u_* = \left(\frac{\tau_0}{\rho} \right)^{1/2} \quad (2.29)$$

in which τ_0 is the surface momentum flux of shear stress ($\text{kg m}^{-3} \text{s}^{-2}$). u , θ and q are obtained from the atmospheric model. d_0 and z_{0m} can be determined as a function of LAI, vegetation height and wind speed. z_{0h} is obtained by:

$$z_{0h} = \frac{z_{0m}}{\exp(kB^{-1})} \quad (2.30)$$

in which B is the Station number (-). Referred is to Su et al. (2001) for the description of the model to determine kB^{-1} .

The stability functions Ψ_m and Ψ_h are, depending on the conditions, either the Monin-Obukov Similarity (MOS) functions or the Bulk Atmospheric Boundary Layer Similarity

(BAS) functions for an unstable atmosphere. For a stable atmosphere alternative equations are applied. See for a more elaborate description Su (2002).

The friction velocity, the sensible heat flux and the Obukhov stability length are obtained by solving the system of non-linear equations in an iterative process. H is constrained by the so called wet H_{wet} and dry H_{dry} limits, determined for a wet and dry pixel in the image. It is assumed that under dry conditions $\rho\lambda E = \rho\lambda E_{dry} = 0$ and thus $H_{dry} = R_n - G$. For wet conditions evaporation is assumed to occur at the potential rate $\rho\lambda E_{wet}$ and is determined with the Penman-Monteith equation (Equation 2.15). Per pixel then the relative evaporation Λ_r (-) can be determined:

$$\Lambda_r = 1 - \frac{H - H_{wet}}{H_{dry} - H_{wet}} \quad (2.31)$$

from which the instantaneous evaporative fraction Λ (-) at satellite overpass can be calculated:

$$\Lambda = \Lambda_r \frac{\rho\lambda E_{wet}}{R_n - G} \quad (2.32)$$

Assuming that the instantaneous evaporative fraction (Λ^{inst}) represents the daily evaporative fraction (Λ^{day}):

$$\Lambda^{inst} = \Lambda^{day} \quad (2.33)$$

the instantaneous as well as the daily latent heat can be calculated:

$$\rho\lambda E^{day} = \Lambda (R_n^{day} - G^{day}) \quad (2.34)$$

$$\rho\lambda E^{inst} = \Lambda (R_n^{inst} - G^{inst}) \quad (2.35)$$

Main assumptions

- $F = 0$, $S_t = 0$, $A_h = 0$
- $\Lambda_{inst} = \Lambda_{day}$
- MOS, BAS theories are applicable

2.4.2 MOD16 evaporation

The MODIS evaporation data set (MOD16A2) was downloaded through the MODIS Land Product Subsets site¹ (ORNLDAAC, 2012). It provides global evaporation and potential evaporation at 1km² spatial resolution, at 8-day, monthly and annual intervals for the period 2000-2010. Input to the algorithm are daily meteorological reanalysis data from NASA's MERRA GMAO (GOES-5), land cover from the MOD12Q1 product (Friedl et al., 2002), albedo from the MCD42B2 and MCD43B3 products Lucht et al., 2000, Jin et al., 2003 (uses MOD12Q1) and FPAR and LAI data from the MOD15A2 product (Myneni et al., 2002). Landuse based parameterization of the stomatal conductance thresholds is based on literature, listed in the Biome-Property-Look-Up-Table (BPLUT).

Algorithm

The evaporation is estimated with the algorithm of Mu et al. (2007), which is based on the Penman-Monteith equation (Equation 2.15).

In the adapted algorithm (Mu et al., 2011) interception from the wet canopy and saturated soil evaporation are taken into account, using Penman (1948).

Total evaporation is calculated as the sum of soil evaporation (E_S) and transpiration (E_T) (Equation 2.36). Interception is not explicitly taken into account (see Equation 1.1).

$$E = E_S + E_T \quad (2.36)$$

E_S and E_T are both determined using Equation 2.15, with total available energy for evaporation ($A = R_n - G$) partitioned between the soil (A_S) and the canopy (A_C) (W m^{-2}). R_n is calculated according to:

$$R_n = (1 - \alpha) R_g + (\epsilon_a - \epsilon_s) \sigma (273.15 + T_a)^4 \quad (2.37)$$

The difference with Equation 2.2 is the determination of the net long wave radiation, the second term on the right hand side. R_{lu} depends on air temperature instead of surface temperature and R_{ld} is assumed to be fully absorbed.

The atmospheric and surface emissivity are given by:

$$\epsilon_a = 1 - 0.26 \cdot e^{(-7.77 \cdot 10^{-4} \cdot T_a^2)} \quad (2.38)$$

$$\epsilon_s = 0.97 \quad (2.39)$$

In the 2010 version G is taken into account

G is assumed to be negligible, so the available energy for the turbulent fluxes is given by R_n . The energy is linearly partitioned over the bare soil and vegetated surface, depending on the vegetation cover fraction f_c :

$$\begin{aligned} A_C &= f_c \cdot A \\ A_S &= (1 - f_c) \cdot A \end{aligned} \quad (2.40)$$

where A_C (W m^{-2}) is the energy available for transpiration and A_S (W m^{-2}) the energy available at the soil surface. f_c is determined from the enhanced vegetation index (EVI):

$$f_c = \frac{EVI - EVI_{\min}}{EVI_{\max} - EVI_{\min}} \quad (2.41)$$

with $EVI_{\max} = 0.95$ and $EVI_{\min} = 0.05$, the EVI for dense green vegetation and bare soil respectively. f_c can take on values between 0 and 1. For areas with full vegetation cover the evaporation is fully determined by transpiration. The EVI is defined as:

$$EVI = N \cdot \frac{\rho_{\text{NIR}} - \rho_{\text{red}}}{\rho_{\text{NIR}} - C_1 \cdot \rho_{\text{red}} - C_1 \cdot \rho_{\text{blue}} + B} \quad (2.42)$$

where $\rho_{\text{NIR/red/blue}}$ is the surface reflectance of the near infra red (NIR), red and blue bands respectively, $N = 2.5$ is the gain factor, $B = 1$ is the canopy background adjustment that addresses non-linear, differential NIR and red radiant transfer through the canopy and $C_1 = 6$ and $C_2 = 7.5$ are the coefficients of the aerosol resistance term (see Huete et al., 2002).

For r_a a constant value of 20 s m^{-1} is applied for both canopy and soil. The effective surface resistance to transpiration (r_c) is the reciprocal of the canopy conductance (g_c). Mu et al. (2007) determine the canopy conductance from the mean potential stomatal conductance per unit leaf area g_{stomata} , multiplied by the Leaf Area Index (LAI). A Jarvis-like approach is used to constrain g_c by multiplication with linear functions of minimum air temperature ($T_{\min}$) and vapour pressure deficit (D). respectively, see Equation 2.43. See for comparison Equation 2.16 and Equation 2.17.

$$g_c = g_{\text{stomata}} \cdot \text{LAI} \cdot f(T_{\min}) \cdot f(D) \quad (2.43)$$

1 http://daac.ornl.gov/get_data.html Version 2010. As from September 2014, MOD16A2 Version 2013 is available through NASA Land Processes Distributed Active Archive Centre (https://lpdaac.usgs.gov/data_access)

The constraining functions $f(T_{\min})$ and $f(D)$ are given by:

$$f(T_{\min}) = \begin{cases} 0.1 & \text{if } T_{\min} \leq T_{S_{\min,close}} \\ \frac{T_{\min} - T_{\min,close}}{T_{\min,open} - T_{\min,close}} & \text{if } T_{\min,close} < T_{S_{\min}} < T_{\min,open} \\ 1 & \text{if } T_{\min} \geq T_{S_{\min,open}} \end{cases} \quad (2.44)$$

$$f(D) = \begin{cases} 0.1 & \text{if } D \geq D_{close} \\ \frac{D_{close} - D}{D_{close} - D_{open}} & \text{if } D_{open} < D < D_{close} \\ 1 & \text{if } D \leq D_{open} \end{cases} \quad (2.45)$$

where the index ‘open’ indicates the threshold value for $T_{\min}$, respectively D , for unconstrained g_c ; ‘close’ indicates the threshold values for fully constrained conditions. For the threshold values, Mu et al. (2007) used biome specific values for twelve biome classes. Soil water availability is thought to be reflected in D and is not included in a separate constraining function. $g_{stomata}$ is taken constant for all biome types.

The bulk surface resistance to soil evaporation r_{soil} can be conceptualised as the sum of a diffusive resistance across the water-air interface ($r_{soil,1}$) and a diffusive resistance from the interface in the soil, through the dry top layer, to the soil surface ($r_{soil,2}$) (van de Griend and Owe, 1994) In Mu et al. (2007)’s algorithm, r_{soil} is approximated with:

$$r_{soil} = 107 \cdot \frac{1}{\left(\frac{273.15 + T_a}{293.15} \right)^{1.75} \cdot \frac{101300}{P_a}} \quad (2.46)$$

which is 107 s m^{-1} for standard meteorological conditions ($P_a = 101300 \text{ Pa}$ and $T_a = 20^\circ\text{C}$).

The aerodynamic resistance for soil evaporation $r_{a,soil}$ is calculated as a parallel resistance to convective (r_{ah}) and radiative (r_{ar}) heat transfer:

$$r_{a,soil} = \frac{r_{ah} \cdot r_{ar}}{r_{ah} + r_{ar}} \quad (2.47)$$

r_{ar} is determined by:

$$r_{ar} = \frac{\rho c_p}{4\sigma T_a^3} \quad (2.48)$$

r_{ac} is assumed to be equal to r_{soil} .

The soil evaporation is determined from the ‘potential’ soil evaporation $E_{S,PM}$, as calculated with Equation 2.15 and $r_s = r_{soil}$, $r_a = r_{a,soil}$, multiplied by a function of relative humidity r (%) and vapour pressure deficit $e_s - e$ (-):

$$E_S = E_{S,PM} \cdot \left(\frac{r}{100} \right)^{\frac{(e_s - e)}{100}} \quad (2.49)$$

Main assumptions

- $A = R_n$
- $E_I = 0$
- $r_a = \text{constant} = 20 \text{ m s}^{-1}$
- $g_s = \text{constant}$
- $g_c \neq f(\theta)$, $g_c = f(T_{\min}, D)$

For a wet soil the phase change water - water vapour takes place at or adjacent to the soil surface and $r_{soil} \approx r_{soil,1}$, which is comparable with the stomatal resistance.

2.4.3 EARS

The EARS evaporation product determines the components of the surface energy balance at daily temporal resolution and a spatial resolution of approximately 5 km by 9 km (2000-2005) and 3 km by 7 km (from 2005 on) in the study area. The evaporation product is entirely based on remotely sensed visible and thermal infrared images of the Meteosat Visible Infra-Red Imager (MVIRI) instrument on board the Meteosat First Generation (MFG) satellites (until 2005) and the Spinning Enhanced Visible Infra-Red Imager (SEVIRI) instrument flying on the Meteosat Second Generation (MSG) satellites (from 2005 on). Both satellites have a geostationary orbit at 0° longitude. MVIRI has three spectral channels in the visible and thermal infra-red spectrum, with a spatial resolution of 5km for the IR channels, 2.5km for the visible channel, at the equator. SEVIRI has twelve spectral channels, of which two narrow-band channels in the visible spectrum (3km sampling resolution), one broad-band channel in the visible spectrum (1km resolution), and 9 channels in the near to thermal infra-red spectrum (3km sampling resolution). The following description is summarized from Rosema et al. (2008).

Algorithm

The surface energy balance forming the basis of the EARS algorithm is:

$$R_n = F + H + \rho\lambda E \quad (2.50)$$

G is assumed to be negligible on a daily basis. S_t and A_h (see Equation 2.1) are not considered. Opposed to the energy balance of WACMOS (Equation 2.23) and the Penman-Monteith equation in MOD16, in the EARS algorithm F is taken into account. Here it is referred to as photosynthetic energy consumption.

The algorithm has two ‘modes’, one for clear conditions and one for clouded conditions. Furthermore, correction of the net radiation is applied to account for the effect of partial cloudiness. Cloud cover between 9h and 15h is determined based on hourly TIR and VIR images. A pixel is classified as cloudy if the planetary albedo α_0^{noon} (-) and the planetary temperature at noon T_0^{noon} (K) exceed certain thresholds related to 10 day minimum α_0 and T_0^{noon} , midnight planetary temperature $T_0^{midnight}$ (K) and air temperature at the top of the atmospheric boundary layer T_a (K). The threshold values depend on day of the year, latitude and local time, and on the extraterrestrial solar radiation R_{se} . See further Rosema et al. (2008).

During clear days, the daily latent heat flux is calculated as the residual in the daily energy balance. Daily sensible heat is calculated from the instantaneous sensible heat at noon and the instantaneous and daily net radiation. During clouded conditions, when no reliable images of the earth’s surface are available, it is assumed that the Bowen ratio equals the Bowen ratio of the last clear day. $\rho\lambda E$ then can be determined from the known available energy.

Clear days

Instantaneous R_n at noon as well as daily R_n are calculated according to Equation 2.2. The instantaneous global radiation on a horizontal plane at the earth’s surface at noon R_g^{noon} is determined with Equation 2.51.

$$R_g^{noon} = \tau \cdot I_S \cdot \cos(\beta) \quad (2.51)$$

where τ is the atmospheric transmittance (-), I_S is the solar radiation at the top of the atmosphere normal to the solar beam (W m^{-2}) and β is the solar zenith at local noon (rad). Here a constant value is applied for I_S , whereas in fact it is a function of the day of the year,

see Equation 2.4. For clear conditions τ is calculated with an adapted Kondratyev model of the form:

$$\tau = 0.77 \cdot f(\alpha, 0.65 \cdot \tau') \quad (2.52)$$

where $\tau'(-)$ is the optical depth. The coefficients 0.77 and 0.65 in Equation 2.52 account for daily variation of the transmittance and for the difference in transmittance of wavelengths outside the visible window. α can be determined from the planetary albedo $\alpha_0(-)$, i.e. the albedo as sensed at the sensor, and τ' :

$$\alpha = f(\alpha_0, \tau') \quad (2.53)$$

For τ' one value is applied for the entire image, which is determined using Equation 2.53 for a pixel with 'known' α . For this purpose it is assumed that the image contains dense forest, where $\alpha = 0.07$. The pixel with the lowest 10-daily minimum planetary albedo in the image (assuming cloud-free conditions during at least one out of ten days) is assumed to coincide with this surface. From the daily value of α_0 at this pixel and the 'known' α , τ' can be calculated. Subsequently τ and α can be calculated for each pixel using Equation 2.52 and Equation 2.53.

Daily global radiation is obtained by integration of Equation 2.51 over dt from sunrise to sunset, in agreement with Equation 2.6.

The upward (R_{lu}) and downward (R_{ld}) components of the long wave radiation are calculated with Equation 2.7 and Equation 2.8, respectively. T_a in the formula for R_{ld} (Equation 2.8) is the air temperature at the top of the atmospheric boundary layer, not the the air temperature close to the surface as in Section 2.1.1. ϵ_s is taken constant at 0.9. ϵ_a is calculated with the empirical Brunt equation:

$$\epsilon_a = 0.58 + 2.73 \cdot q^{0.5} \quad (2.54)$$

For the specific humidity q monthly mean values (climate data) are applied.

The net long wave radiation is rewritten to split it in a so called radiative sensible heat term H_r and a climatic net long wave radiation $R_{n,l,c}$. Rearranging the equation for the net long wave radiation:

$$\begin{aligned} -R_{l,n} &= \epsilon_s \sigma T_s^4 - \epsilon_s \sigma T_a^4 + \epsilon_s \sigma T_a^4 - \epsilon_s \epsilon_a \sigma T_a^4 \\ &= \epsilon_s \sigma (T_s^4 - T_a^4) + \epsilon_s (1 - \epsilon_a) \sigma T_a^4 \end{aligned} \quad (2.55)$$

$$\begin{aligned} -R_{l,n} &\approx 4\epsilon_s \sigma \bar{T}^3 (T_s - T_a) + \epsilon_s (1 - \epsilon_a) \sigma T_a^4 \\ &\approx H_r + R_{l,n,c} \end{aligned} \quad (2.56)$$

where $\bar{T} = (T_s + T_a) / 2$ is the mean of the surface and atmospheric temperature. Since both H_r and the turbulent sensible heat flux H depend on the temperature difference between the surface and the top of the atmosphere, they are combined in one term, H_{tot} .

The instantaneous sensible heat is calculated from the thermal images at noon, according to:

$$\begin{aligned} H_{tot} &= H + H_r \\ &= c_h \cdot \bar{u} (T_s - T_a) + 4\epsilon_s \bar{T}^3 (T_s - T_a) \\ &= C_H (T_s - T_a) \end{aligned} \quad (2.57)$$

where c_h is the heat transfer coefficient ($\text{W m}^{-2} \text{s K}^{-1}$) and $\bar{u}$ the temporal mean wind speed at the top of the atmospheric boundary layer (m s^{-1}). c_h is a function of the vegetation cover f_c , see below. A fixed value of 5 m s^{-1} is used for the wind speed. C_H can be

This step is not required. It was introduced to include thermal inertia, which eventually was left out of the algorithm

The minus sign is used to be consistent with the signs in 3.18 and 3.20 of Rosema et al. (2008)

seen as a thermal conductivity ($\text{W m}^{-2} \text{K}^{-1}$).

Scaling from the instantaneous to the daily H is done according to Equation 2.58.

$$H^{day} = R_n^{day} \cdot \frac{H^{noon}}{R_n^{noon}} \quad (2.58)$$

The evaporative fraction Λ is related to the Bowen ratio B as follows: $\Lambda = 1 / (1 + B)$. It is shown that $\Lambda_{noon} \approx \Lambda_{day}$ on clear days, but not on clouded days, see Appendix D.

This originates from the following:

$$B \equiv \frac{H}{\rho \lambda E} \quad (2.59)$$

$$\frac{B}{1+B} = \frac{H}{\rho \lambda E + H} = \frac{H}{R_n - F} \quad (2.60)$$

where B is the Bowen ratio. Assuming that the Bowen ratio at noon is representative for the daily value:

$$B^{noon} = B^{day} \quad (2.61)$$

In addition to F , G and S_t (and A_h) were already neglected in Equation 2.50. Although the daily averaged fluxes might be close to zero, the midday fluxes are not. So even if Equation 2.61 applies, Equation 2.62 is probably not.

and assuming that $R_n - F \approx R_n$:

$$\frac{H^{noon}}{R_n^{noon}} = \frac{H^{day}}{R_n^{day}} \quad (2.62)$$

The daily mean surface temperature T_s and air temperature at the top of the atmospheric boundary layer T_a are derived from the thermal images of the planetary temperature T_0 at noon and at midnight. Firstly, T_a is obtained by linear extrapolation of the scatter plot of T_0^{noon} and $T_0^{midnight}$ to the point where $T_0^{noon} = T_0^{midnight}$. This is the point where according to perfect heat transfer the following applies:

$$T_s^{noon} = T_s^{midnight} = T_a \quad (2.63)$$

See for a graphical explanation Rosema et al. (2008). Having determined T_a , T_s can be calculated with Equation 2.64.

$$T_s - T_a = \frac{k}{\cos \beta_m} (T_0 - T_a) \quad (2.64)$$

Dry pixel: $\rho \lambda E = 0$, $F = 0$

with k the atmospheric correction factor (-) and β_m the satellite zenith angle (rad). k is obtained for the driest pixel in the image and applied to the entire image. For the driest pixel it is assumed that the latent heat is zero and the amount of energy used for photosynthesis is negligible as well. The energy balance then reduces to:

$$R_n = H = C_H (T_s - T_a) \quad (2.65)$$

In fact R_n and C_H are functions of T_s as well. Equation 2.66 is solved either iteratively or with average T_s from the preceding days.

For the dry pixel k is determined by solving Equation 2.64 and Equation 2.65:

$$\frac{R_n}{C_H} = \frac{k}{\cos \beta_m} (T_0 - T_a) \quad (2.66)$$

F is the part of the incoming radiation that is used for photosynthetic electron transport and CO_2 fixation. It is estimated with:

$$F = \eta (1 - \alpha) R_g f_c \quad (2.67)$$

where η is the daily average photosynthetic light use efficiency (-) and f_c is the vegetation cover fraction (-) based on previous (not daily) values. η is based on Rosema et al. (1998). f_c is determined by the so called relative evaporation f_E :

$$f_c = f_E = \frac{\rho \lambda E}{\rho \lambda E_p} \quad (2.68)$$

in which $\rho\lambda E_p$ is the potential latent heat flux, approximated with:

$$\rho\lambda E_p \approx 0.8R_n \quad (2.69)$$

With known R_n , H and F , the latent heat flux $\rho\lambda E$ is subsequently determined from the energy balance:

$$\rho\lambda E = R_n - F - H \quad (2.70)$$

Clouded conditions

Under clouded conditions, when no (reliable) images of the earth's surface are available, an alternative algorithm is applied. R_g is calculated according to Equation 2.51, yet with a Kubelka-Munk based equation to estimates τ :

$$\alpha_0 = (1 - \tau)^2 \alpha_{Cb} + \tau^2 \alpha \quad (2.71)$$

with α_{Cb} the albedo of cumulus nimbus clouds (0.92).

Clouds are assumed to 'trap' the long wave radiation within the atmospheric boundary layer, and net long wave radiation is assumed to be zero.

$$R_{l,n} = 0 \quad (2.72)$$

Since T_s and thus H cannot be determined during clouded conditions, it is assumed that β is equal to the β on the last cloud free day. H and $\rho\lambda E$ can then be obtained with Equation 2.73 and Equation 2.74, respectively.

According to Rosema et al. 2008 the Bowen ratio depends on moisture conditions and as long as they don't change β is constant.

$$H = \frac{\beta}{1 + \beta} \cdot R_n \quad (2.73)$$

$$\rho\lambda E = \frac{1}{1 + \beta} \cdot R_n \quad (2.74)$$

Main assumptions

- $G = 0$
- $S_t = 0$
- $A_h = 0$
- $\epsilon_0 = 0.9$
- $\bar{u} = 5 \text{ m s}^{-1}$
- $\beta^{noon} = \beta^{day}$
- $\left(\frac{H}{R_n}\right)^{day} = \left(\frac{H}{R_n}\right)^{noon}$, here $F=0$
- F is a function of f_c or relative evaporation in fact
- Under cloud cover: $R_{l,n} = 0$
- β under cloud cover = β at the last clear day
- window contains pixel with $\rho\lambda E = 0$
- window contains pixel with $\alpha = 0.07$
- at least once per 10 days no clouds to calculate α
- $\rho\lambda E_p \approx 0.8R_n$

STUDY AREA AND DATA

The selection of the study area was lead by the availability of required data and the accessibility to knowledge about the catchment. The study area consists of the Ourthe catchment upstream from Tabreux and five eddy covariance (EC) measurement sites in the vicinity of the catchment. The latter are used for validation purposes mainly. Figure 3.1 shows the locations of the catchment, EC sites, meteorological and discharge stations and pluviometers. In this chapter first a description of the main characteristics of the catchment and the five EC sites will be given. This is followed by an overview of the data that were used and a description of how they were processed.

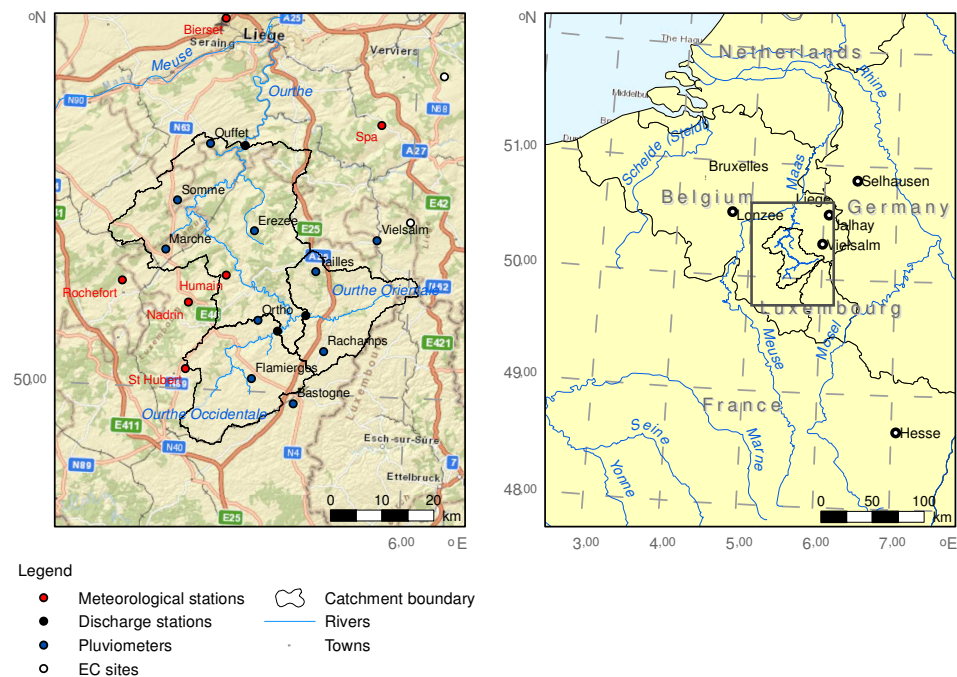


Figure 3.1: Overview of the study area. Left: the Ourthe catchment and close surroundings on top of a topographic map. The catchment under study is the Ourthe upstream from Tabreux. The subcatchments are the Ourthe Orientale catchment in the east and the Ourthe Occidentale in the south. The location of the meteorological stations (red dots), pluviometers (blue dots), closest eddy covariance (EC) measurement sites (white dots) and the discharge stations (unnamed black dots) are shown as well. Right: the location of the Ourthe in the Meuse basin, in the south-eastern part of Belgium. Furthermore the location of all five EC sites is shown: Selhausen (crops) in Germany, Lonze (crops), Jalhay (mixed forest), Vielsalm (mixed forest) in Belgium, and Hesse (deciduous forest) in France.

3.1 OURTHE CATCHMENT

The Ourthe catchment upstream from Tabreux - hereafter referred to as the Ourthe catchment - is situated in the south-eastern part of Belgium, as shown in Figure 3.1. The Ourthe is an important tributary of the river Meuse, with which it confluent in Liège, roughly 30 km downstream of Tabreux, the outlet of the catchment under study. The catchment area is 1609 km². In the upper part of the catchment the Ourthe consists of the branches Ourthe

"The Ourthe is a fast rising river, which makes this river, through its hydrological behaviour and location (close to the Dutch border) the most important Meuse tributary for flood risks in the Netherlands. It is therefore important to forecast discharges of the Ourthe" (Min I&M)

the Occidentale (flowing south-north) and Ourthe Orientale (flowing east-west). The area of the subcatchments of the Ourthe Occidentale and Orientale is 386 km² and 318 km², respectively. The two branches confluence near the city Nisramont, just before a reservoir used for drinking water storage and hydropower production, the Nisramont reservoir. The storage capacity is $3 \cdot 10^6$ m³ (Berger, 1992). After the reservoir the Ourthe flows, meandering, in north-eastern direction to Tabreux, joined by several relatively small tributaries on its way. The catchment is characterized by four geographic regions, namely, from south to north, the Ardennes, the Calestienne, the Famenne and the Condroz, with distinct geology, soils, landscape, and land use, see Figure 3.2. This figure shows the main characteristics of the Ourthe catchment, with in Figure 3.2.f the geographic regions. The elevation in the catchment ranges from 660 m in the Ardennes, the highest elevations being the plateau des Tailles in the mid-east of the catchment and the plateau de Saint-Hubert at the south-western boundary, to 110 m a.s.l. in the Calestienne at the outflow. The average gradient of the Ourthe is $3.7 \cdot 10^{-3}$ (Berger, 1992). The river valley is narrow where the Ourthe flows through limestone and sandstone rock (Ardennes, Calestienne, Condroz), and wider when flowing through the softer schists (Famenne). Discharges can rise quickly during flood waves, which is attributed to the relatively steep gradient, steep slopes and impermeable grounds (Berger, 1992). Inundations occur (regularly) in winter, in the valleys of the Famenne and Calestienne, in the lower part of the Ardennes and in the higher reach of the Ourthe Occidentale (Duchateau and Pironet, 2006).

3.1.1 Climate

The climate in the study area is temperate Atlantic, classified as Cfb according to the Köppen-Geiger classification. The mean wind speed is between 3.5 and 4 ms⁻¹, with a dominant wind direction from south to west. The catchment average annual precipitation in the period 2000-2010 is 1020 mm y⁻¹, ranging between 795 mm y⁻¹ (78% of the mean) in 2003 and 1234 mm y⁻¹ (121%) in 2001, see Figure 3.3d. The mean monthly precipitation is highest in July and August, lowest in April and June, but there is no strong seasonality. The inter-annual variability of the monthly precipitation is rather high, with values between 10% and 250% of the mean, as visible in Figure 3.3c. Figure 3.4 shows the spatial variation of the precipitation in the catchment. Annual precipitation sums range from 880 mm y⁻¹ in Ouffet (280 m a.s.l., north boundary catchment) to 1280 mm y⁻¹ in Tailles (460 m a.s.l.), the inter-annual variability differs per location. Generally, precipitation is influenced by the elevation and aspect, with higher precipitation at higher elevations on south-westerly facing slopes (plateaux of Tailles and Saint-Hubert). The station at Plateau des Tailles does show high annual sums. For Saint-Hubert there is no precipitation data available, so that catchment average precipitation, and especially precipitation in the Ourthe Occidentale subcatchment might be underestimated.

Air temperature data is available from eight stations in and around the catchment, yet time series of most stations are discontinuous. Figure 3.3a and Figure 3.3b show the monthly temperature for the period 2000-2010 at Bierset, north from the catchment. The mean monthly temperature ranges from 2.8°C in January to 18.6°C in July. On average there are 23 days with frost per year. Daily temperatures in the period 2000-2010 range from -11.2°C (January 2009) to 29.9°C (July 2006). The air temperature is highly influenced by the distance to the coast and elevation, with more gentle temperature fluctuations near the coast and lower temperature at higher elevations (0.6°C decrease per 100 m, according to the KMI (2014)). In winter however, the temperature in higher lying areas can be higher due to temperature inertia at elevations between 200 m and 400 m (Calestienne and Famenne depression) (KMI, 2014). Compared to Bierset, the other stations have a similar temporal pattern, with a structural difference of 1.2°C to 3°C in monthly mean temperature for the highest located station (Saint-Hubert, 560m). The spatial variation of the mean temperature of the months July and December 2002 are shown in Figure 3.5. In the same figure, right panel, the temperature is plotted against elevation. The mean lapse

chosen because of data availability

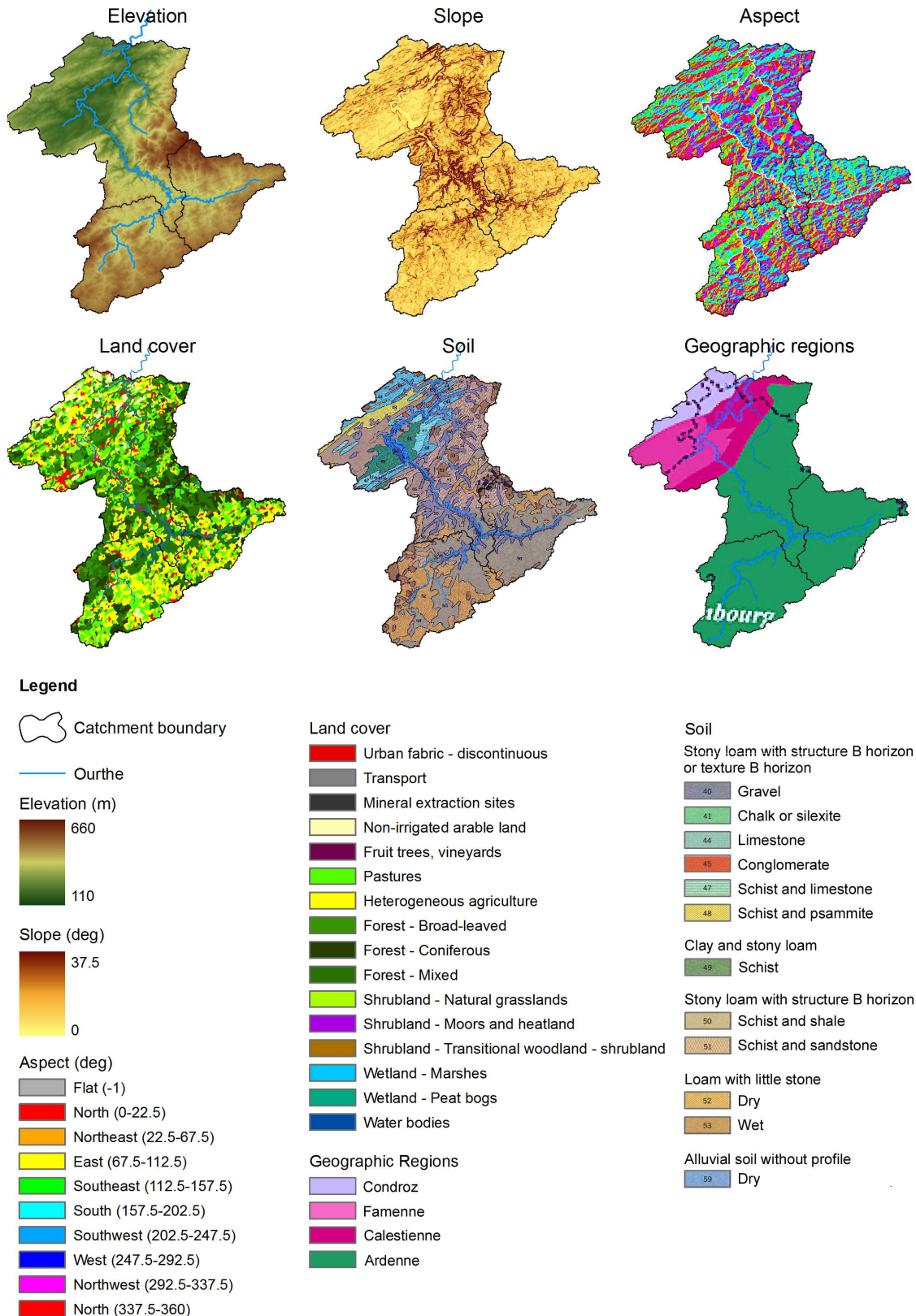


Figure 3.2: Catchment characteristics, from the top left figure, clockwise: a) elevation, b) slope, c) aspect, d) land cover (source: Corine 2006), e) soil (source: eusoils) and f) geographic regions (source: SRBG).

rate is 0.76 in July and 0.63°C in December, but is rather variable throughout the year. The temperature at Nadrin and station 002 is usually lower than expected given the elevation.

3.1.2 Flow regime

The mean discharge in the 2000-2010 period is 23 m³ s⁻¹ in the Ourthe at Tabreux, and 4.9 m³ s⁻¹ and 6.2 m³ s⁻¹ at the outflows of the Ourthe Orientale and Ourthe Occidentale, respectively. There is a clear seasonality with high flows in winter (peak flow in February) and low flows in summer and autumn (minimum between June and October), with a small peak in August, see Figure 3.3e. The mean annual discharge per unit area is ca 10% higher for both subcatchments than for the entire catchment. On a monthly basis this percentage varies, with higher values in winter and lower (negative) in summer. From the two subcatchments the Ourthe Orientale has a higher discharge per unit area in February, March and August. In the remaining months the Ourthe Occidentale dominates.

The annual runoff coefficient is 0.44, 0.44 and 0.51 for the Ourthe, the Ourthe Orientale and Ourthe Occidentale respectively, showing the importance of evaporation in the annual water balance: more than 56% of the annual precipitation evaporates. The inter-annual variability of the runoff coefficient is in the order of 14% for the Ourthe and Ourthe Occidentale, and slightly higher for the Ourthe Orientale (for smaller variability in Q and P in this subcatchment). In the runoff coefficient based on monthly values, the growing season is clearly visible: the coefficient for the Ourthe ranges from 0.92 in winter, to 0.13 in summer, with relatively stable values during January - April, and gradually changing values during the growing season, see Figure 3.3g. For the two subcatchments winter values are more variable and for the Ourthe Occidentale the maximum coefficients exceeds 1.

The variability in Q is larger than in P in all subcatchments.

The explanation for the high runoff coefficient of the Ourthe Occidentale catchment can not obviously be found in physical differences between the catchments: land cover, topography and geology of the two subcatchments are rather similar. However, as written above, we don't have precipitation data for the plateau de Saint-Hubert, where precipitation is likely to be much higher than measured at the lower lying precipitation stations. Precipitation in the Ourthe Occidentale catchment therefore is most likely underestimated. Since this is a relatively small part of the entire catchment, the effect on the runoff coefficient of the entire catchment is much smaller, but present.

3.1.3 Anomalies

Spring and summer temperatures are relatively stable in the study period. Average spring temperature deviates in 2007, caused by an extremely warm April, and in 2006 and 2010, when spring is relatively cold. Deviating summers are 2003, with high average temperature in the summer months, 2006 with an extremely hot July, and the summer of 2000, which was relatively cold. Autumn and winter temperatures are much more variable. Autumns of 2005 and especially 2006 are much warmer than average, whereas 2007 and 2010 are cold. Relatively cold winters occur in 2008/2009 and 2009/2010, and an extremely cold winter month in 2010. The winters of 2006/2007 is relatively hot.

Annual precipitation ranges between 122% and 78% of the mean annual precipitation, for both 2000-2006 and 2000-2010, with the most extreme years (for the annual values) in the 2000-2006 period. 2003, 2005 and 2006 (Jul and Aug, although extremely wet May) are relatively dry years. 2001 (Apr and Sep) and 2002 (Feb) are wet.

3.1.4 Landscape and geology

The largest and highest part of the catchment belongs to the Ardenne plateau. The Ourthe Orientale and Ourthe Occidentale are fully located in this region. The average elevation is 550 m a.s.l, causing a relatively cold and rainy climate. The Ardenne plateau is dominated

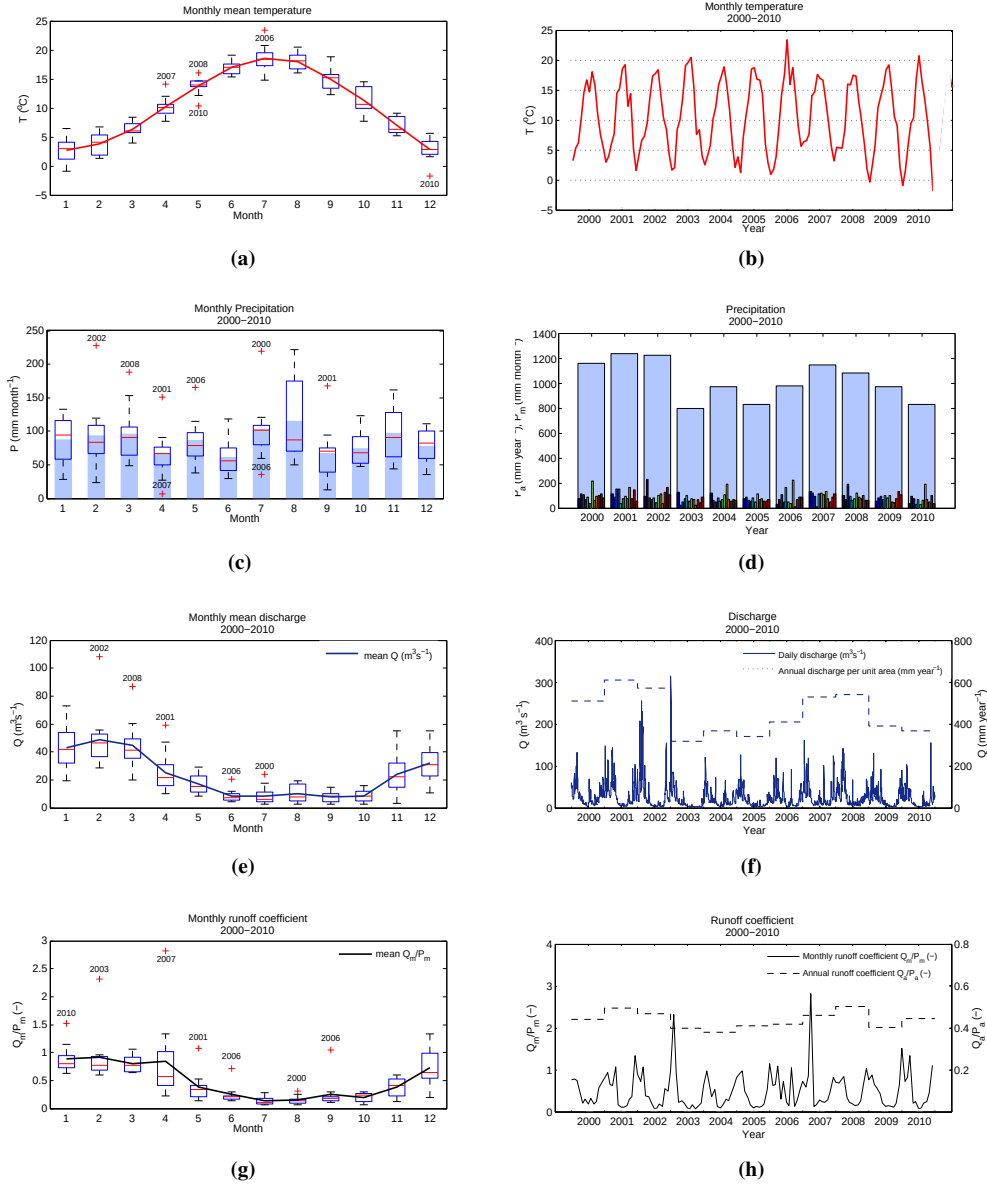


Figure 3.3: Temperature (a and b), precipitation (c and d), discharge (e and f) and runoff coefficient (g and h) for the Ourthe catchment in the period 2000-2010. Temperatures are recorded at Bierset, north of the Ourthe catchment. Precipitation is the catchment average precipitation, based on measurements at ten stations in the catchment. Discharge and discharge per unit area as measured at Tabreux. The runoff coefficient Q/P (-) is based on the shown discharge and precipitation. The left panel shows the statistics (boxplot) of a) the monthly mean temperature (°C), c) monthly precipitation (mm month⁻¹), e) monthly mean discharge (m³ s⁻¹) and g) runoff coefficient based on monthly discharge per unit area (mm year⁻¹) and monthly precipitation. Shown are the median (red line), 25% and 75% quantiles (blue box) and extreme events (red crosses). The latter are labelled with the year of occurrence. The right panel shows the full time series of the parameters at the relevant interval for the specific parameter.

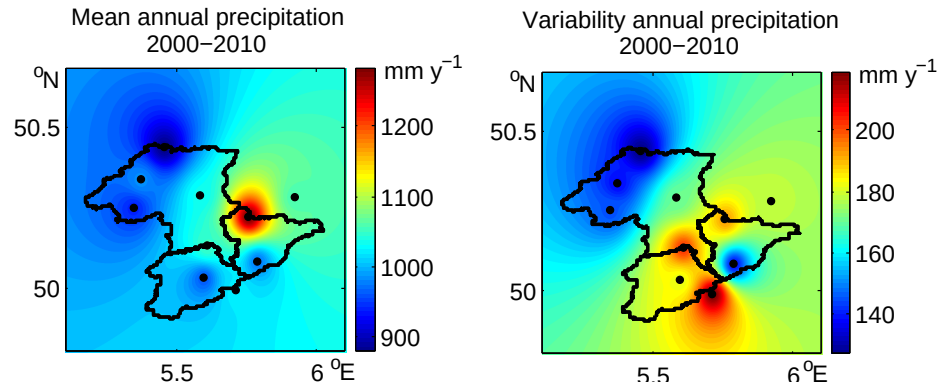


Figure 3.4: Spatial distribution of the annual precipitation in the period 2000-2010, based on the precipitation observations at ten stations. Although here an inverse distance distribution with power two is shown, the catchment mean precipitation is calculated based on Thiessen polygons. Left: the mean annual precipitation. Right: the standard deviation of the annual precipitation. Especially the upstream part of the catchment has a relatively high variability, apart from Rachamps.

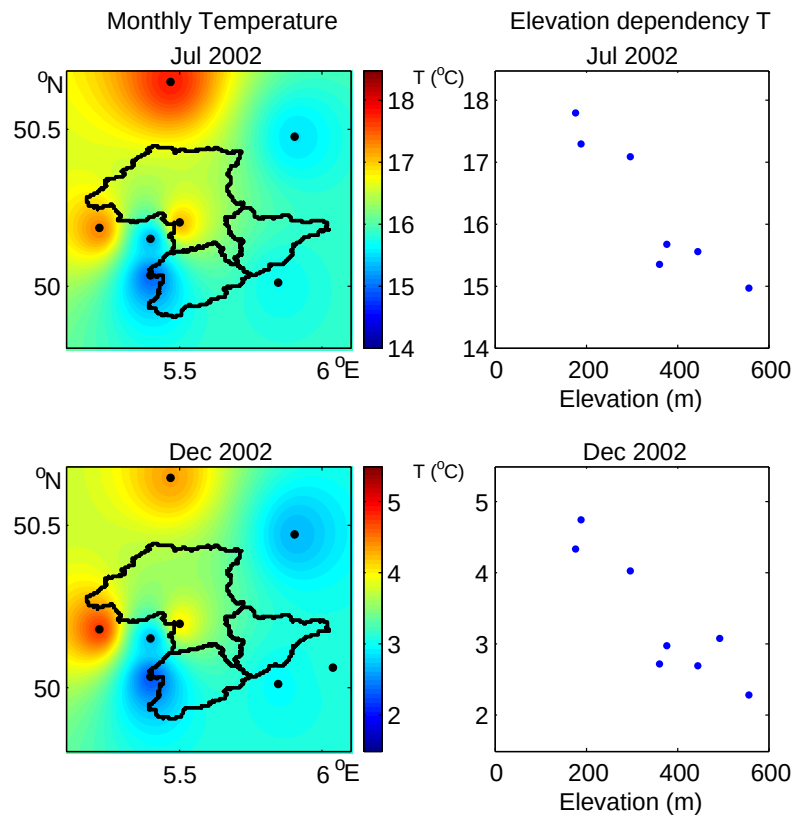


Figure 3.5: Spatial distribution of the monthly mean air temperature in July (top) and December (bottom), based on the temperature observations at seven (eight) stations (black dots in the left figures). The right panel shows the temperature as a function of elevation for the same months. The temperature difference between the warmest and coldest stations is about 3°C in summer as well as in winter. There is an elevation dependency visible, yet not all variation in monthly temperature can be explained by it.

by early Devonian quartzites (sandstone) and slates (SGBD). Steep sided river valleys with narrow alluvial planes have been carved out in the furthermore flat or rolling plateau, as visible in Figure 3.2.b. Soils in the Ardennes are mainly loamy soils with structure B horizon. Apart from some higher areas in the south of the catchment, soils have a stony content. In the south the stones exist of schist and shale, further north schist and sandstone. Drainage is usually good, but soils are shallow (FAO, 2014). Parts of the higher plateau, plateau des Tailles and plateau de Saint-Hubert, have wet loamy and peat soils (poor drainage). Going to the north-west, the Ourthe passes the geologically younger (Mid-Devonian) Calestienne - a narrow limestone formation with a hilly topography. In Figure 3.2.a and b, it is clearly visible as a relatively flat band at a lower elevation than the Ardennes and with steep slopes towards the approximately 100 m lower lying Famenne region. A karst system is created, with permanent, underground rivers at up to 70 m below the surface. Soils are shallow, very stony loams, with a limestone and schist stony content. Water infiltrates quickly in the permeable limestone and soils are dry and warm up quickly in spring. According to Denis (1992) the climate in the area is slightly warmer than in the surroundings due to the thermal characteristics of the limestone. However, cold air masses from the Ardennes can accumulated in the depression, causing a temperature inversion and a cool climate (FAO, 2014). The gradually sloping (south-west to north-east) area north-west of the Calestienne, is part of the Famenne depression. This depression consists of relatively homogeneous late-Devonian schists, which are soft (frost-sensitive) and easily erodible. The floodplains are wide, the river bed is narrow, see the characteristic flat, low area on both sides of the Ourthe just before the river bends to the east (Figure 3.2.b). The soils here originate from alluvial deposits: stony-loamy and - uniquely in the catchment – clayey soils. Parallel to the river there is a band with stony loam with gravel (FAO, 2014). The clayey parts (platic clay) are poorly drained, and consecutively saturate or dry out completely (Denis, 1992). The highest part of the Famenne (northern part, bordering the Condroz) has stony-loamy soils with a schist stony content. This part of the catchment still frequently inundates (Duchateau and Pironet, 2006) The higher region forming the north-western boundary of the catchment is part of the Condroz relief. It is characterized by a succession of parallel late-Devonian psammite (hard sandstone) ridges and plateaus, and early Carboniferous limestone depressions, oriented south-west - north-east. There is one depression and one ridge recognizable within the catchment boundaries. The soil in the depression is stony-loamy with a limestone stony content. The ridge has a stony-loamy soil with psammite and schist. Furthermore there are small areas with moderately dry loam soils (loess deposit). (Duchateau and Pironet, 2006, , SRBG, Berger, 1992, .)

3.1.5 Land cover

The population density in the Ourthe catchment is, with 50 inhabitants per km², low. Built-up area covers 5.6% of the total surface area, with the highest population density in Marcheen-Famenne. The rural area is approximately equally divided between forests (44%) and agriculture (48%). Where hillslopes are too steep and where soils are unsuitable for agriculture (nutrient poor, too stony, shallow or wet), forest are located, compare Figure 3.2.b and d. Four agricultural land cover classes are distinguished in the CORINE data base, see Appendix B, with as main difference the areal extent of the agricultural plots and heterogeneity of the landscape. The crops in the rotation system in the majority of the catchment are winter wheat and maize in most years. Sometimes maize is replaced by sugar beet. In the northern part of the catchment (and in the Condroz region, stretching to the north) the crops in the rotation system are maize and summer barley.

An overview of the characteristics of the Ourthe catchment and the subcatchments is given in Table 3.2 and Figure 3.6.

Quartzites: hard, non-foliated metamorphic rock, originating from pure quartz sandstone. Slates (NL: leisteen): foliated metamorphic rock, originating from shale.

Shale (NL: schalie): fine-grained, fissile and laminated clastic sedimentary rock composed of clay minerals and silt.

Claystone (NL: kleisteen): fine-grained, non-fissile, sedimentary rock consisting primarily of compacted and hardened clay

Schist: medium-grade metamorphic rock having more than 50% platy and elongated minerals, exhibiting schistosity.

Sandstone: clastic sedimentary rock composed mainly of sand-sized minerals or rock grains (porous)

Caves of Hotton and Barvaux

Table 3.1: Crops cultivated under a rotational system in the Ourthe catchment. In the majority of the catchment winter wheat and fodder maize predominate. In the northern part of the catchment (part of the Condroz region) sugar beet and summer barley are cultivated as well.

Rotation system in the Ourthe catchment						Cropping pattern in 2002
Year	Majority catchment		Condroz region			
2000	sb	ww	mf	sb	sb	
2001	ww	mf	sba	mf	ww	
2002	mf	ww	mf	sb	sb	
2003	ww	sb	sba	mf	ww	
2004	mf	ww	mf	sb	sb	
2005	ww	mf	sba	mf	ww	
2006	mf	ww	mf	sb	sb	
2007	ww	mf	sba	mf	ww	
2008	mf	ww	mf	sb	sb	
2009	ww	mf	sba	mf	ww	
2010	po	ww	mf	sb	sb	

sb = sugar beet (orange), ww = winter wheat (red), mf = fodder maize (yellow), po = potatoes, sba = summer barley (cyan), fallow (grey).

The crop is the harvested crop in the specified year. The spatial resolution (1km²) is rather coarse for the spatial extent of the agricultural parcels (10-100s ha) and crop types should be seen as an estimate.

Data from: Wattenbach, M. *Crop data for Europe*. Available through the GHG-Europe Database, accessed April 2014. (<http://gaia.agraria.unitus.it/> > GHG Europe > Data > Others data > Crop data for Europe).

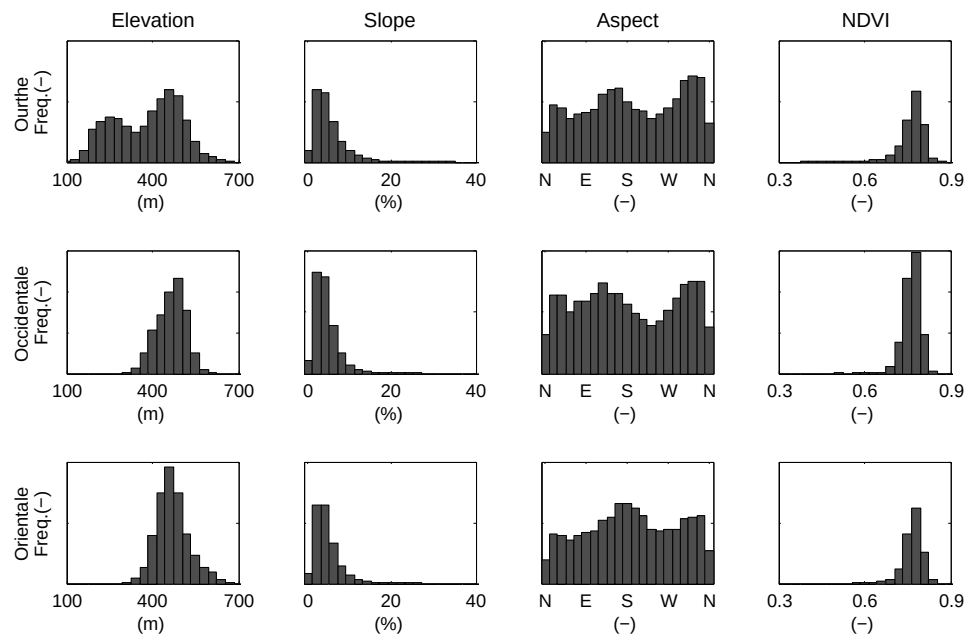


Figure 3.6: Catchment characteristics of the Ourthe and of the subcatchments Ourthe Occidentale and Ourthe Orientale. The histograms show the distribution of the elevation, slope, aspect and NDVI. The main difference between the entire catchment and the two subcatchments is the elevation, as far as it concerns topography and NDVI.

Table 3.2: Summary of the characteristics of the Ourthe Catchment and subcatchments

	Ourthe	Ourthe downstream part	Ourthe Occidentale	Ourthe Orientale
NDVI (-)				
μ	0.77	0.77	0.77	0.77
min	0.38	0.38	0.50	0.56
max	0.86	0.86	0.85	0.85
σ	0.04	0.05	0.03	0.03
Elevation (m)				
μ	380	312	466	470
min	110	110	297	301
max	663	661	597	663
σ	114	101	500	561
Slope (%)				
μ	4.9	5.4	4.2	4.3
min	0	0	0	0
max	33	33	26	26
σ	3.6	4.1	2.7	2.9
Aspect ($^{\circ}$)				
μ	193	198	182	189
min	0	0	0	0
max	360	360	360	360
σ	106	108	108	100

3.2 EDDY COVARIANCE MEASUREMENT SITES

In this section the characteristics of the EC sites that are used in this research are given, and the heterogeneity of the land surface within the grid cell of the remote sensing products is discussed. Theoretical background of the EC measurement technique is described in Chapter 2.

The EC sites are chosen such that as many different land cover classes as possible are represented, within the same climatic region as the catchment (warm temperate fully humid with warm summer (Cfb)). Five EC sites fulfil the requirements, the characteristics of which are summarized in Table 3.3. The furthest EC site is located at a distance of 200km from the catchment. The land use classes that are covered are cropland (Lonzee and Selhausen), deciduous broad-leaved forest (Hesse) and mixed forest with coniferous and broad-leaved trees (Vielsalm and Jalhay). Although grassland and shrubland are importantly present in the Ourthe catchment as well, there are no suitable EC sites available with these land uses. The elevation of the EC sites ranges from 102m to 486m, which is within the range of the elevation of the study area. Although all sites have a Cfb climate, a considerable difference in annual mean temperature and annual precipitation between the sites is present. The mean temperature ranges from 6°C to 10°C, precipitation from 700mm y⁻¹ to 1200mm y⁻¹, Jalhay and Vielsalm being the coldest and wettest of the five sites and located at the highest elevations.

Figure 3.7 shows the grid cells of the RS products that comprise the EC sites, on top of the CORINE 2006 land cover map (EEA, 2012). Due to the different projections and spatial resolution of the RS products, the area of the specific grid cells deviates rather much, in both surface area and location. EARS covers a much larger (about 70 times) and more heterogeneous area per grid cell than the other two products. The WACMOS and MOD16 grid cells partially overlap. At Vielsalm and Selhausen the latter two products have more than 1/2 of their area in common; for Lonzee, Jalhay and Hesse the overlapping area is less than 1/3 of the surface area.

As can be seen from Figure 3.7, the land cover in the direct vicinity of the EC towers is relatively homogeneous - a prerequisite for reliable EC measurements, see Section 3.3. Within the WACMOS and MOD16 grid cells the land cover is rather homogeneous as well - in terms of main land cover class, crop types might differ within the area - with more than 70% of the land cover class of the specific EC tower site, except for WACMOS at Jalhay, see Table 3.4. The WACMOS grid cell at Jalhay comprises almost 50% transitional woodland, probably outside the footprint of the EC tower. Since this is woodland in development (increasing NDVI over the years, see the PCA analysis in Chapter 4) the evaporation might deviate from the evaporation from the mixed forest as measured at the EC site, merely in absolute sense. The EARS grid cells cover a wide range of land cover classes at all sites, with the proportion of the EC site land cover ranging between 35% and 82%. For Hesse agriculture is dominant in the grid cell with 54% of the total area, and not the broad-leaved forest of the EC site. The proportion of agriculture in the Vielsalm grid cell is with 35% considerable as well, next to the mixed forest. The grid cell covering the Jalhay site has a rather distinct land cover composition, with transitional woodland (as in the WACMOS grid cell) and a large area with peat bogs. Both are expected to have an evaporation pattern that deviates from the evaporation as measured at the EC site, but no direct evaporation measurements are available to confirm this. Noteworthy furthermore is the site Selhausen. The land cover of the EARS grid cell for this site consist for 25% of an open-pit lignite (brown coal) mine, having completely absent vegetation. Evaporation in this part thus is solely soil (and interception) evaporation, which is likely to be much lower than evaporation of vegetated areas, with smaller time scales.

In a point scale study relating E_{EC} measurements with evaporation determined with the Penman-Monteith equation with seasonally varying canopy conductance, it was found that Jalhay forest has a much lower surface resistance against evaporation compared to the other forest sites. This might indicate that the forest is relatively young and thus behaving similar to transitional woodland.

Table 3.3: Site characteristics EC measurements

Site	Lon (°)	Lat (°)	Elevation (m)	Climate	Mean T (°C)	Mean P (mm y ⁻¹)	Landuse	Canopy height (m)
Vielsalm	50.306	5.997	450	Cfb	7.0	1150	Mixed forest (Beech, Douglas fir)	27;35
Jalhay	50.564	6.073	486	Cfb	6.0	1200	Mixed forest	5-7 ^a
Hesse	48.674	7.066	300	Cfb	9.2	885	Deciduous broadleaf forest (Beech)	14
Lonzee	50.552	4.745	165	Cfb	10.0	750	Cropland rotation (ww/sp/ww (m)/sb) ^c	0 - 0.8 ^b
Selhausen	50.871	6.450	102	Cfb	9.9	698	Cropland rotation (ww/ww/sb) ^c	0 - 0.8

^a Estimation, based on EC tower height. ^b Estimation, based on crop type. ^c ww = winter wheat, sp = seed potato, sb = sugar beet, m=mustard.

Source: All EC data is downloaded from <http://www.europe-fluxdata.eu/>. Ancillary information is gained from the Fluxnet and Carbo-Europe (through bgc-jena) databases and from publications. Principal investigator (PI), database (db), references (ref) per site: Vielsalm: Vincke, C. (PI), Fluxnet and Carbo-Europe (db). Jalhay: Francois, L. (PI), Euroflux (db). Hesse: Granier, A. and Longdoz, B. (PI), Fluxnet and Carbo-Europe (db). Lonzee: Moureaux, C. (PI), Fluxnet (db), Suleau et al. 2011, Aubinet et al. 2009, Moureaux et al. 2006 (ref). Selhausen: Schneider, K. (PI), Euroflux (db), Schmidt et al. (2012) (ref).

Table 3.4: Land cover per grid cell of the E_{RS} products, for the grid cells comprising the EC towers. The percentage of the main land cover classes are given in black. In grey the percentages of the subclasses are given.

Fluxtower	RS product	Agriculture (%) (Crops, Pasture)		Forest (%) (BroadL., Conif, Mixed)		Woodland (%)	Built up (%)	Other classes (%, class)
Vielsalm (mixed forest)	MOD16	14.5	(13.0, 1.5)	85.5	(-, -, 85.5)			
	WACMOS	26.5	(26.5, -)	73.5	(-, -, 73.5)			
	EARS	31.3	(8.7, 22.6)	63.0	(-, 19, 44)		4.8	0.6 natural grassland, 0.2 wetland
Jalhay (mixed forest)	MOD16			95.8	(-, 70.0, 25.8)			4.3 natural grassland
	WACMOS			54.0	(-, 54.0, -)	46.0		
	EARS			61.1	(5.2, 37.1, 18.8)	13.3		2.4 natural grassland, 23.2 wetland
Hesse (broad-leaved forest)	MOD16			100.0	(100.0, -, -)			
	WACMOS			100.0	(100.0, -, -)			
	EARS	54.0	(33.4, 20.5)	39.3	(34.5, 3.1, 1.7)	0.8	3.8	2.2 open pit mining
Lonzee (agriculture)	MOD16	79.8	(79.8, -)				20.3	
	WACMOS	100.0	(100, -)					
	EARS	81.7	(79.9, 1.8)	1.1	(1.1, -, -)		17.2	
Selhausen (agriculture)	MOD16	100.0	(100, -)					
	WACMOS	100.0	(100, -)					
	EARS	51.3	(48.4, 2.9)	10.5	(8.1, -, 2.3)	4.8	8.5	24.9 open pit mining

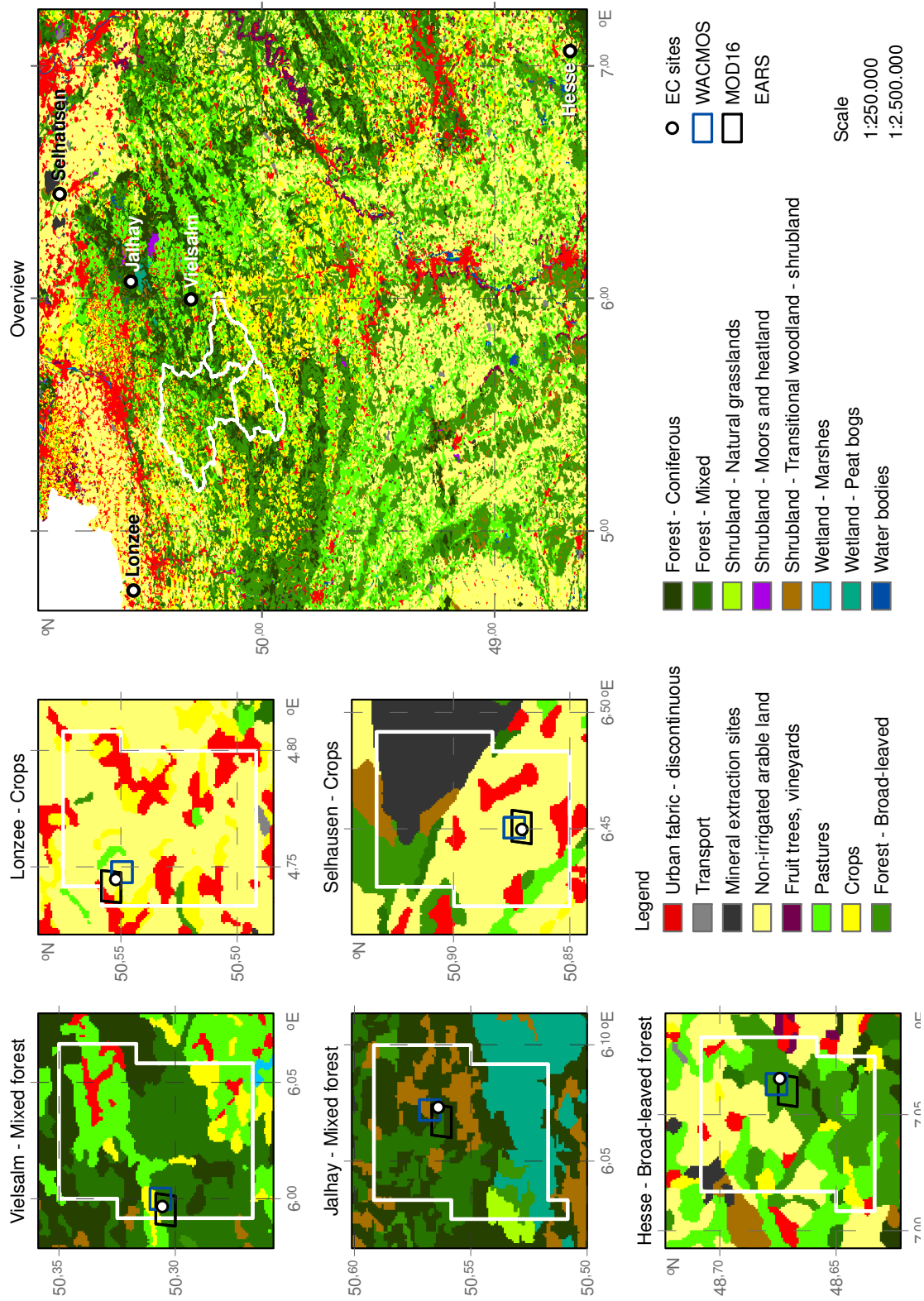


Figure 3.7: Land cover and EC towers. Left: Grid cell coverage of the RS evaporation products per EC tower, in white EARS, blue WACMOS and black MOD16 (sinusoidal projection). Right: Overview of the study area, EC tower location (white dots) and the Ourthe catchment and subcatchments (white line)

3.3 DATA

In the overview of the study area (Figure 3.1), the location of the EC measurement sites and the meteorological and precipitation stations are shown.

3.3.1 Meteorological data, precipitation, discharge

Meteorological data of temperature, wind speed, relative humidity and global radiation is available for 1992-2005 on a daily basis, and from 2001-2011 on an hourly, but discontinuous, timestep. Precipitation data from ten stations, measured with pluviometers, is available on an hourly basis for the period 1992-2011. Areal interpolation of the precipitation data has been performed with Thiessen polygons. Given the strong dependency on elevation and aspect applying Cokriging interpolation should yield a more accurate representation of the spatial distribution of precipitation. Discharge measurements from three stations (Tabreux, Ortho and Mabompre) are used in this study, available for the same period as the precipitation data.

3.3.2 Eddy covariance data

The eddy covariance data used in this research is available at L2, L3 and L4 quality level, depending on the variable, measurement site and period. L2 data is quality checked and flagged only, whereas L4 data is quality checked, QAQC (quality assurance and quality control) filtered (friction velocity (u^*) and spikes), gap-filled and flagged. For validation of the remotely sensed evaporation products the half hourly $\rho\lambda E$ data is used, where available at L4 level. Daily evaporation is calculated from the positive half hourly LE data .

Evaporation if $\rho\lambda E \geq 0$,
condensation if $\rho\lambda E < 0$

3.3.3 NDVI

MODIS 16-Day Composite L3 Global 250m SIN Grid [Collection 5] NDVI data is used in this research to examine the vegetation dynamics in the study area. The data is downloaded through the MODIS Land Product Subsets site (ORNLDAAC, 2012) for the period 2000-2010.

Both available products are used, namely the MOD13Q1 and MYD13Q1, which are based on the daily MODIS L2G surface reflectance from the MODIS sensors onboard the Terra and Aqua platforms, respectively. The advantage of a composite product is the minimization of bad quality data due to e.g. clouds and snow. Within the 16 day period per pixel the highest high quality data (quality based on cloud contamination and viewing geometry) is selected and assumed to be representative for the entire 16 day period. MOD13 and MYD13 are processed 8 days out of phase to provide a quasi-8 day temporal frequency (Solano et al., 2010).

In the analysis of the vegetation dynamics, the NDVI time series of the MOD13 and MYD13 products are combined into one data set. For the timing of the data the start date of the composite period is applied, resulting in above mentioned quasi-8 day temporal frequency. The actual interval between two subsequent NDVI values can in principle be 16 days. Quality assessment data from the pixel reliability layer have been used to value the data. Indications are: good quality (0), marginal data but useful (1), snow/ice cover (2) and cloudy (3). Data contaminated with snow, ice or cloud cover are removed.

VEGETATION DYNAMICS

4.1 INTRODUCTION

Although the exact role vegetation plays in evaporation and water partitioning in the catchment is unclear, at least at small spatial scales the dynamics in vegetation should be reflected in the dynamics of transpiration, interception and soil evaporation, and potentially in total evaporation. In this chapter the temporal and spatial dynamics of vegetation in the catchment are examined, based on multiyear timeseries of NDVI and supported by CORINE land cover data. The objective of the analysis is to get insight into the (observable) temporal variability in vegetation 'greenness' (integrating vegetation density, cover, leaf area, 'health') of both natural vegetation and agricultural crops, and to eventually distinguish areas within the catchment with similar temporal dynamics.

4.2 METHOD

The NDVI is an index that is commonly used to monitor or study a large variety of phenomena related to vegetation, e.g. vegetation density, vegetation cover, leaf area index, phenological changes, biomass production, transpiration and droughts, and vegetation composition (e.g. Deering (1978), Carlson and Ripley (1997), Glenn et al. (2007)). See Chapter 2 for the theoretical background of this vegetation index. Satellite derived NDVI data often have a rather high noise-to-signal ratio and might be contaminated by cloud cover or snow, despite quality checks of the data. See e.g. Hmimina et al. (2013), who studied the potential of MODIS derived NDVI to predict phenology, amongst others at the EC sites Lonze and Hesse (also used in this thesis). To acquire information on vegetation dynamics from the NDVI signal usually smoothing algorithms or model fitting are applied. 'Phenological metrics' such as 'start of the season' and 'start senescence', or longer term trends are subsequently extracted from the smoothed signal, see de Beurs and Henebry (2010) for an overview of methods. In this research a principal component analysis (PCA) on the multi year NDVI data has been performed. PCA is a transformation technique that can be applied on - among others - image time series, with as main purpose to reduce the dimensionality of the data set, while retaining as much of the information (variance) present in the original data. The structure of the data can thus more easily be analyzed. Applying PCA on NDVI time series provides an alternative to phenological metrics for tracking the response of the vegetated land surface to variability resulting from climate and other land surface changes (Hall-Beyer, 2003). It does not directly derive phenological metrics, but rather distinguishes areas that have a particular temporal pattern in common. At the same time noise can be separated from the signal (see below). Furthermore, apart from seasonal changes, non-cyclic changes of the surface are extracted as well, if present in the dataset (Hall-Beyer, 2003). Several studies (e.g. Hall-Beyer, 2003, Hirose et al., 1996, Eastman and Fulk, 1993) successfully applied PCA on NDVI data to identify similarities in phenology, to classify land cover or to detect change. In the following sections first the theory of PCA is explained, followed by a description of the way the results can be interpreted, and of the data that were used. Dataset $X=(X_1, \dots, X_p)^T$ is the time series of p NDVI images of size $n_1 \cdot n_2 = n$. Variable X_i thus is the i -th image in the time series.

4.2.1 Principle of Principal Component Analysis

PCA is a linear transformation of the variables of the data set $X = (X_1, \dots, X_p)^T$ into the data set $Y = (Y_1, \dots, Y_p)^T$, consisting of uncorrelated variables Y_i , called the principal components, see Figure 4.1.

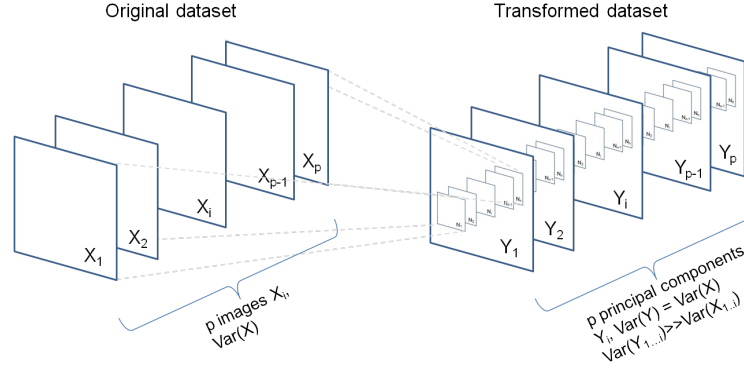


Figure 4.1: Principle of PCA for time series analysis

Each of the principal components (Y_i) is a linear combination of the mean-centered variables of data set X , given by:

$$\begin{aligned} Y_1 &= \gamma_{1,1} \cdot (X_1 - \bar{X}_1) + \dots + \gamma_{1,i} \cdot (X_i - \bar{X}_i) + \dots + \gamma_{1,p} \cdot (X_p - \bar{X}_p) \\ Y_2 &= \gamma_{2,1} \cdot (X_1 - \bar{X}_1) + \dots + \gamma_{2,i} \cdot (X_i - \bar{X}_i) + \dots + \gamma_{2,p} \cdot (X_p - \bar{X}_p) \\ &\dots \\ Y_p &= \gamma_{p,1} \cdot (X_1 - \bar{X}_1) + \dots + \gamma_{p,i} \cdot (X_i - \bar{X}_i) + \dots + \gamma_{p,p} \cdot (X_p - \bar{X}_p) \end{aligned} \quad (4.1)$$

or in matrix form:

$$Y = \Gamma^T (X - \bar{X}) \quad (4.2)$$

with $\bar{X}_i$ the mean of X_i and $\Gamma = (\gamma_1, \dots, \gamma_p)$, with $\gamma_i = (\gamma_{i,1}, \dots, \gamma_{i,n})^T$ the weight vector with the weights or coefficients of principal component i . The weights are chosen such that the first component accounts for the greatest possible variance of the data set, with each successive principal component having a smaller variance and being perpendicular (independent) to the former principal components, while $\gamma_{i,1}^2 + \gamma_{i,2}^2 + \dots + \gamma_{i,n}^2 = 1$. The weight vectors fulfilling these requirements are the eigenvectors of the covariance matrix S of the original data set:

$$\text{Var}(X) = S = \frac{1}{n-1} (X - \bar{X}) (X - \bar{X})^T = \Gamma \Lambda \Gamma^T \quad (4.3)$$

$$\text{Var}(Y) = \text{Var}(\Gamma^T (X - \bar{X})) = \Gamma^T S \Gamma = \Lambda \quad (4.4)$$

in which Γ is the orthogonal matrix with the eigenvectors γ_i of S in its columns, and $\Lambda = \text{diag}(\lambda_1, \dots, \lambda_p)$ the diagonal matrix with the corresponding eigenvalues λ_i .

So, the transformation vector with the coefficients of Y_i is the i -th eigenvector (γ_i) of the covariance matrix of X . The variance accounted for by Y_i is the corresponding eigenvalue (λ_i), ordered such that $\lambda_i < \lambda_{i-1}$ for all i . The explained variance by principal component

i is usually expressed as the fraction of the total variance (ψ_i). See further Jolliffe (2005) and Härdle and Simar (2007). The following equations hold for principal component i , Y_i :

$$\begin{aligned}
 \text{Var}(Y_i) &= \lambda_i \\
 \text{Var}(Y_1) &\geq \dots \geq \text{Var}(Y_p) \\
 \text{Var}(X) &= \text{Var}(Y) = \sum_{j=1}^p \lambda_j \\
 \psi_i &= \frac{\lambda_i}{\sum_{j=1}^p \lambda_j} \\
 \bar{Y}_i &= 0 \\
 \text{Cov}(Y) &= 0
 \end{aligned} \tag{4.5}$$

There are as many principal components as there are variables in the original data set, yet by neglecting lower order components containing only a small part of the total variance (smallest eigenvalues), the dimensionality of the data set can be reduced. The data is mean-centered on all variables to ensure that the data cloud is centered on the origin of the principal components. In the case of time series analysis, the images are normalized (z-score) to prevent that images with a much larger dynamic range (variance) due to a few contaminated pixels (Jolliffe, 2005, Hirosawa et al., 1996) or due to the season (summer images vs winter images) (Hall-Beyer, 2003) will dominate the first few principal components. Mathematically this means that the PCA is performed on the correlation matrix instead of on the covariance matrix of X . Equation 4.1 - Equation 4.5 hold, yet with $(X - \bar{X})$ replaced by N , the normalized data set.

4.2.2 Interpretation

Since the principal components are derived from the NDVI data itself, interpretation requires additional information of the site's characteristics that could explain differences in NDVI, such as vegetation cover and land use, climate, geography. The above mentioned studies (Hall-Beyer, 2005, Hirosawa et al. 1996, Eastman and Fulk, 1993) found that the first principal component (Y_1) represented 'accumulated greenness', as Hirosawa et al. (1996) call it, or vegetation density. The corresponding coefficient had positive, consistent elements over the entire period; Y_1 thus resembled the temporal mean NDVI. Y_2 was found to represent the main seasonal variability of vegetation in all of the above mentioned studies, except for Hall-Beyer (2003), where seasonality was captured by Y_3 . Succeeding components discriminate smaller areas which have a particular seasonality in common; interannual differences caused by e.g. climatic anomalies; sensor degradation. Later components can capture anomalies in smaller areas or short time scales, isolating potential noise, but also indicating small areas behaving differently from their surroundings (Hall-Beyer 2003, Eastman and Fulk 1993). Depending on the length of the time series, change of the land surface is found to be uncovered as well (Hirosawa et al., 1996).

Applying PCA to spatial subsets with shared characteristics (e.g. equal land cover), more detailed information about these areas can be derived (Hirosawa et al., 1996). Also, when instead of multiyear a single year analysis is performed, annually changing temporal patterns in vegetation (agricultural practice) are likely to be distinguished.

4.2.3 Application to the Ourthe Catchment

The time series used in this study consists of all available good quality quasi 8 day composite MODIS NDVI images, clipped for the Ourthe catchment only, for the years 2000-2006. Referred is to Section 3.3 for further details on the data. In total 128 images are used, with 10, 11, 17, 28, 21, 25, respectively 16 images for the individual years.

The first six principal components and associated coefficients are analysed. For interpretation the spatial (in the images of Y) and temporal (in the coefficients Γ) patterns are

visually compared with known geographical (land cover, elevation, aspect, slopes, geology) and climatological patterns in the catchment. This data can be found in Chapter 3. Furthermore, the land cover specific NDVI patterns are used in the interpretation of the observed patterns. These patterns are simple statistics of the NDVI data per land cover type.

4.3 RESULTS AND DISCUSSION

In this section first some statistics of the NDVI time series are shown, giving insight in the gross temporal and spatial variability and its sensitivity to land cover. Subsequently the results of the PCA are shown and discussed.

4.3.1 NDVI statistics

The unsmoothed, quality checked time series for the years 2000-2006 of the mean NDVI for the entire catchment and per land use class for the vegetated areas are shown in Figure 4.2 and Figure 4.3, respectively. In Figure 4.2 the seasonal pattern with high NDVI in summer (max = 0.85) and low NDVI in winter (min = 0.55) is clearly visible. In summer in most years there is a small dip in NDVI between June and August. The month in which maximum NDVI is reached is rather variable over the years, between May and September. 2000 and 2003 deviate from the other years, with maximum NDVI early in the growing season. Although the total range of NDVI values is large (about 0.9), the areal extent of non-vegetated areas (having low NDVI values) is small and the range of values of 95% of the pixels is in the order of the seasonal range of the spatial mean NDVI (0.3). Looking at the vegetation specific NDVI, the top panel in Figure 4.3 compares the NDVI of different forest types. The main difference between deciduous broadleaf (311, green line) and coniferous forest (312, yellow) clearly is the amplitude of the seasonal pattern, with higher NDVI for deciduous broadleaf forest in the growing season and lower NDVI in winter. The summer time NDVI of transitional woodland-shrub (324, orange/red) increases over the seven year period and approaches the NDVI pattern of mixed forest (313, orange) towards the end of the period. The middle panel shows the mean NDVI for the agricultural areas. Permanent pastures (231, green), agriculture with complex cultivation patterns (242, cyan) and agriculture with significant areas with natural vegetation (243, blue) have - for the spatial average - a similar temporal NDVI pattern, with a dip in June, July or August. The mean NDVI of non-irrigated arable land (211, red) is much lower in all seasons, and the temporal pattern is different from that of the other agricultural areas, with generally lower values in the end of the growing season. The main difference between the agricultural classes is the extent of the area under a rotation system (>75% for 211, <75% for 24x), the area with permanent crops and natural vegetation (<25% for 211, <75% for 243, undefined for 242) and urban fabric (up to 30% for 24x, sporadic for 211), see Appendix B. In the bottom panel the temporal pattern of permanent pastures (231, green) is compared with natural grasslands (321, orange) and green sports and leisure facilities (142, blue). Notable is the much lower values for natural grasslands at the end of the growing season and in winter. The natural grasslands in the catchment are concentrated in the Famenne, north-east of Marche-en-Famenne and originate from abandoned pastures (Ministère de la Région Wallonne, DGRNE) and are (partly) used as military training field. The differences between pastures and natural grasslands can be attributed to vegetation density and species (in summer NDVI doesn't discriminate the two classes), vegetation species, and grazing or mechanical harvesting in pastures (231). The steep dip in summer NDVI in 2006, visible even for the forests, might be attributed to the extremely hot and dry July month (see Section 3.1.1).

In Figure 4.4 the spatial variability of the multi year (2000-2006) mean and standard deviation of the NDVI are shown. Figure 4.5 gives the seasonal differences, based on the same period. Compare the spatial pattern in mean NDVI with the main topographic features and land cover in Figure 4.6. Notice that the spatial resolution of the NDVI product allows for the discrimination of the major roads in the catchment, visible as the relatively straight yellow lines (low NDVI) in the figure of the mean NDVI. The big green/blue spot in the lower north-western part of the catchment is the city Marche-en-Famenne. Areas with a relatively low mean NDVI and high standard deviation are associated with non-irrigated arable land (crop rotations and temporal pastures), visible mainly in the northern

The numerical code for the land cover classes refers to the CORINE land cover classes. In Appendix B a specification of the classes is given.

The class 'non-irrigated arable land' seems to imply that other agricultural areas are irrigated. This is unlikely to be practised in areas similar to the Ourthe. Pers. comm. G. Regele (2014)

part of the catchment as blue-ish to yellow spots in the mean NDVI figure. Relatively high mean NDVI, but especially high standard deviation, occurs for deciduous forests, see the relatively large red to yellow spots in the figure of the standard deviation. Lastly, needleleaf forest is generally recognizable by a high mean NDVI with relatively low temporal variability. In Figure 4.5 needleleaf and broadleaf forests are clearly distinguished in the winter (highest values for needleleaf forest) and summer (highest values for broadleaf forest) figures. Generally, the main land cover classes can be distinguished in the (seasonal) mean NDVI figures. The spatial heterogeneity has a rather small scale.

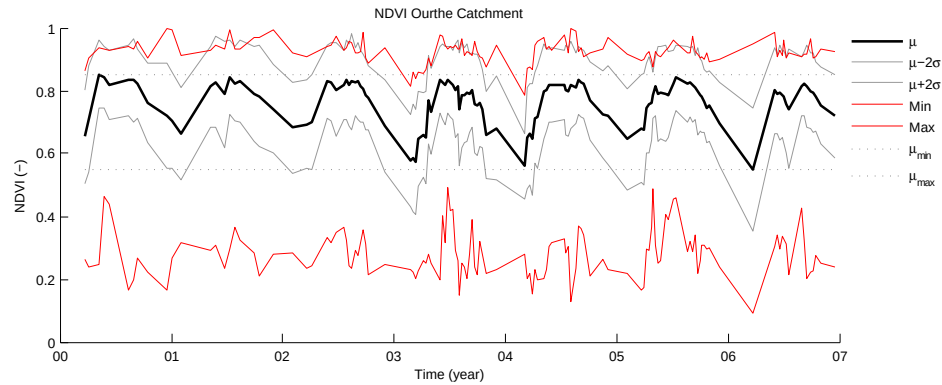


Figure 4.2: NDVI time series for the Ourthe catchment for the period 2000-2006. Shown are the spatial mean NDVI (μ , thick black line), the ca. 95% confidence bounds ($\mu \pm 2\sigma$, in grey) and the spatial minimum and maximum NDVI values (red lines). Furthermore, the grey dashed lines show the minimum and maximum values of the catchment average NDVI, shown as a reference in Figure 4.3.

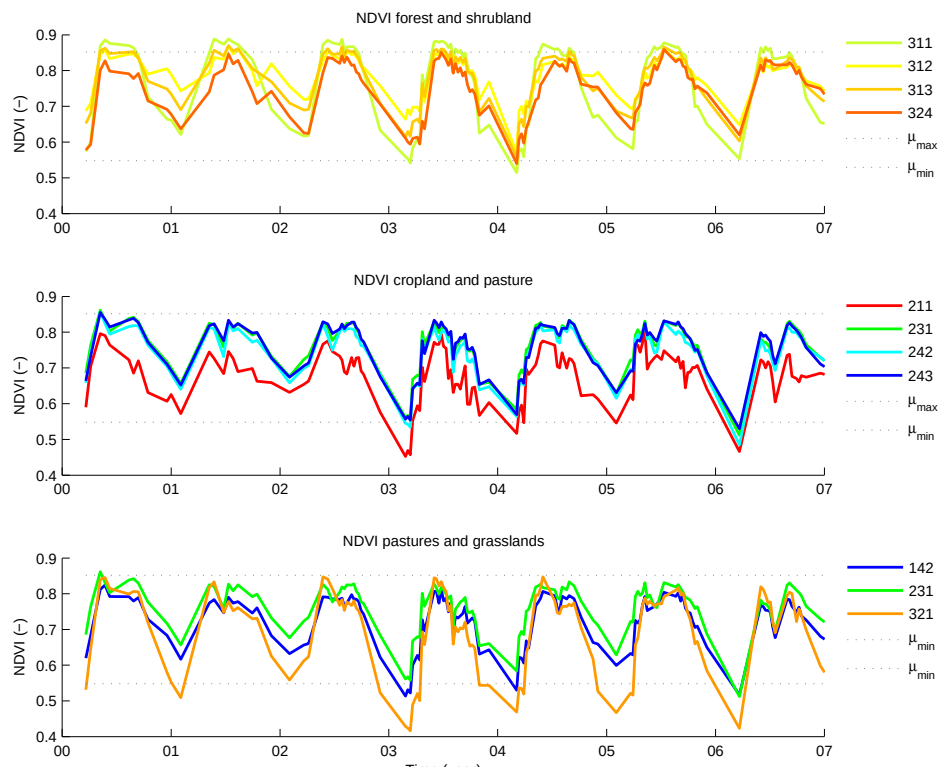


Figure 4.3: Mean NDVI per land use class for the period 2000-2006. Top: forest and transitional woodland. Middle: agriculture. Bottom: pastures and grasslands. The gray dashed lines mark the minimum and maximum NDVI values of the catchment average NDVI.

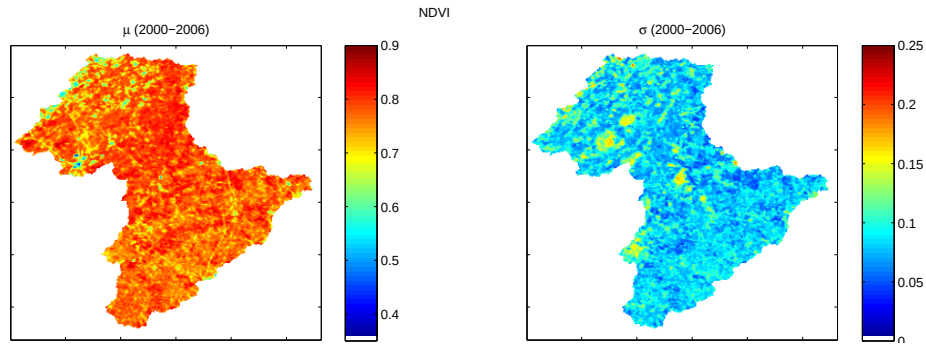


Figure 4.4: NDVI statistics of the Ourthe catchment: mean (left) and standard deviation (right) of the NDVI over the period 2000-2006. The NDVI data that is used is the nan-filtered, quality checked data.

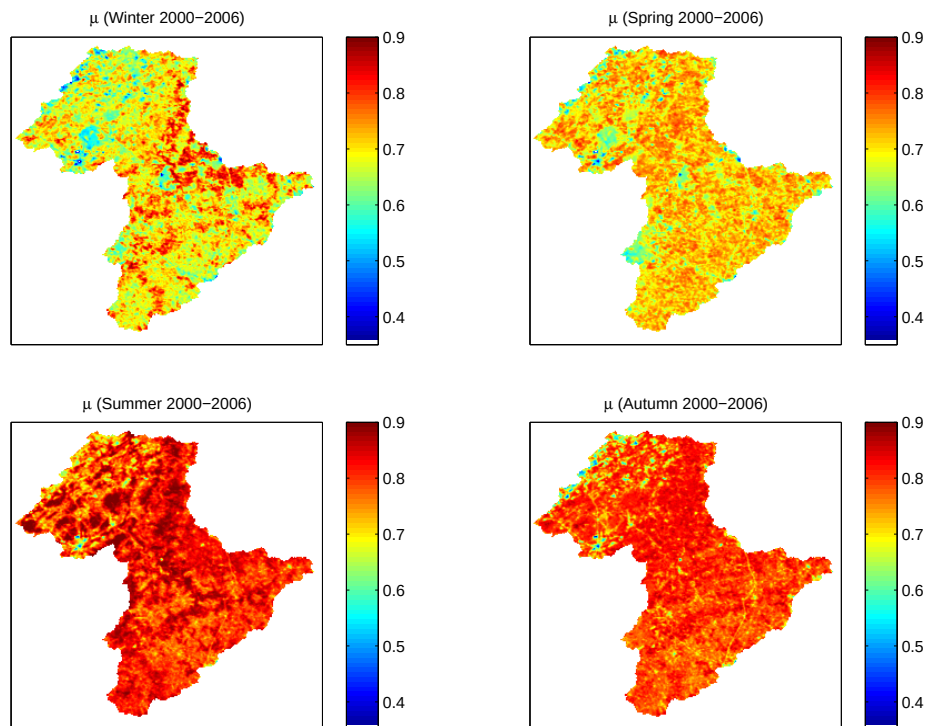


Figure 4.5: Seasonal mean NDVI in the Ourthe catchment for the winter, spring, summer and autumn period for the years 2000-2006.

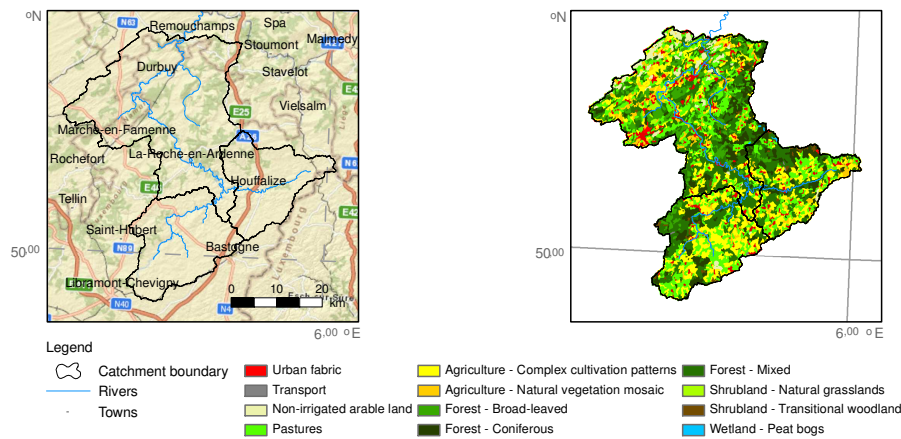


Figure 4.6: Characteristics of the Ourthe catchment. Left: topography, right: land use, repeated from Chapter 3

4.3.2 Principal component analysis

The first six principal components ($Y_1 - Y_6$) and associated coefficients (weighing factors $\gamma_1 - \gamma_6$) of the multiyear (2000-2006) data set are shown in Figure 4.7 and Figure 4.8, respectively. Interpretation of the components is performed based on the comparison of the temporal pattern of the coefficients with the temporal pattern in NDVI, and of the spatial pattern of the images of the principal components with the spatial patterns of the main land use classes, geology and topographical features in the catchment. The latter are shown in Figure 3.2. The value per pixel for principal component Y_i (i.e. the elements in $Y_i = (y_{i,1}, \dots, y_{i,n})$) is called the score. For the six principal components, the distribution over the land cover classes is given for pixels with a very high ($y_{i,j} \geq 2\sigma$), high ($\sigma \leq y_{i,j} < 2\sigma$), medium ($-\sigma < y_{i,j} < \sigma$), low ($-2\sigma < y_{i,j} \leq -\sigma$) and very low score ($y_{i,j} \leq -2\sigma$), with σ the standard deviation of the elements of Y_i (the mean of the elements is zero by definition), see Figure 4.14-Figure 4.19, right panel. Although the histograms are somewhat distorted by the uneven distribution of all pixels over the land cover types (i.e. for some land cover classes there are only a few pixels, so they won't appear clearly in the histograms), the comparison of the histograms over all score subgroups, does show which land cover types dominate which scores, if any. Additionally, the distribution of the scores over the pixels per land cover class is given for the first six principal components in Appendix C, Figure C.1-Figure C.6, showing the land cover dependency from a different perspective. In the left panel of Figure 4.14-Figure 4.19, the normalized coefficient is compared with the normalized time series of the mean NDVI of all pixels, and of the pixels with a very high and a very low score, to explain which aspects of the NDVI pattern are distinguished. The variables are normalized using z-scores for comparison. The selection for the subgroups ($y_{i,j} \geq 2\sigma$ and $y_{i,j} \leq -2\sigma$) allows for the discrimination of the most extreme patterns, but it should be noted that it concerns less than 5% of the pixels, and is not always the optimal set.

Coming to the interpretation of the results, the first six principal components together explain about 65% of the variation of the 128 NDVI images. The spatial patterns of the first three principal components (Y_1 - Y_3) show a strong relation with the main land cover classes, which can be seen by comparing Figure 4.7 and Figure 4.6. This also appears from the histograms in Figure 4.14a to Figure 4.16a. The three lower components (Y_4 - Y_6) seem to capture differences within the land use classes mainly, as visible in the histograms in Appendix C, where the distribution of the scores (y_i) per land cover class is centered around zero for all classes.

The spatial pattern of Y_1 and Y_2 agrees well with the mean, respectively the standard deviation of the NDVI: compare Figure 4.4 and the two top left figures in Figure 4.7. Together they explain more than 50% of the total variability of the NDVI data. As described in Section 4.3.1 the mean NDVI and thus Y_1 mainly distinguishes between differences in vegetation cover and/or canopy density, as was found by e.g. Hall-Beyer (2003), Hirose et al. (1996) and Eastman and Fulk (1993) as well. Forested sites have the highest scores, non- or sparsely vegetated areas have low scores. The temporal fluctuations of γ_1 , the coefficient of the first principal component, indeed are small, meaning that all time steps have approximately the same weight.

The second and third principal components are associated with differences in seasonality between land use types. Y_2 discriminates between areas with a large difference in summer and winter NDVI (high negative Y_2 score) and areas with a different (high positive Y_2 score) or less strong (small negative Y_2 scores) seasonality. As written, the spatial pattern of Y_2 scores resembles the spatial pattern of the standard deviation of the multi-year NDVI. The temporal pattern of the coefficient, mirrored in the x-axis, agrees well with the temporal pattern of the mean NDVI of pixels with a low Y_2 score. This is mainly deciduous broad-leaved forest (311) and the parts of mixed forest (313) where broad-leaved trees dominate the signal, see Figure 4.15b. Natural grasslands (321), transitional woodland - shrub (324)

*mirrored because here the
coefficient corresponds with
negative scores*

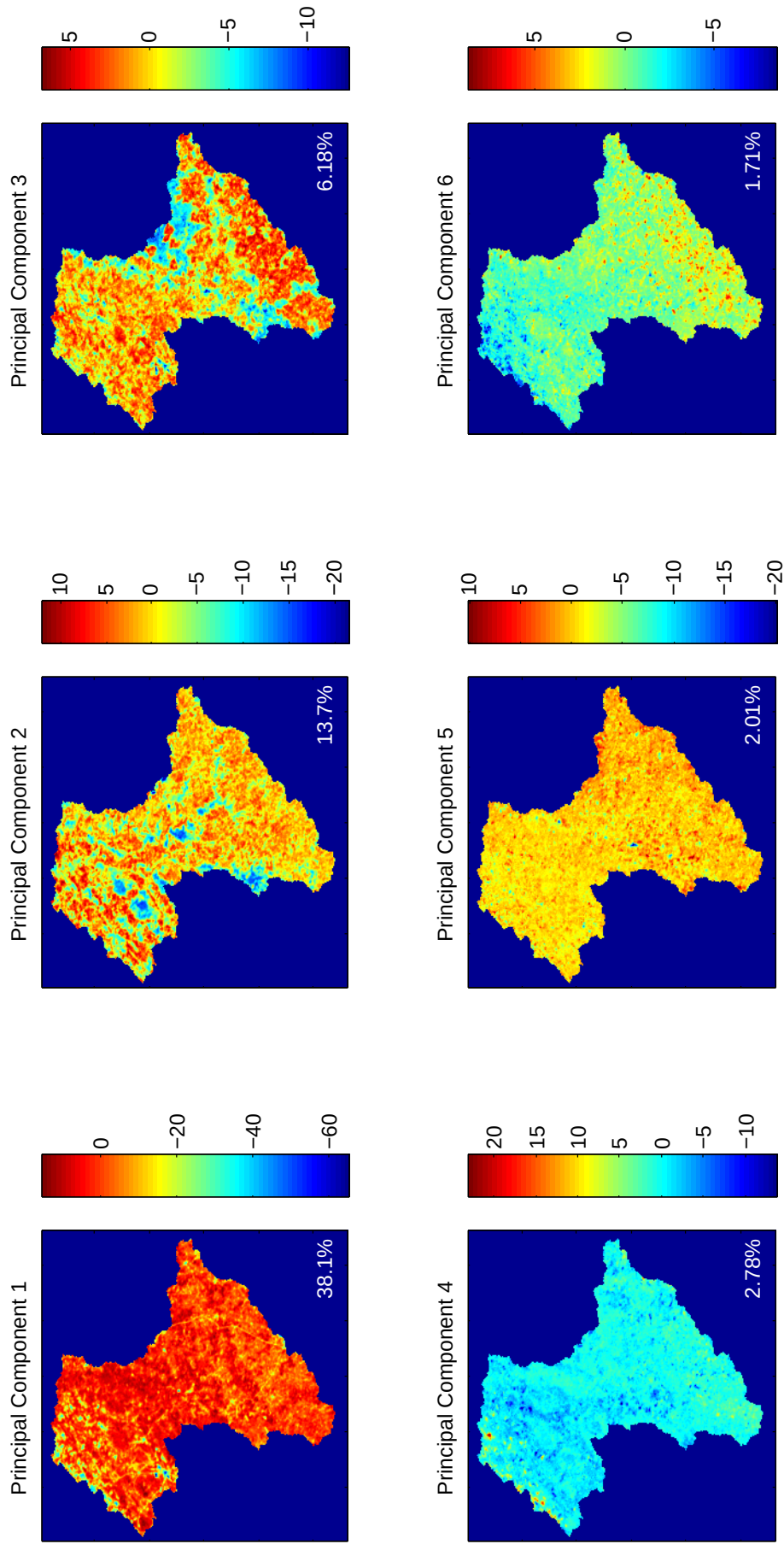


Figure 4.7: Principal component scores of the first six principal components for the multi year (2000-2006) NDVI data set. The scores themselves have no physical meaning, but areas with similar scores relate more closely to each other than to other areas in the catchment. In other words: areas with similar scores for a certain component have those aspects of the NDVI pattern that are captured by the specific component in common.

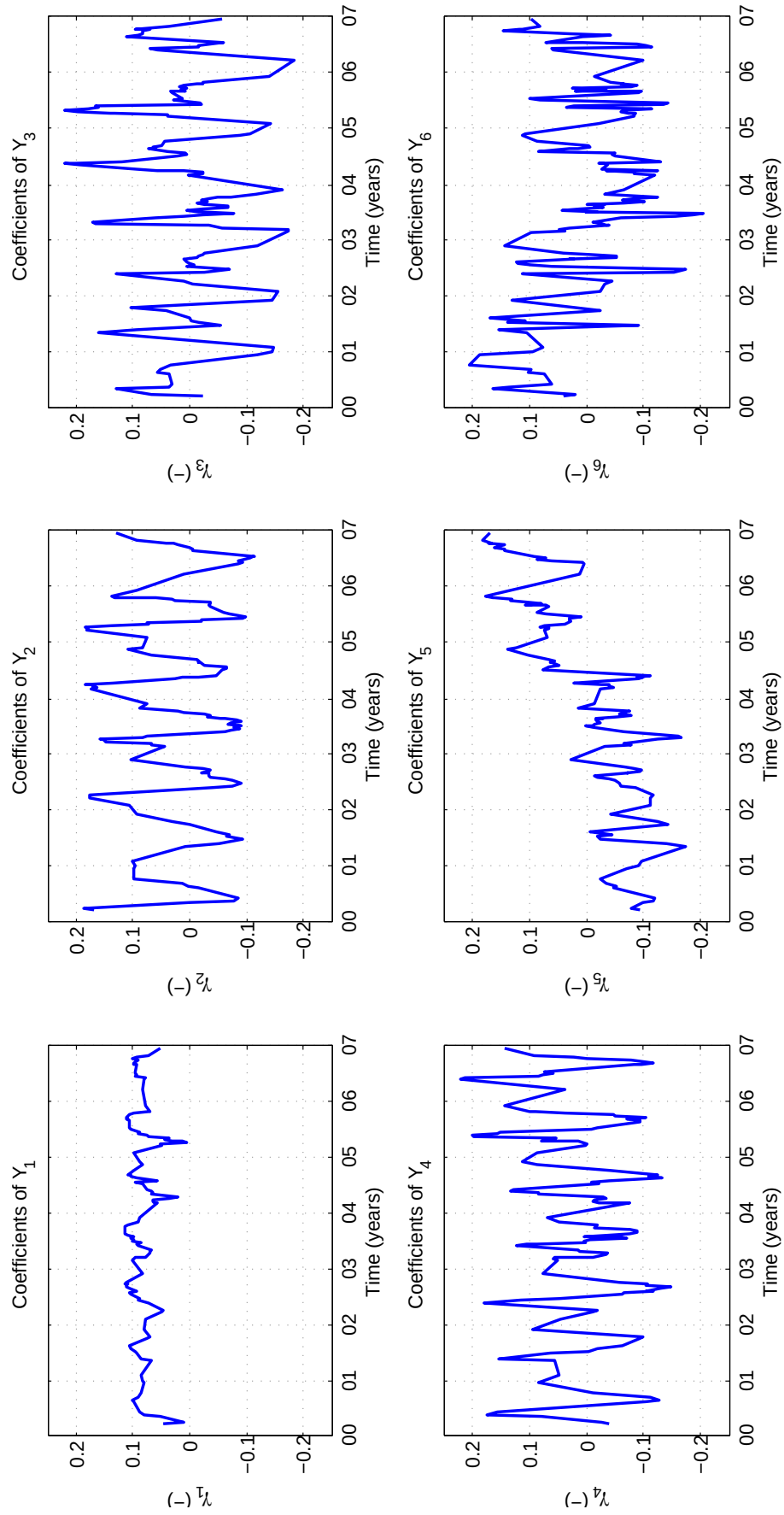


Figure 4.8: Coefficients of the first six principal components of the multiyear (2000-2006) NDVI data set.

and wetland (412) have dominantly negative Y_2 scores as well, yet these land use types have a small share in the total land cover. The pixels with a high positive score are parts of the agricultural area, especially grassland (231) and heterogeneous agriculture (242 and 243). The mean NDVI pattern of these pixels is characterized by a dual seasonality, with increasing NDVI earlier in spring and later in autumn and a minimum in mid summer (around July), see the middle figure in Figure 4.15a. The difference between the temporal behaviour of the entire catchment and the low negative scoring pixels (and γ_2) is small.

The third principal component distinguishes between areas predominantly occupied with agriculture (high Y_3 score) and areas with coniferous forest and transitional shrubland-woodland (low Y_3 score). The coefficient shows a dual seasonality, with the first peak in May/June and a second peak in August / September, agreeing with the NDVI pattern of the highest scoring pixels (see Figure 4.16a). The peak in May/June is generally much higher than the second peak. 2003 deviates in the sense that there are three peaks. 2006 deviates as well, with a later (due to data availability) and larger minimum in March and a smaller subsequent spring peak value.

All agricultural classes are represented in the highest scoring pixels, supplemented with natural grassland (321) and deciduous forest (311). Agriculture was importantly present in the high scoring pixels in Y_2 as well, yet here accompanied by pixels with coniferous forests. In Figure 4.9 the temporal NDVI pattern of high scoring pixels in Y_2 and Y_3 are compared. The timing of the mean seasonality of the two groups is similar, but the amplitude is generally smaller for pixels with a high Y_2 score (black line) than for pixels with a high Y_3 score (red line). Especially in winter the NDVI is lower for high scoring Y_3 pixels, and the spring maximum is higher. In summer 2003 and 2005 - dry years, different crops in 2003 - the differences are smaller. If we look at the spatial distribution of the high Y_2 and high Y_3 scores within the agricultural land cover classes (Figure 4.11), a rather clear partitioning is visible. Pixels with high Y_2 scores and coinciding smaller amplitude in the seasonal pattern of NDVI are located in the north-western part of the catchment; pixels with high Y_3 scores (higher amplitude) in the south-east. The pattern exists for the individual agricultural subclasses as well, including permanent pastures. The same analysis we can perform for deciduous forests with dominantly low Y_2 scores and high Y_3 scores. From this it appears that broad-leaved forest in the more downstream area has a higher amplitude in the seasonal NDVI pattern than broad-leaved forest at the higher elevated areas in the Ardennes. Differences are relatively small though, and probably related to overstorey and/or understorey composition. The NDVI time series of the low scoring pixels of Y_3 has a single seasonality with relatively low amplitude, lagging behind the pattern of the coefficient with about two months (apart from the deviating spring of 2006), see Figure 4.16a.

The fourth principal component merely distinguishes between pixels within the same land use class, with the highest contrast in the agricultural classes: 211, 231, 242, 243, in leisure facilities (142) and to a lesser extent in mixed forest (313). The coefficient shows a dual season with high weights from late spring to early summer, low summer values, and the second peak in winter, see Figure 4.17a. Comparing the coefficient of Y_3 and Y_4 , the timing of maximum and falling weights in spring/early summer is approximately equal (Y_4 lags a little behind), but during the rest of the year the pattern of Y_4 is opposed to the pattern of Y_3 . As such, the fourth principal component distinguishes areas (agriculture mainly) with maximum NDVI early in the growing season (between the second week in May and the first week in June $\pm$ one week) from areas with maximum NDVI later in the growing season (between the last week in June and the first week of September $\pm$ one week) or with a dual seasonality. The time series of the NDVI of the pixels with a high, respectively low Y_4 score is given in Figure 4.10. There is a strong geographical pattern in the Y_4 scores (see for entire catchment: Figure 4.7, for agriculture: Figure 4.11), with dominantly higher scores along the northern boundary of the catchment, that is in the Condroz region, and along the boundary in the south(-east). Areas with low Y_4 scores are located more or less in the middle of the catchment, at lower elevations. This contrast in mainly agricultural land most likely indicates differences in crop types and related crop emergence and yellowing

or harvest dates. Since crop rotation is applied in the Ourthe catchment and especially in the Condroz region (land cover class 211 has a rotational system in $> 75\%$ of the area), the relatively regular pattern in NDVI is surprising. The early, single season agrees with the crop development of winter wheat. The single season later in summer could be any other crop that is generally cultivated in the Ourthe region. The temporal pattern of the latter is less regular and might reflect different successive crops. The geographical distribution of crop types is potentially related to elevation (higher Y_4 values in the lower laying areas) and therewith temperatures and suitability of the land for certain types of agriculture. Available - coarse - information on harvested crops in this region (see Chapter 3) does not confirm an elevation dependent pattern however.

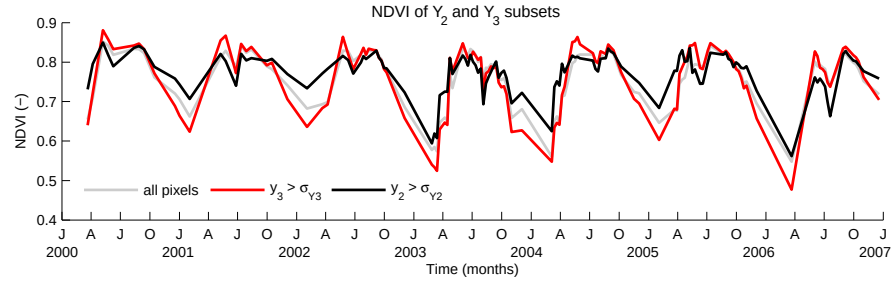


Figure 4.9: Comparison of the mean NDVI of pixels with a high Y_2 score ($y_2 > \sigma_{Y2}$), a high Y_3 score ($y_3 > \sigma_{Y3}$) and the mean NDVI of the entire catchment. The geographical pattern is shown in Figure 4.11.

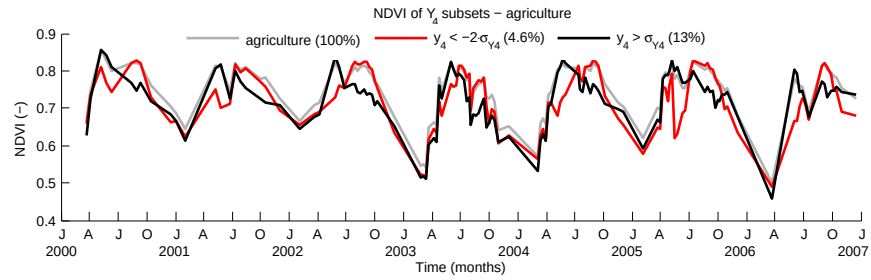


Figure 4.10: Comparison of the seasonality of the NDVI of Y_4 subsets for agriculture. Shown are the mean NDVI of pixels with $y_4 > \sigma_{Y4}$ (red), pixels with $y_4 < -2 \cdot \sigma_{Y4}$ (black) and the mean NDVI of all agricultural areas (grey). The geographical pattern is shown in Figure 4.11.

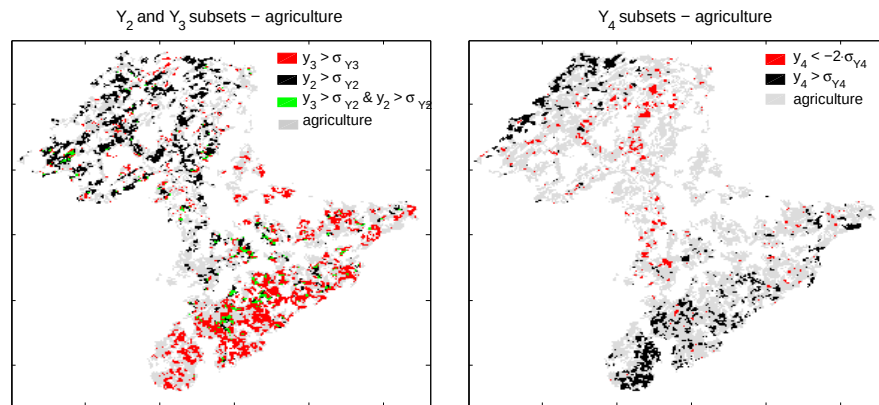


Figure 4.11: Spatial pattern of agriculture subsets. Left: distribution of Y_2 (black) and Y_3 (red) subsets with scores $> \sigma$, for agriculture (grey). In green pixels with both $y_2 > \sigma_{Y2}$ and $y_3 > \sigma_{Y3}$ are shown. Right: distribution of Y_4 subsets with $y_4 > \sigma_{Y4}$ (black) and $y_4 < -2\sigma_{Y4}$ (red), having a single season. The north-western part of the catchment generally has a smaller seasonal amplitude in NDVI than the south-eastern part of the catchment, where the elevation is higher.

Y_5 merely distinguishes between pixels within the land use classes as well, except for transitional woodland (324) and wetland (412), which have only scores in the higher reach, and urban fabric (112), dominant in the medium range Y_5 values. The coefficient γ_5 shows an increasing trend, which is also visible in the NDVI pattern of pixels with positive Y_5 scores (412, 324 completely, parts of 313, 312, 242, 231, 211). Figure 4.12 shows the mean NDVI of all pixels and of the subclasses with low, respectively high Y_5 scores in one figure to allow for a better comparison. For woodland the majority of the pixels have a rather stable summer and winter NDVI over the seven year period. A subset of this class, coinciding with the highest Y_5 scores, has an increasing summer NDVI over the period 2000-2004, which is stabilized in the last years of the study period. Comparing 30m resolution Landsat images in different years, these pixels (red pixels in Figure 4.7e.) indeed seem to develop to more forest-like areas with higher NDVI. For wetland the NDVI is stable in all years, and the spread in Y_5 scores is relatively small. Also the highest scoring wetland pixels do not show an obvious trend. For pixels with a high negative score (all vegetated land use classes are represented in the $<-2\sigma_{Y_5}$ subset apart from natural grassland and wetlands; forests have the highest share) an opposite trend is visible. High negative scores are isolated in relatively small spots (blue in Figure 4.7) and seem to be deforested in the 2000-2006 period based on the Landsat images. These spots represent 2% of the total forested area.

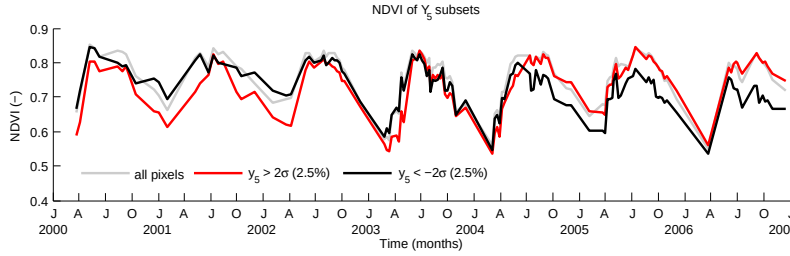


Figure 4.12: Comparison of the mean NDVI of the entire catchment and pixels with a high (2.5% of all pixels) and low (2.5%) Y_5 score.

Remarkable for the sixth principal component is the dominance of heterogeneous agriculture (242) for high Y_6 scores, located in the south-east, and arable land (211) for low Y_6 scores in the north. The coefficient fluctuates around a longer term trend of decreasing weights in the period 2000-2003, then has relatively high weights at the end of 2004, low weights at the end of 2005 and high again in 2006. The mean NDVI pattern of the agricultural area in the catchment deviates in 2003 (see above). High scoring pixels in Y_6 have even lower late summer NDVI compared to other years. Two explanations can be given: 1) a different crop was cultivated in this year, namely sugar beet in stead of fodder maize. 2) 2003 was an anomalous warm year, with high temperatures in spring and summer. Compared to 2001 and 2002 it was dry as well. This can have a different effect on different crop types, or the anomalies are less severe in certain regions.

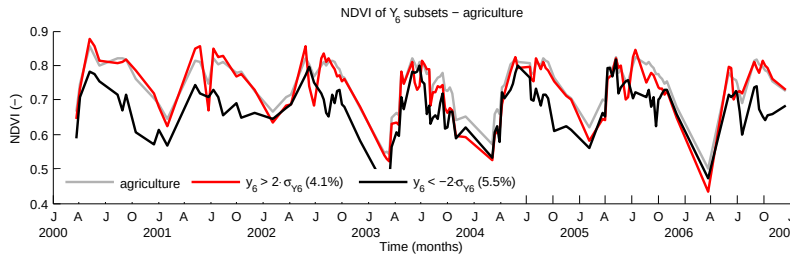


Figure 4.13: Comparison of the NDVI of Y_6 subsets of agriculture. Shown are the mean NDVI of pixels with $y_6 > \sigma_{Y_6}$ (red), pixels with $y_6 < -2 \cdot \sigma_{Y_6}$ (black) and in grey all agricultural pixels.

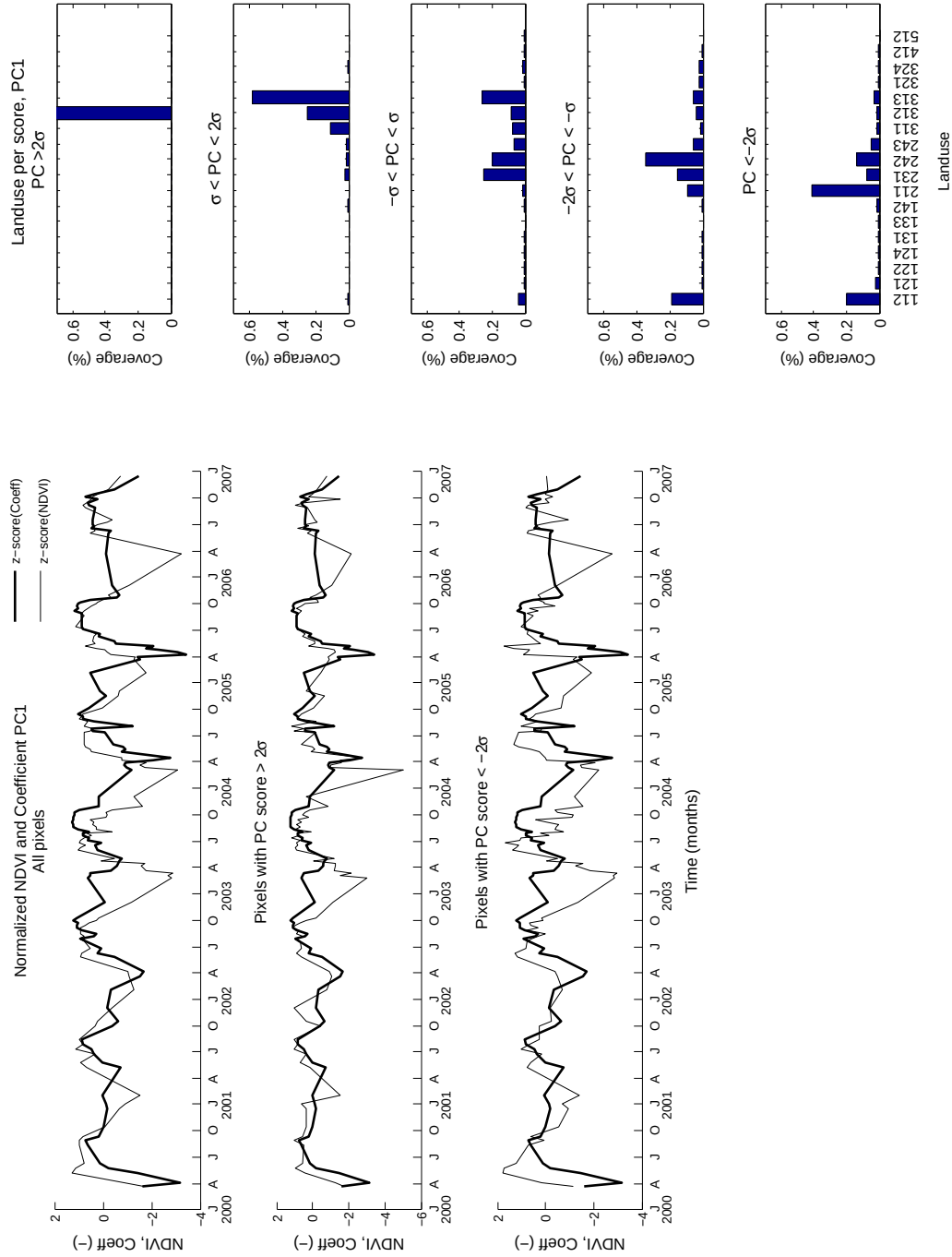


Figure 4.14: Characteristics per Y_1 score subclass. Right: comparison of the normalized coefficient of PC1 and temporal pattern of the mean NDVI for the entire catchment (top), mean NDVI for the pixels with a Y_1 score $> 2\sigma$ (middle) and pixels with a Y_1 score $< -2\sigma$. Left: histograms showing the distribution of the pixels within a PC score subclass over the land use classes. The fraction (-) is shown on the y-axis (shown maximum = 0.6), the code for the land uses is given on the x-axis. 1** = artificial land cover, 2** = agricultural land cover, 31* = forest, 32* = shrubland, 412 = wetland and 512 = open water. The full legend is given in Section 3.1

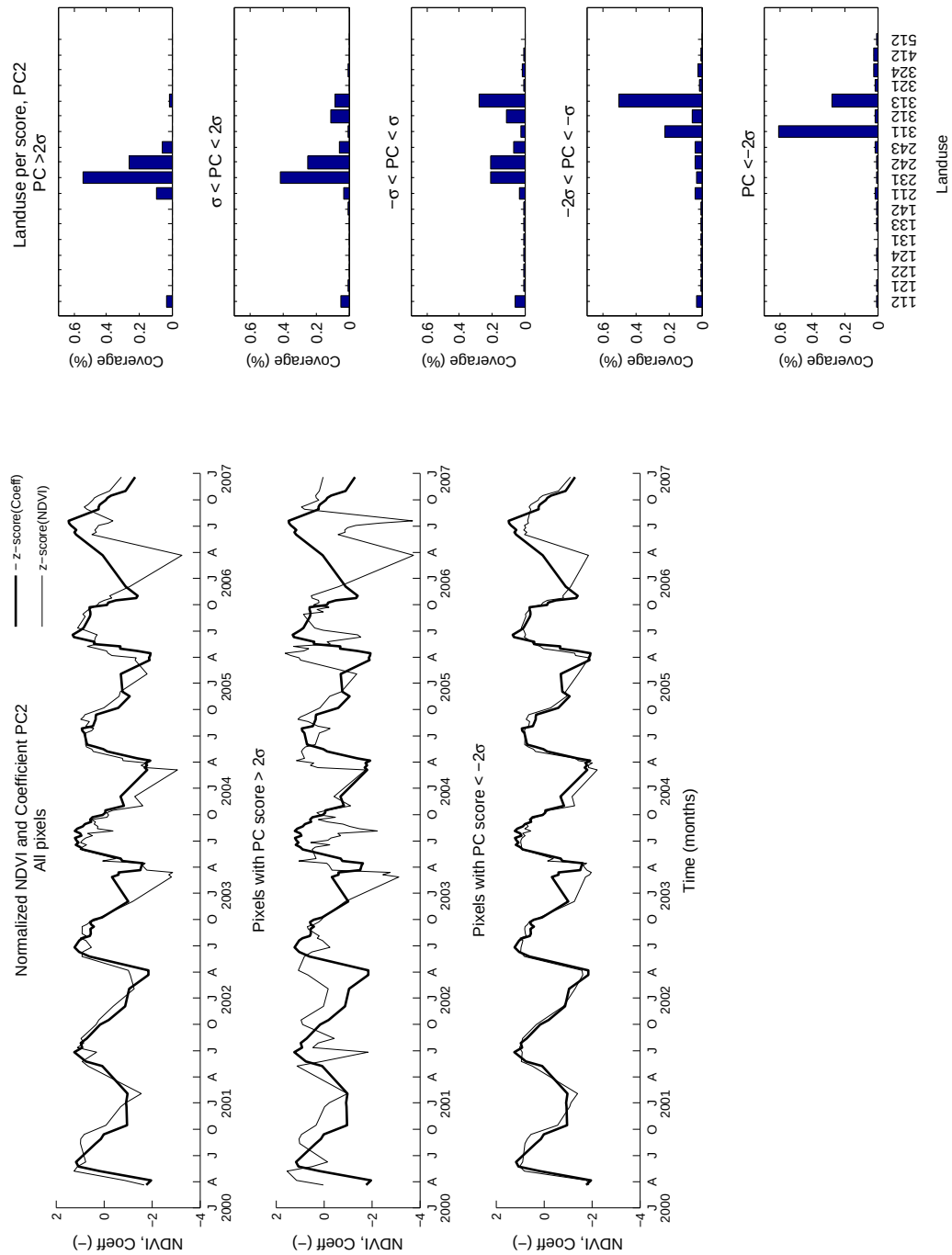


Figure 4.15: Characteristics per Y_2 score subclass.

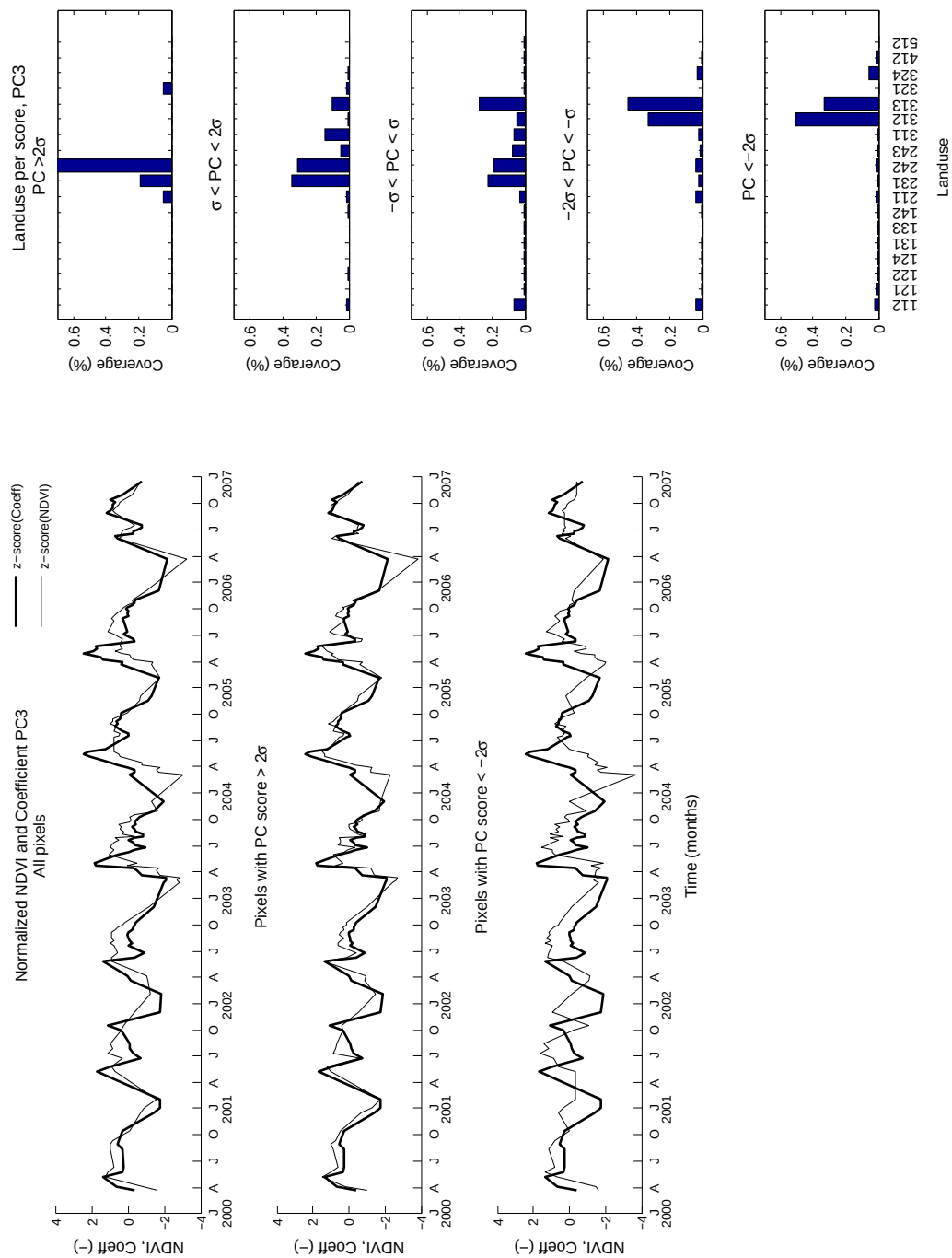


Figure 4.16: Characteristics per Y_3 score subclass.

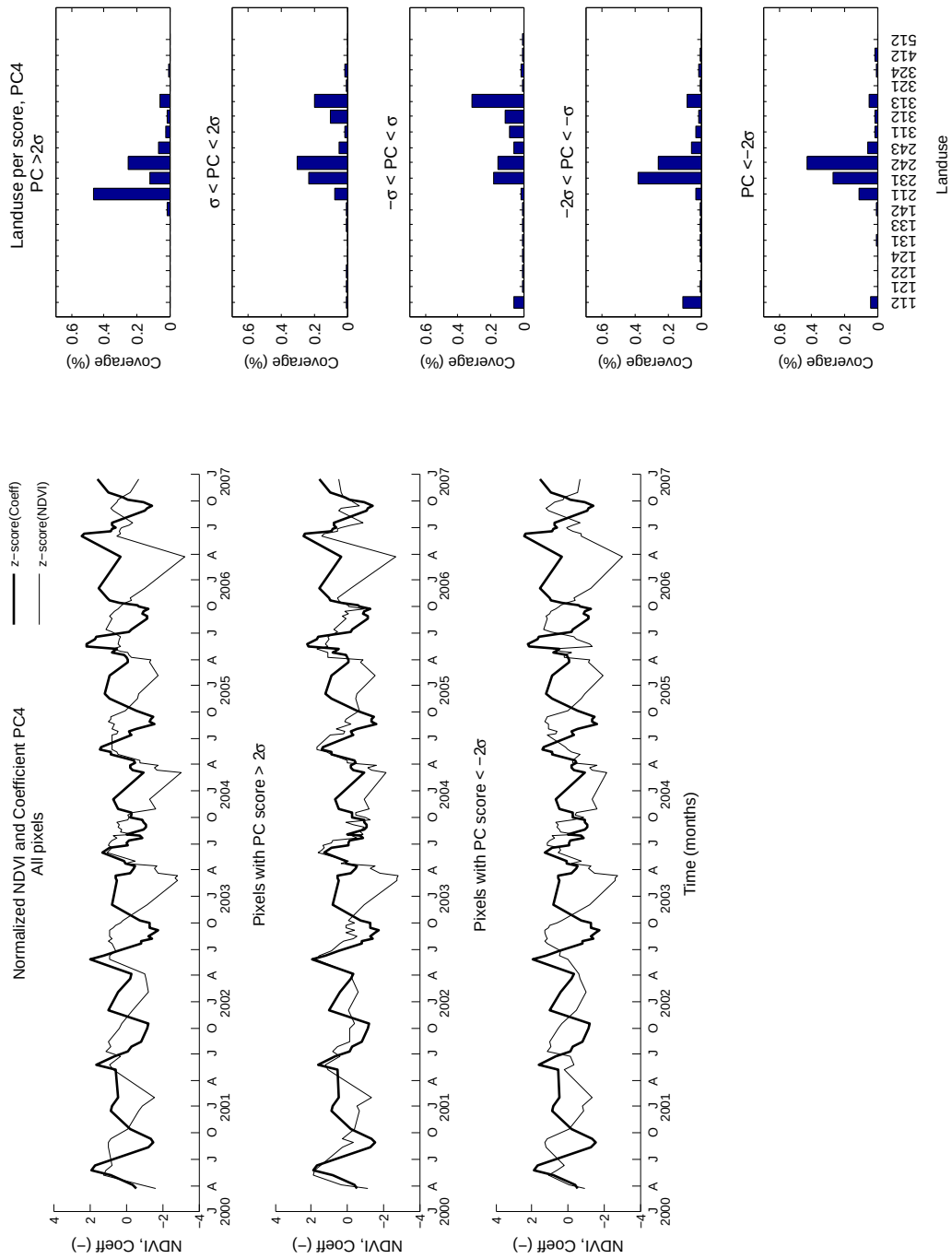


Figure 4.17: Characteristics per Y_4 score subclass.

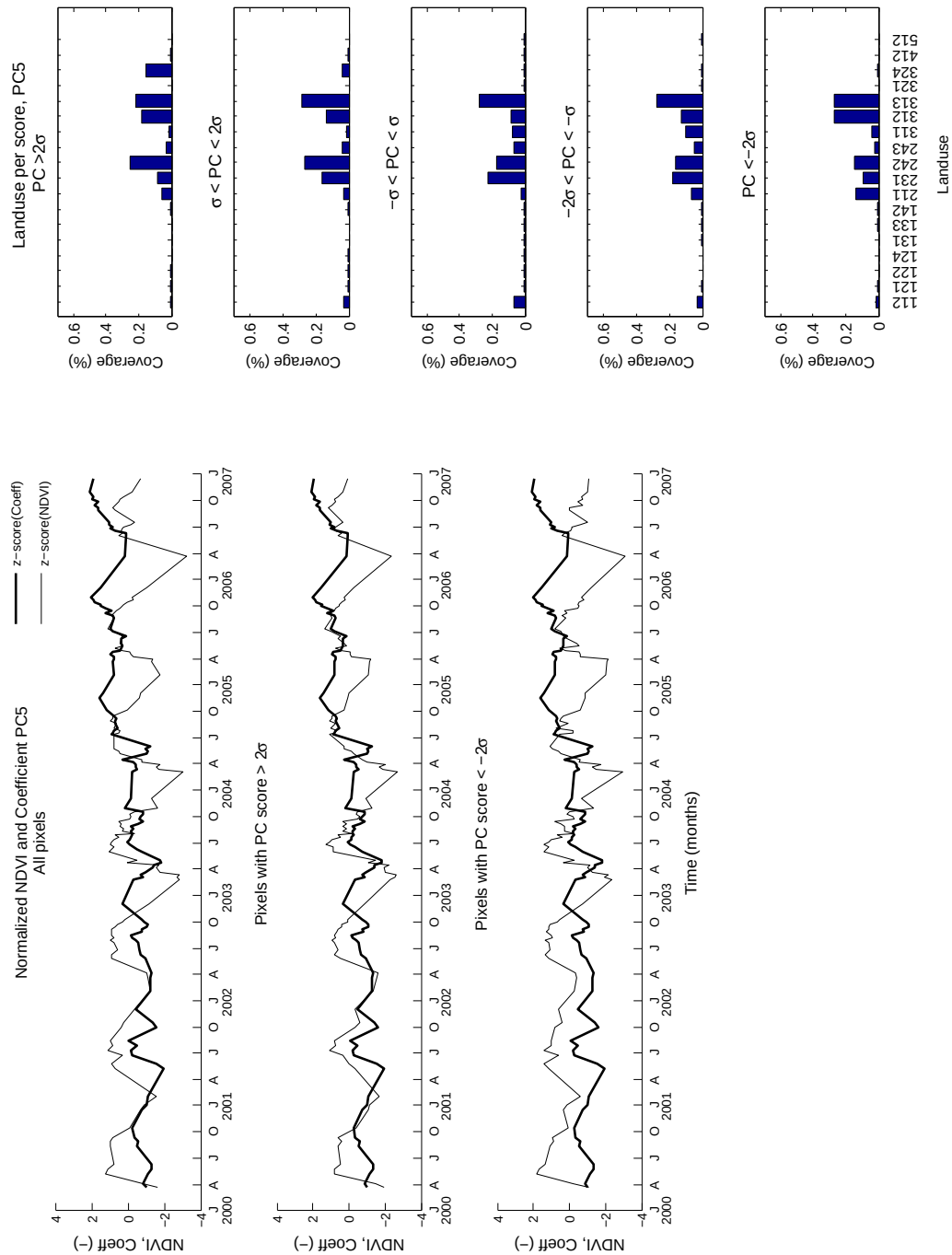


Figure 4.18: Characteristics per Y_5 score subclass.

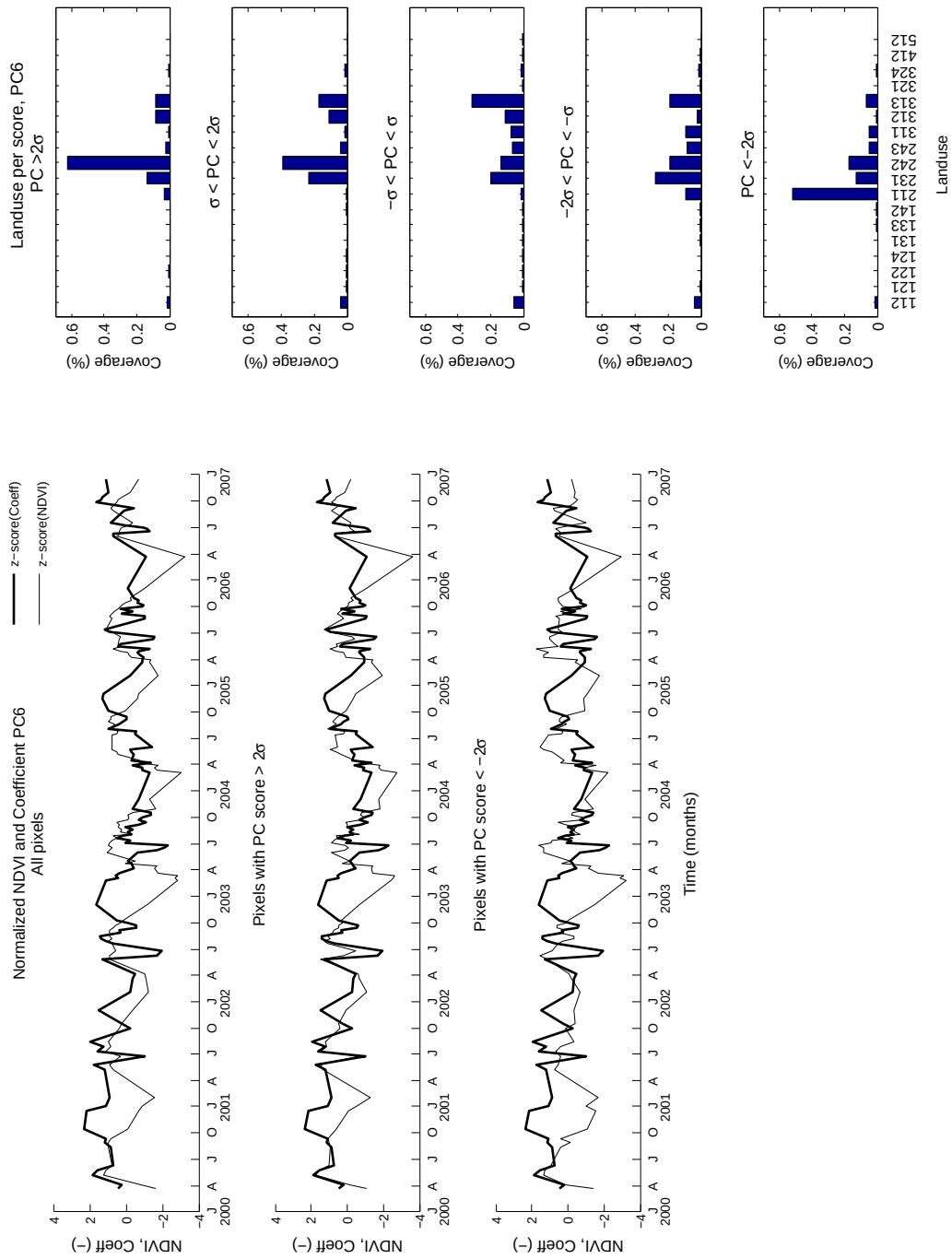


Figure 4.19: Characteristics per Y_6 score subclass.

4.4 CONCLUSION

From the principal component analysis differences in vegetation dynamics between as well as within the CORINE land cover types became apparent. The main difference in the temporal dynamics in the Ourthe catchment exists between deciduous vegetation (broad-leaved forests, natural grasslands, parks) and agricultural crops and pastures. The 'green up' from the first subset generally sets in later in the growing season and senescence starts earlier than is the case for the agricultural plots. The 'average' behaviour of the agricultural pixels, comprising permanent pastures; relatively homogeneous patches with annual crops under rotation; and patchworks of smaller areas with annual or permanent crops, pastures or natural vegetation, is a dual season, with a first peak in May/June and a second in August/September. Within the subset of agricultural pixels, areas in the downstream part of the catchment generally have a smaller seasonal amplitude in NDVI (smaller spring peak, higher winter values) than areas in the upstream part of the catchment, as appeared from comparison of the second and third principal component. Furthermore, near the northern boundary (Condroz region) and in the south-west of the catchment, agricultural areas with a single season are located, with a maximum NDVI relatively early in the season (May/June). Since no detailed information is available on crop types, sowing and harvest dates, the exact reason for the differences is not known. It seems to - at least partially - be related to elevation, yet whether crops behave differently at higher elevations or whether there is a geographical relation with crop types is not known.

For broad-leaved forest there is a geographical difference in the temporal NDVI pattern as well, appearing from the second and third principal components. Forests in the downstream part of the catchment generally have a higher amplitude in NDVI, mainly due to low winter values, than deciduous forest in the higher areas of the Ardennes. The timing of bud burst, leaf development and senescence is equal, that is, within the same 16 day period of the NDVI composite.

Concerning longer term trends or changes in the vegetation in the catchment, slowly increasing NDVI is observed for a subset of transitional woodland - shrub. Furthermore clear cuts were captured of relatively small areal extent (2% of the forested area).

The spatial resolution of 250 m allows for the discrimination of the main land cover types and the larger agricultural parcels. Where agricultural land is mixed or bordered with small patches of natural vegetation, the NDVI signal will be mixed. A larger limitation of the analysis is the temporal accuracy of the NDVI time series, i.e. 16 days. Information on the actual retrieval dates is available but was not used in the analysis for practical reasons. Timing differences between crop types are probably captured, yet differences in phenological events due to elevation and therewith temperature gradients in the catchment may not be.

In the next chapter the remotely sensed evaporation products are validated. Comparison of the geographical patterns of temporal differences in NDVI with patterns in evaporation can potentially shine a brighter light on the relevance of the vegetation dynamics in evaporation.

Mean NDVI of subcatchments is not different from mean NDVI of the entire catchment.

Table 4.1: Characteristics of principal components 1-6

In bold the principal component number (Y), the explained variance (ψ) and the main characteristic that is distinguished by the principal component is given. In normal text the specific characteristic of the pixels with a high, respectively a low score (y) is given, along with the dominant land cover in these subsets.

Y	ψ	Distinguishes	
		Subset with high y	Subset with low y
1	38.1%	Vegetation density High vegetation density Forests (3xx)	Sparsely vegetated areas or low density Discontinuous built up area (1xx)
2	13.7%	Seasonality differences between land cover classes Dual seasonality and/or rel. high late autumn and winter NDVI Maxima: 1) in spring (early Apr-early Jun) 2) in late summer (end Aug - half Sep (Oct in 2001)) Agriculture (231, 242, 243) - subset in downstream part catchment, coniferous forest (312, 313) - few pixels, distributed over catchment	Single season with high contrast between summer and winter NDVI Maximum in summer (June/July) Deciduous forest (311, 313) and natural vegetation (32x, 412, 1xx)
3	6.18%	Seasonality differences between (and within) land cover classes Dual seasonality and rel. high amplitude. Maxima: 1) spring (early-end May (mid Jun in 2006)) 2) late summer (end Aug (mid Oct in 2001)) Agriculture (231, 242, 243) - upstream part catchment has highest y_3 ; deciduous forest (311, 312) - subset in downstream part catchment	Single season with relatively low contrast between summer and winter Maximum in summer (June/July) Coniferous forest (312, 313); transitional woodland/shrub (324); wetland (412)
4	2.78%	Seasonality differences within land cover classes Single season Maximum in late spring (April-May) Agriculture (211, 242, 243) - possibly winter wheat and rapeseed; shrubland (32x)	Single season (predominantly) Maximum in late summer (July-September) Agriculture (211, 242, 243) - possibly maize, sugar beet, potato; built up area
5	2.01%	Differences in summer NDVI trend Increasing or stable summer NDVI over 2000-2006 Transitional woodland/shrub (324); wetland (412)	Decreasing summer NDVI Isolated spots in forest (2%), probably clear cuts
6	1.71%	Interannual differences Relatively low (summer) NDVI in 2003, 2004, 2006 Agriculture subset (242) in south-eastern part catchment	Low NDVI in 2000, 2001, 2006 Agriculture subset (211) in northern part catchment

VALIDATION OF THE REMOTELY SENSED PRODUCTS

5.1 INTRODUCTION

Validation of the remotely sensed evaporation is performed using the ground-based evaporation measurements from the five eddy covariance (EC) towers, for the temporally continuous products MOD16 and EARS, as well as for the temporally discontinuous product WACMOS. For MOD16 and EARS furthermore a validation of the multi-year water balance is performed. Because WACMOS is only available for clear days a water balance analysis is not suitable for this product. In the next sections the validation metrics are explained.

5.2 METHOD

5.2.1 Validation against EC measurements

The remotely sensed evaporation estimates (E_{RS}) of the three products MOD16, WACMOS and EARS, are validated for the study area against the EC measurements of evaporation (E_{EC}).

The E_{RS} time series of the pixels comprising the five EC measurement sites are compared to the E_{EC} time series. The comparison comprises visual inspection of time series plots, a linear regression analysis of E_{RS} on E_{EC} with least squares fitting and an analysis of the bias $d = E_{RS} - E_{EC}$ as a function of time and the magnitude of the evaporation (expressed in E_{EC}). Assuming that the E_{EC} measurements represent the real evaporation within the footprint of the EC site, and that they are representative for the area within the coinciding pixel of the RS products, E_{RS} is a 'good' estimate for the evaporation if the difference between E_{EC} and E_{RS} is within acceptable limits, or if there is a bias that is consistent for all sites and all periods. The performance of the RS products, or the agreement between the two data sources, is quantified using the mean bias ($\bar{d}$), the root mean squared error ($RMSE$, see Equation 5.1) and the limits of agreement ($[l_l, l_u]$), i.e. the limits between which 95% of the differences will lie, given a normal distribution of the differences, see Equation 5.2.

$$RMSE = \sqrt{\frac{1}{n} \sum (E_{RS,i} - E_{EC,i})^2} \quad (5.1)$$

$$\begin{aligned} l_u &\approx \bar{d} + 2 \cdot \sigma_d, \\ l_l &\approx \bar{d} - 2 \cdot \sigma_d \end{aligned} \quad (5.2)$$

where n is the number of data points, l_u and l_l are the upper, respectively the lower limit of agreement, $\bar{d}$ is the mean difference and σ_d is the standard deviation of d .

Furthermore, the coefficients of the linear regression and the coefficient of determination (r^2) - equivalent to the explained variance - as given by Equation 5.3 are used to quantify the strength of the linear relation between E_{EC} and E_{RS} :

$$r^2 = 1 - \frac{\sum (E_{RS,i} - \hat{E}_{RS,i})^2}{\sum (E_{RS,i} - \bar{E}_{RS})^2} \quad (5.3)$$

with $\hat{E}_{RS,i} = a \cdot E_{EC,i} + b$, a and b the coefficients of the linear regression.

The period and temporal resolution of the time series used in the validation is taken equal for all RS products to allow a comparison of the performance. The period is the year 2008, the only year for which WACMOS data is available. The smallest common temporal resolution is 8 days, the resolution of the MOD16 product. For EARS, which has a daily coverage, the 8 day evaporation is straightforwardly calculated as the sum over the 8 day periods given by MOD16. WACMOS however is available on clear days only, resulting in sometimes none or just a few clear days within the 8 day period. Here the 8 day evaporation is taken as the average evaporation at the clear days in the interval - if available -, multiplied by 8. Assuming reduced evaporation on cloudy days, this method implies an overestimation of the 8 day evaporation. Comparison of the performance of the different RS products must thus be done with some caution, and the effect of temporal averaging on the correlation should be considered. For WACMOS and EARS the validation of the evaporation flux has been performed at the smallest temporal resolution available in the respective datasets as well and the self-preservation of Λ (Section 2.4) is examined. The method and results are given in Appendix D.

Cloudy days means less solar energy. Does that necessarily mean reduced evaporation?

Representativeness of the EC measurements

To what extent is 'the same thing' measured?

The representativeness of the EC measurements for the E_{RS} of the different products and the mutual comparability of the E_{RS} products is examined in terms of the size and heterogeneity of their footprints. The area around the EC towers that is included in the grid cells of the RS products and the heterogeneity within are described in Section 3.1. The footprint of the EC measurements - ranging between 0.1-1 km², see Chapter 2 - is generally somewhat smaller than the grid cells of the products with the highest resolution, namely WACMOS and MOD16 (ca 1 km²). If however the area within the grid cells comprising the EC sites have a similar land cover (although elevation, aspect, slope might be as important, they are not examined), the EC measurements are considered representative for the remote sensing estimates. For MOD16 and WACMOS this is the case for all sites apart from Jalhay for WACMOS. The WACMOS pixel for Jalhay has a large area transitional shrub-/woodland, whereas the dominant land cover around the site is mixed forest. For the site Selhausen there is no WACMOS data available for the grid cell comprising the measurement site, and the analysis is performed for the upwind neighbouring pixel. The grid cells of EARS are too large (ca. 21 km²) and too heterogeneous compared to the footprint of the EC measurements. Especially at Jalhay and Selhausen $E_{RS,EARS}$ is expected to deviate rather much from E_{EC} , due to the distinct land uses within the grid cell boundaries (wetland and a brown coal mining respectively). Table 5.1 summarizes the above. Despite this issue of representativeness, the validation is performed for all sites and RS products - it is the best we have.

The upwind pixel is the pixel in the wind direction that is dominant in the entire measurement period, not specified per day

5.2.2 Validation at catchment scale

The temporally continuous products MOD16 and EARS are validated based on the catchment scale evaporation for the period 2000-2006 (2010 for EARS). Firstly the MOD16 and EARS time series of the 8-day evaporation, spatially averaged over the Ourthe catchment

Table 5.1: Comparability EC sites and RS grid cells

	EC site	WACMOS	MOD16	EARS
Land cover	Dominant	Representative land cover = x, or deviating as specified		
Vielsalm	mixed forest	x	x	31% agriculture
Jalhay	mixed forest	46% shrub-/woodland	x	23% wetland
Hesse	broad-leaved forest	x	x	54% agriculture
Lonze	agriculture	x	x	x
Selhausen	agriculture	x	x	25% open pit mining
Areal extent	0.1-1 km ²	1 km ²	1 km ²	21 km ²

and the Ourthe Orientale and Occidentale subcatchments, are compared mutually, with the catchment average NDVI and with the potential evaporation. The NDVI, reflecting vegetation green-up and senescence/leaf fall, and the potential evaporation, giving the available energy for evaporation, are expected to ‘form an envelope’ around the ‘actual’ evaporation.

Secondly an analysis of the water balance for the Ourthe catchment and two subcatchments is performed. The annual water balance for year i yields:

$$\left(\frac{\Delta S}{\Delta t}\right)_i = P_i - Q_i - E_i \quad (5.4)$$

with $\Delta S/\Delta t$ (mm y⁻¹) the annual storage change, P (mm y⁻¹) the annual precipitation, Q (mm y⁻¹) the annual discharge per unit area and E the evaporation (mm y⁻¹). For estimated fluxes this becomes:

$$\left(\frac{\Delta S}{\Delta t}\right)_i = \hat{P}_i - \hat{Q}_i - \hat{E}_i + \varepsilon_{flux,i} \quad (5.5)$$

with $\varepsilon_{flux,i}$ (mm y⁻¹) the sum of the unknown error in the individual fluxes in year i :

$$\varepsilon_{flux,i} = (P - \hat{P})_i + (Q - \hat{Q})_i + (E - \hat{E})_i \quad (5.6)$$

Catchment scale precipitation is estimated with precipitation data at the ten stations, interpolated using Thiessen polygons, see Section 3.3. For evaporation obviously the remotely sensed estimates are applied: $E_{RS,MOD16}$, respectively $E_{RS,EARS}$. Discharge estimates per unit area involve discharge measurements and the determination of the catchment area. There is no data available on the storage change over time. Although in many studies the annual storage change is assumed to be negligible - driven by data limitations - the storage might change in wet or dry years. E.g. Wang and Alimohammadi (2012) show that the interannual variability of storage change is more significant than that of evaporation especially under water limited conditions, but for energy limited conditions it applies as well. Over multiple years however, the storage change is assumed to be zero:

$$\frac{1}{n} \sum_{i=1}^n \left(\frac{\Delta S}{\Delta t}\right)_i = 0 \quad (5.7)$$

So if we have a sufficiently long time series we can determine the annual average ε_{flux} with:

$$D_i = \hat{P}_i - \hat{Q}_i - \hat{E}_i = \left(\frac{\Delta S}{\Delta t}\right)_i + \varepsilon_{flux,i} \quad (5.8)$$

Storage change was estimated from the water balance with remotely sensed evaporation estimates from Zhang et al. (2010), based on a MOD16-like algorithm).

combined with Equation 5.7:

$$\frac{1}{n} \sum_{i=1}^n D_i = \frac{1}{n} \sum_{i=1}^n \varepsilon_{flux,i} \quad (5.9)$$

The study period comprises 7 years (11 years for EARS), with annual precipitation ranging between 122% and 78% of the mean precipitation, for both 2000-2006 and 2000-2010, with the most extreme years (for the annual values) in the 2000-2006 period. See Section 3.1.3 for further details. It is assumed that for this period Equation 5.7 holds. Furthermore, to be able to use Equation 5.9 for validation of the remotely sensed evaporation estimates, it is assumed that the largest error is in $\hat{E}$, although the error in $\hat{P}$ may be substantial as well.

5.3 RESULTS AND DISCUSSION

In this section the results of the validation of WACMOS, MOD16 and EARS are shown and discussed. In the next section, firstly the results of the validation against the EC measurements are described in general terms and in a comparative way. Subsequently, per product, the source of the disagreement between E_{EC} and E_{RS} is discussed. The second section addresses the validation at the scale of the catchment.

5.3.1 Validation against EC measurements

The time serie plots in Figure 5.2 to Figure 5.4 of the 8 day interval evaporation measured at the EC sites (E_{EC}) and the remotely sensed estimates (E_{RS}) from MOD16, WACMOS and EARS respectively, give a first indication of the agreement between E_{EC} and E_{RS} . The statistical summary of the comparison is given in Table 5.2 - Table 5.4. The order of magnitude of the MOD16 estimates generally agrees rather well with E_{EC} , apart from spring/summer at the forested sites Vielsalm (mixed forest) and Hesse (broad-leaved forest), where $E_{RS} > E_{EC}$ with a mean bias of 2.6 and 3.6 mm 8d⁻¹, respectively. Jalhay (mixed forest) deviates from the other two forested sites with $E_{RS} < E_{EC}$ and a mean bias of -6.2mm 8d⁻¹. WACMOS overestimates the evaporation at all EC sites, for the forested sites to a larger extent (mean bias is 24mm 8d⁻¹) than for the agricultural sites (mean bias is 12mm 8d⁻¹). EARS overestimates the evaporation at the forested sites as well with a mean bias of 8mm 8d⁻¹. For the agricultural sites the order of magnitude of the EARS estimate agrees rather well with that of E_{EC} (mean bias = 0.27mm 8d⁻¹).

Looking at the correlation plots in Figure 5.5 to Figure 5.7 for the individual sites and in Figure 5.8 for the ensemble of sites, the strength of the linear relationship and consistency of this relationship for the different sites can be seen. For a moment taking the coefficient of determination as a measure for the performance of the RS evaporation estimates, and neglecting the influence of the number of data points, the following can be seen: the performance of MOD16 is relatively high with r^2 's of around 0.80 for all sites apart from Vielsalm ($r^2 = 0.60$). WACMOS performs generally worse than the other two products, with r^2 's around 0.60 for all sites, apart from Jalhay where it is slightly higher (0.67). EARS performs better at the forested sites than at the agricultural sites with r^2 's around 0.85 for Vielsalm, Jalhay and Hesse against 0.73 and 0.51 for Lonzee and Selhausen, respectively. Especially at Selhausen the performance is low.

Considering the best performing product per site, for the forested sites EARS performs best, closely followed by MOD16 (the MOD16 performance at Vielsalm is much lower than the EARS performance and only slightly higher than WACMOS). For the agricultural sites MOD16 performs best, followed by EARS at Lonzee, by WACMOS at Selhausen.

Looking at the coefficients of the linear regression per product, there is hardly any consistency between the sites: the slope of the linear regression differs rather much per site. There is some agreement in the coefficients of sites within the same land cover class for all RS products: large slopes ($\gg 1$) for the forested sites Vielsalm and Hesse, with increasing overestimation with increasing evaporation; slopes closer to 1 (both >1 and <1) for the agricultural sites with a seemingly more constant error with changing evaporation. Jalhay however deviates, with a slope in between that of the other forested sites and the agricultural sites for WACMOS and EARS, and an even smaller slope than the agricultural sites for MOD16. The lack of consistency in the linear relation between sites per product can be seen in the performance of the linear regression of the ensemble of stations as well: considerably reduced r^2 compared to the individual sites (see Table 5.2 to Table 5.4). The r^2 of the linear regression of the data points of the ensemble of sites within the same land cover class (last two columns in the Tables) is slightly better than for the ensemble of all stations for WACMOS and EARS, but not for MOD16 because of the distinct behaviour of Jalhay compared to the other forested sites.

Figure 5.9 to Figure 5.11 give further insight into the nature of the difference between E_{RS} and E_{EC} and its dependence on time and magnitude of the evaporation. Furthermore the limits of agreement within which 95% of the errors is located (given by the mean error $\pm 2\sigma$ of the error) is shown. For MOD16 there is a periodicity visible in the error (left side of Figure 5.9), which is not obviously related to the magnitude of the evaporation (see right side of Figure 5.9). For values of E_{EC} lower than $5 \text{ mm } 8\text{d}^{-1}$ or so, the error is relatively small, but for larger values the error is random between the limits of agreement. For the forested sites Vielsalm and Hesse the error is highest from the beginning to halfway the growing season. At the agricultural sites the difference fluctuates. The limits of agreement, even without applying confidence intervals, cover a range that is wider than the range of the evaporation itself for Vielsalm and Hesse. For Jalhay and the agricultural sites the error is in the order of half or two third of the maximum evaporation. Potential sources for the temporal difference between $E_{RS,MOD16}$ and E_{EC} are a too high maximum canopy conductance (during the growing season vegetation cover is large and evaporation will originate - at least according to the MOD16 algorithm - mainly from transpiration and interception, not from soil evaporation); constraints on the canopy conductance are not conservative enough; soil water availability limits evaporation, which is not directly taken into account in the algorithm, as described in Chapter 2; the climatological input is not representative for the point scale; albedo and/or LAI is not representative for the point scale due to heterogeneous land cover within the grid cell; and lastly of course the E_{EC} measurement could be erroneous especially at the dense forest sites, but the reason for the temporal behaviour would be unclear. Grid cell heterogeneity could cause a periodical error in leaf area index and albedo due to the presence of agriculture in the grid cell of Vielsalm (although the areal extent of the agricultural area is small) and at the agricultural sites a mixture of sowing (emergence) and harvest dates within the grid cell is a possible source of the periodicity in the error. For Hesse however the land cover in the grid cell is rather homogeneous and cannot explain the periodic difference in E_{EC} and $E_{RS,MOD16}$.

To elucidate the observed difference further, the Penman-Monteith equation is applied with point scale meteorological data of one of the EC sites, Vielsalm. Evaporation is determined for the reference crop with a constant canopy conductance (g_s) of $1/70 \text{ ms}^{-1}$ (this can be considered as potential evaporation), and for a seasonal varying $g_s = g_{s,max} \cdot GSI$. $g_{s,max}$ is determined by equating the Penman-Monteith equation and EC measurements over a multi year time series. The maximum dry day canopy conductance is used as maximum conductance. The seasonal variability is modelled with the growing season index (GSI), see Section 2.1. Results are shown in Figure 5.1. The modelled evaporation at point scale with site specific g_s (E_{PM}) agrees much better with the E_{EC} than $E_{RS,MOD16}$. Since for E_{EP} no daily constraints for evaporation are applied and given the pattern of the black line in Figure 5.1, the seasonal difference between E_{EC} and $E_{RS,MOD16}$ cannot be explained by soil moisture (or otherwise constrained) evaporation. The explanation that is left is a too high net radiation, which may be the result of the spatial scale of the meteorological data, or due to an error in albedo.

$g_{s,max}$ in Figure 5.1 can be seen as a calibrated, site specific parameter. MOD16 takes $g_{s,max}$ constant for all biomes and scales it with LAI for seasonal variability and with additional constraints for daily variability.

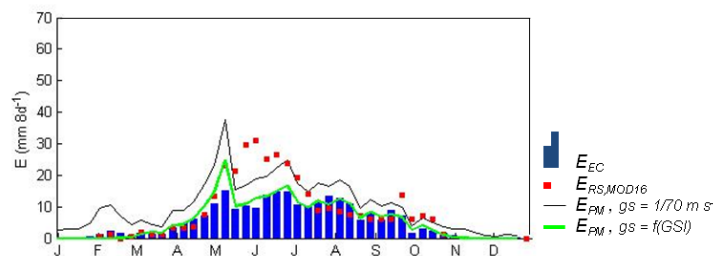


Figure 5.1: Comparison of $E_{RS,MOD}$ (red squares), E_{EC} (blue bars) and modelled evaporation with point scale meteorologic data from the EC site and the Penman-Monteith equation E_{PM} . The black line shows E_{PM} with constant canopy conductance for a reference crop. The green line represents E_{PM} with a seasonal varying canopy conductance and a site specific maximum canopy conductance.

For Hesse the overestimation of $E_{RS,MOD16}$ compared to E_{EC} starts in winter, before soil water availability can be an unaccounted constraint. Here the maximum canopy conductance and how it develops with increasing NDVI is probably an additional source of the discrepancy. The third forested site, Jalhay, shows the drawback of the use of constant canopy conductance in the MOD16 algorithm even more clearly. Evaporation at Jalhay is much higher than at the other forest sites, most likely merely because of a higher canopy conductance than due to meteorological conditions, as was shown by equating the Penman-Monteith equation and measured evaporation and solving for the canopy conductance (not shown). This is partly confirmed by the fact that the performance of WACMOS and EARS is similar for the three forested sites. Only EARS shows indeed higher evaporation for Jalhay than for Vielsalm and Hesse, but this might be caused by the relatively large area wetland within the EARS grid cell as well. A more elaborate analysis of the MOD16 algorithm applied on the point scale data would elucidate the source of the differences further, but that is left out of the scope of this research.

Maximum canopy conductance is influenced by stand age, tree height, leaf age, canopy structure, species.

The difference between $E_{RS,WACMOS}$ and E_{EC} does not show a clear pattern in time, although during the growing season the difference reaches - but is not restricted to - higher values than outside the growing season, see Figure 5.10. This applies to both forested and agricultural sites. For Vielsalm and Hesse the difference seems to be proportional to the magnitude of evaporation. The limits of agreement - in Figure 5.10 constant with the magnitude of the evaporation - could be taken linearly increasing, probably leading to slightly narrower limits over the full reach. The constant limits for Vielsalm and Hesse are $52\text{mm } 8\text{d}^{-1}$ and $50\text{mm } 8\text{d}^{-1}$ apart, respectively, which is more than 3 and 2.5 times the measured range in actual evaporation (E_{EC}). Taking the limits varying with time would, by visual inspection, not reduce the distance between them to acceptable values. For the three other sites, Jalhay, Lonze and Selhausen, there is no obvious relation with the magnitude of the evaporation: about 95% of the differences between $E_{RS,WACMOS}$ and E_{EC} lies between $25 \pm 22\text{mm } 8\text{d}^{-1}$ ($E_{EC,max} = 16\text{mm } 8\text{d}^{-1}$) for Jalhay, between $17 \pm 15\text{mm } 8\text{d}^{-1}$ ($E_{EC,max} = 30\text{mm } 8\text{d}^{-1}$) for Lonze and for Selhausen between $8.6 \pm 18\text{mm } 8\text{d}^{-1}$ ($E_{EC,max} = 35\text{mm } 8\text{d}^{-1}$). The limits of agreement thus are unacceptably wide apart. Looking further into the origin of this limited agreement, the validation of the WACMOS product is extended to the original daily interval of the product, and to the assumption of self-preservation of the evaporative fraction (Λ) at the days with cloud free overpass times. The analysis is shown in Appendix D, here the main findings are given. Due to the averaging over the 8 day interval the strength of the linear relation between $E_{RS,WACMOS}$ and E_{EC} is higher (higher coefficient of determination) for the 8 day interval than for the daily interval, except for Vielsalm. The agreement between $E_{RS,WACMOS}$ and E_{EC} however improves by using the daily interval for the forested sites: the distance between the limits of agreement is approximately twice the maximum observed E_{EC} at Vielsalm and Hesse, and about equal to the maximum E_{EC} at Jalhay. For the agricultural sites the limits of agreement are in the same order of magnitude as was the case for the 8-day interval. This reflects the introduced error by taking evaporation on cloud free days as an estimate for cloudy days. Although better, also for the daily interval the limits of agreement are unacceptable. Analysis of instantaneous and daily Λ shows that Λ_{day} is slightly underestimated by Λ_{inst} on days WACMOS images are available, in correspondence with what is found in literature for completely clear days. The limited agreement between $E_{RS,WACMOS}$ and E_{EC} and the large overestimation of $E_{RS,WACMOS}$ at the forested sites thus is not explained by the invalidity of the self-preservation of Λ . Comparison of the Λ contained in the WACMOS product and Λ derived from EC measurements does shed light on the overestimation and limited agreement between $E_{RS,WACMOS}$ and E_{EC} : the correlation of Λ_{WACMOS} and Λ_{EC} is poor (on average $r^2 = 0.06$ for the forested sites and 0.01 for the agricultural sites), and $\Lambda_{WACMOS} \gg \Lambda_{EC}$ for the forested sites. Timmermans (pers.com., May 2014) attributes this discrepancy to an error in stability functions in the WACMOS version 1.0 algorithm. The stability functions are applied in the calculation of the sensible heat flux H and evaporation at the wet limit, see Section 2.4. The magnitude and ‘direction’ of the error depend on the atmospheric con-

ditions. Although net radiation and the ground heat flux are variables in the determination of Λ as well, it is expected that the main source of the error in Λ is the erroneous stability functions.

As described in Equation 5.2.1 the EC measurements are not considered to be representative for the EARS evaporation estimates, due to the discrepancy in size and heterogeneity of the area that is covered by one grid cell and the footprint of the EC measurements. Applying limits of agreement is not relevant for this product, explaining the difference between E_{EC} and $E_{RS,EARS}$ is. Noteworthy in Figure 5.11 is the magnitude of the mean difference, which is much higher for the forested sites than for the agricultural sites, and the periodicity of the difference. The periodicity in the difference at Lonze originates from the periodicity in E_{EC} , which is not sensed by the EARS product (see also Figure 5.4). The periodicity in E_{EC} is caused by combined crop development (sugar beet) and meteorological conditions. Although more than 80% of the EARS grid cell comprising the Lonze EC site is agriculture, crop types and sowing dates will deviate within the EARS grid cell, and the combined effect is a more smoothed evaporation. In Selhausen, apart from the 'smoothing' of the periodicity of evaporation, the EARS estimate seems to be shifted forward in time: at the beginning of the growing season $E_{RS,EARS}$ is lower than E_{EC} , but declines later in the growing season. This might be the effect of the presence of the brown-coal mine in the EARS grid cell. Concerning the overestimation of $E_{RS,EARS}$ at forested sites in general, this might be caused by the fact that the surface temperature that is sensed, is the temperature at the top of the canopy. The EARS algorithm does not account for the distribution of the temperature from the top of the canopy to the soil, and the sensible heat flux is likely to be underestimated (Foppen, pers. com., September 2014). Moreover, in the EARS algorithm the ground heat flux G , the physical heat storage S_t and energy advection A_h are neglected in the energy balance. Energy for photosynthesis F is used in Equation 2.50, but neglected in the calculation of the daily sensible heat Equation 2.58 as well. Although the daily average G , S_t and A_h may be close to zero, the instantaneous values at noon most likely are not. In forests especially the storage of heat in the moist air under the canopy is expected to be considerable. Neglecting these terms in Equation 2.58 results in an underestimation of the sensible heat and thus an overestimation of the evaporation.

The magnitude of the effect can not be established for the three forested sites in this study, due to the large area of agriculture (Vielsalm and Hesse) and wetland (Jalhay) in the EARS grid cells. An indication however of EARS giving a landuse specific distorted view, follows from the comparison of the $E_{RS,EARS}$ from the Lonze, Vielsalm and Hesse grid cells. Where the EC measurements show a clear difference in magnitude of the evaporation between the agricultural and forested sites, the EARS estimate of the evaporation in the Vielsalm and Hesse grid cells is as high or even higher than the evaporation from the Lonze grid cell, although the proportion forest is significant in both cells. This should be noted when applying the EARS product to the catchment scale.

Overall, there is limited agreement between the EC measurements and the remotely sensed evaporation estimates, sometimes explainable, yet often it is unclear what is the exact source of the difference.

Table 5.2: Statistics of the comparison of E_{EC} and E_{RS} for MOD16 with 8 day interval.

The first three parameters are the mean of E_{EC} , the mean of E_{RS} and the number of data points n used in the comparison. The next three parameters are the mean bias ($\overline{E_{RS} - E_{EC}}$) and the standard deviation of the bias ($\text{sd}(E_{RS} - E_{EC})$) and the root mean squared error (or difference really) (RMSE) of the two data sets. The parameters concerning the linear regression ($y = \hat{E}_{RS}$, $x = E_{EC}$) and correlation are the last three parameters: slope, intercept and coefficient of determination (r^2)

		MOD16 - 8 day interval					All sites	Forest	Crops
		BEVie	BEJal	FRHes	BELon	DESeh			
$\overline{E_{EC}}$	(mm 8d ⁻¹)	8.16	13.90	7.09	11.87	14.65	10.80	8.61	13.28
$\overline{E_{RS}}$	(mm 8d ⁻¹)	10.77	7.75	10.68	9.15	14.50	11.00	10.23	11.87
n	(-)	25	11	31	29	30	126	67	59
$\overline{(E_{RS} - E_{EC})}$	(mm 8d ⁻¹)	2.60	-6.15	3.59	-2.72	-0.15	0.20	1.62	-1.41
$\text{sd}(E_{RS} - E_{EC})$	(mm 8d ⁻¹)	6.18	3.90	4.65	3.37	5.65	5.77	6.19	4.81
RMSE	(mm 8d ⁻¹)	6.59	7.19	5.81	4.28	5.56	5.75	6.35	4.97
slope ($d\hat{E}_{RS}/dE_{EC}$)	(-)	1.52	0.68	1.35	0.73	1.06	0.90	0.95	0.94
intercept	(mm 8d ⁻¹)	-1.63	-1.64	1.10	0.53	-1.02	1.30	2.07	-0.65
r^2	(-)	0.60	0.81	0.80	0.89	0.79	0.64	0.49	0.79

Table 5.3: Statistics of the comparison of E_{EC} and E_{RS} for WACMOS with 8 day interval

		WACMOS - 8 day interval					All sites	Forest	Crops
		BEVie	BEJal	FRHes	BELon	DESeh			
$\overline{E_{EC}}$	(mm 8d ⁻¹)	9.66	13.84	7.17	15.34	15.16	11.87	8.99	15.24
$\overline{E_{RS}}$	(mm 8d ⁻¹)	40.90	39.11	26.94	32.14	23.72	30.25	32.70	27.38
n	(-)	16	11	35	23	30	115	62	53
$\overline{(E_{RS} - E_{EC})}$	(mm 8d ⁻¹)	31.24	25.27	19.77	16.80	8.56	18.38	23.71	12.14
$\text{sd}(E_{RS} - E_{EC})$	(mm 8d ⁻¹)	13.08	11.07	12.41	7.44	9.22	12.88	13.12	9.37
RMSE	(mm 8d ⁻¹)	33.71	27.39	23.25	18.31	12.47	22.41	27.05	15.28
slope ($d\hat{E}_{RS}/dE_{EC}$)	(-)	2.80	1.62	2.25	1.14	1.07	1.15	2.11	1.10
intercept	(mm 8d ⁻¹)	13.85	16.66	10.81	14.58	7.49	16.59	13.74	10.62
r^2	(-)	0.58	0.67	0.63	0.61	0.57	0.36	0.61	0.53

Table 5.4: Statistics of the comparison of E_{EC} and E_{RS} for EARS with 8 day interval

		EARS - 8 day interval					All sites	Forest	Crops
		BEVie	BEJal	FRHes	BELon	DESeh			
$\overline{E_{EC}}$	(mm 8d ⁻¹)	5.81	11.10	6.30	9.89	12.63	8.92	6.92	11.31
$\overline{E_{RS}}$	(mm 8d ⁻¹)	13.91	19.31	13.95	12.37	10.84	13.36	14.86	11.58
n	(-)	44	18	42	42	45	191	104	87
$\overline{(E_{RS} - E_{EC})}$	(mm 8d ⁻¹)	8.10	8.20	7.64	2.48	-1.79	4.44	7.94	0.27
$\text{sd}(E_{RS} - E_{EC})$	(mm 8d ⁻¹)	7.23	6.39	7.20	5.51	7.17	7.87	7.02	6.74
RMSE	(mm 8d ⁻¹)	10.80	10.29	10.44	5.98	7.31	9.02	10.57	6.70
slope ($d\hat{E}_{RS}/dE_{EC}$)	(-)	2.15	1.45	1.92	1.02	0.76	1.04	1.75	0.85
intercept	(mm 8d ⁻¹)	1.45	3.24	1.88	2.28	1.23	4.09	2.74	2.00
r^2	(-)	0.86	0.86	0.84	0.73	0.51	0.53	0.83	0.58

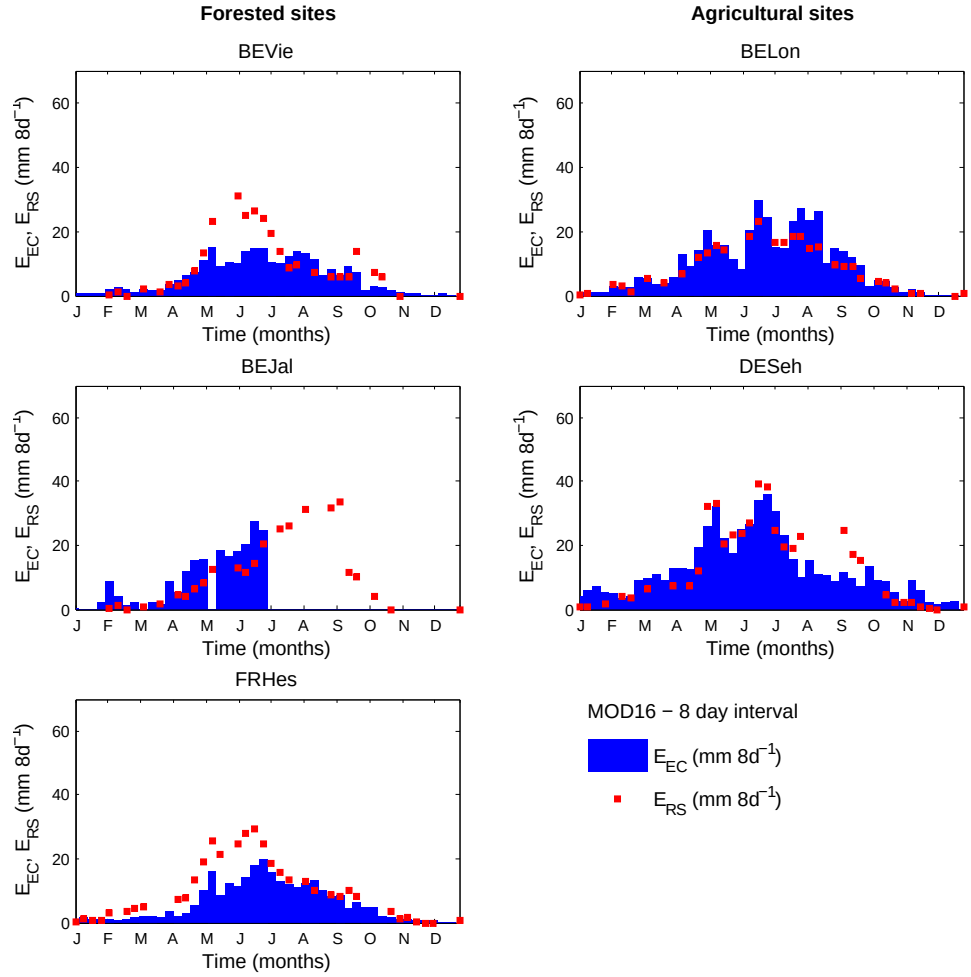


Figure 5.2: Comparison of the time series of E_{EC} (blue bars) and $E_{RS,MOD16}$ (red squares) for the 8 day interval for the year 2008 and for the five EC sites. E_{EC} (mm 8d⁻¹) is the evaporation measured at the specific EC site, summed over the 8 day interval. $E_{RS,MOD16}$ (mm 8d⁻¹) is the MOD16 evaporation estimate for the 8 day interval. On the left side the forested sites Vielsalm (mixed forest), Jalhay (mixed forest) and Hesse (broad-leaved forest). On the right the agricultural sites Lonzee and Selhausen.

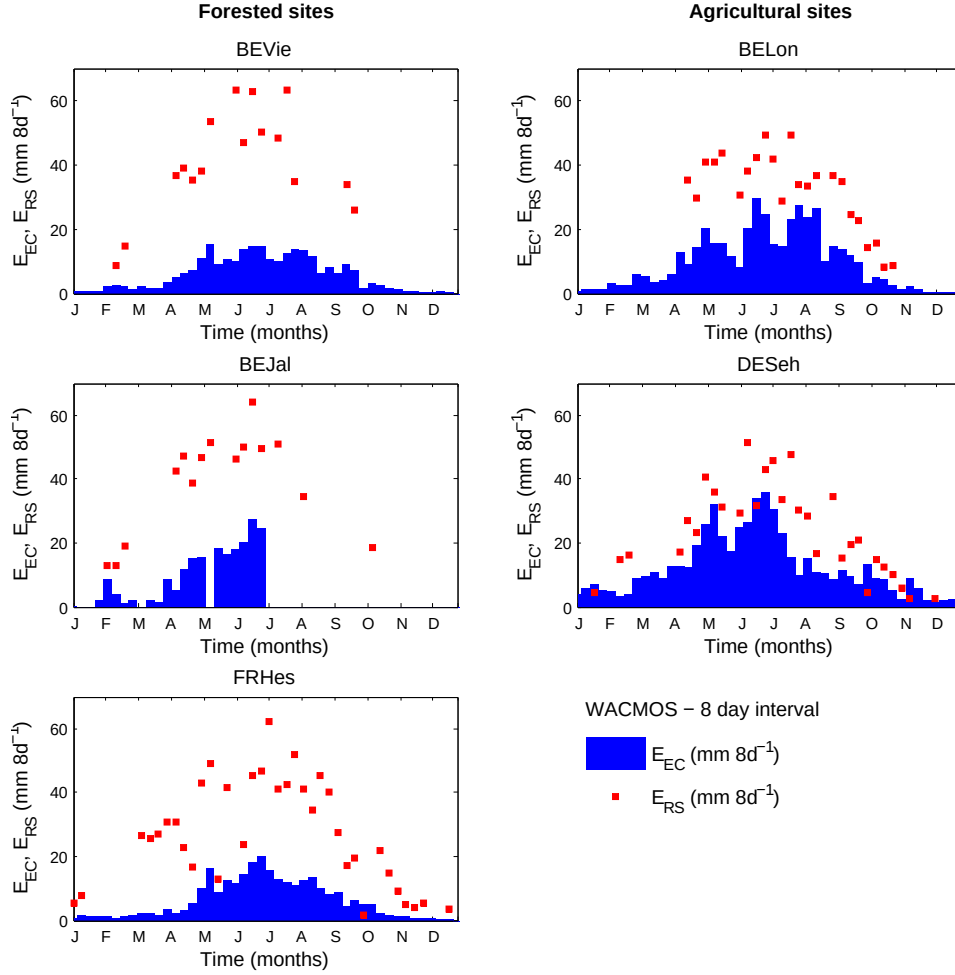


Figure 5.3: Comparison of the time series of E_{EC} (blue bars) and $E_{RS,WACMOS}$ (red squares) for the 8 day interval for the year 2008 and for the five EC sites. E_{EC} ($\text{mm } 8\text{d}^{-1}$) is the evaporation measured at the specific EC site, summed over the 8 day interval. $E_{RS,WACMOS}$ ($\text{mm } 8\text{d}^{-1}$) is the mean clear day WACMOS evaporation estimate of the pixel in which the EC tower is located (upwind pixel for Selhausen) integrated over the 8 day interval. On the left side the forested sites Vielsalm (mixed forest), Jalhay (mixed forest) and Hesse (broad-leaved forest). On the right the agricultural sites Lonze and Selhausen.

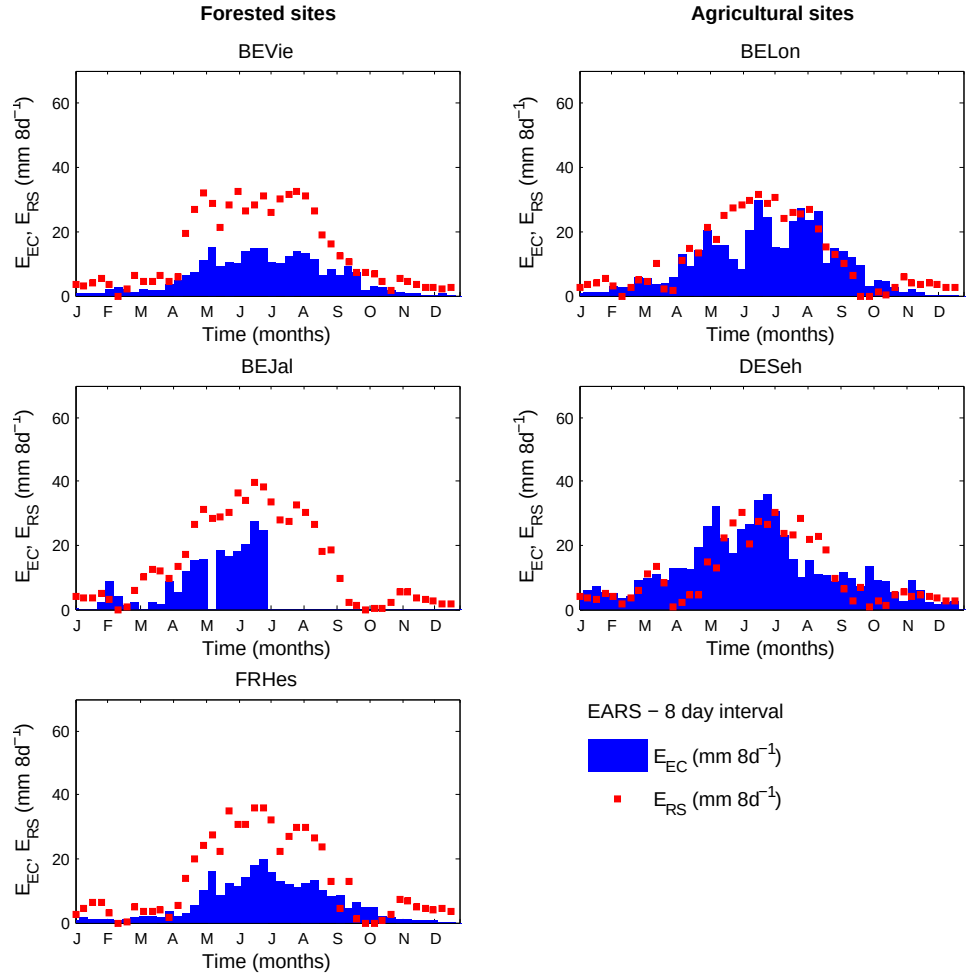


Figure 5.4: Comparison of the time series of E_{EC} (blue bars) and $E_{RS,EARS}$ (red squares) for the 8 day interval for the year 2008 and for the five EC sites. E_{EC} ($\text{mm } 8\text{d}^{-1}$) is the evaporation measured at the specific EC site, summed over the 8 day interval. $E_{RS,EARS}$ ($\text{mm } 8\text{d}^{-1}$) is the EARS evaporation estimate of the pixel in which the EC tower is located, summed over the 8 day interval. On the left side the forested sites Vielsalm (mixed forest), Jalhay (mixed forest) and Hesse (broad-leaved forest). On the right the agricultural sites Lonze and Selhausen.

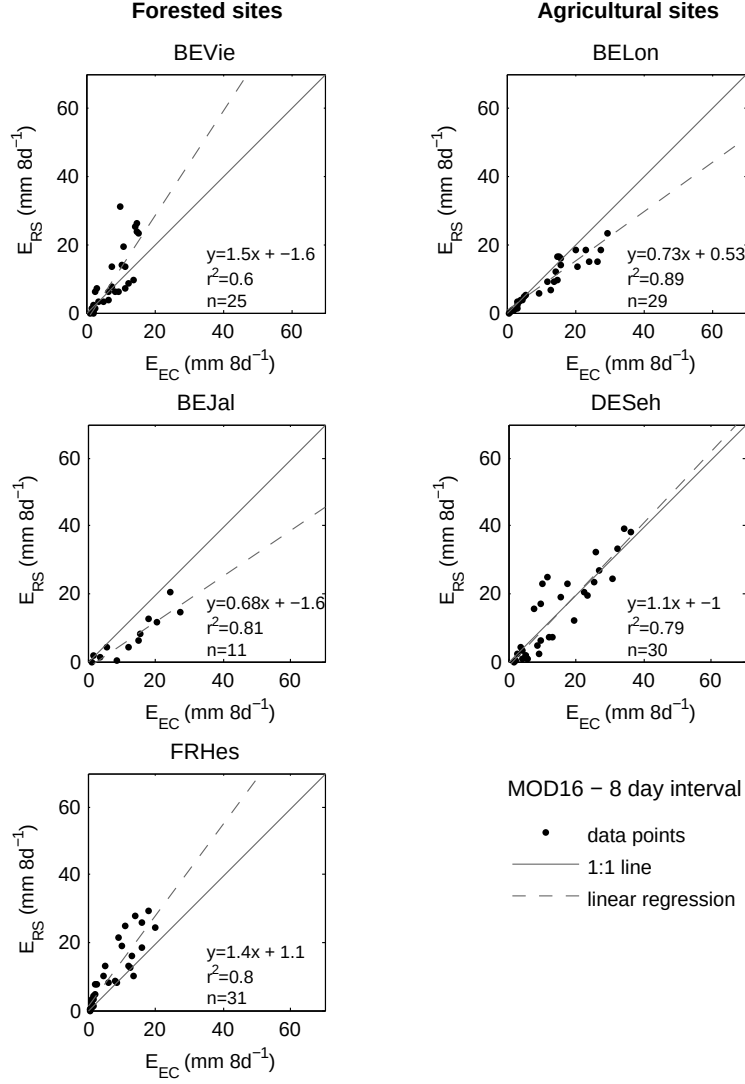


Figure 5.5: Correlation of E_{EC} and $E_{RS,MOD16}$ (mm 8d⁻¹) for the eight day interval for the year 2008 and for the five EC sites. $E_{RS,MOD16}$, the MOD16 evaporation estimate for the pixel in which the EC tower is located, on the y-axis, and E_{EC} of the specific station on the x-axis. The solid black line is the 1 : 1 line. The dotted black line shows the linear regression of the data points. The equation and r^2 of this line are given in the lower right corner of all figures.

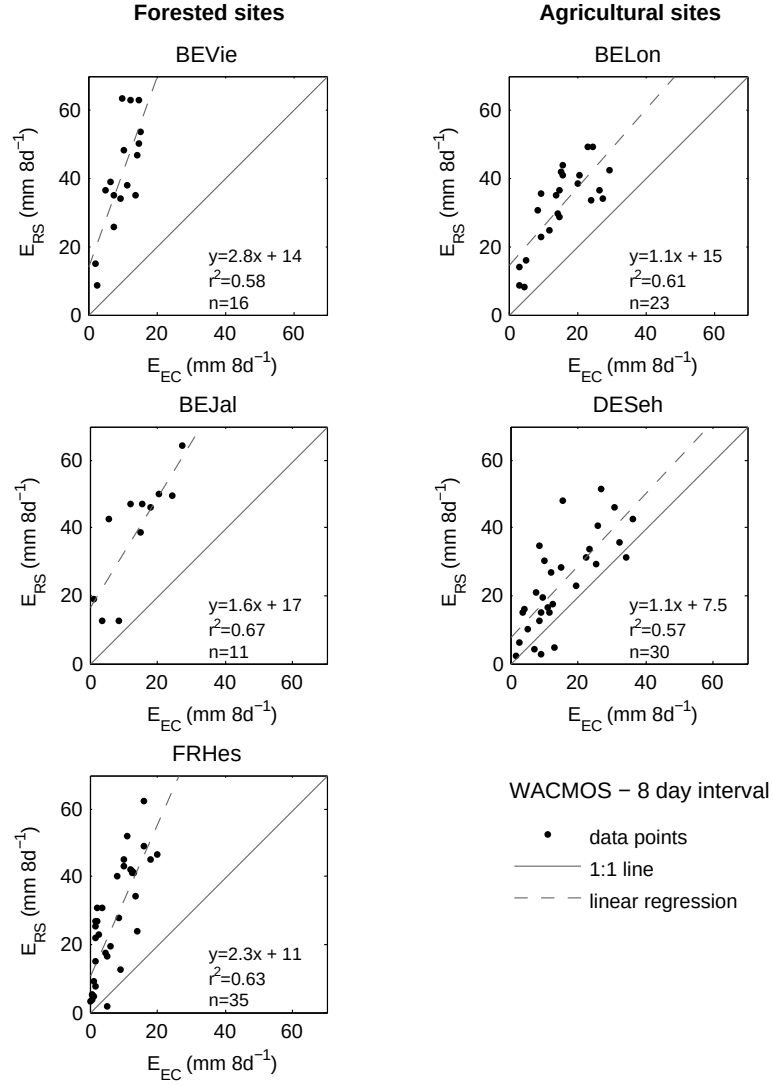


Figure 5.6: Correlation of E_{EC} and $E_{RS,WACMOS}$ (mm 8d⁻¹) for the eight day interval for the five EC sites. $E_{RS,WACMOS}$, the WACMOS evaporation estimate of the pixel in which the EC tower is located (the upwind pixel for Selhausen), on the y-axis, and E_{EC} of the specific site on the x-axis. The solid black line is the 1 : 1 line. The dotted black line shows the linear regression of the data points. The equation and r^2 of this line are given in the lower right corner of all figures.

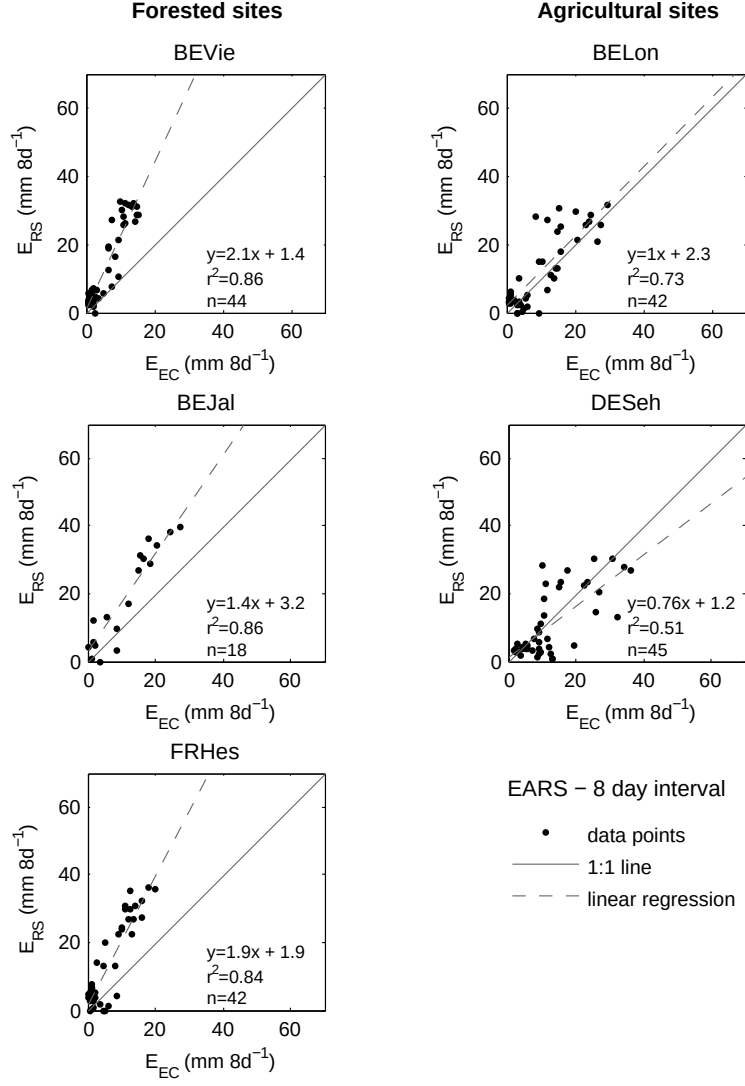


Figure 5.7: Correlation of E_{EC} and $E_{RS,EARS}$ (mm 8d⁻¹) for the eight day interval for the five EC sites. $E_{RS,EARS}$, the EARS evaporation estimate for the pixel in which the EC tower is located, on the y-axis, and E_{EC} of the specific station on the x-axis. The solid black line is the 1 : 1 line. The dotted black line shows the linear regression of the data points. The equation and r^2 of this line are given in the lower right corner of all figures.

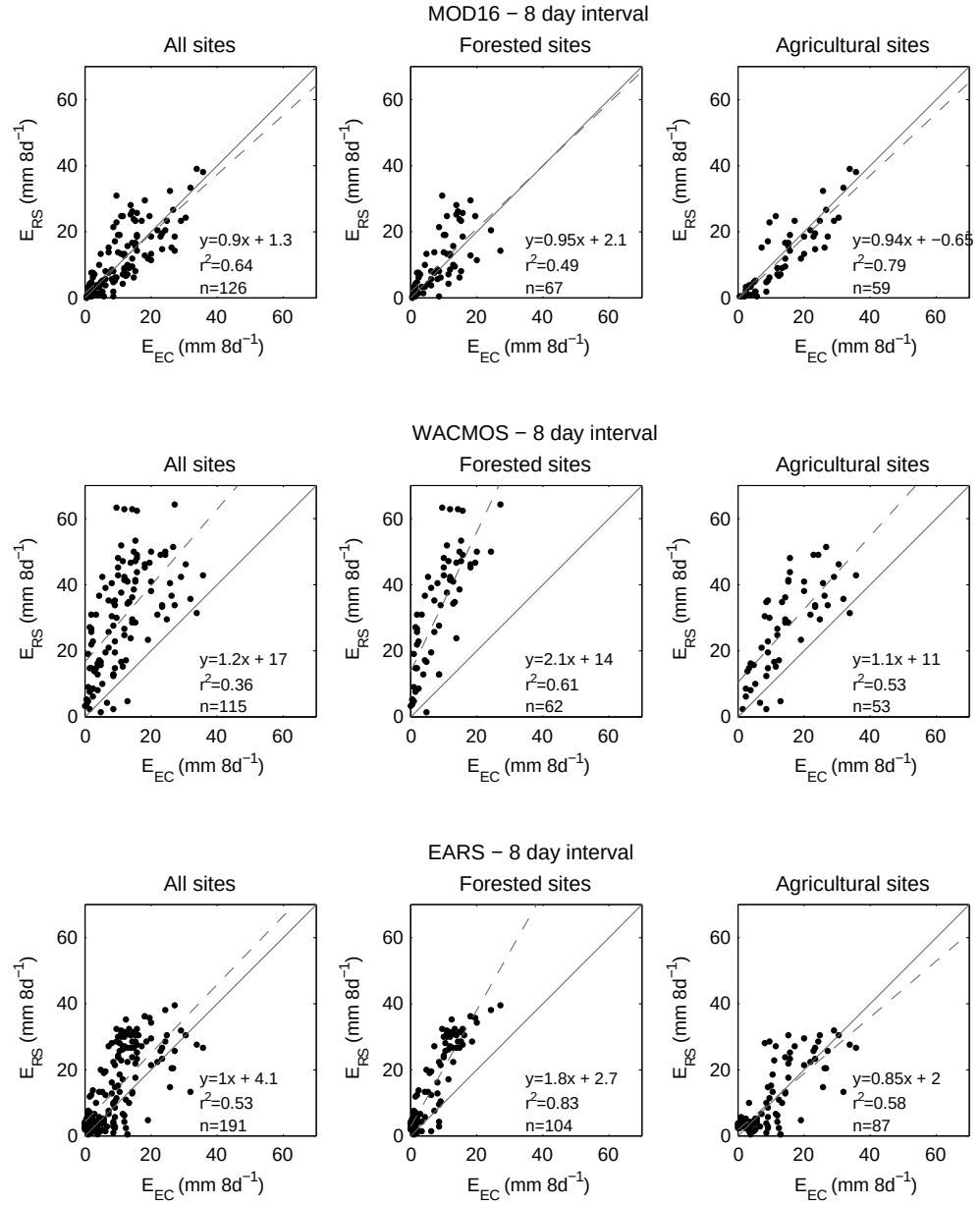


Figure 5.8: Correlation of E_{EC} and E_{RS} for the ensemble of stations - 8 day interval. From top to bottom: correlation plots for MOD16, WACMOS and EARS evaporation estimates. From left to right: correlation for the data of all EC sites, for the data of the forested sites (Vielsalm, Jalhay and Hesse), and for the data of the agricultural sites (Lonze and Selhausen).

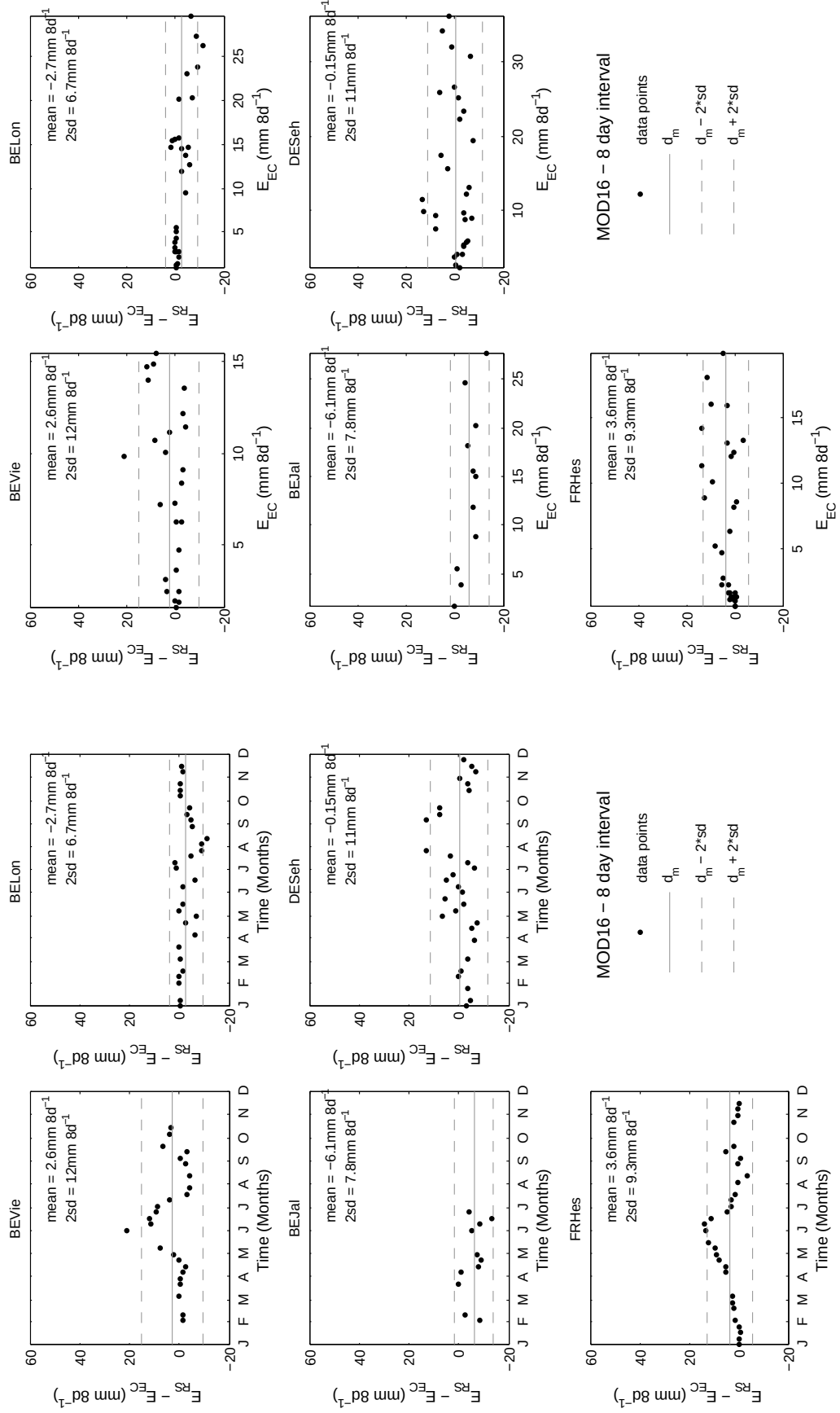


Figure 5.9: Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for MOD16 and the 8 day interval

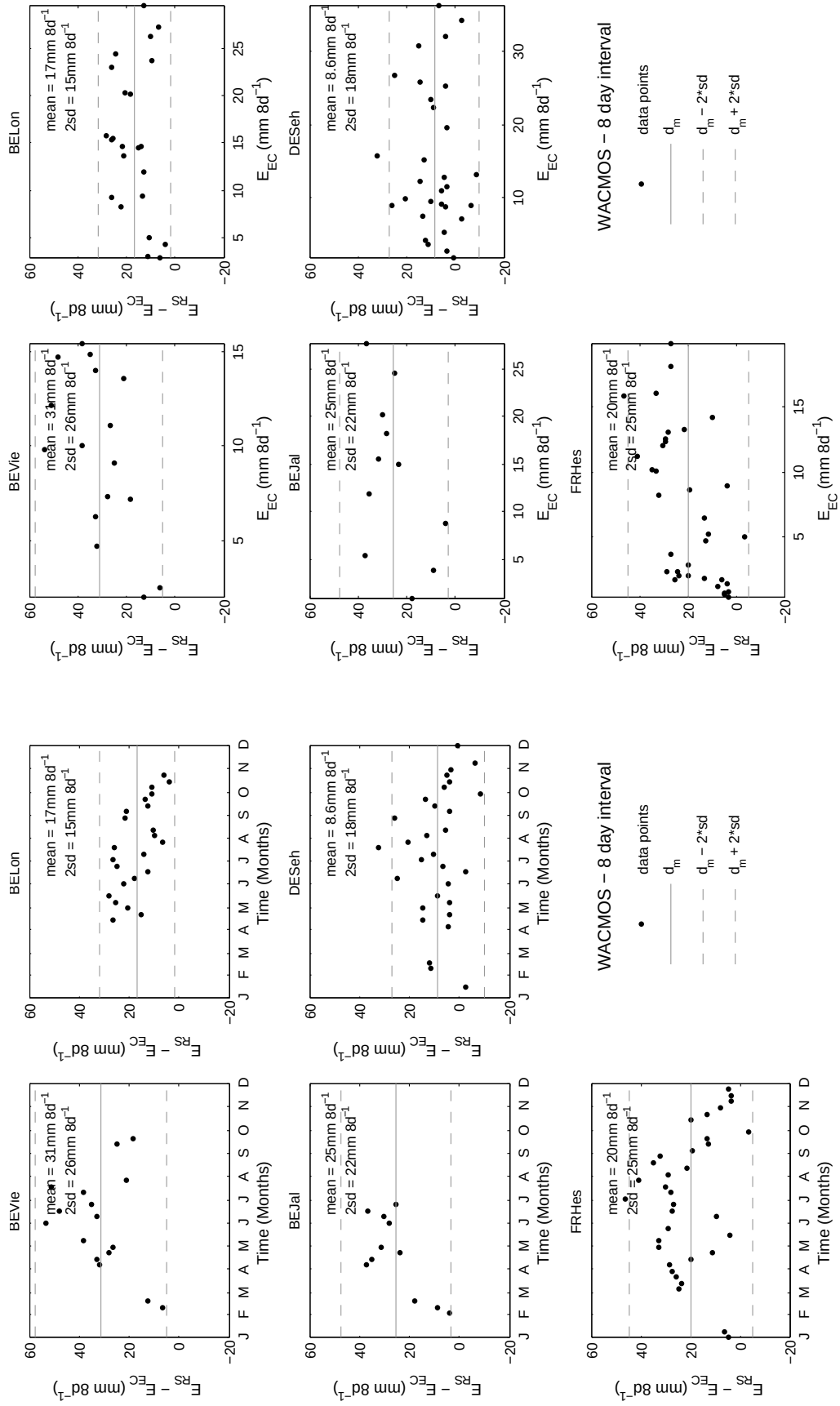


Figure 5.10: Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for WACMOS and the 8 day interval.

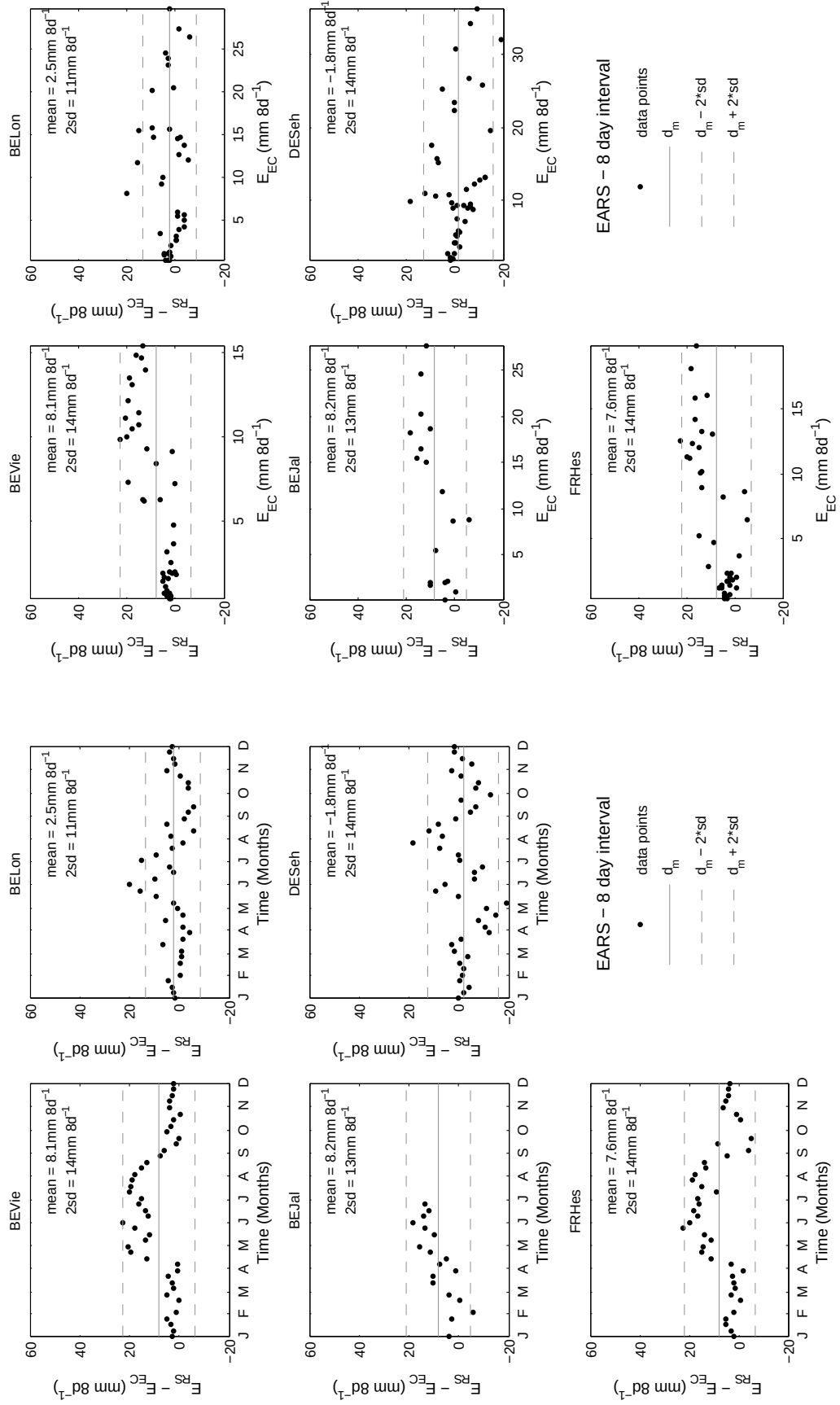


Figure 5.11: Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for EARS and the 8 day interval.

5.3.2 Validation at catchment scale

The spatially mean 8-day evaporation in the period 2000-2006 for the Ourthe catchment according to EARS and MOD16 is compared in Figure 5.12. The spatially averaged NDVI is shown along. $E_{RS,MOD16}$ is smaller in summer and winter, but in the transition seasons this is not always the case. In some years the springtime increase in evaporation sets in earlier according to MOD16, but is generally less steep and the maximum evaporation is reached later. The decrease in autumn clearly sets in later according to MOD16, and $E_{RS,MOD16}$ remains larger than $E_{RS,EARS}$ in this season. Generally, the growing season according to MOD16 seems shifted forward in time compared to the EARS estimate, with approximately 16 days. However, the main summertime peaks in $E_{RS,MOD16}$ and $E_{RS,EARS}$ coincide in the majority of the years, although the relative magnitude is different. Comparison of the pattern in remotely sensed evaporation with the spatially mean NDVI does not give much more clarity on the validity of the evaporation products, than that the seasonality in NDVI comprises the seasonal pattern in evaporation: the springtime increase in NDVI occurs more or less at the same time as the increase in evaporation (both $E_{RS,EARS}$ and $E_{RS,MOD16}$); the decrease in evaporation at the end of the growing season precedes the decrease in NDVI. MOD16 is almost equal to the potential evaporation ($E_{p,RS,EARS}$) in this period in most of the years. $E_{RS,EARS}$ is generally smaller than ($E_{p,RS,EARS}$). This suggests that it is mainly the meteorological conditions that limits evaporation in this season, not the presence of green vegetation. The evaporation in the subcatchments Ourthe Orientale and Ourthe Occidentale is similar to that of the Ourthe catchment.

$E_{RS,MOD16}$, $E_{RS,EARS}$ and NDVI increase around the same time in spring. In summer $E_{RS,EARS} > E_{RS,MOD16}$. In autumn $E_{RS,EARS}$ decrease sets in first, followed by $E_{RS,MOD16}$, which equals E_p , and much later NDVI.

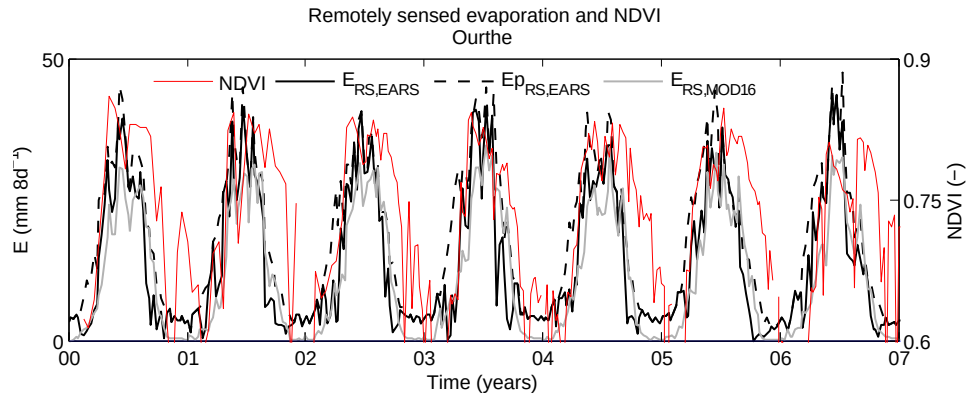


Figure 5.12: Evaporation (mm/8d) for the Ourthe Catchment for the period 2000-2006. In black $E_{RS,EARS}$ is shown, in gray $E_{RS,MOD16}$. The black dashed line shows the potential evaporation from the EARS product. The spatially averaged quasi 8 day composite NDVI is shown in red.

The annual water balance of the Ourthe catchment and subcatchments for the period 2000-2006 (2010 for EARS) is given in Table 5.5 for the Ourthe catchment, in Table 5.6 and Table 5.7 for the Ourthe Orientale and Ourthe Occidentale respectively. Firstly comparing the annual evaporation, $E_{RS,MOD16}$ is structurally smaller than $E_{RS,EARS}$, with a difference of up to 15% of the annual precipitation. Furthermore the temporal variation and the relative magnitude of the mean evaporation per (sub)catchment is different for the two products. The difference between the two products per catchment is clearly visible in Figure 5.13, which shows the cumulative daily fluxes ($E_{RS,EARS}$ in green, $E_{RS,MOD16}$ in yellow). The barplots in Figure 5.14 give a better view on the water balance of the individual years. Notable here is the relatively constant annual value of evaporation over the years compared to the variability in annual precipitation and runoff, for both $E_{RS,EARS}$ and $E_{RS,MOD16}$. The seasonality of the rainfall is not considered, so it is unclear whether the annual pattern is representative for the growing season. If it is, it means that there are other factors limiting evaporation than water availability, even in the driest years (2003 and 2005), or that there is enough storage in the catchment to overcome these years. MOD16 might not 'see' the

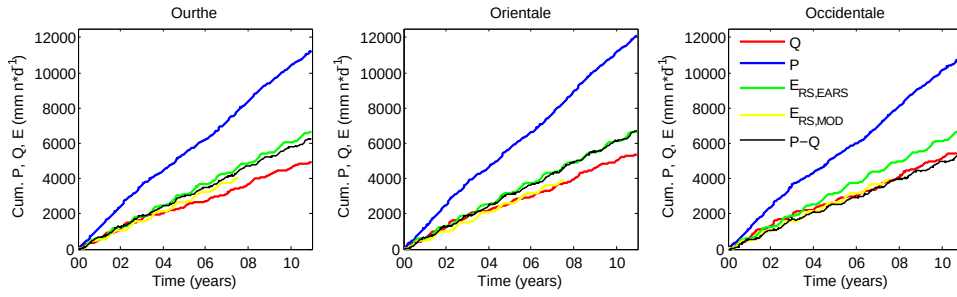


Figure 5.13: Cumulative water balance of the Ourthe (top) and the subcatchments Ourthe Orientale (middle) and Ourthe Occidentale (bottom), for the period 2000 - 2010.

limited water availability if it is not reflected in the air temperature and vapour pressure deficit or in the LAI, yet EARS should. In Figure 5.12 it can be seen that in 2003 the peak evaporation is relatively high according to EARS, but that the evaporation decreases steeply rather early in the season. This steep decrease is missed by $E_{RS,MOD16}$. Note that in 2004 $E_{RS,MOD16}$ is much lower than in 2003, whereas $E_{RS,EARS}$ is higher.

The fraction E/P is more or less constant for high precipitation and increases linearly with decreasing rainfall for lower precipitation. The same pattern is visible for $E_{RS,MOD16}$ and $E_{RS,EARS}$. How this fraction would logically behave in dry years is not obvious and applying the Horton index might be suitable to get more insight into 'logicality' of E_{RS} in dry years. The Horton index is the fraction of evaporation from the plant available water, see (Troch et al., 2009), which should approach 1 in dry years, given the governing climate in the Ourthe catchment. However, information on the separation of the river discharge in the fast and base flow components is required. For now, the logicality of the variability of $D_i = P_i - Q_i - E_i$ over the years should do the job.

For the period 2000-2006 and the EARS-based water balance the annual D (absolute and as percentage of P) is relatively small and fluctuating around zero for the Ourthe (mean $D = -32\text{mm y}^{-1}$, mean $D/P = -4\%$) and Ourthe Orientale (-12mm y^{-1} , -2%), but always negative or zero (2002) for the Ourthe Occidentale (-129mm y^{-1} , -15%). D in the water balance with the MOD16 estimates is much larger and predominantly positive with an annual mean D of $+50\text{mm y}^{-1}$ (4%) for the Ourthe, $+92\text{mm y}^{-1}$ (8%) for the Ourthe Orientale and -24mm y^{-1} (-4%) for the Ourthe Occidentale.

Based on these figures the EARS evaporation estimates seem reasonable for the Ourthe and Ourthe Orientale: D_i yields the largest negative values for the driest years, indicating water storage depletion. The highest positive D_i occurs in 2007, a relatively wet year. In the remaining years negative and positive D_i do not correlate completely with dry and wet years, but the relative magnitude is smaller than $\pm 5\%$ of the annual precipitation. In Figure 5.13 it can be seen that the cumulative evaporation follows the cumulative difference between rainfall and discharge with a small deviation in the Ourthe, and rather nicely in the Ourthe Orientale. For the Ourthe Occidentale however the deviation is considerable, and the always negative D_i cannot be explained by an accumulating ground water storage depletion. It thus indicates an overestimation of the evaporation, an overestimation of the discharge, unaccounted groundwater flow into the subcatchment (sandstone lithology), or an underestimation of the precipitation. Comparing the three catchments, the annual evaporation according the EARS estimate for the Ourthe Occidentale is slightly higher than, but with a pattern similar to the mean evaporation of the whole catchment. For the Ourthe Orientale the temporal pattern over the years is slightly different. Based on the catchment characteristics, see Section 3.1, there are no obvious reasons for an additional overprediction of $E_{RS,EARS}$ in the Ourthe Occidentale subcatchment: the percentage agriculture is slightly higher to the cost of forest compared to the whole catchment and the Orientale subcatchment, which is more likely to result in a reduced overprediction compared to the other catchments; NDVI of all catchments is approximately equal; topographic character-

The period 2000-2010 shows a similar picture: -32mm y^{-1} (4%) for the entire catchment, 5mm y^{-1} (-0.01%) for the Ourthe Orientale and -124mm y^{-1} (14%) for the Ourthe Occidentale subcatchment.

istics (aspect, slope, elevation) and geology are similar for the two subcatchments. What is different is the runoff coefficient, which is relatively high for the Occidentale (0.51) compared to the Ourthe and Ourthe Orientale (both 0.44), with relatively low precipitation (mean annual P of 987mm y^{-1} versus 1020mm y^{-1} and 1099mm y^{-1} respectively) and high runoff (503mm y^{-1} , 450mm y^{-1} and 489mm y^{-1} respectively). Underestimation of the precipitation in the Occidentale subcatchment - no precipitation data is used from the higher areas in this subcatchment - explains both the large negative D_i and the (too) high runoff coefficient for this specific subcatchment.

Although different, based on the annual water balance the MOD16 estimates seem reasonable for the Ourthe catchment and Ourthe Occidentale subcatchment. For the Ourthe Orientale however, there is too much water accumulating over the years. Furthermore, since the precipitation in the Ourthe Occidentale is most likely underestimated by not including rainfall data from the higher elevations, D in the water balance for the Ourthe Occidentale will be higher than shown in Figure 5.13 and will turn positive as well. This given, the MOD16 product structurally underestimates the evaporation (assuming correct discharge data) leading to too much water in the catchment.

Table 5.5: Annual water balance for the Ourthe catchment

Year		2000	2001	2002	2003	2004	2005	2006	2007	2008	2009	2010
P	(mm y^{-1})	1159	1234	1225	795	970	831	977	1146	1080	970	830
Q	(mm y^{-1})	509	609	571	317	369	340	409	527	540	390	369
$E_{RS,EARS}$	(mm y^{-1})	595	642	614	616	621	589	619	567	593	608	557
$E_{RS,MOD16}$	(mm y^{-1})	511	520	542	569	528	545	507	-	-	-	-
$P - Q - E_{RS,EARS}$	(mm y^{-1})	55	-16	41	-138	-20	-97	-51	52	-53	-28	-96
$P - Q - E_{RS,MOD16}$	(mm y^{-1})	139	105	112	-91	73	-53	61	-	-	-	-

Table 5.6: Annual water balance for the Ourthe Orientale subcatchment

Year		2000	2001	2002	2003	2004	2005	2006	2007	2008	2009	2010
P	(mm y^{-1})	1182	1297	1282	843	1082	924	1029	1238	1218	1060	929
Q	(mm y^{-1})	552	659	612	337	409	389	423	571	633	401	390
$E_{RS,EARS}$	(mm y^{-1})	629	662	632	614	643	566	594	562	586	619	551
$E_{RS,MOD16}$	(mm y^{-1})	495	506	524	563	506	527	494	-	-	-	-
$P - Q - E_{RS,EARS}$	(mm y^{-1})	1	-24	39	-108	30	-31	12	104	-1	41	-12
$P - Q - E_{RS,MOD16}$	(mm y^{-1})	135	132	146	-57	168	7	111	-	-	-	-

Table 5.7: Annual water balance for the Ourthe Occidentale subcatchment

Year		2000	2001	2002	2003	2004	2005	2006	2007	2008	2009	2010
P	(mm y^{-1})	1141	1219	1221	746	875	772	948	1122	1069	1008	736
Q	(mm y^{-1})	617	682	591	342	438	350	417	632	608	459	393
$E_{RS,EARS}$	(mm y^{-1})	612	636	627	641	633	605	635	560	574	611	551
$E_{RS,MOD16}$	(mm y^{-1})	507	512	530	558	514	532	500	-	-	-	-
$P - Q - E_{RS,EARS}$	(mm y^{-1})	-88	-99	2	-237	-196	-183	-104	-70	-113	-63	-208
$P - Q - E_{RS,MOD16}$	(mm y^{-1})	18	25	100	-154	-77	-110	31	-	-	-	-

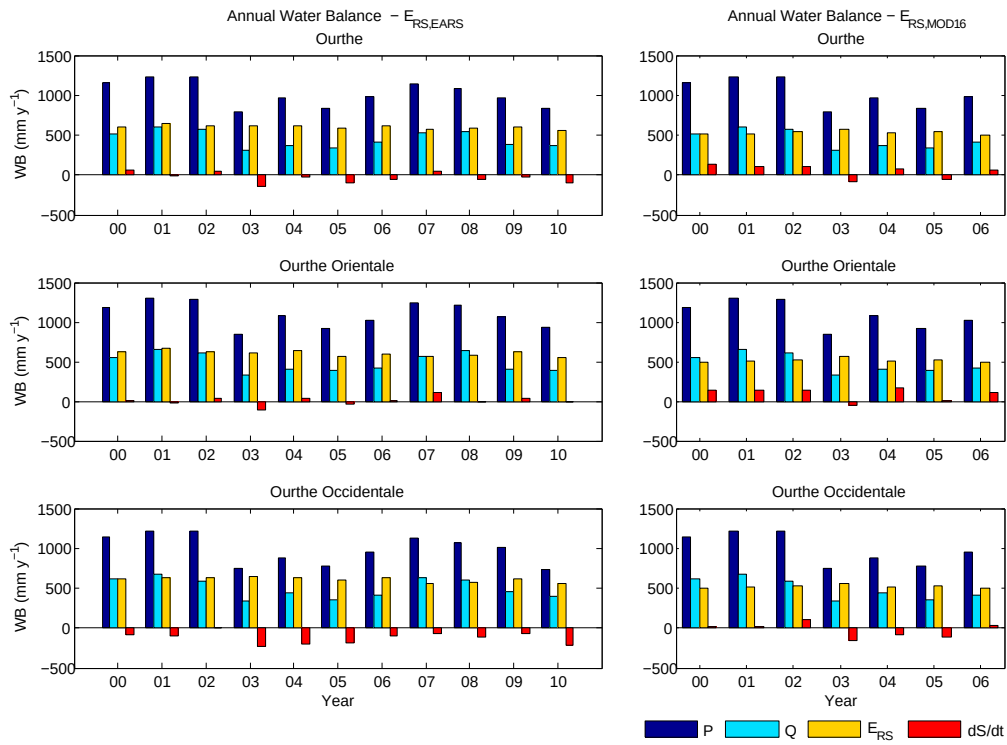


Figure 5.14: Annual water balance of the Ourthe (top) and the subcatchments Ourthe Orientale (middle) and Ourthe Occidentale (bottom), with MOD16 evaporation estimates, for the years 2000 - 2006.

5.4 CONCLUSIONS

The three remotely sensed evaporation products are validated against eddy covariance measurements at five sites in the vicinity of the catchment. The agreement between the remotely sensed evaporation estimates and the ground based eddy covariance measurements is generally poor for all products. The distance between the limits of agreement (even without taking confidence intervals into account) ranges from half to several times the range of measured evaporation (E_{EC}).

WACMOS largely overestimates evaporation compared to the eddy covariance measurements. The version 1.0 product has troubles in estimating the evaporative fraction (Λ), likely because of an acknowledged error in the description of the stability functions. Given the extremely poor correlation between Λ_{RS} and Λ_{EC} , the correlation between $E_{RS,WACMOS}$ and E_{EC} is surprisingly good, but generally worse than for MOD16 or EARS. The WACMOS product is not further used in this research.

The MOD16 algorithm is merely a model with distributed land surface (observed) and meteorological (modelled) input. Estimating the sensible heat flux from the land surface temperature - although here also the accuracy depends on the applied parametrization of the relation, as in the Penman-Monteith equation - seems to be a more independent observation. However, the order of magnitude of the MOD16 estimates agrees relatively well with the order of magnitude of the eddy covariance evaporation measurements. The difference between E_{EC} and $E_{RS,MOD16}$ is, especially for the forested sites, periodical, in the growing season. By comparing Penman-Monteith evaporation based on point scale meteorological data with the remotely sensed and eddy covariance observations, it was shown that the difference in E_{EC} and $E_{RS,MOD16}$ is not caused by limitations of the algorithm itself, but most likely by the scale of the input data. Although the scale of the remotely sensed input in the model is relatively small, the meteorological data applies to much wider scales. This could cause a discrepancy between the ground based measurements and estimates at the scale of the grid cell. At larger scales however, pixels probably compensate for each other. An important limitation of the algorithm is the estimation of the canopy conductance. In the MOD16 algorithm g_s is constant for all biome types, multiplied by the maximum LAI. At the five sites $g_{s,max}$ deviates rather much (more than maximum LAI), even within the same land cover classes. Jalhay is the strongest example hereof. The fact that soil moisture availability is not taken into account in the MOD16 algorithm was not shown to be a problem. It should be mentioned that precipitation in 2008 was 'normal'. In dryer years there could be an unaccounted effect of soil moisture stress. At the catchment scale, a water balance analysis has been performed for the period 2000-2006, based on the remotely sensed evaporation, precipitation from ten ground stations and discharge data. The average annual storage change in this period is 4% of the annual precipitation for the Ourthe, 8% for the Ourthe Orientale and -4% for the Ourthe Occidentale. Precipitation in the Ourthe Occidentale is known to be underestimated due to a missing station at higher elevations. This means that the average annual storage is slightly higher than indicated above for the Ourthe catchment and especially for the Ourthe Occidentale, meaning accumulating water in the catchment.

For the EARS evaporation estimates, because the scale of the eddy covariance evaporation is not representative for the size of the grid cells of the EARS evaporation product, the limits of agreement between E_{EC} and E_{RS} are not considered relevant. Observed periodical differences for the agricultural sites can logically be explained by heterogeneity of the land cover within the grid cells, but this is not examined in detail with additional data. From the validation at the forested sites it became apparent that evaporation at forested sites is overestimated at most of the days during the growing season. This is attributed to the fact that 1) the temperature distribution within the canopy is not taken into account and 2) in the determination of the daily sensible heat flux from the ratio of the instantaneous sensible heat to the available energy, the heat storage in and under the canopy S_t , the ground heat flux G and the energy used for photosynthesis F are not taken into account. This results

in an underestimation of the daily sensible heat and thus an overestimation of the latent heat. The temporal trend however is captured, reflected in the relatively high correlation between E_{EC} and $E_{RS,EARS}$ for the forested sites. At the catchment scale, the annual storage change in the period 2000-2006 is on average -4% of the precipitation for the entire catchment, -2% for the Ourthe Orientale and -15% for the Ourthe Occidentale. Given the underestimation of precipitation in the Ourthe Occidentale, these figures are acceptable. A positive mean bias as observed at the eddy covariance sites for especially forested areas apparently falls within the accuracy of the other data sources.

Based on the results of the validation at catchment scale EARS performs slightly better in terms of water balance closure, yet the MOD16 estimates are acceptable as well. The benefit of the EARS product is that the evaporation is determined on a daily time step, although on a relatively low spatial resolution. Mainly for this reason the EARS product is chosen to be applied as forcing in the conceptual rainfall-runoff model in the next step of the research.

WATER BALANCE MODEL

6.1 INTRODUCTION

In the previous chapters i) the spatio-temporal vegetation dynamics were explored and areas with similar temporal dynamics could be distinguished, and ii) three remotely sensed evaporation estimates were validated, with the conclusion that the EARS evaporation estimates seem to be acceptable to be used further. This gives us two sources of ancillary data related to/on evaporation to be explored in water balance models.

The main questions to be answered in this chapter are:

- How realistic is the modelled evaporation flux in water balance models when we apply the conventional procedure to describe the evaporation flux (Equation 1.2)?
- Can we get a more realistic representation of the internal model processes, (i.e. better water partitioning in the catchment), if we apply a more realistic evaporation in our models?

Seem acceptable, i.e. the requirements of water balance closure over multiple years and no (large) unexplainable differences between the temporal pattern of point scale measurements and the remotely sensed estimates.

6.2 METHOD

The remotely sensed evaporation estimates are assumed to be a realistic representation of the real evaporation flux. To prevent having to make any assumption on how to model evaporation at the scale of our catchment at this first step, the remotely sensed estimates are used to directly force a water balance model. As such the water partitioning between the catchment and the atmosphere is imposed to the model. Parameters that define the partitioning in the fast and slow runoff components still need to be determined/calibrated. In this first step of exploring the remotely sensed data as forcing, the lumped modelling approach has been adopted for simplicity and the most direct use of $E_{RS,EARS}$. Given its relatively coarse spatial resolution versus the spatial scale of vegetation (Chapter 4) and landscape (Chapter 3) no direct link can be made with the vegetation dynamics. This would however be an interesting combination for better understanding the spatio-temporal heterogeneity in evaporation in the catchment. The issue of scale can be circumvented by an appropriate semi-distributed modelling approach, in which total evaporation is split in its components (E_T , E_I and E_S) at smaller spatial scales than provided by the EARS estimates, linking the two scales. However, in the scope of this thesis, we restrict ourselves to the first logical step: direct forcing of a lumped water balance model with remotely sensed evaporation. The model forced with ‘actual’ evaporation is referred to as $FLEX^E$, see below. The model performance of $FLEX^E$ is compared to the performance of two models with the same model structure but with potential evaporation as forcing as in the conventional procedure. In $FLEX^{Ep}$ potential evaporation is calculated with the Penman equation from point scale meteorological measurements. $FLEX^{Ep,RS}$ uses potential evaporation from the EARS product, which is determined as $0.8R_n$ (Section 2.4.3), see Table 6.1 for an overview. We don’t know exactly how good the remotely sensed estimates are, since we can only validate them on longer time scales and to limited spatial scales. Model performance therefore is measured with information in streamflow data, see below.

It is hypothesized that especially the seasonality in streamflow simulation will be improved by forcing the model to ‘generate’ a more realistic evaporation. Furthermore parameters are expected to be better identifiable, giving more confidence in the internal model processes.

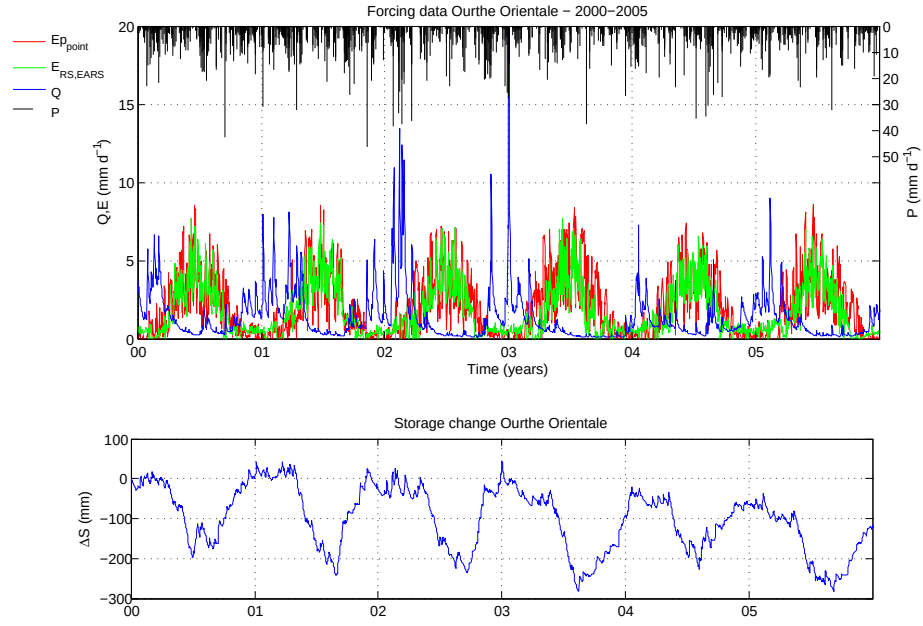


Figure 6.1: Forcing for the Ourthe Orientale in the period 2000-2005. The top panel shows the specific discharge Q (blue), EARS evaporation $E_{RS,EARS}$ (green), Penman evaporation with meteorological data from Saint Hubert E_{p_point} (red) and catchment average precipitation P (black bars). The bottom panel shows the cumulative difference in P , $E_{RS,EARS}$ and Q . In Appendix E this figure is given for the period 2000-2010, from which it can be seen that during this period ΔS does go back to zero.

6.2.1 Forcing data and subcatchment

The water balance models are applied to the Ourthe Orientale subcatchment. In this subcatchment the cumulative storage reaches zero within the 10 year period for which data is available (Figure 6.1), whereas the Ourthe Occidentale subcatchment, and the entire catchment to a lesser extent, slowly ‘empty’ in the 2000-2010 period (Section E.1). Although the error was considered to be acceptable in Chapter 5, if we have the choice we can as well apply the model where we know that the water balance is closed (whether this is for the right reasons is still an unsolved question).

The time step used in the model is daily, in agreement with the time step of the remotely sensed evaporation data ($E_{RS,EARS}$). Although some of the detail of the within-day variations in the discharge is lost, we didn’t want to manipulate the $E_{RS,EARS}$ time series to hourly values, thus adding unknown detail.

Precipitation is measured with pluviometers. Measurements are averaged over the catchment using Thiessen polygons. Potential evaporation is calculated with the Penman equation with point scale meteorological input data from station Saint Hubert. This station is considered to best capture the catchment averaged meteorological data. In the FLEX^E model, the catchment average $E_{RS,EARS}$ time series is determined from resampled grid cells of the original satellite input data to 1/120 degrees resolution. Chapter 3 provides an overview and location of the data.

The period used for calibration of the model is 2001-2004, containing a succession of a relatively wet, dry and average year, see Section 3.1.1 and Section 3.1.3. 2000 (wet) and 2005 (dry) are used for validation of the model. As spin-up time the first year in the timeseries is duplicated.

6.2.2 Model structure

The model structure applied in this study is of the FLEX type (Fenicia et al., 2006). The applied reservoirs are in agreement with the work of Euser et al. (2014). The configuration of the reservoirs and the connecting fluxes are slightly different. Four model variants are considered. The simplest model (variant 0) has 8 (FLEX^E) and 9 ($\text{FLEX}^{Ep(RS)}$) parameters. In the three other variants combinations of two additional processes are considered: a threshold for fast runoff and non-linearity of the stage-discharge relation of the fast runoff component. The first is motivated by the fact that water in the catchment is available for vegetation, but not for streamflow generation. Table 6.1 gives an overview of the model variants.

The model consists of three reservoirs: an unsaturated reservoir (S_u), a fast response reservoir (S_f) and a slow response reservoir (S_s), see Figure 6.2.

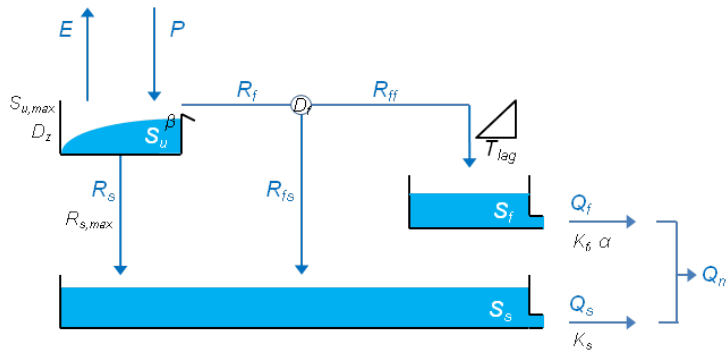


Figure 6.2: Structure of the lumped models FLEX^{Ep} and FLEX^E . Fluxes are shown in blue, model parameters in black and stocks in white.

The conceptualization of the storage distribution and fluxes into and out of the unsaturated reservoir determines the water partitioning. Evaporation, i.e. the sum of interception, transpiration and soil evaporation, originates from this reservoir. In FLEX^E the evaporation is forced to mimic the catchment averaged $E_{RS,EARS}$. In FLEX^{Ep} and FLEX^{Ep,RS} evaporation is modelled using the commonly applied conceptualization after Feddes et al. (2001), see Equation 6.1.

$$E = E_p \cdot \min \left(1, \frac{S_u}{S_{u,max}} \frac{1}{L_p} \right) \quad (6.1)$$

with E_p the potential evaporation (mm d^{-1}), $S_{u,max}$ the maximum storage in the unsaturated reservoir (mm) and L_p (-) the fraction of $S_{u,max}$ below which evaporation is moisture constrained.

Precipitation (P) is partitioned into infiltration (P_i) (Equation 6.2) replenishing S_u , and fast runoff (R_f) (Equation 6.2). The partitioning depends on the soil moisture availability and optionally a threshold value ($D_z \cdot S_{u,max}$) as expressed in the runoff coefficient ρ (Equation 6.4).

$$P_i = (1 - \rho) \cdot P \quad (6.2)$$

$$R_f = \rho \cdot P \quad (6.3)$$

$$\rho = \begin{cases} \left(\frac{S_u - D_z \cdot S_{u,max}}{S_{u,max} - D_z \cdot S_{u,max}} \right)^\beta & \text{if } P > 0 \quad \cap \quad S_u > D_z \cdot S_{u,max} \\ 0 & \text{if } P \leq 0 \quad \cup \quad S_u \leq D_z \cdot S_{u,max} \end{cases} \quad (6.4)$$

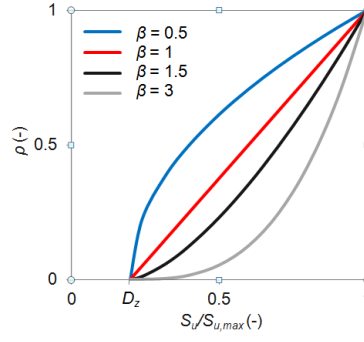


Figure 6.3: Runoff coefficient for different values of β

where the shape factor β (-) accounts for the distribution of the saturation threshold in the catchment, and D_z (-) is a constant fraction of $S_{u,max}$ (Figure 6.3). R_f is routed to either the fast response reservoir as runoff (R_{ff}) (Equation 6.5) or to the slow response reservoir representing preferential recharge (R_{fs}) (Equation 6.6), depending on the constant factor D_f (-).

$$R_{ff} = (1 - D_f) \cdot R_f \quad (6.5)$$

$$R_{fs} = D_f \cdot R_f \quad (6.6)$$

From the unsaturated reservoir water percolates to the slow response reservoir at a maximum rate $R_{s,max}$ (mm d^{-1}), linearly constrained by the relative soil moisture content (Equation 6.7).

$$R_s = \frac{S_u}{S_{u,max}} \cdot R_{s,max} \quad (6.7)$$

The routing of Q_{ff} is represented by a transfer function and the recession of the fast response reservoir, placed in series. The transfer function is a simple convolution, with linearly increasing weights and T_{lag} time steps (d). It accounts for the delay in the system before entering the fast response reservoir and controls the simulation of the rising limb of the hydrograph, whereas the reservoir mainly controls the declining limb (Fenicia et al., 2006). Discharge from the fast response reservoir Q_f (mm d^{-1}) is modelled using a (non-)linear relation between storage and discharge with timescale K_f (d) and optionally a shape factor α (-) (Equation 6.8).

$$Q_f = \left(\frac{S_f}{K_f} \right)^\alpha \quad (6.8)$$

Routing of the slow response components of the discharge is represented by a linear storage-discharge relation for the slow response reservoir only, with timescale K_s (d) (Equation 6.9).

$$Q_s = \frac{S_s}{K_s} \quad (6.9)$$

Total modelled discharge Q_m finally is calculated as the sum of Q_f and Q_s .

Table 6.1: Model variants with different forcing and D_z and α

Model	Forcing		Variant	Parameter range	
	Method	Spatial scale		D_z	α
FLEX ^E	$E_{RS,EARS}$	catchment	0	0	1
FLEX ^{Ep}	$E_p = \text{Penman}$	point	1	[0,1]	1
FLEX ^{Ep,RS}	$E_p = 0.8 R_n$	catchment	2	0	[0,3]
			3	[0,1]	[0,3]

6.2.3 Model evaluation

Model performance is evaluated using the multi-objective optimization approach (Gupta et al., 1998), targeting different aspects of the hydrograph. The best performance that can be achieved by the model, given the available data, is visualized by the Pareto-optimal front. The objective functions are defined such, that the closer the value is to zero, the better is the performance of those aspects of the hydrograph targetted by the objective function. Thus, the closer the Pareto front is to the origin of the objective function space, the better is the overall model performance (Fenicia et al., 2007). The objective functions applied here all are based on the Nash-Sutcliffe efficiency (NSE) (Nash and Sutcliffe, 1970) of signatures of the hydrograph. Equation 6.10 gives the general form of the objective function for the time series of signature X , with n the number of observations, i the specific time step, subscripts m and o denote the modelled and observed values and the overline indicates the average of the n observations.

$$F_X = 1 - NSE_X = \frac{\sum_{i=1}^n (X_{o,i} - X_{s,i})^2}{\sum_{i=1}^n (X_{o,i} - \bar{X}_{o,i})^2} \quad (6.10)$$

The following signatures and associated objective functions are used:

- F_Q , with the NSE of the flow Q . F_Q mainly targets the performance under high flow conditions;
- $F_{\log Q}$, using the logarithm of the flow, which is more sensitive to the low regime;
- F_{FDC} , with the NSE of the flow duration curve (FDC) targeting the flow variability, while disregarding time dependency of the flow;
- $F_{Q_{seas}}$, with the NSE of the 30-day moving average of the flow, which captures the seasonal variation of the flow. This objective function is chosen to test the hypothesis that the seasonality of the water balance components is better captured in FLEX^E than in FLEX^{Ep} (after Yokoo et al. (2008), Ye et al. (2012), although they used the long term mean seasonal flow (the regime curve)).

6.2.4 Calibration

The total number of parameters in FLEX^{Ep} is 10, or 8 when the non-linear fast reservoir and the soil moisture threshold in the runoff coefficient are left out ($\alpha = 1$ and $D_z = 0$). FLEX^E has 9, respectively 7 parameters. The model parameters are calibrated using the first two objective functions, i.e. F_Q and $F_{\log Q}$. A stepwise model calibration has been performed, using Monte Carlo sampling to sample the parameter space. Firstly, after a few iterations of the sampling algorithm, the parameter range for K_s and D_f was narrowed

using visual inspection of the hydrographs of the values of the best performing models, on a semi-log and linear scale respectively. Determining K_s with a proper recession analysis is not straightforward, because of the year round precipitation. Subsequently, in an iterative process with 10.000 to 50.000 parameter sets per iteration, the parameter space of the remaining parameters was narrowed, based on identifiability of the parameters. Since in the $FLEX^E$ model the actual evaporation is applied as forcing, the storage of the unsaturated reservoir can turn negative for certain combinations of $S_{u,max}$, D_z and $R_{s,max}$, in dry years especially. Parameter sets not fulfilling the constraint:

$$S_u \geq 0 \quad (6.11)$$

are considered non-behavioural and removed.

An overview of all model parameters and the initially applied parameter ranges are given in Table 6.2.

Table 6.2: Initial parameter ranges for the lumped models

Parameter		Initial range	
		min	max
$S_{u,max}$	(mm)	100	1000
β	(-)	0	5
D_z	(-)	0	1
D_f	(-)	0	1
$R_{s,max}$	(mm d ⁻¹)	0	5
T_{lag}	(d)	1	10
K_f	(d)	1	10
K_s	(d)	1	200
α	(-)	0.1	3
L_p	(-)	0.4	1.5

6.2.5 Validation

For validation of the model is based on split-record validation, with time series of 1 year before and 1 year after the calibration period. Furthermore two additional objective functions F_{FDC} and $F_{Q_{seas}}$ are used.

6.3 RESULTS AND DISCUSSION

6.3.1 Parameter identifiability

The assessment of parameter identifiability is shown in Section E.2 for the first calibration iteration (that is with the initial parameter range, see Table 6.2). If a parameter is not identifiable, we should pose questions to the realism of the model or to the objective functions we use in model calibration and validation. It is expected that the parameters in FLEX^E are better identifiable, since i) there is one degree of freedom less (no L_p), and ii) evaporation is imposed, which puts a constraint on minimum S_u , affecting $S_{u,max}$, β and, if applied, on D_z .

Figure E.4 and Figure E.5 show the dotted plots of the parameters against the objective functions F_Q and F_{logQ} respectively for all parameters sets and for the behavioural sets fulfilling Equation 6.2.4. $S_{u,max}$ and $R_{s,max}$ are constrained by the condition $S_u \geq 0$ in the four model variations. When no threshold for quick runoff is applied (FLEX^{E,0} and FLEX^{E,2}), β is constrained on the left (lower) side for the behavioural parameter sets as well, to values larger than 0.5 with α (FLEX^{E,1}) and 0.6 without (FLEX^{E,1}). Furthermore a lower left boundary for $S_{u,max}$ is found than for the models with $D_z = 0$. In other words, a smaller storage capacity is required if part of the storage is available for evaporation only, which makes sense. The exact values of the constraints resulting from Equation 6.2.4 are slightly different for F_{logQ} and F_Q , but are not conflicting. K_s seems to ‘favour’ extremely low values according to both F_Q and F_{logQ} , which in combination with high splitter values (D_f) leads to unrealistic(?) highly fluctuating baseflow and summer baseflow higher than the observed discharge. Since the behaviour of the groundwater depletion in summer is rather different in different years (compare 2001 and 2003), it was decided not to perform a recession analysis, and to constrain K_s and D_f based on well behaving parameter sets. The range of K_s is [45,55]; $D_f < 0.1$.

For FLEX^{Ep} and FLEX^{Ep,RS} $R_{s,max}$ is well identifiable for all model variants. $S_{u,max}$ has, as opposed to FLEX^E, no identified constraint on the lower boundary. But the parameter range can be narrowed to lower values than the initial. Applying D_z comes at the cost of the identifiability of $S_{u,max}$, β and L_p . When additionally a non-linear fast flow reservoir is applied the identifiability of the mentioned parameters is improved. For FLEX^{Ep,RS} T_{lag} is better defined than in FLEX^{Ep}. For K_s the same applies as for FLEX^E.

In subsequent iterations the parameter ranges have been further reduced until the values given in Section E.2.

6.3.2 Model performance in terms of objective functions

In Figure E.9 the model performance for the model configurations with (0) and without (3) the soil moisture threshold for fast runoff and non-linearity of the fast reservoir are compared for FLEX^E, FLEX^{Ep} and FLEX^{Ep,RS}. Furthermore the influence of applying a smaller parameter range for K_s together with a smaller D_f , in order to get a more stable (and more realistic?) slow flow component, is shown.

Adding $D_z > 0$ and $\alpha \neq 1$ results in a slight improvement with respect to F_Q for FLEX^E and FLEX^{Ep} and a higher improvement for FLEX^{Ep,RS}. For FLEX^E the value for F_{logQ} slightly decreases. The performance with respect to F_{fdc} increases for all model variants by adding D_z and α . For FLEX^E interestingly there is at the same time a decrease in the ability to simulate the seasonal discharge, which is not the case for the two other models.

Constraining the parameter range for K_s and restricting D_f to smaller values for a more gradual slow flow component, has no consequences with respect to any of the performance measures for FLEX^E. For FLEX^{Ep,0}, the performance with respect to F_{fdc} decreases. A slightly reduced performance for F_{fdc} and a considerable reduction of the combined performance (front) of F_{logQ} and F_Q is visible for FLEX^{Ep,RS,0} and FLEX^{Ep,RS,3}. It should be mentioned here, that the number of iterations in the calibration process is not equal for all

models, nor is the 'convergence speed' towards the optimal model. For some model configurations the optimal set - based on data, model structure and objective functions - might not have been found. This seems especially to be the case for $\text{FLEX}^{Ep,3}$ with free K_s and $\text{FLEX}^{Ep,RS,3}$ with constrained K_s . In the following only the models with constrained K_s and D_f are considered. Referred is to Appendix E for the results of the parameter ranges following from calibration.

Figure 6.5 compares the performance of FLEX^E , FLEX^{Ep} and $\text{FLEX}^{Ep,RS}$ for the model variants with constrained K_s . The figures are similar for model variants 0 and 3. The models forced with potential evaporation largely outperform the model forced with $E_{RS,EARS}$, with respect to all performance indicators except for the flow duration curve. The difference in performance for the models forced with potential evaporation is relatively small. Forcing with Penman evaporation with data from a single station results in lower (better) values for F_Q and F_{logQ} , but especially F_{seas} , than forcing with potential evaporation based on the catchment average remotely sensed $0.8R_n$.

6.3.3 Comparison of the evaporation

The difference in $E_{RS,EARS}$ and the evaporation as modelled with FLEX^{Ep} (blue) and $\text{FLEX}^{Ep,RS}$ is considerable (Figure 6.6). In most years, $E_{RS,EARS}$ has a shorter season (later start) of high evaporation and higher maximum evaporation than the evaporation estimates of FLEX^{Ep} and $\text{FLEX}^{Ep,RS}$. Evaporation modelled with FLEX^{Ep} and $\text{FLEX}^{Ep,RS}$ equals the potential evaporation from autumn (around October) until spring (around May), and is moisture constrained (at least in the model) in summer. $E_{RS,EARS}$ is during summer almost equal to the potential evaporation. Only in 2003, which had an extremely hot summer, and a relatively low annual precipitation (not an extremely dry summer), a period of reduced evaporation is observed for $E_{RS,EARS}$. It should be noted here, that interception was not taken into account in $\text{FLEX}^{Ep,RS}$, although the importance of this flux is acknowledged, see e.g. Savenije (2004), Gerrits et al. (2010). In FLEX^E it is taken into account in a lumped way, under the assumption that the time scale of interception is less than one day. Forest floor interception however might have longer time scales. In winter the potential as well as the actual evaporation according to EARS is relatively high compared to the point-scale Penman evaporation estimates. The high winter $E_{RS,EARS}$ values are preceded by a dip at the end of the growing season (October), caused by something different than energy availability (not visible in $Ep,RS,EARS$). Evaporation from the two models forced with potential evaporation mainly differs in winter, when evaporation according to FLEX^{Ep} is much lower than $\text{FLEX}^{Ep,RS}$ (despite the strangely behaving clear days).

6.3.4 Comparison of the discharge

Looking at the hydrographs of the seasonal Figure 6.7 and daily Figure 6.10 flow, the most notable difference between the models forced with potential and actual evaporation occurs during spring (April, May), which is the start of the growing season (see for the growing seasons Chapter 4). At this time in the year the flow as modelled by FLEX^E largely overestimates the observed flow. In 2002 and 2004 $\text{FLEX}^{Ep,(RS)}$ overestimates the flow in this period as well, although to a lesser extent, but in 2003, which had a relatively dry winter/spring this is not the case. Furthermore, the first peak runoff generation at the end of the growing season (September/October) is underestimated by FLEX^E , whereas it is overestimated by $\text{FLEX}^{Ep,(RS)}$.

The pareto-optimal parameter sets for FLEX^E generally have a different slow flow behaviour than $\text{FLEX}^{Ep,RS}$, with lower, more stable baseflow. Whether this indeed reflects a more realistic behaviour should be further investigated. In summer 2003, the hottest year in the simulated period, the lowest discharge occurs, which with the current combination of parameters D_f , $R_{s,max}$ and K_s is always overpredicted, in all models.

From the perspective of the model the spring time overprediction in $FLEX^E$ is caused by too much storage in the unsaturated reservoir (Figure 6.9) and potentially by too high baseflow as well. For $FLEX^{Ep}$ and $FLEX^{Ep,RS}$, for which evaporation increases earlier in spring (Figure 6.6), soil moisture storage is already reduced and a large part of the precipitation is stored in the unsaturated reservoir. The underestimation of the flow at the end of summer for $FLEX^E$ is caused by too little soil moisture, so that precipitation is stored and only little runoff is generated.

6.3.5 Where could it go wrong?

Clearly there is a discrepancy between i) the evaporation as modelled by the models forced with potential evaporation and the remotely sensed evaporation estimates, and ii) the ability to mimic the observed discharge. In terms of the chosen objective functions the models forced with potential evaporation have a (much) higher performance during both the calibration and the validation period. It was hypothesized however that at least the simulation of the seasonality of the streamflow data would be better with $FLEX^E$ than in $FLEX^{Ep(RS)}$. Now the question is what does this mean? Which of the evaporation estimates is a more realistic representation of reality?

If $E_{RS,EARS}$ is right

First assuming that the remotely sensed evaporation estimates are right, the most obvious reason for a better performance of $FLEX^{Ep}$ and $FLEX^{Ep,RS}$ is that parameters compensate for an erroneous representation of catchment processes (the issue of equifinality, (e.g. Beven, 1996)). The more degrees of freedom in the model the easier it is to mimic a certain modelling objective. $FLEX^{Ep}$ and $FLEX^{Ep,RS}$ have one extra parameter, namely L_p . Furthermore, $S_{u,max}$ is completely free, whereas in $FLEX^E$ the unsaturated storage is forced to be large enough to supply water for evaporation. In the calibration process L_p and $S_{u,max}$ were initially left free on a wide parameter range. In the iteration steps the parameter range could be narrowed, yet the parameters are not fully identifiable and compensate for each other. Indeed, in the pareto-optimal set, low values for $S_{u,max}$ are associated with lower values for L_p . Fixing L_p , e.g. at 0.5 ((Savenije, 1997)) should lead to a better identifiability of the remaining parameters. In fact the interesting question is not so much why do $FLEX^{Ep}$ and $FLEX^{Ep,RS}$ perform much better. The relevant question is what explains the bad performance of $FLEX^E$ in spring and autumn, still under the assumption that $E_{RS,EARS}$ is right.

In $FLEX^E$, although the initial parameter ranges could be narrowed, especially D_f , $R_{s,max}$, β and α and D_z , not all parameters are fully identifiable. This is reflected in the wideness of the pareto-optimal front as well. $S_{u,max}$ in $FLEX^E$ is not necessarily better identifiable than in the models based on potential evaporation, yet the range of optimal parameters has higher values. Only for these higher values evaporation can take place under nearly unconstrained conditions, as the EARS estimates indicate. To get a better parameter identifiability we should fix some of the parameters related to the fast and slow flow response. From recession analysis, K_s cannot be straightforwardly determined, since precipitation occurs year round. In summer, when $E > P$, the recession is more or less constant in most years, but this seems more associated with a form of fast flow than the actual groundwater flow, given the summer 2003 behaviour. The low flow simulation was assumed to be best represented by narrowing the range of K_s to (45d-55d) and $D_f < .1$ or $.2$, as described, based on a limited number of model runs. Yet, afterwards, the value for K_s is potentially too low. Increasing the response time should go simultaneously with decreasing percolation and/or preferential recharge. Apart from fixing parameters, it could also be worthwhile to remove parameters by reducing complexity. The model we started the modelling exercise with is a relatively complex model, in which many assumptions have been made on the dominant processes. Although this model was based on prior research

on this catchment (Euser et al., 2014), it could be elucidating to start with a simpler model and stepwise add complexity if required.

Based on the model results of the current model, we should think of processes that can explain a water ‘transfer’ from spring, through summer, to autumn. Especially the observed water ‘excess’ in spring, right at the start of the growing season, is an interesting observation. If we think in terms of storages, there should be a place in the catchment where water is stored, but cannot come to runoff. In principle the soil moisture threshold D_z is a candidate, which, together with preferential recharge, fits in the theory of the two water worlds: water for streamflow generation and water for vegetation in a compartmented system (Brooks et al. (2009), McDonnell (2014)). In the currently applied model, with its 9 degrees of freedom, the value for D_z could not be identified based on the calibration process in which D_z and $S_{u,max}$ compensate for each other. Fixing $S_{u,max}$ could simply be done by determining the maximum observed range in the cumulative storage in the period 2000-2010, which is more or less in the line with Gao et al. (2014), although here actual evaporation can be used, taking into account interannual storage transfer. Having $S_{u,max}$ fixed, D_z should be better identifiable. Percolation from the unsaturated to the slow reservoir doesn’t fit in the theory of the two water worlds, and its removal from the model should be considered. From the current model results it is difficult to predict whether the above mentioned adaptations can solve the overprediction of the streamflow in spring. Probably it mainly affects the streamflow in autumn, when soil moisture has to be supplied. In autumn the flow is already underestimated, which might get worse.

Another possibility is that not less, but more complexity should be introduced by taking spatial information of catchment heterogeneity into account. Looking at the results of the principal component analysis (Chapter 4), in the third principal component a clear pattern is visible in the Ourthe Orientale subcatchment, splitting the subcatchment in roughly two areas with different phenology or growing season. The component distinguishes areas with vegetation with a dual season (agriculture) with increasing NDVI earlier in spring, and areas with a single season with a relatively small contrast between summer and winter (dominantly needleleaf forest). Differences in phenology or growing season should be reflected in transpiration and potentially result in spatial differences in water partitioning. Lumping these areas could result in an erroneous stream flow simulation. As mentioned, the relatively coarse spatial resolution of the EARS product does not allow for a direct comparison between vegetation dynamics and evaporation dynamics at the appropriate scale. This could be circumvented by a semi-distributed modelling approach. It however requires assumptions on the magnitude of evaporation from different components. And although the pattern in NDVI is known for the subareas, which should be reflected in the pattern of transpiration at least at the start of the season, erroneous assumptions on the relative magnitude of different vegetation classes seem ‘dangerous’ and are maybe better calibrated. In topography driven models (Savenije, 2010) the catchment is semi-distributed based on landscape characteristics in hydrological similar units. This allows for a better, landscape unit specific, process representation, in a parsimonious way. It is suggested to combine the topography based classification of dominant processes with the vegetation patterns derived in Chapter 4. Since vegetation is closely related to the landscape, and the spatial resolution of the vegetation data is relatively high, patterns partially overlap.

If FLEX is right

If the model is right, to a certain degree at least, the overestimation of the streamflow in spring and the underestimation in autumn are a sign that the temporal pattern in the evaporation estimates of EARS is not correct: the spring increase in evaporation should start earlier and summer and maximum evaporation should be lower. An overprediction of the summer evaporation fits in the point scale observation that the EARS estimates for forests are too high, especially mid summer. Since about 44% of the catchment is covered by forests, there should be an effect of this. Since the multi annual water balance seems to be correct given the data on precipitation and specific discharge, somewhere in the catchment

or at some time, the overprediction of forest evaporation should be compensated for by an underestimation of evaporation. In theory this could be at the start of the season, as indicated by the model results. From the point scale comparison however there is nothing to confirm this. Furthermore, the spring increase in catchment average NDVI occurs at about the same time as the increase in evaporation. The spatial variation of the EARS evaporation in relation to the average NDVI per pixel has not been examined. What is notable from the point scale comparison as well as from the catchment scale temporal evaporation pattern is a dip in evaporation lasting a few days in the period September-November and January-February, which is not observed in the EC measurements. This was not yet analysed in the validation study of Chapter 5. Figure D.2 shows the time series of E_{EC} and $E_{RS,EARS}$ for the daily measurements. Clouded and cloud free days are indicated. It appears that in winter, on clear days, according to the parametrization of the sensible heat in the EARS algorithm, all available energy is transferred into sensible heat and latent heat is zero. This is not in agreement with the EC measurements of evaporation. Here thus an error seems to be introduced. The magnitude of the error is restricted to the number of clear days (not that many) and magnitude of wintertime evaporation (not that high). Other possible reasons for a time lag in springtime increase and autumn decrease could be related to the changes in the atmosphere and at the surface, affecting atmospheric transmissivity, surface roughness, albedo, wind speed. How exactly this influences the EARS evaporation estimates is not examined at this stage. This would be an interesting follow-up study.

6.4 CONCLUSION

The evaporation modelled with the conventional conceptualization of the evaporation flux ($FLEX^{Ep}$), is crucially different from the EARS remotely sensed evaporation estimates (E_{RS}). The seasonality has a different timing with especially a much later increase in spring according to $E_{RS,EARS}$. Furthermore, where according to ($FLEX^{Ep}$) evaporation is moisture constrained during most of summer, $E_{RS,EARS}$ equals the potential evaporation rate during the majority of the time. Evaporation modeled with $FLEX^{Ep}$ and $FLEX^{Ep,RS}$ mainly differs in winter.

The model performance with respect to the objective functions F_Q and F_{logQ} used in the calibration, and for two additional objective functions F_{fdc} and F_{seas} is much better for the models forced with potential evaporation $FLEX^{Ep}$ and $FLEX^{Ep,RS}$ than for the model forced with actual evaporation $FLEX^E$. The hypothesis that especially the seasonality in streamflow can be better modelled with the remotely sensed evaporation estimates as forcing can be rejected. To the contrary: especially in spring and autumn model predictions deviate rather much from the observed flow.

Between $FLEX^{Ep}$ and $FLEX^{Ep,RS}$ differences are relatively small, but $FLEX^{Ep}$ performs better than $FLEX^{Ep,RS}$ as well. Parameter identifiability should be evaluated more formally, but for now it can be concluded that there is no significant difference in parameter identifiability between the models, even though the degrees of freedom in $FLEX^{Ep}$ and $FLEX^{Ep,RS}$ are with one extra parameter higher. There is a difference in the optimal values for $S_{u,max}$, β , and $R_{s,max}$, with generally higher values for the first two and lower for the latter in $FLEX^E$.

Whether the remotely sensed evaporation estimates are good and the (parametrisation of the) conceptual water balance is wrong or whether the remotely sensed evaporation estimates are wrong cannot be fully confirmed based on the current results. Possible lines of improvements are suggested.

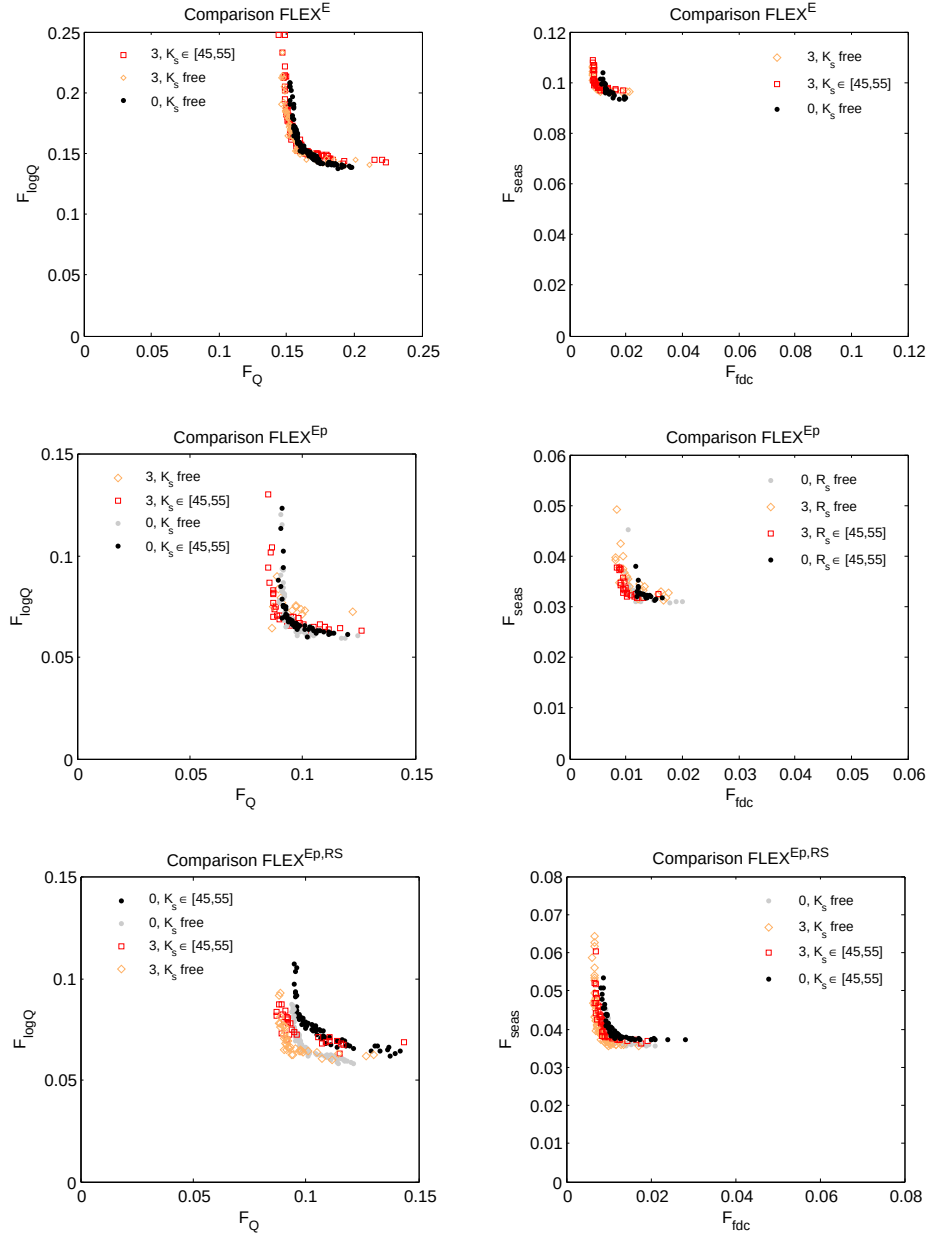


Figure 6.4: Model performance with respect to F_Q , F_{logQ} , F_{fdc} and F_{seas} for $FLEX^E$ (top), $FLEX^{Ep}$ (middle) and $FLEX^{Ep,RS}$ (bottom) for model variants without (0) and with (3) soil moisture threshold and non-linear fast reservoir. The models are calibrated on F_Q and F_{logQ} . The parameter sets in the left and right panel are not equal, but the independent pareto optimal sets according to the two shown performance measures.

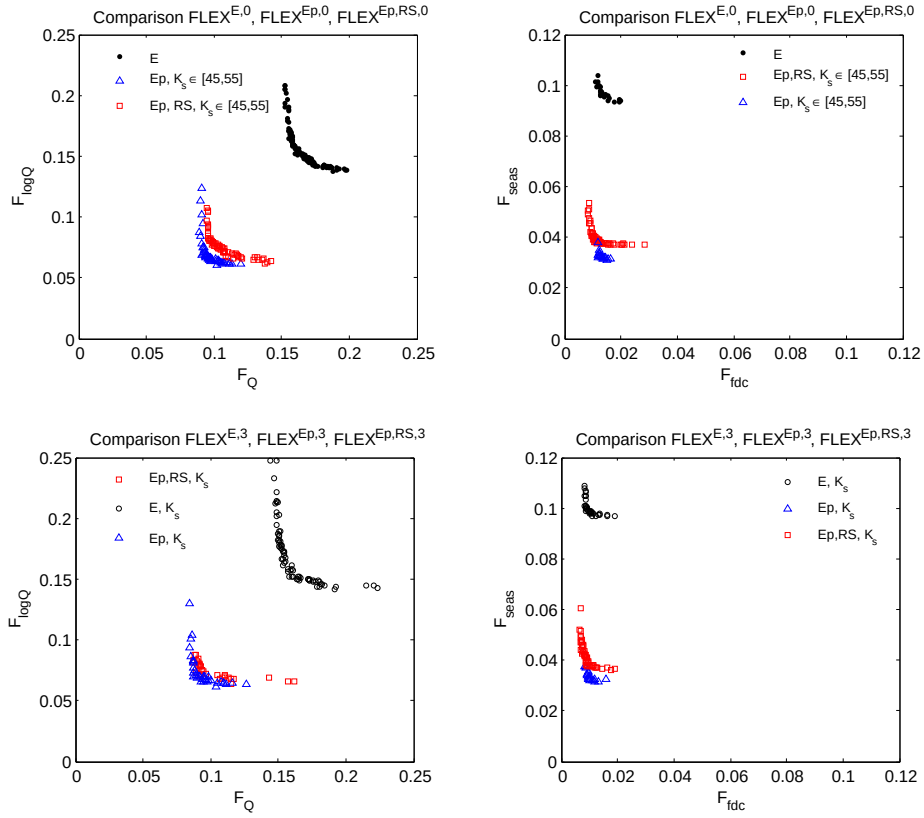


Figure 6.5: Model performance of $FLEX^E$ (black dots), $FLEX^{Ep}$ (blue triangles) and $FLEX^{Ep,RS}$ (red squares) with respect to objective functions F_{logQ} and F_Q (left) and F_{seas} and F_{fdc} (right), for the model variants without (top) and with (bottom) soil moisture threshold for runoff and non-linear fast reservoir. K_s and D_f are constrained based on the behaviour of the slow flow component.

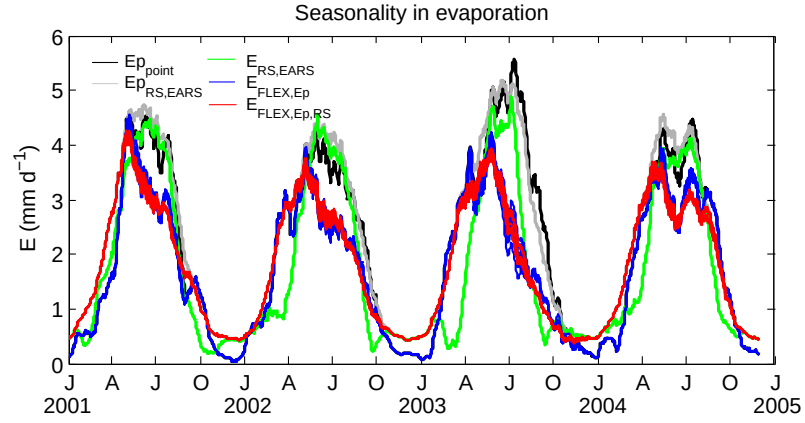


Figure 6.6: Seasonality in evaporation (30-days moving average), according to EARS (green), $\text{FLEX}^{Ep,0}$ (blue) and $\text{FLEX}^{Ep,RS,0}$ (red). The potential evaporation used as input in the latter two models is shown along in black and grey respectively.

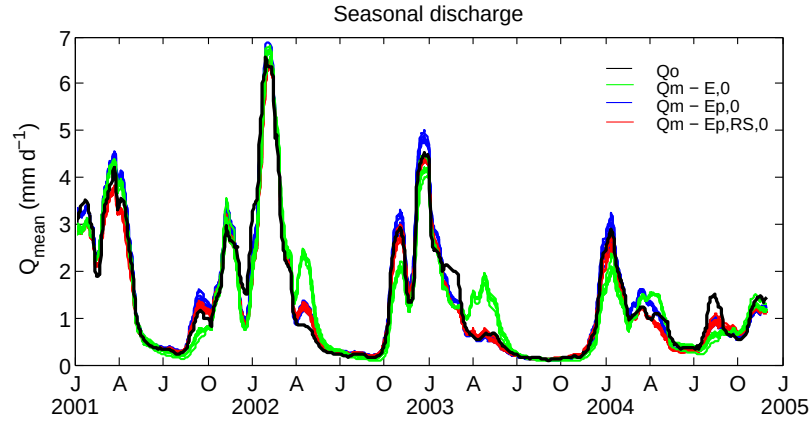


Figure 6.7: Hydrograph with mean monthly flow, calculated as the 30-days moving average, showing the seasonal fluctuations in the discharge. Shown are Q_m of 7 Pareto-optimal parameter sets based on F_Q and $F_{\log Q}$ for $\text{FLEX}^{E,0}$ (green), $\text{FLEX}^{Ep,0}$ (blue) and $\text{FLEX}^{Ep,RS,0}$ (red) and the observed discharge Q_o (black).

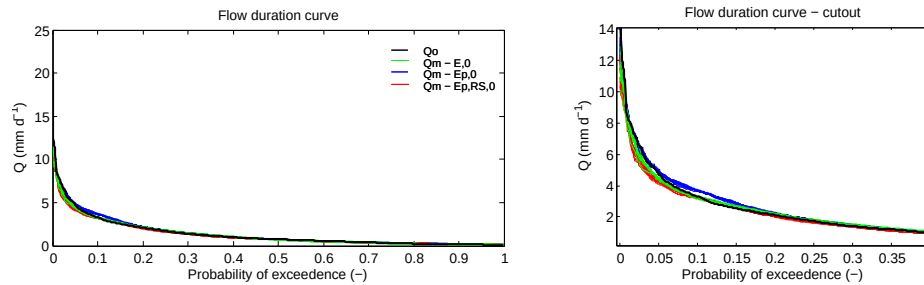


Figure 6.8: Flow duration curve (FDC) for $\text{FLEX}^{E,0}$ (green), $\text{FLEX}^{Ep,0}$ (blue) and $\text{FLEX}^{Ep,RS,0}$ (red) compared to the flow duration curve of the observed discharge (black). Left: full spectrum of flows. Right: cutout of the FDC at the transition between high and low flows, where the divergence from the observed flow and the difference between the models is highest.

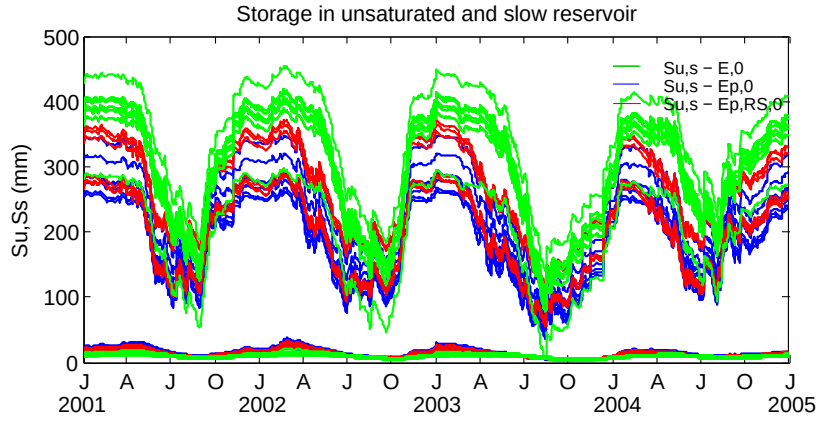


Figure 6.9: Storage in the catchment in the unsaturated reservoir and in the groundwater reservoir. Shown are S_u and S_s of 7 Pareto-optimal parameter sets based on F_Q and $F_{\log Q}$ for FLEX E,0 (green), FLEX Ep,0 (blue) and FLEX Ep,RS,0 (red).

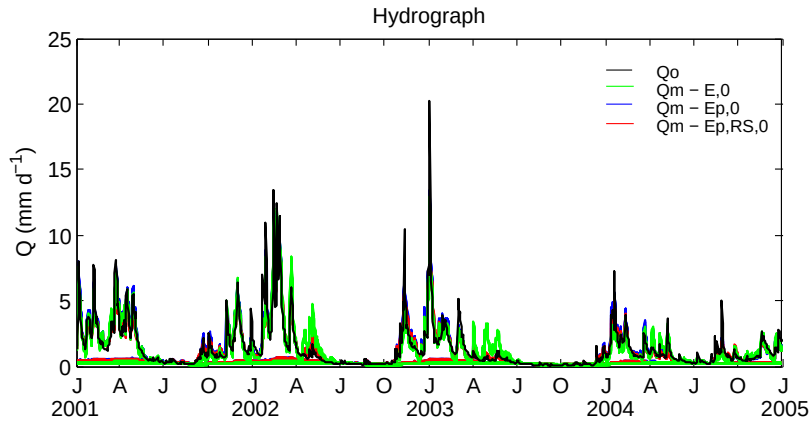


Figure 6.10: Hydrograph with the daily discharge. Shown are Q_m of 7 Pareto-optimal parameter sets based on F_Q and $F_{\log Q}$ for FLEX E,0 (green), FLEX Ep,0 (blue) and FLEX Ep,RS,0 (red) and the observed discharge Q_o (black).

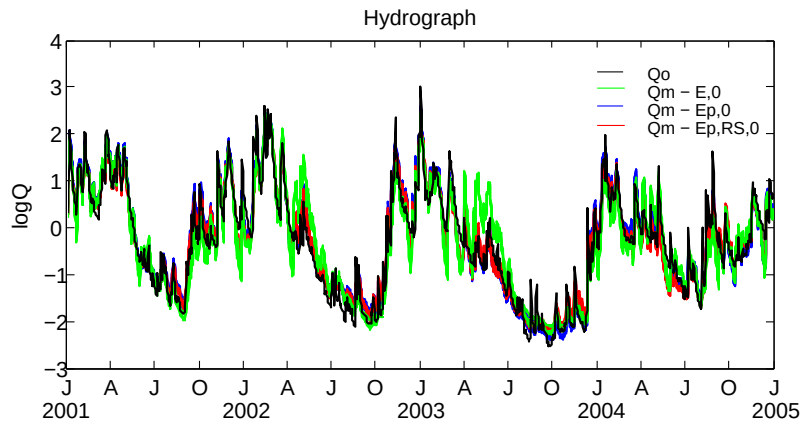


Figure 6.11: Hydrograph of the logarithm of the discharge. Shown are Q_m of 7 Pareto-optimal parameter sets based on F_Q and $F_{\log Q}$ for FLEX E,0 (green), FLEX Ep,0 (blue) and FLEX Ep,RS,0 (red) and the observed discharge Q_o (black).

CONCLUSIONS AND OUTLOOK

7.1 CONCLUSION

In this thesis the use of remotely sensed data to increase our understanding of catchment scale evaporation and water partitioning is examined.

Motivated by our understanding of the control of vegetation on evaporation at the point scale, we investigated how variable vegetation is in the catchment, in both space and time. With a principal component analysis on MODIS NDVI time series, areas with a similar temporal variability in NDVI could be distinguished. The most important variability in the catchment is associated with phenology and agricultural activities. Areas with an increasing trend in NDVI were identified as well, but have a relatively small areal extent. For hydrological implications the observed variability in phenology and growing seasons are probably most relevant.

Motivated by our lack of understanding of catchment scale evaporation, three remotely sensed evaporation products, namely EARS, MOD16 and WACMOS, have been analysed for their use as ancillary data. A validation study resulted in the selection of the EARS evaporation product as a sufficiently reliable and practically suitable product to be explored further. MOD16 performed reasonably, but tends to underestimate the catchment scale evaporation, potentially related to the fact that interception is not taken into account. WACMOS (SEBS) was shown to have an extremely poor correlation between the remotely sensed evaporative fraction and the evaporative fraction determined from eddy covariance measurements.

To test the realism of the commonly applied procedure to calculate evaporation in conceptual rainfall-runoff models, and to see whether the internal water partitioning can be better represented if a more realistic evaporation is applied, a comparative modelling exercise is performed. Lead by the spatial and temporal resolution of the EARS evaporation estimates, a lumped conceptual model was applied on a daily timestep. It was shown that the evaporation modelled with the conventional conceptualization of the evaporation flux ($FLEX^{Ep}$), is fundamentally different from the EARS remotely sensed evaporation estimates (E_{RS}). The seasonality has a different timing, and where according to ($FLEX^{Ep}$) evaporation is moisture constrained during most of summer, E_{RS} equals the potential evaporation rate during the majority of this period. Contrary to our hypothesis, in terms of streamflow simulation, $FLEX^{Ep}$ outperforms the model in which the remotely sensed evaporation estimates - considered as a more realistic estimate of the catchment scale evaporation - are imposed ($FLEX^E$). Especially in spring and summer the streamflow simulations of $FLEX^E$ deviate considerably from the observed streamflow. Furthermore, parameter identifiability was shown not to be significantly better for $FLEX^E$ than $FLEX^{Ep}$.

These results indicate that either the EARS product is not as accurate as we expected, or that the lumped conceptual model does not represent the dominant processes occurring in the catchment. Assuming the latter, this can be explained by i) a too complex model with too many degrees of freedom as the limited parameter identifiability indicates, ii) an erroneous representation of the dominant processes or a wrong hypothesis on the occurring dominant processes and iii) the level of aggregation of evaporation and other hydrological processes is too high and the model too simple to be representative for the heterogeneous catchment. Possible lines of improvements are suggested.

7.2 OUTLOOK

We believe that forcing a model with (reliable) remotely sensed evaporation, gives insight into the realism of the internal model processes, without having to make assumptions on evaporation functioning. The next step would be to test hypotheses concerning the evaporation process itself.

At this point, the two ancillary data sources, namely i) spatio-temporal dynamics in vegetation at a relatively small scale and ii) the EARS remotely sensed evaporation product at coarser spatial scales, have not yet been combined in one modelling approach. We suggest to do this in a topography driven semi-distributed model, combined with the vegetation patterns and associated temporal dynamics derived in Chapter 4. Topography driven semi-distributed models are said to better represent the dominant processes in the catchment (with the conventional procedure of evaporation modelling that is). The information on differences in phenology and agricultural growing season in the (sub)landscape can be used to more accurately model the influence of temporal dynamics in vegetation on evaporation. The use of the remotely sensed estimates in this approach however is not as straightforward as in the lumped model.

Validation of remotely sensed evaporation is difficult due to the issue of scale: the grid size of the remotely sensed products is usually larger than the scale at which we can measure evaporation. Although we have reasons to believe that the EARS evaporation estimates are a better representation of the catchment scale evaporation than the evaporation determined with water balance models, there are some unclarified issues. Investigating the influence of seasonally changing surface and atmospheric characteristics for often constant parametrisations in remotely sensed estimates is recommended. Even more strongly, it is recommended to look further into the overestimation of evaporation in forested areas. Especially the common practice of neglecting the ground heat flux and the heat storage change in/under the canopy in the calculation of the instantaneous evaporative fraction should be reconsidered.

EARS EWBMS ALGORITHM

A.1 MFG VERSUS MSG

The remotely sensed visible and thermal infrared images input to the EARS algorithm until 2005 are from the Meteosat Visible Infra-Red Imager (MVIRI) instrument on board the Meteosat First Generation (MFG) satellites. From 2005 on images from the Spinning Enhanced Visible Infra-Red Imager (SEVIRI) instrument on board Meteosat Second Generation (MSG) satellites are used. The most important difference between the MVIRI and SEVIRI images is the spatial resolution of the images and the number and band width of the spectral channels. The consequence for the evaporation estimates is shown in Figure A.3, where the annual potential evaporation ($0.8R_n$) and actual evaporation based on the MVIRI (MFG) and SEVIRI (MSG) images are compared for the year 2005. Table A.1 gives the statistical summary of the comparison.

Table A.1: EARS based on MFG vs MSG

		MFG 2005 (mm y ⁻¹)	MSG 2005 (mm y ⁻¹)	MFG-MSG (mm y ⁻¹)	Rel. Diff (%)
$E_{RS,EARS}$	Ourthe	579.1	588.7	-9.6	-1.6
	Ourthe Orientale	589.6	565.9	23.7	4.1
	Ourthe Occidentale	591.9	605.0	-13.0	-2.2
$E_{PRS,EARS}$	Ourthe	784.3	820.0	-35.7	-4.5
	Ourthe Orientale	776.7	806.6	-29.9	-3.8
	Ourthe Occidentale	778.5	807.9	-29.4	-3.7

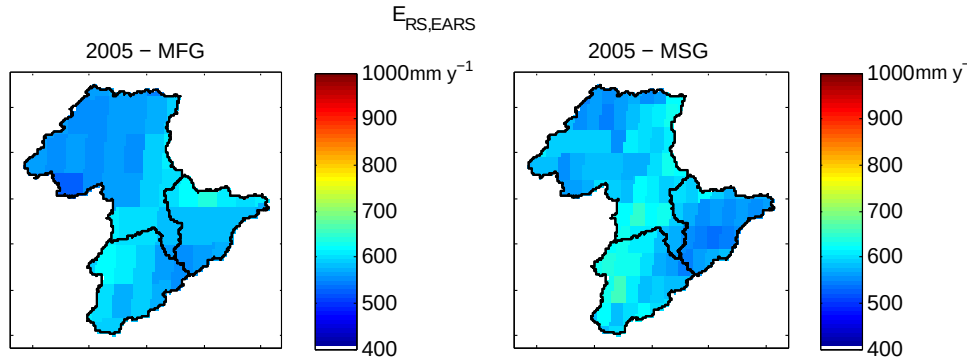


Figure A.1: EARS annual evaporation estimate for the Ourthe catchment in 2005. Left: MVIRI (MFG) images as input data. Right: based on SEVIRI (MSG) images.

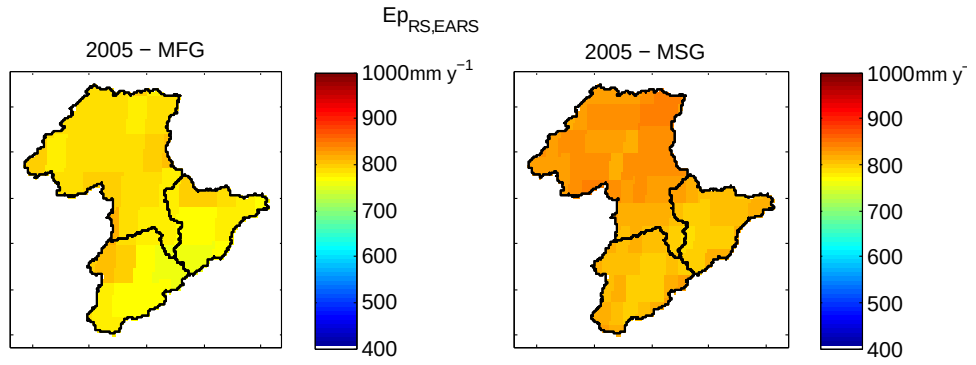


Figure A.2: EARS annual potential evaporation estimate (calculated as $0.8R_N$) for the Ourthe catchment in 2005. Left: MVIRI (MFG) images as input data. Right: based on SEVIRI (MSG) images.

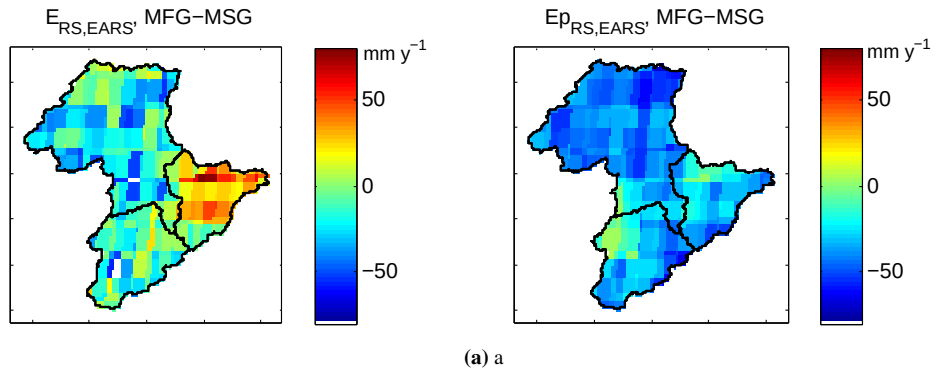


Figure A.3: Difference in EARS annual evaporation estimate based on MVIRI (MFG) and SEVIRI (MSG) images in 2005. Left: Difference in potential evaporation ($E_{p,RS,EARS}$). Right: Difference in evaporation ($E_{RS,EARS}$).

CORINE LAND COVER CHARACTERIZATION

In this research information on the land cover in the study area originates from the CORINE 2006 land cover database at scale 1:100.000 (EEA, 2012). Table B.1 gives the description and relevant features of the land cover classes that are present in the study area. Special attention is given on the presence and type of vegetation per class.

Table B.1: Land cover classes in the CORINE land cover database, for the selection of land covers present in the study area. Summarized from Bossard et al. (2000)

Code	Main class	Subclass	Description
1xx	Artificial areas		
112	Built up area	Discontinuous urban fabric	Areas mainly occupied by dwellings and building used by administrative/public utilities or collectivities, including their connected areas, associated with vegetated areas and bare soil which occupy discontinuous but significant surfaces.
121	Built up area	Industry, Commerce	Areas mainly occupied by industrial activities, including associated lands and access infrastructure.
122	Built up area	Transport	Motorways and railways, including associated installations (stations, embankments), with a minimum width of 100m.
124	Built up area	Airports	Airport installations: runways (concrete, grass-surfaced), buildings and associated lands (grasslands mainly)
131	Built up area	Mineral extraction sites	Artificial areas mainly occupied by open-pit extraction of construction material and other minerals
133	Built up area	Construction sites	Spaces under construction development, earthworks, soil or bedrock excavations.
14x	Artificial non-agricultural vegetated areas		Areas voluntarily created for recreational use.
141	Artificial non-agricultural vegetated areas	Vegetated urban areas	Artificial non-agricultural vegetated areas >25 ha, situated within or in contact with urban fabrics.
142	Artificial non-agricultural vegetated areas	Sport and leisure facilities	Camping grounds, sports grounds, leisure parks, etc.
2xx	Agricultural areas		
211	Agricultural areas	Non-irrigated arable land	Lands with >75% of the area under a rotation system used for annually harvested plants and temporally fallow lands (<3 years), which are not irrigated. Cultivation can be on open field, under plastic or under glass.
222	Agricultural areas	Fruit trees and berry plantations	Parcels planted with fruit trees or shrubs: single or mixed fruit species and fruit trees with permanently grassed surfaces.
231	Agricultural areas	Pastures	Permanently used (>5 years) pastures for grazing (mainly) or fodder production. Dense grass cover of floral composition, dominated by graminacea (grasses).
24x	Agricultural areas	Heterogeneous agricultural areas	Areas of annual crops associated with permanent crops on the same parcel, annual crops cultivated under forest trees, juxtaposed areas of annual crops, meadows and permanent crops, landscapes in which crops and pastures are intimately mixed with natural vegetation or natural areas.
242	Agricultural areas	Complex cultivation patterns	Juxtaposition of small parcels of diverse annual crops, pasture and/or permanent crops, eventually with scattered houses or gardens.
243	Agricultural areas	Natural vegetation mosaic	Land principally occupied by agriculture, with significant areas of natural vegetation.

Table B.1 – continued

Remarks	Surface area specification
Between 30 to 80 % of the total surface should be impermeable.	Vegetated patches <25 ha; vegetated surface area <30%; 30% - 80% impermeable area (if >80% 111)
Includes water retention dams >25ha; agricultural farms >25 ha.	Vegetated patches <25 ha;
Includes compounds of crossroads >25 ha; vegetated areas <25 ha	Vegetated patches <25 ha;
Includes adjacent grass areas and dispersed trees and shrubs; within the buffer zone of airport	not specified
Includes flooded gravel pits and temporary mining pools <25 ha; line vegetation belts if part of buffering/protective zones	not specified
	not specified
Includes parks, lawns, flower beds, gardens; woods in urban fabric; zoological and botanical gardens. City parks are excluded (242)	Vegetated patches >25 ha; >50 % of total area is vegetated
	not specified
Rotational system. Includes cereals, legumes, fodder crops (including grass), root crops, flowers, fruit trees; multi-year plants; semi-permanent crops; drained arable land. Permanent pastures are excluded (231)	Individual plots with a surface area of several ha to tens or hundreds of ha; >75% of parcels under rotational system, annually harvested
Permanent crops, not under rotation system.	not specified
Permanent crops. Includes natural or sown herbaceous species; unimproved or lightly improved meadows; abandoned arable land used as pastures (after 3 years).	<10% - 20% scattered trees and shrubs; <25% arable land; <50 % dispersed trees and shrubs in wooded meadows.
Includes hobby and city gardens; free space >25 ha surrounded by discontinuous built up area (else 112); mixed parcels of annual crops; parcels of grassland; parcels of arable land <25 ha; parcels of permanent crops <25 ha; scattered houses when built up parcels cover <30 % of patchwork area.	<30 % urban fabric (else 112); <75 % of the area is cultivated under a rotation system (else 211)
Includes parcels of arable land <25 ha; orchards, vineyards and berry plantations <25 ha; rests of natural forest, trees, shrubs <25 ha; water bodies; scattered heaps of stones; sporadically houses;	<75% agricultural land (else 21x, 22x, 23x); <75% natural vegetation (else 3xx).

Table B.1: Land cover classes in the CORINE land cover database – continued

Code	Main class	Subclass	Description
31x	Forests		Areas occupied by forests and woodlands with a vegetation pattern composed of native or exotic trees, which can be used for the production of timber or other forest products. Trees are higher than 5 m with a canopy closure >30%.
311	Forest	Broad-leaved forest	Forest with predominantly broad-leaved trees.
312	Forest	Coniferous forest	Forest with predominantly coniferous trees.
313	Forest	Mixed forest	Forests in which neither broad-leaved nor coniferous species dominate.
321	Shrubland	Natural grasslands	Low productivity grassland, often situated on rough, uneven ground. Vegetation is herbaceous (<150 cm high, gramineous species are prevailing), developed under minimum human interference (no mowing, fertilization, stimulation).
322	Shrubland	Moors and heathland	Vegetation with low and closed cover, dominated by bushes, shrubs and herbaceous plants.
324	Shrubland	Transitional woodland-shrub	Bushy or herbaceous vegetation with scattered trees. Can represent woodland degradation or forest regeneration/recolonisation.
41x	Wetlands		Areas flooded or liable to floodings during a great part of the year with a specific vegetation coverage of low shrub, semi-ligneous or herbaceous species.
411	Wetlands	Marshes	Non-forested, low-lying land usually flooded in winter, and more or less saturated by water all year round.
412	Wetlands	Peat bogs	Peatland consisting mainly of decomposed moss and vegetable matter. May or may not be exploited.
512	Water	Water bodies	Natural or artificial stretches of water.

Table B.1 – continued

Remarks	Surface area specification
Includes transitional woodland areas when canopy closure is >50 % and average breast diameter >10 cm; clear-cuts.	>30 % canopy closure; >75 % broad-leaved trees
Includes coniferous wooded land; non-evergreen coniferous trees woodland of larch trees (<i>Larix spp.</i>); clear-cuts;	>30 % canopy closure; >75 % coniferous trees
	>30 % canopy closure; <25 % canopy closure by coniferous species; <25 % canopy closure by deciduous species; patches of same species <25 ha
Includes protected areas; karstic areas; military training fields; areas of shrub formations and scattered trees; grass formations of inundated alluvial plains; small sport airports with non-concreted or asphalted runways which are not used for agriculture or forestry.	>75% of vegetated area is herbaceous vegetation; <25 % shrubs and trees; <25 % bare rock surface; no fallow land (is 211)
Vegetation includes heather, briars, gorse, laburnum, etc ¹ .	<30 % tree-like species in heathland under recolonizing process (else 324).
Includes young broad-leaved and coniferous wood species with herbaceous vegetation and dispersed solitary trees or small forests; agricultural lands under recolonization process with forest trees; young plantations; open clear-felled or regeneration areas with regrowth during transition stage lasting <5-8years; damaged forest areas with more than 50% dead trees; wooded fen, bog and transitional bog.	>30 % forest trees in agricultural lands under recolonization process; 30% - 50% trees or small forests (<25 ha) in natural grasslands (321, 31x); >10 % scattered trees on bare rocks.
Includes fens and transitional bogs without peat deposition or on peaty ground <30 cm thick, with specific vegetation composed of reeds, bulrushes, rushes, sedges, tall herbs and sphagnum hummocks, often with alder or willows and other water plants ² ; water-fringe vegetation of reed beds, sedge communities, fen-sedge beds, tall rush swamps, riparian cane formations.	not specified
Drained and wooded peat bogs are excluded.	not specified
	Water body >25 ha; groups of small lakes (<25 ha) with: a) combined total area >25 ha and b) >75 % of the surface is free water.

PRINCIPAL COMPONENT ANALYSIS

C.1 DISTRIBUTION PRINCIPAL COMPONENT SCORES PER LANDUSE CLASS

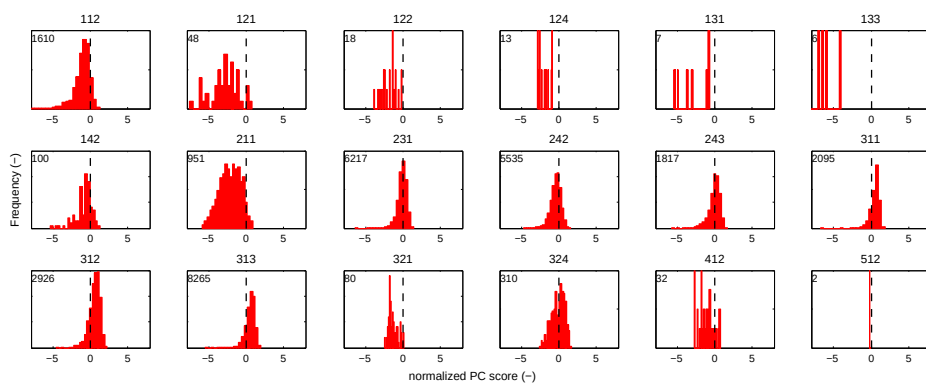


Figure C.1: Distribution of the normalized score of PC1 per land use class. The score is plotted on the x-axis. On the y-axis the frequency of the score is given. The number of pixels per land use class is shown in the upper left corner. See Section 3.1 for the codes of the land uses.

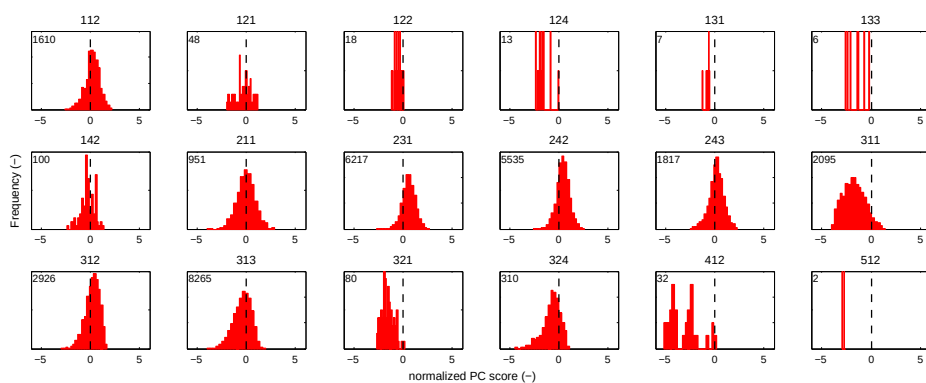


Figure C.2: Distribution of the normalized score of PC2 per land use class.

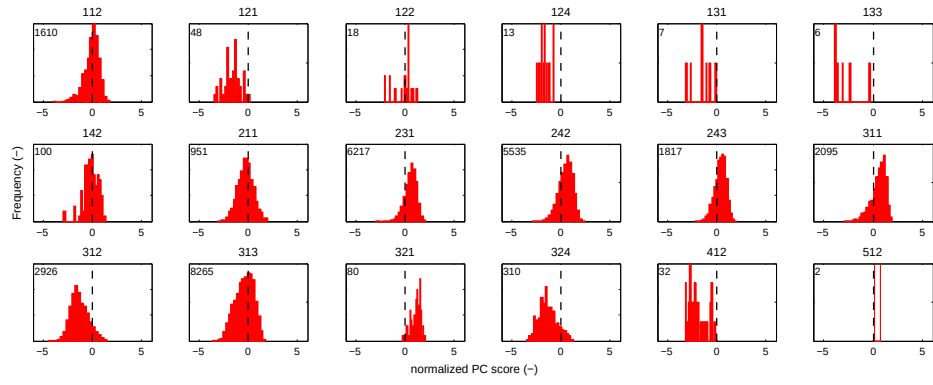


Figure C.3: Distribution of the normalized score of PC3 per land use class.

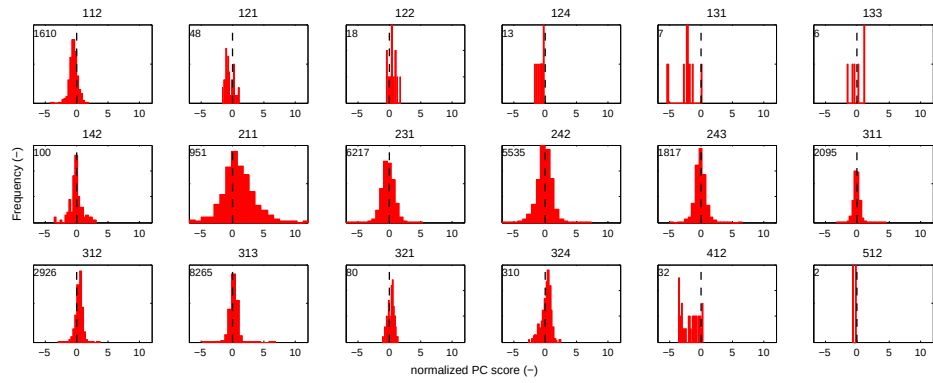


Figure C.4: Distribution of the normalized score of PC4 per land use class.

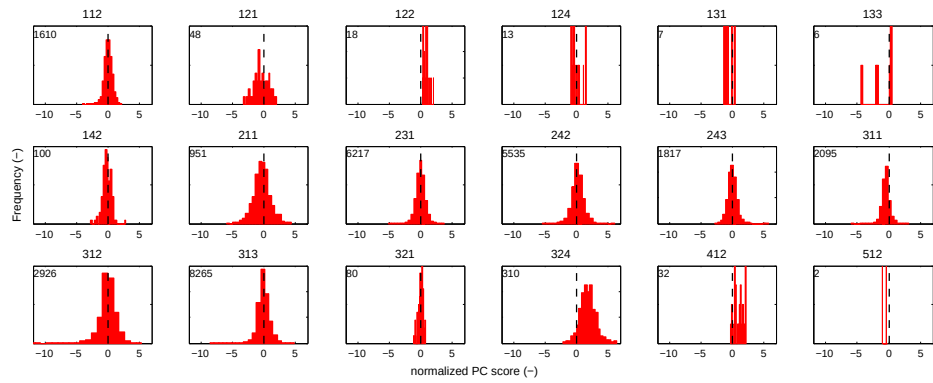


Figure C.5: Distribution of the normalized score of PC5 per land use class.

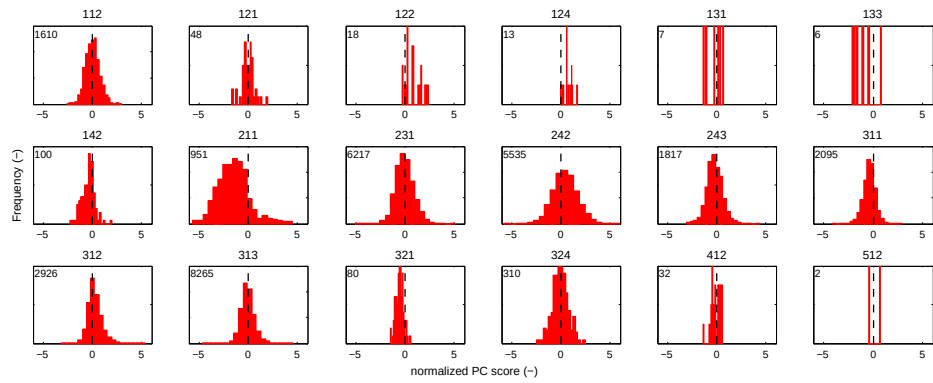


Figure C.6: Distribution of the normalized score of PC6 per land use class.

Table C.1: Characteristics of principal components 1-3

Y	ψ	Temporal pattern of γ_i	Coinciding γ_i subset	Distinguishes:	Characteristics and dominant land cover of high scoring pixels	Land cover	Characteristics and dominant land cover of low scoring pixels	Land cover
1	38.1%	No seasonality	high and mean	Vegetation density	High vegetation density: Forests	31x	Sparsely vegetated or low density: Discontinuous built up area ($f_c = 20\%-30\%$); natural grasslands (low winter NDVI); arable land (low summer NDVI); wetlands (low winter NDVI).	1xx 321 211 412
2	13.7%	Single season, maximum in summer (June/July)	- low	Seasonality differences between land cover classes	Dual seasonality and/or rel. high late autumn and winter NDVI. Maxima in spring (early April-early June) and late summer (end August - half September (Oct in 2001)) Agriculture (downstream part catchment has dominantly higher Y_2 scores ($y_2 > \sigma$), coinciding with slightly higher winter NDVI, than agriculture at higher elevations ($0 < y_2 < \sigma$)) Coniferous forest (few pixels, distributed over catchment)	231 242 243 312 (313)	Single season with high contrast between summer and winter NDVI. Maximum in June-July Deciduous forest (almost all pixels); Natural grasslands; Transitional shrubland/woodland; Wetlands; Discontinuous built up area	311 (313) 321 324 412 1xx
3	6.18%	Dual season, maxima in spring (early-end May (mid June in 2006)) and late summer (end August (mid October in 2001))	high	Seasonality differences between (and within) land cover classes	Dual seasonality and rel. high amplitude. Maxima as γ Agriculture (upstream part catchment has slightly higher amplitude; difference between $y_3 > 0$ and $y_3 > \sigma$ small) Deciduous forest (subset in downstream part catchment, higher amplitude (both higher summer and lower winter NDVI)) Natural grassland	231 242 243 311 (313) 321	Single season with relatively low contrast between summer and winter maximum in June/July Coniferous forest (includes almost all pixels); Transitional shrubland/woodland Wetland Non vegetated areas	312 (313) 324 412 121 123,

Table C.2: Characteristics principal components 4-6

Y	ψ	Temporal pattern of γ_i	Coinciding subset	Distinguishes:	Characteristics and dominant land cover of high scoring pixels	Land cover	Characteristics and dominant land cover of low scoring pixels	Land cover
4	2.78%	Dual 'season' Maxima in spring (end May-early June) and winter (early December) Summer and autumn pattern opposite of γ_3	high	Seasonality differences within land cover classes	Single season, maximum in late spring (April-May)		Single season (predominantly), maximum in late summer (July-September)	
					Agriculture (probable crops: winter wheat, rapeseed)	211, 231, 242, 243	Agriculture (probable crops: maize, sugar beet, potato)	211, 231, 242, 243
					Natural grasslands	321	Sparsely vegetated areas	112, 123
					Transitional shrubland/woodland	324	Wetland	412
5	2.01%	Fluctuation around increasing annual mean γ . Highest increase in 2004	high	Contrast between longer term NDVI trends	Increasing or stable annual mean NDVI		Decreasing annual mean NDVI	
					Transitional shrubland/woodland (stable or increasing annual mean NDVI)	324	Isolated spots in forest (2%) potentially clear cuts	31x
					Wetland (stable annual mean NDVI)	412		
6	1.71%	Longer term NDVI oscillation Decline from 2000-2003, high values in 2004 and 2006 and low in 2003 and 2005	high	Interannual differences	Relatively low values in 2003, 2004, 2006 Agriculture - subset in south-eastern part catchment Sparsely vegetated areas	242	Low values in 2000, 2001, 2006 Agriculture - subset in northern part catchment Natural grasslands	211 321

VALIDATION REMOTELY SENSED EVAPORATION

Next to the validation for 8 day averaged evaporation estimates, the EARS and WACMOS products are validated at the respective original time steps of the evaporation estimates. Furthermore some additional checks have been performed to further explain observed discrepancies in the validation. Method and results are shown in the next sections.

D.1 METHOD

The relevant temporal resolution for EARS is one day. For WACMOS the daily evaporation as well as the instantaneous evaporation at time of satellite overpass (noon) are validated. Furthermore, based on the results of the WACMOS validation, namely $E_{RS,WACMOS} > E_{EC}$, it was decided to evaluate the self-preservation of the evaporative fraction Λ on the days WACMOS images are available. Peng et al. (2013) showed for a large variety of biomes that the self-preservation of Λ is true for clear sky conditions and instantaneous values between 12.00 and 13.00 local time (LT), and reasonably well for instantaneous values between 11.00 and 14.00LT. For cloudy conditions however, he found that Λ_{inst} is more variable, so that the assumption does not hold. Satellite overpass time of the satellites used in the WACMOS product is around noon. The availability of data means cloud free conditions during overpass times. However, full day cloud free conditions are not guaranteed by that. The effect of partly cloudy conditions is not fully clear and depends on the exact conditions. The results of Peng et al. (2013) suggest an overestimation of the daily Λ during (partly) cloudy conditions and consequently an overestimation of the daily evaporation. The representativeness of the instantaneous evaporative fraction for the daytime evaporative fraction thus is investigated for the days at which WACMOS images are available, based on EC measurements. Furthermore the WACMOS Λ is compared with the Λ derived from the EC measurements.

The time series of the EC measurements of $\rho\lambda E$ and H are based on the 30min measurements, see Section 3.3. The instantaneous values of the evaporative fraction and the latent heat flux at time of satellite overpass are based on the hourly average. Since ground heat flux (G) measurements are not available for all sites, the relation $H + \rho\lambda E = R_n + G$ is used to calculate Λ , after Peng et al. (2013) :

$$\Lambda = \frac{\rho\lambda E}{R_n - G_0} = \frac{\rho\lambda E}{H + \rho\lambda E} \quad (D.1)$$

As described in Section 2.3, the turbulent fluxes are generally underestimated with 10% to 30% in total by the eddy flux measurements, as appears from structurally unclosed energy balances. According to e.g. Twine et al. (2000) the Bowen ratio is sustained however, which would legitimate the replacement of $R_n - G$ by $H + \rho\lambda E$, despite the underprediction of the turbulent fluxes. The daily Λ , Λ_{day} is calculated from the 24h (0:00-24:00LT) average turbulent fluxes, according Equation D.2:

$$\Lambda_{day} = \frac{\overline{\rho\lambda E_{0:00-24:00}}}{\overline{H_{0:00-24:00}} + \overline{\rho\lambda E_{0:00-24:00}}} \quad (D.2)$$

refer to section Algorithm with description of algorithms

self-preservation of Λ :

$$\Lambda_{inst} = \Lambda_{day}$$

Peng et al. (2013) uses Equation D.1 to correct the turbulent fluxes, as suggested by Twine et al. (2000). But applying the utmost right hand side of Equation D.1 directly on the uncorrected fluxes yields the same results for Λ .

Peng et al. (2013) use daytime average (8:00-17:00LT) Λ , which was calculated in this study as a reference as well, see Equation D.3.

$$\Lambda_{daytime} = \frac{\overline{\rho\lambda E_{8:00-17:00}}}{\overline{H_{8:00-17:00}} + \overline{\rho\lambda E_{8:00-17:00}}} \quad (D.3)$$

D.2 VALIDATION AGAINST EC MEASUREMENTS 1-DAY INTERVAL

Table D.1: Statistics of the comparison of E_{EC} and E_{RS} for WACMOS with daily interval. The first four parameters are the slope, intercept, coefficient of determination (r^2) and root mean square error (RMSE) of the linear regression with n data points. The last five parameters are the mean of E_{EC} and E_{RS} respectively, the root mean square difference of E_{RS} and E_{EC} , the mean difference ($\overline{E_{RS} - E_{EC}}$) and the standard deviation of the difference $sd(E_{RS} - E_{EC})$.

		WACMOS - daily interval					All sites	Forest	Crops
		BEVie	BEJal	FRHes	BELon	DESeh			
$\overline{E_{EC}}$	(mm d ⁻¹)	1.73	2.70	1.28	2.66	2.31	1.97	1.45	2.44
$\overline{E_{RS}}$	(mm d ⁻¹)	5.16	5.52	3.56	4.19	3.22	3.81	4.04	3.60
n	(-)	31	6	90	54	86	267	127	140
$\overline{(E_{RS} - E_{EC})}$	(mm d ⁻¹)	3.44	2.82	2.28	1.53	0.92	1.84	2.59	1.15
$sd(E_{RS} - E_{EC})$	(mm d ⁻¹)	1.59	1.86	1.80	1.37	1.51	1.79	1.81	1.49
RMSE	(mm d ⁻¹)	3.77	3.29	2.90	2.04	1.76	2.56	3.15	1.88
slope	(-)	1.98	0.88	1.77	0.86	0.75	0.89	1.73	0.81
intercept	(mm d ⁻¹)	1.74	3.13	1.30	1.89	1.50	2.05	1.52	1.61
r^2	(-)	0.65	0.39	0.51	0.46	0.39	0.33	0.54	0.42

Table D.2: Statistics of the comparison of E_{EC} and E_{RS} for EARS with daily interval.

		EARS - daily interval					All sites	Forest	Crops
		BEVie	BEJal	FRHes	BELon	DESeh			
$\overline{E_{EC}}$	(mm d ⁻¹)	0.74	1.78	0.84	1.29	1.63	1.14	0.85	1.47
$\overline{E_{RS}}$	(mm d ⁻¹)	1.80	3.09	1.98	1.72	1.54	1.80	1.97	1.63
n	(-)	341	46	298	304	320	1309	685	624
$\overline{(E_{RS} - E_{EC})}$	(mm d ⁻¹)	1.07	1.31	1.13	0.43	-0.09	0.66	1.11	0.16
$sd(E_{RS} - E_{EC})$	(mm d ⁻¹)	1.01	0.96	1.04	0.96	1.10	1.15	1.02	1.07
RMSE	(mm d ⁻¹)	1.47	1.62	1.54	1.05	1.11	1.32	1.51	1.08
slope	(-)	1.82	1.09	1.57	0.80	0.64	0.88	1.55	0.70
intercept	(mm d ⁻¹)	0.46	1.16	0.66	0.69	0.49	0.80	0.65	0.60
r^2	(-)	0.76	0.75	0.69	0.58	0.43	0.46	0.72	0.48

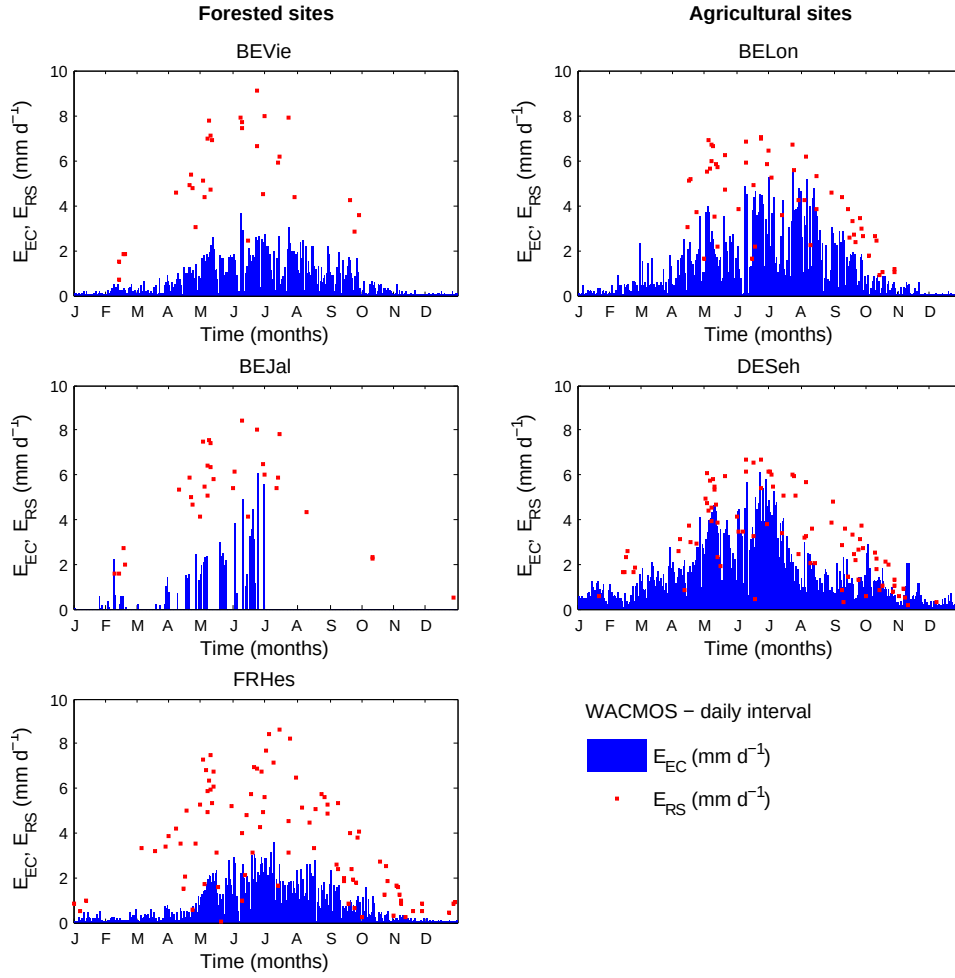


Figure D.1: Comparison of the time series of E_{EC} (blue bars) and $E_{RS,WACMOS}$ (red squares) for the daily interval for the year 2008 and for the five EC sites. On the left side the forested sites Vielsalm (mixed forest), Jalhay (mixed forest) and Hesse (broad-leaved forest). On the right the agricultural sites Lonzee and Selhausen.

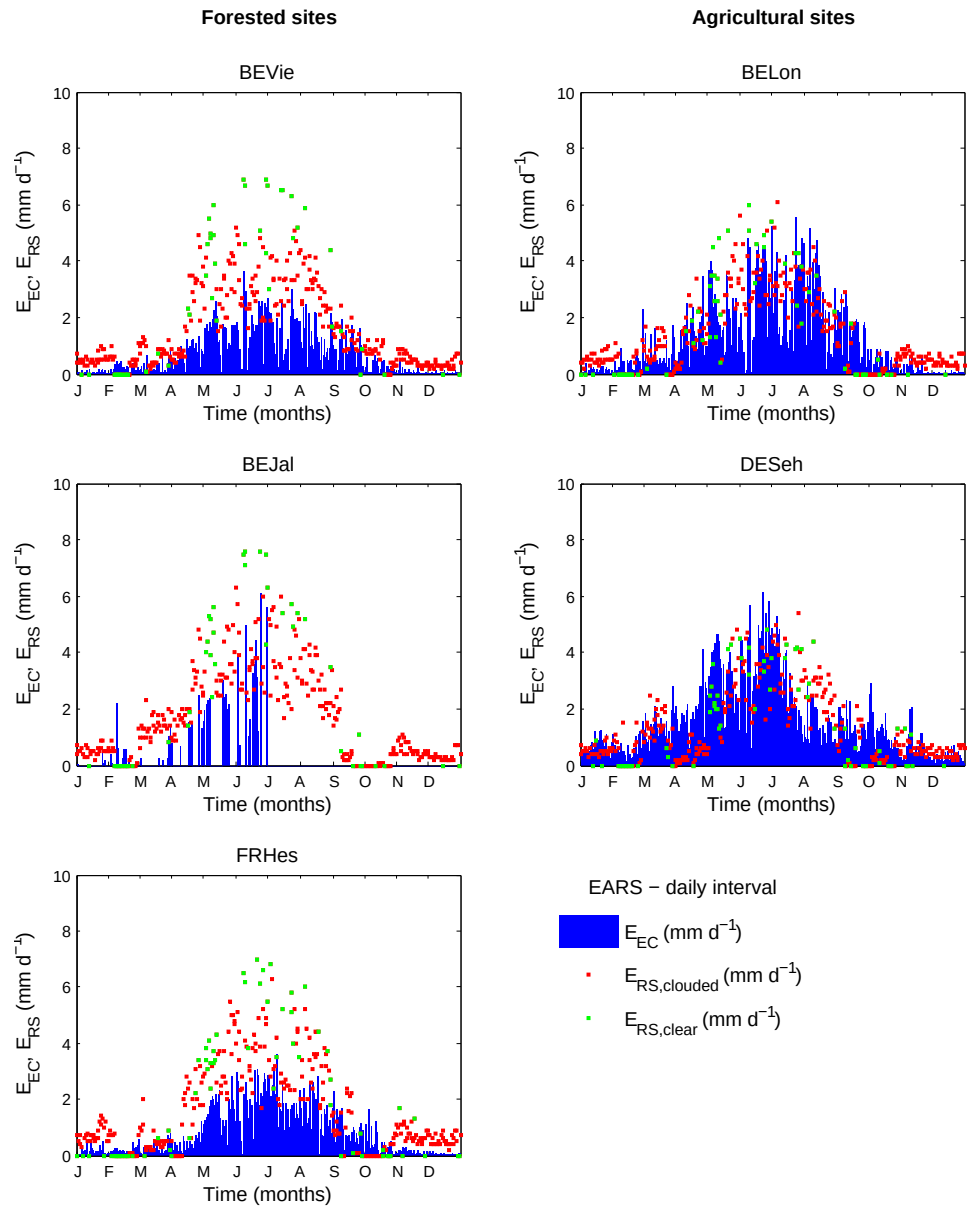


Figure D.2: Comparison of the time series of E_{EC} (blue bars) and $E_{RS,EARS}$ at clear (red squares) and at clouded days (green squares) for the daily interval for the year 2008 and for the five EC sites. On the left side the forested sites Vielsalm (mixed forest), Jalhay (mixed forest) and Hesse (broad-leaved forest). On the right the agricultural sites Lonzee and Selhausen.

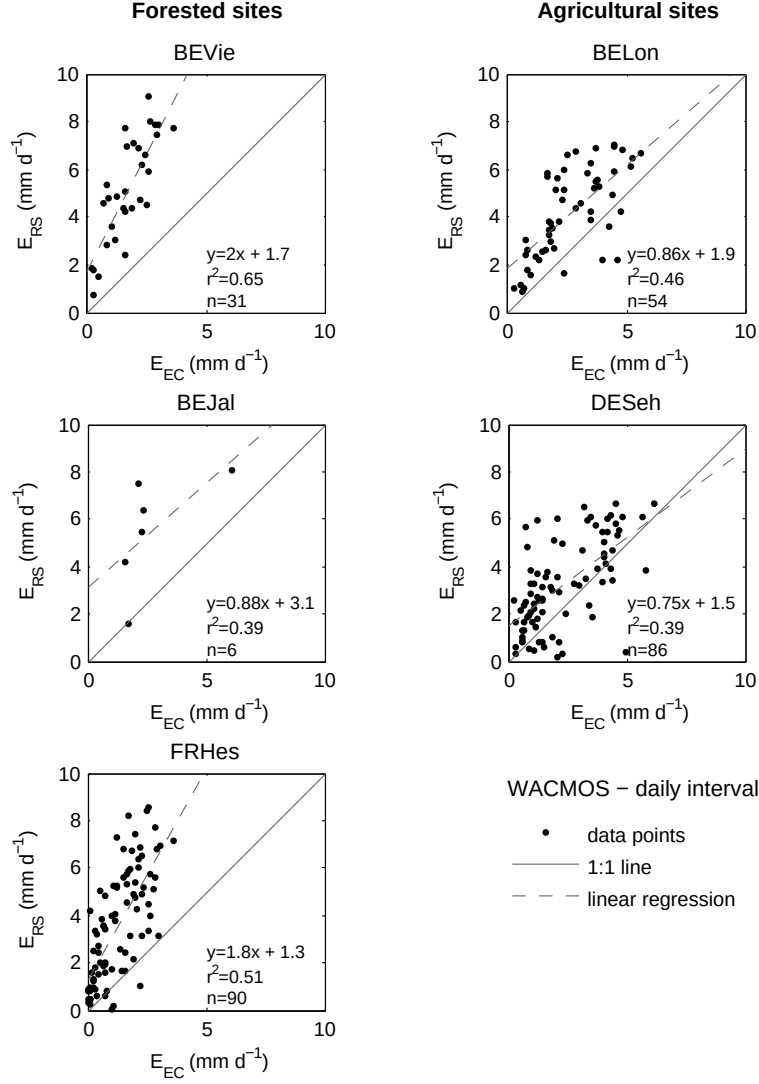


Figure D.3: Correlation of E_{EC} and $E_{RS,WACMOS}$ (mm d⁻¹) for the daily interval for the year 2008, and for the five EC sites. $E_{RS,WACMOS}$, the WACMOS evaporation estimate of the pixel in which the EC tower is located (the upwind pixel for Selhausen), on the y-axis, and E_{EC} of the specific site on the x-axis. The solid black line is the 1 : 1 line. The dotted black line shows the linear regression of the data points. The equation and r^2 of this line are given in the lower right corner of all figures.

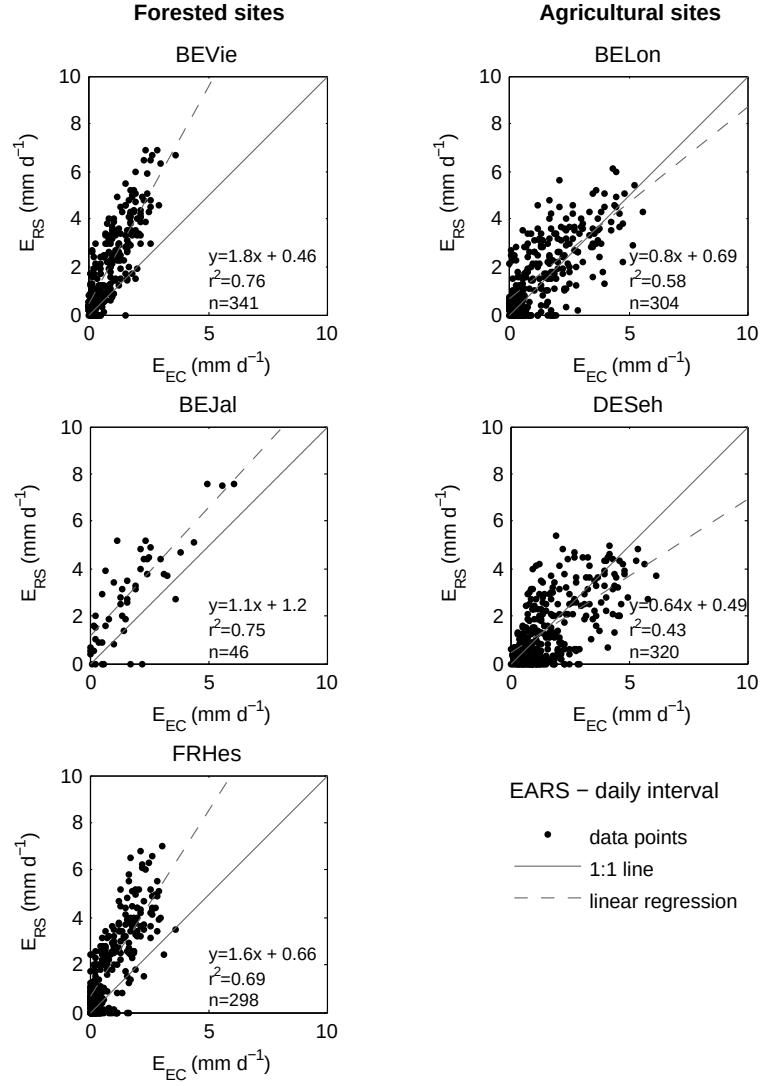


Figure D.4: Correlation of E_{EC} and $E_{RS,EARS}$ (mm d⁻¹) for the daily interval for the year 2008, and for the five EC sites. $E_{RS,EARS}$, the EARS evaporation estimate for the pixel in which the EC tower is located, on the y-axis, and E_{EC} of the specific station on the x-axis. The solid black line is the 1 : 1 line. The dotted black line shows the linear regression of the data points. The equation and r^2 of this line are given in the lower right corner of all figures.

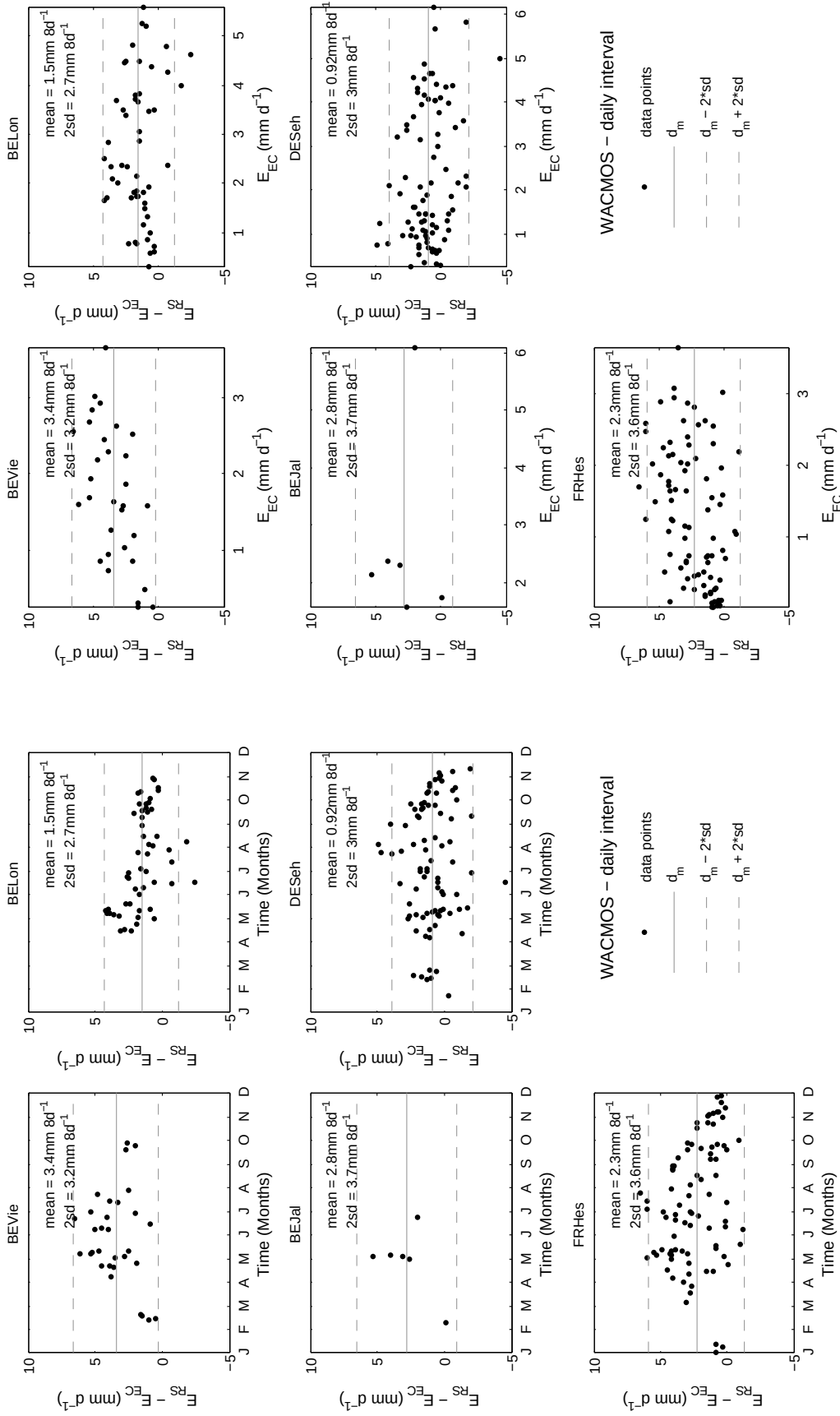


Figure D.5: Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for WACMOS and the daily interval

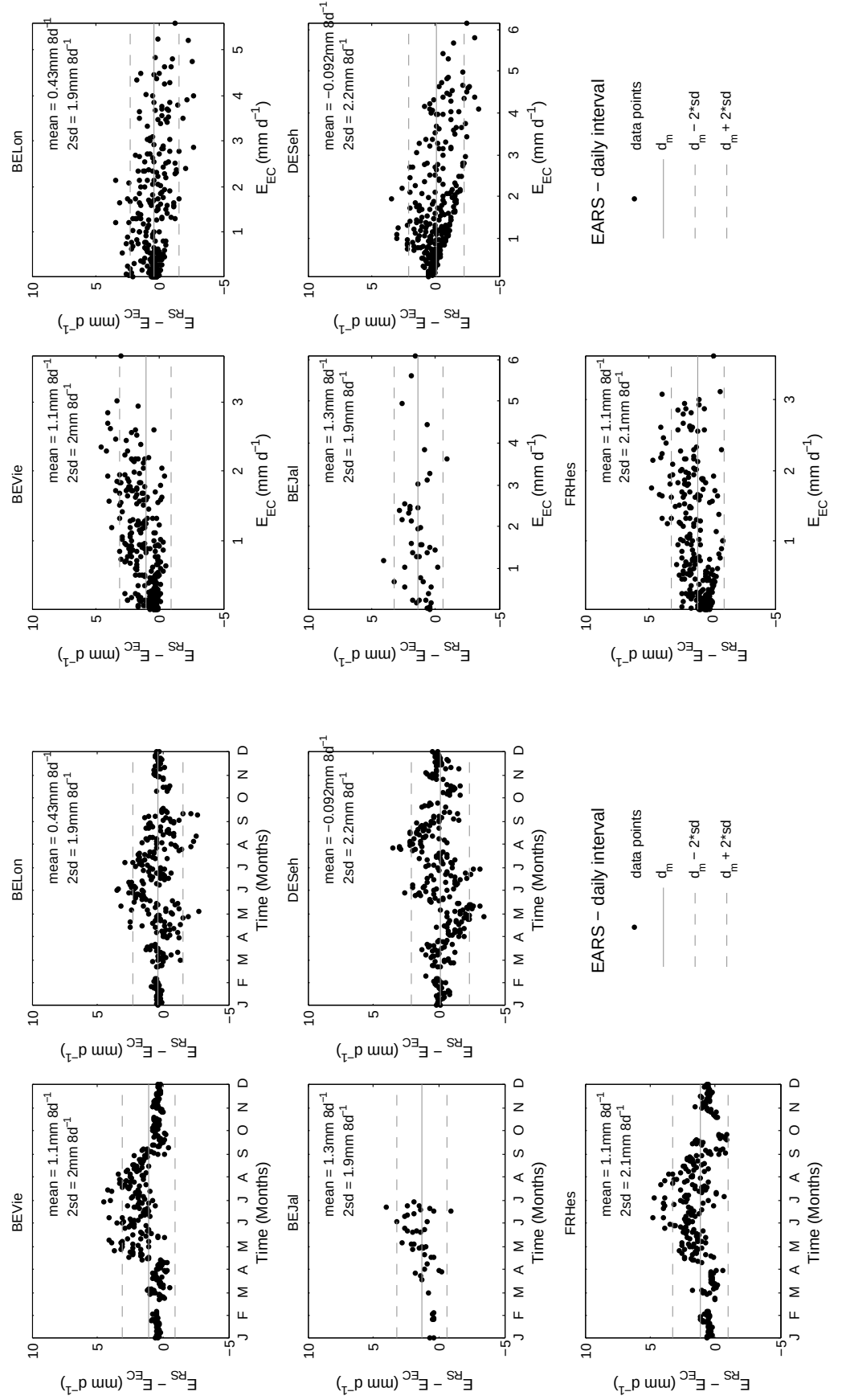


Figure D.6: Difference E_{RS} and E_{EC} plotted against time (left) and against E_{EC} (right) per station, for EARS and the daily interval

D.3 EVAPORATIVE FRACTION

D.3.1 *Self-preservation of Λ*

Figure D.7 shows the correlation between the instantaneous Λ (Λ_{inst}) at time of satellite overpass (12:00LT) and the daily Λ (Λ_{day}) for the year 2008, for the five EC sites. Λ_{day} is generally higher than the value at noon. For clear-sky satellite overpass times - days on which WACMOS images are available - Λ_{day} is much closer to Λ_{inst} , see the red dots in Figure D.7. The same analysis is performed for the daytime Λ . Results are comparable and not shown. Statistics for the days for which WACMOS images are available are summarized in Table D.3. Λ is consequently underestimated for all sites. The positive mean bias for Selhausen is caused by a large deviation at a single day. Excluding this value reduces the mean bias to -0.135, with a standard deviation of 0.193. The underestimation is in correspondence with literature, see e.g. (Peng et al., 2013). Thus, using the instantaneous Λ at 12.00LT at clear sky (not necessarily a completely clear day) to approximate daily Λ would cause an underestimation of daily evaporation.

Table D.3: Comparison Λ_{inst} and Λ_{day}

EC site	Λ_{inst}	Λ_{day}	$\Lambda_{inst} - \Lambda_{day}$		RMSE	n
	mean	mean	mean	sd		
Vielsalm	0.288	0.332	-0.044	0.046	0.063	30
Jalhay	0.427	0.657	-0.283	0.490	0.554	17
Hesse	0.408	0.519	-0.111	0.450	0.461	89
Lonze	0.723	0.846	-0.122	0.312	0.332	54
Selhausen	0.770	0.735	0.035	1.613	1.605	89

D.3.2 *Comparison $\Lambda_{RS,WACMOS}, \Lambda_{EC}$*

By comparing Λ contained in the WACMOS product with Λ determined from the EC measurements (Figure D.8), it is shown that there is hardly any agreement between the two. The coefficient of determination for the linear relation has values ranging from 0.15 to 0.00022 for the forested sites and are almost zero for the agricultural sites. This figure also shows the source of the overestimation of $E_{RS,WACMOS}$ for the forested sites compared to E_{EC} : $\Lambda_{RS,WACMOS} \gg \Lambda_{EC}$.

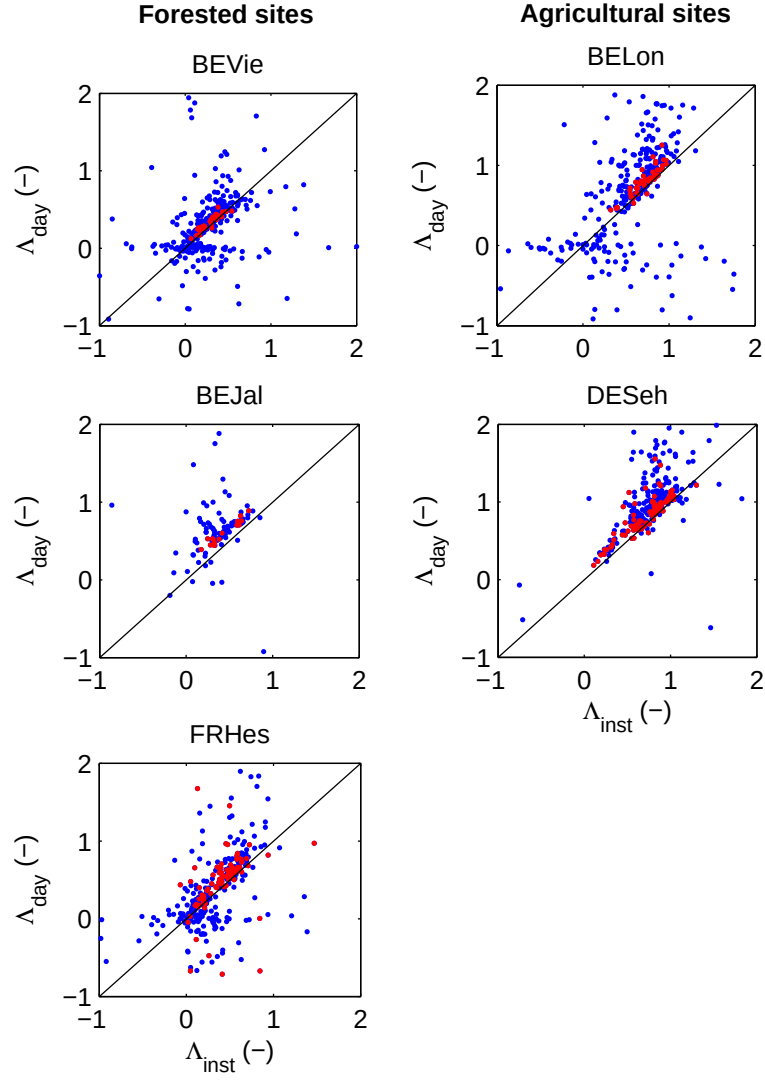


Figure D.7: Correlation of Λ_{inst} and Λ_{day} , according to the EC measurements for the five EC sites. Λ_{day} is shown on the y-axis, Λ at time of satellite overpass at the x-axis. The blue dots show the data on all days in 2008. In red the days on which WACMOS images are available are shown, having cloud-free overpass times. The solid black line is the 1 : 1 line. The dashed line is the linear regression line of the data points, with s = slope, y_0 = intersect, r^2 = correlation coefficient. For stations 1, 3 and 5 one out-layer was removed.

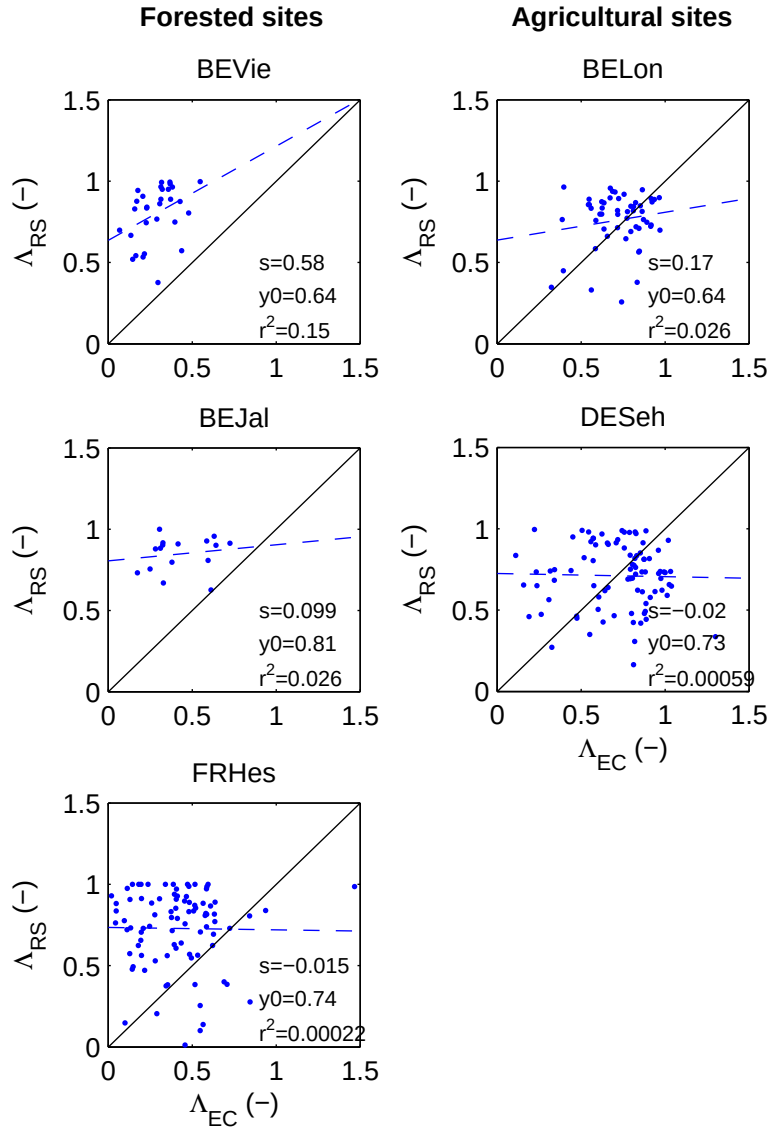


Figure D.8: Correlation of $\Lambda_{RS,WACMOS}$ and Λ_{EC} for the forested sites (left) and for the agricultural sites. For Λ_{EC} the instantaneous value is used. The solid black line is the 1 : 1 line. The dashed line is the linear regression of the data points, with s = slope, y_0 = intercept, r^2 = correlation coefficient.

MODEL RESULTS

E.1 FORCING - DECISION ON CATCHMENT

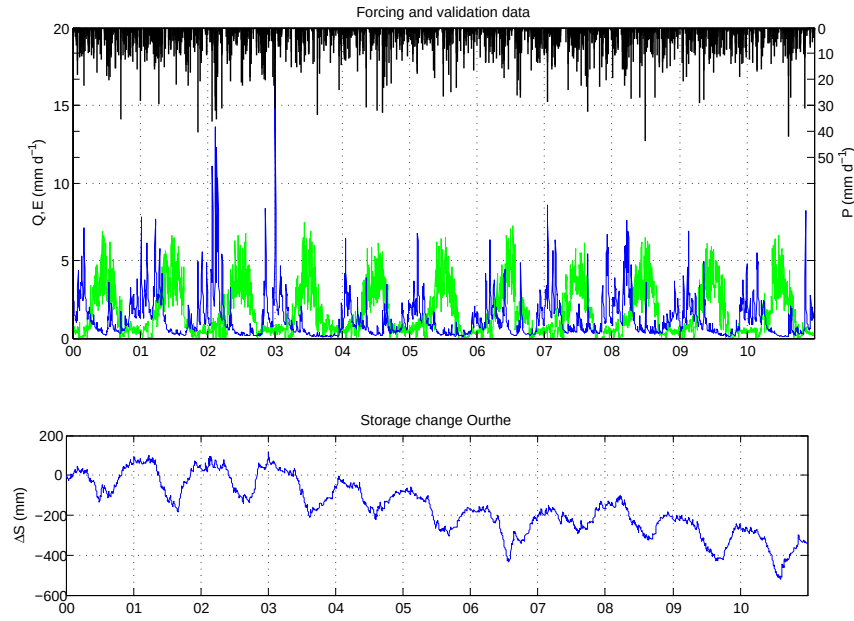


Figure E.1: Forcing for the Ourthe in the period 2000-2010. The top panel shows the specific discharge Q (blue), EARS evaporation $E_{RS,EARS}$ (green), and catchment average precipitation P (black bars). The bottom panel shows the cumulative difference in P , $E_{RS,EARS}$ and Q . There is a net outflow of water in catchment, according to the presented data, mainly caused by a net outflow in the Ourthe Occidentale subcatchment.

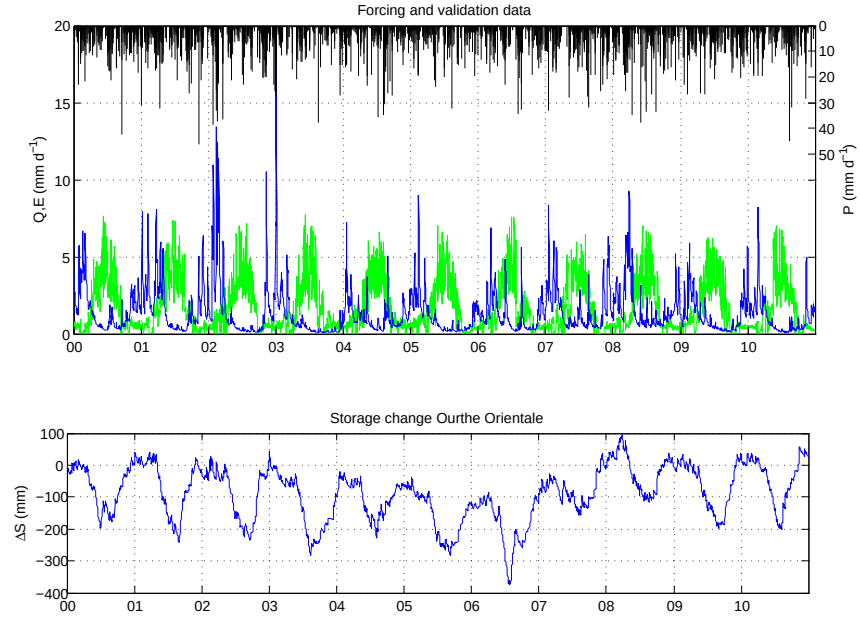


Figure E.2: Forcing for the Ourthe Orientale in the period 2000-2010. The top panel shows the specific discharge Q (blue), EARS evaporation $E_{RS,EARS}$ (green) and catchment average precipitation P (black bars). The bottom panel shows the cumulative difference in P , $E_{RS,EARS}$ and Q .

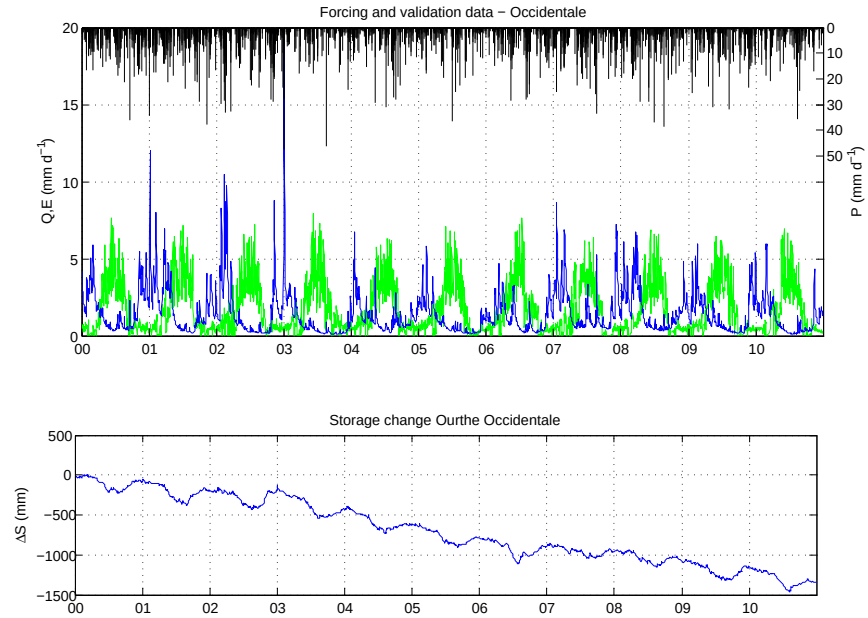


Figure E.3: Forcing for the Ourthe Occidentale in the period 2000-2010. The top panel shows the specific discharge Q (blue), EARS evaporation $E_{RS,EARS}$ (green) and catchment average precipitation P (black bars). The bottom panel shows the cumulative difference in P , $E_{RS,EARS}$ and Q . There is a considerable net outflow of water in this subcatchment, according to the presented data.

E.2 MODEL CALIBRATION

Table E.1: Posterior parameter range for models with $D_f > 0$ and $\alpha \neq 1$

Parameter		FLEX ^{E,3}		FLEX ^{Ep,3}		FLEX ^{Ep,RS,3}	
		min	max	min	max	min	max
$S_{u,max}$	(mm)	240	600	180	500	200	500
β	(-)	2	5	1	5	1	5
D_z	(-)	0	0.4	0	0.6	0	0.5
D_f	(-)	0	0.25	0	0.1	0	0.1
$R_{s,max}$	(mm d ⁻¹)	0	0.5	0.1	0.5	0	0.4
T_{lag}	(d)	1	2	1	2	1	2
K_f	(d)	3.8	10	2.7	9	1	10
K_s	(d)	45	55	45	55	45	55
α	(-)	0.8	1.3	0.8	1.3	0.7	1.5
L_p	(-)			0.4	0.9	0.4	0.9

Table E.2: Posterior parameter range for models with $D_f = 0$ and $\alpha = 1$

Parameter		FLEX ^{E,0}		FLEX ^{Ep,0}		FLEX ^{Ep,RS,0}	
		min	max	min	max	min	max
$S_{u,max}$	(mm)	295	600	200	500	200	500
β	(-)	3	5	2	5	2.5	5
D_z	(-)	0	0	0	0	0	0
D_f	(-)	0	0.2	0	0.1	0	0.1
$R_{s,max}$	(mm d ⁻¹)	0	0.35	0	0.5	0	0.5
T_{lag}	(d)	1	2	1	2	1	2.5
K_f	(d)	3	8	2	7	1	10
K_s	(d)	45*	55*	45	55	45	55
α	(-)	1	1	1	1	1	1
L_p	(-)			0.4	0.9	0.4	0.8

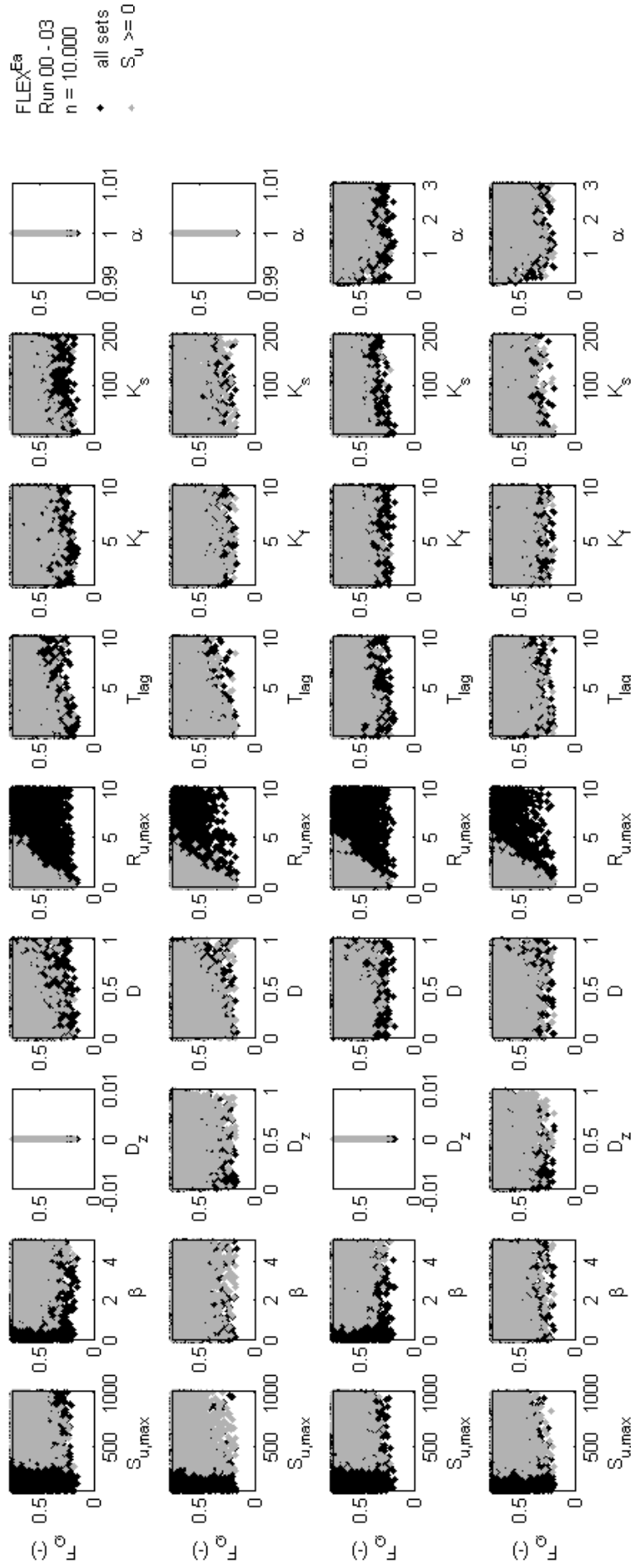


Figure E.4: Parameter identifiability for FLEX^E after a preliminary run with the initial parameter range, for objective function F_Q . In black all 10,000 parameter sets are shown, in grey only the parameter sets fulfilling the constraint $S_u \geq 0$. From top to bottom: 1) the unsaturated zone threshold for fast runoff and the non-linearity of the fast reservoir are left out ($D_z = 0$ and $\alpha = 1$); 2) $D_z \in [0, 1]$, $\alpha = 1$; 3) $D_z = 0$, $\alpha \in [0.1, 3]$; 4) $D_z \in [0, 1]$, $\alpha \in [0.1, 3]$.

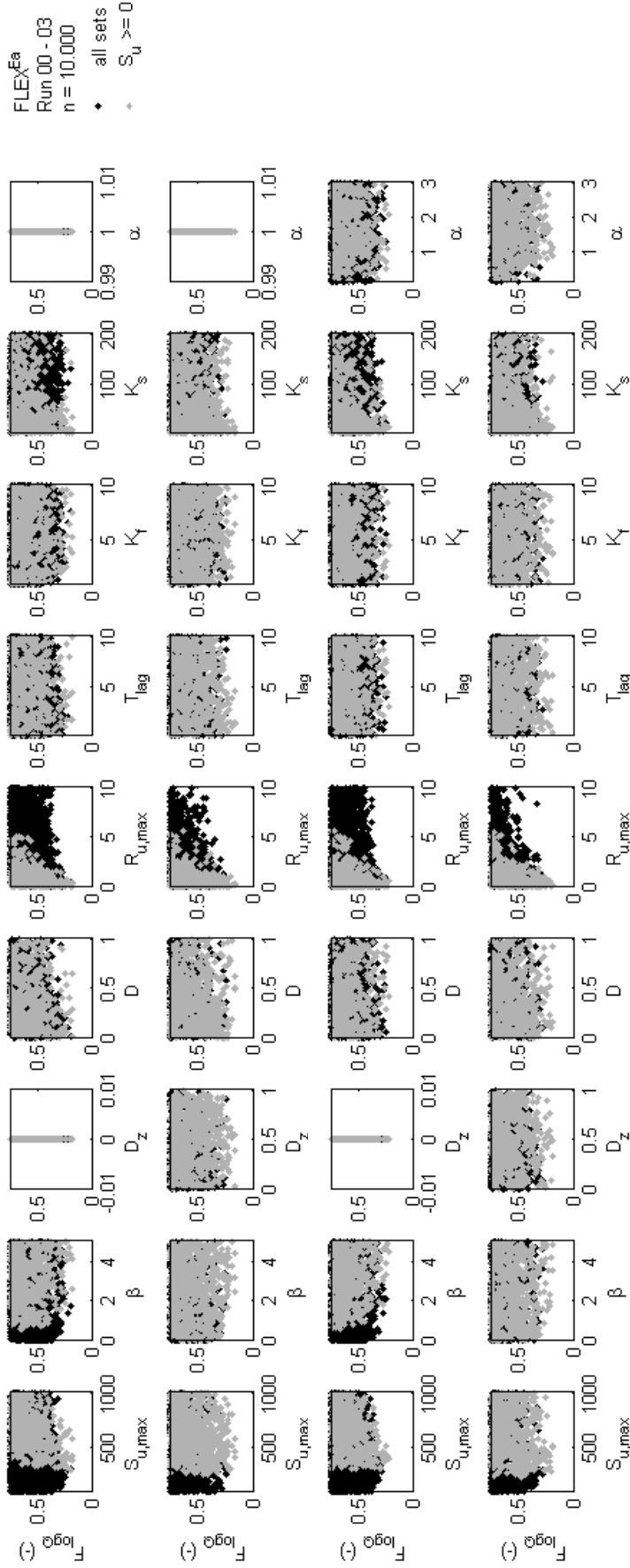


Figure E.5: Parameter identifiability for FLEX^{Ea} after a preliminary run with the initial wide parameter range, for objective function F_{logQ} . In black all 10,000 parameter sets are shown, in grey only the parameter sets fulfilling the constraint $S_u \geq 0$. From top to bottom: 1) the unsaturated zone threshold for fast runoff and the non-linearity of the fast reservoir are left out ($D_z = 0$ and $\alpha = 1$); 2) $D_z = [0, 1]$, $\alpha = 1$; 3) $D_z = 0$, $\alpha = [0.1, 3]$; 4) $D_z = [0, 1]$, $\alpha = [0.1, 3]$.

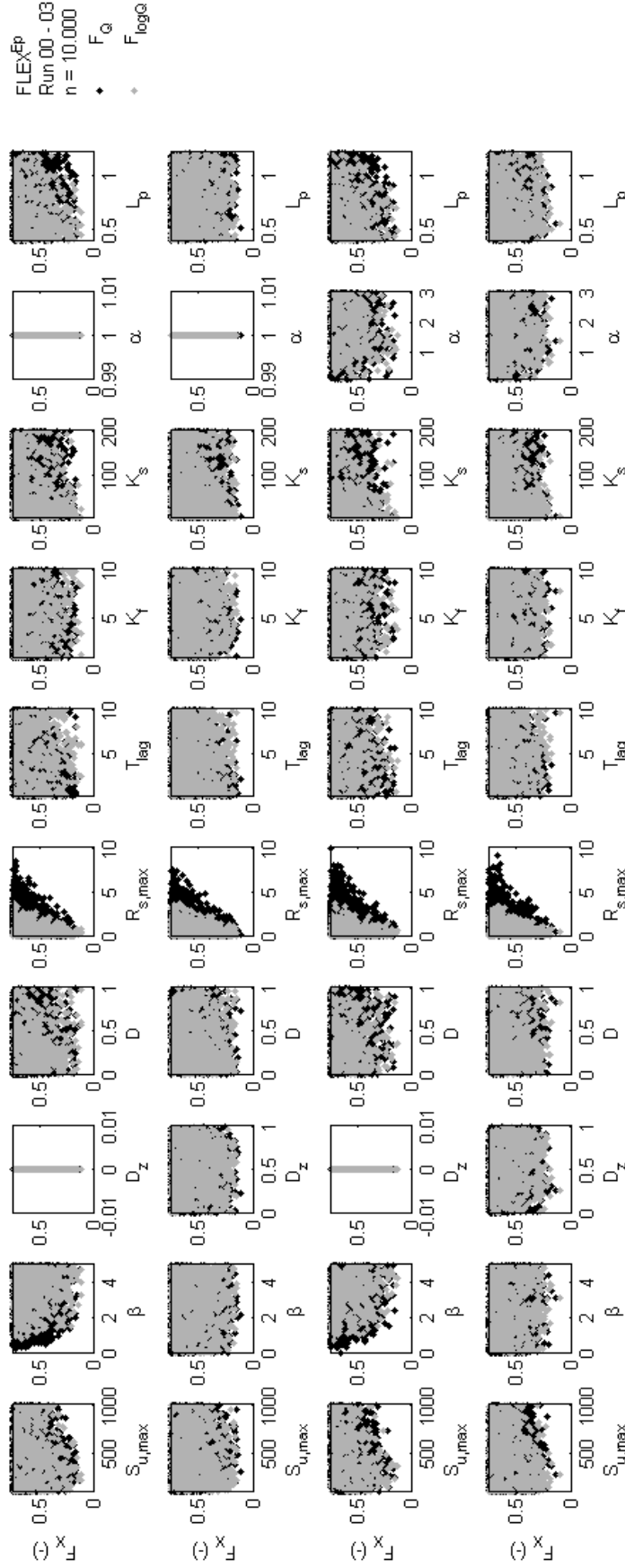


Figure E.6: Parameter identifiability for FLEX^{Ep} after a preliminary run with the initial parameter range, for objective functions $F_{\log Q}$ in grey and F_Q in black. From top to bottom: 1) the unsaturated zone threshold for fast runoff and the non-linearity of the fast reservoir are left out ($D_z = 0$ and $\alpha = 1$); 2) $D_z = [0, 1]$, $\alpha = 1$; 3) $D_z = 0$, $\alpha = [0.1, 3]$; 4) $D_z = [0, 1]$, $\alpha = [0.1, 3]$.

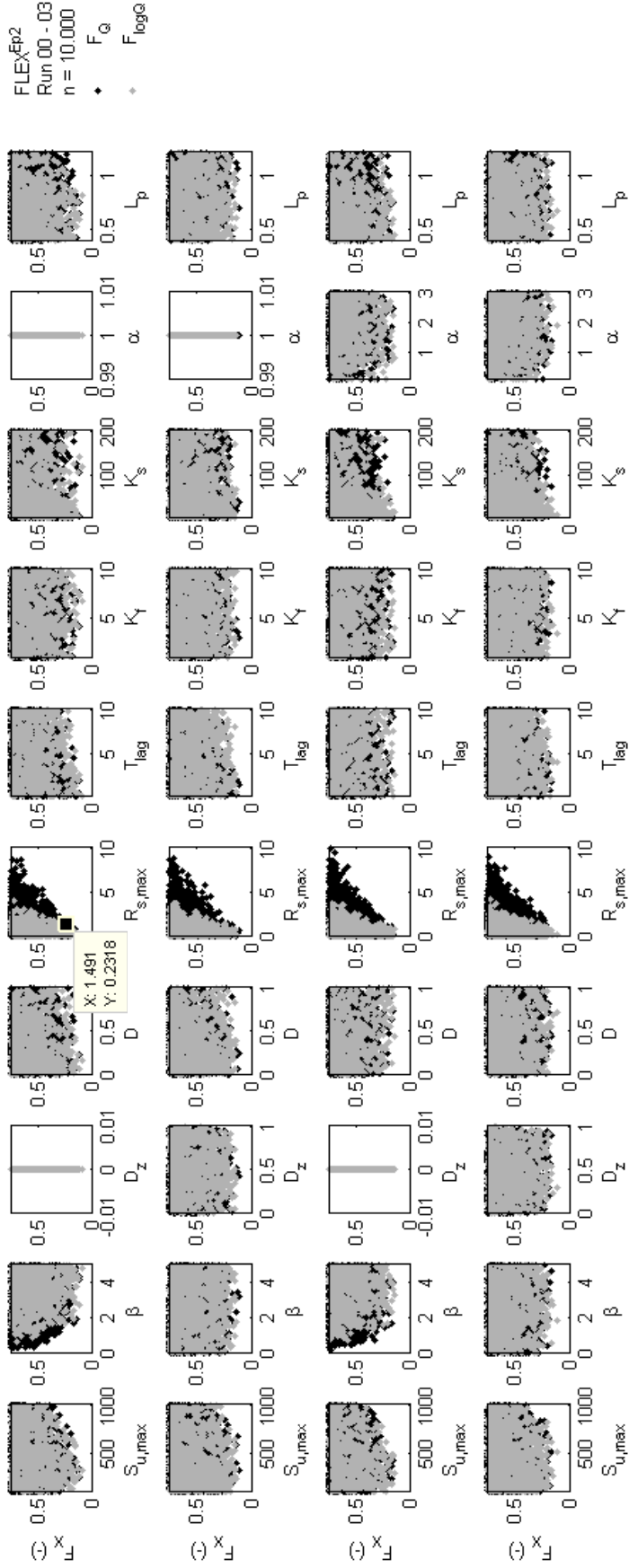
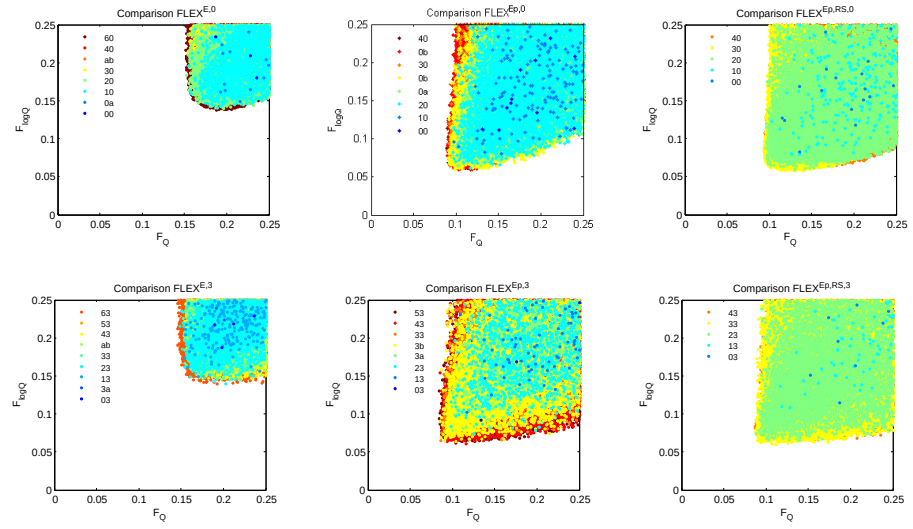


Figure E.7: Parameter identifiability for FLEX^{Ep} with $E_p = E_{p,rs}$ after a preliminary run with the initial parameter range, for objective functions F_{logQ} in grey and F_Q in black. From top to bottom: 1) the unsaturated zone threshold for fast runoff and the non-linearity of the fast reservoir are left out ($D_z = 0$ and $\alpha = 1$); 2) $D_z = [0, 1]$, $\alpha = [0, 1]$; 3) $D_z = [0, 1]$, $\alpha = [0, 1]$; 4) $D_z = [0, 1]$, $\alpha = [0, 1, 3]$.



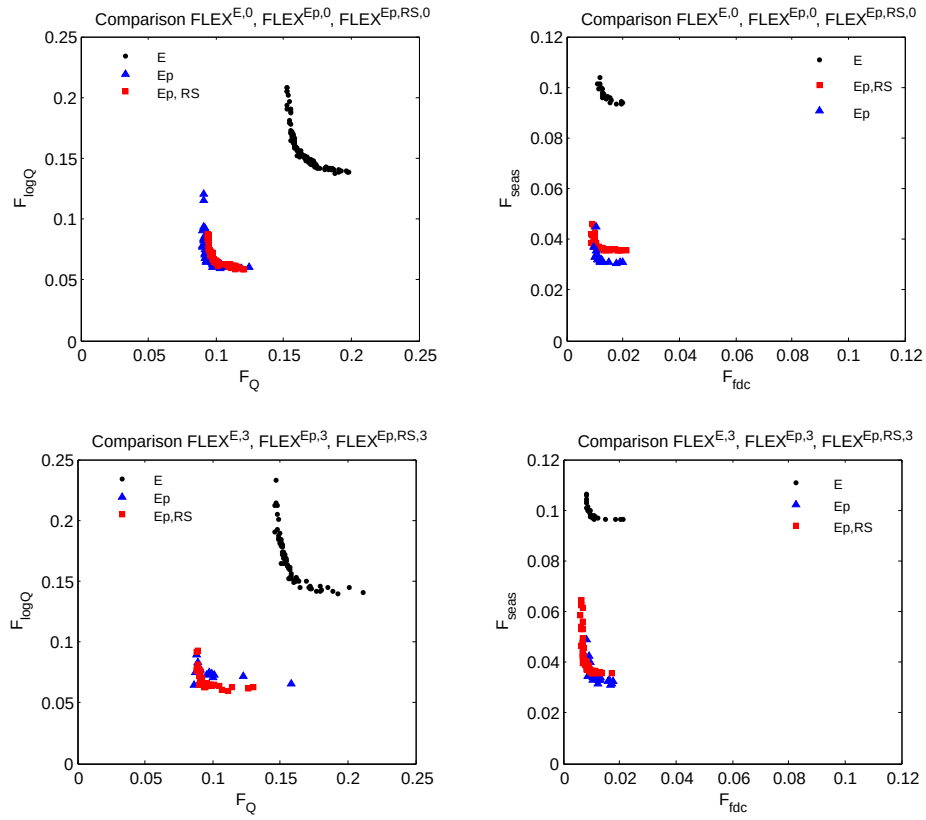


Figure E.9: Model performance of $FLEX^E$ (black dots), $FLEX^{Ep}$ (blue triangles) and $FLEX^{Ep,RS}$ (red squares) with respect to objective functions F_{logQ} and F_Q (left) and F_{seas} and F_{fdc} (right), for the model variants without (top) and with (bottom) soil moisture threshold for runoff and non-linear fast reservoir.

E.3 RESULTS MODEL 0

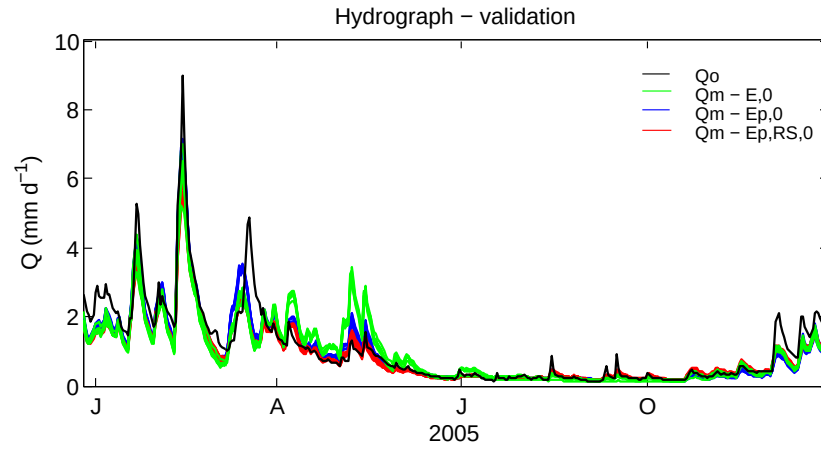


Figure E.10: Hydrograph during validation for model variant 0. Shown are Q_m of 7 Pareto-optimal parameter sets based on F_Q and F_{logQ} for $FLEX^{E,0}$ (green), $FLEX^{Ep,0}$ (blue) and $FLEX^{Ep,RS,0}$ (red) and the observed discharge Q_o (black).

E.4 RESULTS MODEL 3

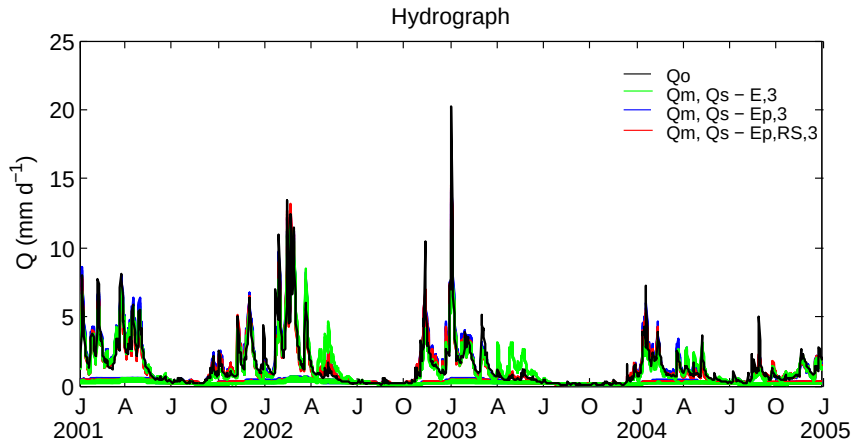


Figure E.11: Hydrograph with mean monthly flow, calculated as the 30-days moving average, showing the seasonal fluctuations in the discharge. Shown are Q_m of 10 Pareto-optimal parameter sets based on F_Q and F_{logQ} for $FLEX^{E,3}$ (green), $FLEX^{Ep,3}$ (blue) and $FLEX^{Ep,RS,3}$ (red) and the observed discharge Q_o (black).

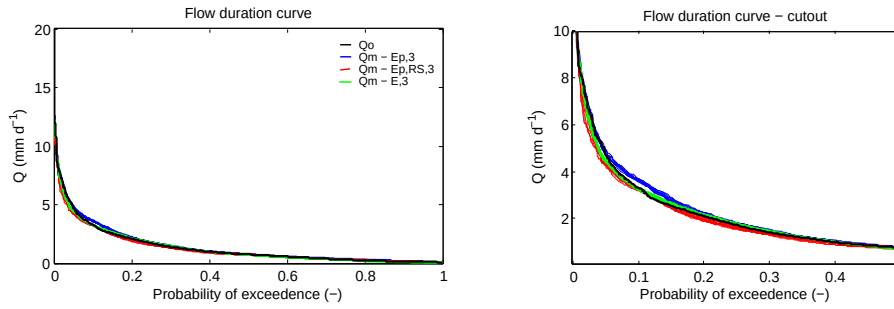


Figure E.12: Flow duration curve (FDC) for $FLEX^{E,3}$ (green), $FLEX^{Ep,3}$ (orange) and $FLEX^{Ep,RS,3}$ (red) compared to the flow duration curve of the observed discharge (black). Left: full spectrum of flows. Right: cutout of the fdc at the transition between high and low flows, where the divergence from the observed flow and the difference between the models is highest.

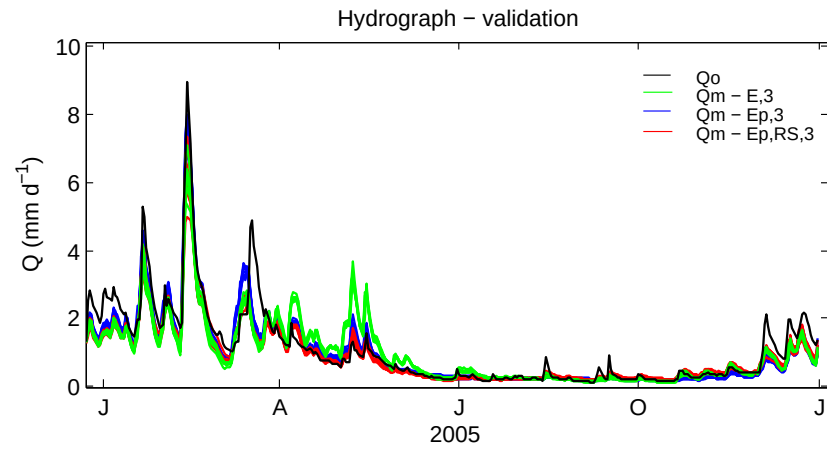


Figure E.13: Hydrograph during validation for model variant 3. Shown are Q_m of 10 Pareto-optimal parameter sets based on F_Q and F_{logQ} for $FLEX^{E,3}$ (green), $FLEX^{Ep,3}$ (blue) and $FLEX^{Ep,RS,3}$ (red) and the observed discharge Q_o (black).

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