Bachelor Thesis

Investigation of soil moisture - atmosphere interactions and their influence on temperature in Europe

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Investigation of soil moisture - atmosphere interactions and their influence on temperature in Europe
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Abstract

In recent research the consideration of soil moisture in land-atmosphere interactions has received increasing attention. These interactions can have direct effects on the climate. Especially for long-term climate forecasts, soil moisture plays an important role, as its memory spans weeks up to months. The aim of this thesis is to examine the effect of soil moisture content (determined by precipitation, evapotranspiration and runoff) on temperature in Europe. The analysis showed that especially in Mediterranean regions temperature is more strongly influenced by soil moisture content. In general, correlations are higher for dryer soils (leading to an increase in temperature). These findings can be used to improve weather forecast skills, especially for such extremes as the 2003 summer heat wave which was partially caused by dry soils because of a precipitation deficit which already started in spring and lasted during the summer.
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1. Introduction

In recent research the consideration of soil moisture in land-atmosphere interactions receives increasing attention. These interactions can have direct effects on the climate. Positive (negative) soil moisture anomalies can lead to positive (negative) evapotranspiration anomalies, and the associated increased (decreased) evaporative cooling can lead to a cooling (warming) of the overlying air (Koster et al., 2010). However, indirect feedbacks with cloud cover and dry air advection may also play a role (Hirschi et al., 2011).

The soil moisture content is controlled by: (i) water input (rainfall and snow melt), (ii) evapotranspiration and (iii) runoff. As the timescales of soil moisture dynamics span weeks to a couple of months (Koster et al., 2010), a precipitation anomaly in the preceding season can still have an effect on the soil moisture content in the actual season. This effect has already been shown by Mueller and Seneviratne (2012); they found a strong relationship between precipitation deficits and the subsequent occurrence of hot extremes in a large fraction of the world. Especially two recent events underline this finding: in summer 2003, a heatwave hit central Europe, while in summer 2010 an extreme heat anomaly impacted North America, Asia, Africa and European Russia. Both heatwaves were preceded by reduced spring precipitation frequency. Beside heat-related deaths, these extremes caused immense economic damage due to crop shortfall and forest fires. Also Vautard et al. (2007) found that the ten warmest European summers over the past five decades were in average preceded by reduced winter and spring precipitation frequency around the Mediterranean.

As the rate of evaporative cooling changes faster with lower soil moisture content, low soil moisture conditions are more sensitive to changes and therefore also lead to higher coupling with atmosphere interactions. Especially when considering that Southern Europe will tend to get drier in the future (Seneviratne et al., 2012), this feedback contributes to amplified extreme events. These increases in climate variability have a greater effect on society than changes in mean climate because it is more difficult to adapt to changes in extremes (Seneviratne et al., 2006).

Given the fact that climate change leads to stronger land-atmosphere interactions as well as the findings that soil moisture dynamics have a long timescale and influences local climate, it is conclusive, that in order to forecast precipitation, air temperature, and other meteorological quantities weeks to months in advance, prediction systems must take advantage of soil moisture-atmosphere feedbacks (Koster et al., 2010).

The main focus of this bachelor thesis is to examine the effect of soil moisture content on evapotranspiration and hence temperature in Europe. For this purpose a conceptual hydrological model from Orth et al. (2013) was applied to simulate soil moisture content and evapotranspiration. These modeled data were then compared with observed temperature.
Additionally it was focused on the two heat summers in 2003 and 2010 to study if the investigated relationships change under drier conditions. This scenario may occur more frequently in the future because of climate change.

2. Methodology

2.1.1. Simple water balance model

In order to study the influence of soil moisture on climate, a simple water balance model introduced by Orth et al. (2013) was used. It is based on the following equation:

\[ sm_{n+\Delta t} = sm_n + (P_n - E_n - Q_n) \Delta t \]  

(1)

The time step \( \Delta t \) is defined as 1 day in this study. \( sm_{n+\Delta t} \) is the soil moisture content (in mm) at time \( n+\Delta t \). It depends on precipitation \( P_n \), evapotranspiration \( E_n \), runoff \( Q_n \) (all in mm day\(^{-1}\)) and on soil moisture content \( sm_n \) (mm) of the preceding day (time \( n \)). As the two variables evapotranspiration and runoff depend on soil type and moisture content, two simple formulas were used to capture these dependencies. These will be described in the next sections.

2.1.2. Evapotranspiration

Heat transfer between land and atmosphere can either occur as sensible heat (related to temperature), or latent heat (related to moisture). While oceans have unlimited water, basically all the radiation energy is used for evaporation and is therefore transferred to latent heat (no warming). In deserts, on the other hand, where almost no water is available, all the radiation energy is transferred to sensible heat (except from the reflected parts of the incoming radiation) resulting in warming of the surface. This leads to a strong radiation-coupled temperature variability in the deserts, while the ocean surface temperature remains more constant. In other land areas however, net radiation is transferred partly to latent heat and partly to sensible heat, depending on the soil moisture content (for example as soil moisture increases, plants open more stomata, allowing more photosynthesis, but also more loss of water). The relative amount of net radiation that is used for evapotranspiration is called evaporative fraction and calculated in this model as follows:

\[ \frac{\lambda \rho E}{R} = \beta \left( \frac{sm_n}{c} \right)^\gamma \]  

with \( \gamma > 0 \) and \( \beta \neq 1 \),

(2)
where $\lambda$, the latent heat of vaporization ($2260 \text{ kJ kg}^{-1}$) and $\rho$ the density of water ($\text{kg m}^{-3}$) ensure that $E$ has the same unit as the net radiation $R$ ($\text{W m}^{-2}$). Soil moisture ($\text{sm}$) is scaled with the soil’s water-holding capacity $c$ (290 mm for all locations) so that the function operates on the degree of saturation. The unitless variables $\gamma$ and $\beta$ are soil-dependent parameters. The exponent $\gamma$ ensures that the function is strictly monotonically increasing, while the factor $\beta$ characterizes the maximum possible evaporative fraction (ranging from 0 to 1), preventing evapotranspiration from exceeding net radiation even when water is fully available (Orth et al., 2013). The reason why $\beta$ can never exceed 1 is simply a reflection of the assumption that net radiation provides the energy needed for evaporation. The remaining fraction of the radiation ($1 - \frac{\lambda E}{R}$) is used for sensible heat and therefore for warming of the surface.

### 2.1.3. Runoff

In this model the runoff is assumed to be dependent only on precipitation and soil moisture, thereby neglecting impacts of other variables. The formula is defined as follows:

$$\frac{Q}{P} = \left(\frac{\text{sm}}{c}\right)^\alpha \quad \text{with } \alpha \neq 0,$$

where the unitless, soil-dependent exponent $\alpha$ ensures that the runoff ratio increases monotonically with soil moisture.

### 2.1.4. Function plots

If normalized evapotranspiration and runoff are plotted against the soil moisture content, their typical behavior looks as follows:
Both ratios increase with increasing soil moisture content. But while the evaporative fraction increases rapidly in low soil moisture conditions and stays almost constant afterwards, the runoff fractions show the opposite behavior. To better observe the changes of these ratios per mm change in soil moisture, the derivatives were formed:

\[
\frac{d \frac{ET}{R}}{dSM} = \beta \gamma \left(\frac{SM}{c}\right)^{\gamma - 1}
\]  
\[\text{(4)}\]

\[
\frac{d \frac{Q}{P}}{dSM} = \alpha \left(\frac{SM}{c}\right)^{\alpha - 1}
\]  
\[\text{(5)}\]

When these two equations are plotted, the following dependency can be seen:
The lower the soil moisture content the stronger the change of the evaporative fraction. For soil moisture contents of 100mm or above, there is hardly any change in the slope anymore. The opposite behavior can be seen for the runoff. Because of the fact that the slope of the evaporative fraction increases infinitely with very low soil moisture contents, not representing realistic values, it was assumed in this model that soil moisture content can never fall below 5mm.

From these two graphs two observations can be made.

1) As explained before, the evaporative fraction indicates how much of the radiation is actually used for evapotranspiration. The rest of the radiation is used for warming. Therefore, in regions with lower soil moisture content and hence lower evaporative fractions a larger part of the radiation is transformed into sensible heat and therefore into warming of the surface.

2) With lower soil moisture content the change of the evaporative fraction also gets stronger per mm change in soil moisture (indicated by the slopes). As explained in number 1), this fraction has an effect on the warming of the surface. Therefore, the coupling strength between soil moisture and surface warming is stronger in dryer regions.

This soil moisture – evapotranspiration coupling influences also other factors, such as cloud formation or precipitation change. However, as the main focus of this thesis is to observe the effects on temperature, these factors will not further be considered.
When observing the soil moisture content during the different seasons, the following behavior can be seen: soil moisture content is high during spring, decreases to its lowest value in summer and increases again in autumn. Hence the effect of land-atmosphere feedback due to evapotranspiration should be strongest in summer while in spring and autumn it should be weaker.

2.1.5. Accounting for snow

If precipitation occurs in combination with a mean daily temperature below a threshold of 2°C, snow is accumulated. The amount of snow accumulation is linearly dependent on the temperature, where with 0°C all the precipitation occurs as snow and with 2°C all precipitation occurs as water. Snow melting on the other hand takes place if snow is present and the temperature is above 1°C, whereas the extent of the melting depends linearly on the temperature and is controlled by the degree-day melt factor $f_m$. The melted amount is added to the precipitation of that day. Therefore the runoff function (3) is not only scaled by precipitation but also by snow melt.

2.2. Data

All employed data sets cover entire Europe with a resolution of 0.5°x0.5°. The investigated time period is 1984-2011.

2.2.1. Meteorological data

Meteorological data were used from 1984-2011. Precipitation and temperature variables were obtained from E-OBS data set Version 9, while net radiation data were satellite-derived from the National Aeronautics and Space Administration (NASA)/ Global Energy and Water Cycle Experiment (GEWEX) surface radiation budget (SRB) project, as well as from the CERES experiment. Since the CERES data only extend from 2000-2012, the SRB data is used from 1984-1999. Moreover, for the period 2000-2007 where both data sets were available, the differences in mean and variability between them were determined, such that the SRB data could be scaled to correspond better with the more accurate CERES data (further information on Orth et al., 2014).
2.2.2. Model parameters

The evaporation parameters (β and γ) were fitted for each grid point in order to maximize the correlation between modeled sensible heat flux and observed temperature. For this purpose only days with soil moisture content below the 5%-quantile were considered, because the soil moisture-ET-temperature coupling is strongest under dry conditions (Orth and Seneviratne, 2014).

The other parameters were constant for all sites. The runoff exponent α was set to be 5.1, the degree-day mallet factor fm was set to 1.35 and water holding capacity c was assumed to constantly be 290 mm over the whole domain (Orth and Seneviratne, 2014).

2.3. Study structure

For this study the mean evaporative fractions, variabilities of the evaporative fraction and temperature variability of the spring, summer and autumn from 1984-2011 were computed at each grid cell. Winter was not considered as snow and frozen soils prevent soil moisture-temperature interaction. Additionally a correlation analysis was conducted to determine the coupling strength of soil-atmosphere interactions. All these modeled values were plotted into a map of Europe to exactly see the regional and temporal differences. In order to find out how land-atmosphere interactions change during extreme events, the two heat summers 2003 and 2010 were compared with the average of all summers (1984-2011). The characteristics of these two heat waves are shortly explained in the proceeding sections.

2.3.1. 2003 Heat wave

Reconstructions based on historical climatology data shows that summer 2003 was by far the hottest summer since at least 1500 AD, with an estimated excess mortality varying between 25 000 and 70 000 deaths in Western Europe (D’Ippoliti et al., 2010). Based on mean surface temperatures, the hottest region was centered over France, northern Italy, western Switzerland and Southern Germany (Rebetez et al., 2006). August mean monthly temperature anomalies reached +6.0 °C at many weather stations in these regions. Across wide areas (especially in central Europe) precipitation was below average from the beginning of the year and stayed largely below the long-term normal values much later (Rebetez et al., 2006).
2.3.2. 2010 Heat wave

The heat wave during the summer 2010 was hottest in June and occurred over large parts of Eastern Europe and western Russia, but also over the Eastern United States, Middle East and Northeastern China. It was partly caused by one of the strongest La Nina events ever observed, which lasted from June 2010 to June 2011 (Mokhof, 2011). Additionally, the western part of Russia was associated with a very prolonged (about two months) blocking of the westward transport in the troposphere of midlatitudes of the Northern Hemisphere (Mokhof, 2011). As a consequence, the drought of 2010 in Russia is estimated as most significant over the past 60 years. It lead to an estimated additional mortality of 58 000 people, a 25-30% drop of annual countrywide crop production and a financial loss of about round $ 15 billion to the Russian government (Barriopedro et al., 2011).

3. Results

3.1. Evaporative fraction

As can be seen from Figures 3-5, this fraction is the highest in spring, decreases in summer and increases again in autumn. Also an increase of the rate from south to north is visible. Especially the Mediterranean regions are different from the rest of Europe. Therefore regional variability exceeds the seasonal variability. North of the Mediterranean, the evaporative fraction is mainly limited by the highest possible evaporative fraction, defined by $\beta$. Especially in Southern Scandinavia and parts of Switzerland, Southern Germany and Austria this highest possible fraction $\beta$ goes up to 1. This is a result of the fact that in these regions with such high $\beta$ also observed precipitation was highest. The dark blue spots (low $\beta$) on the other hand in parts of Switzerland/Austria, east/southeast of the Black Sea and in Norway/Finland are all in mountainous or cold areas, indicating that there are other effects, such as snow cover causing these extremes.

When soil moisture content decreases from spring to summer, the southern parts show a stronger change of the fraction, while the change in the northern parts and central Europe is much weaker. It is important to mention, that the higher homogeneity in spring all over Europe does not mean that soil moisture content is similar everywhere. Instead southern parts have already much lower soil moisture content in spring than the rest of Europe. But because of the fact that the evaporative fraction stays almost constant with higher moisture content, (see Figure 1) almost no difference can be seen. As the slope of the evaporative fraction increases continuously with lower soil moisture content, regions with lower initial
soil moisture content experience a stronger change in the evaporative fraction per mm removal of moisture. In summer then these differences can clearly be seen. In central and northern Europe however, the initial soil moisture content in spring is too high and therefore much more moisture had to be removed in summer to yield the same changes in the evaporative fraction.

Figure 3: Mean evaporative fraction of spring 1984-2011.
Figure 4: Mean evaporative fraction of summer 1984-2011.

Figure 5: Mean evaporative fraction of autumn 1984-2011.
As it did rain less the spring before the heat wave 2003 (Rebetez et al., 2006), spring soil moisture contents were already lower and hence the sensitivity of the evaporative fraction to soil moisture higher. Therefore also the slopes of the evaporative fraction were stronger which resulted in a stronger change of the evaporative fraction from spring to summer. As a result more regions had lower evaporative fractions at this time. Compared to the average summer, these fractions were decreasing in most parts of central Europe, the Balkan Peninsula and France. In western Russia on the other hand, the fractions were lower. When linking these findings with the effect of evaporative cooling explained chapter 2.1.3., then one would expect an above-average temperature in regions, where the fractions were decreasing and a below-average temperature where these fractions were increasing. In fact, temperatures were increasing all over Europe, especially in France. As a consequence, warm temperature anomalies occurred in large parts of Europe. Only western Russia, which had no precipitation deficit in spring, experienced as expected a below-average summer temperature. To summarize, the low evaporative fraction, induced by a reduced precipitation frequency, contributed to the observed temperature anomaly. In fact, Mueller and Seneviratne (2012) suggested a strong relationship between precipitation deficits and the subsequent occurrence of hot extremes in a large fraction of the world. Therefore surface moisture deficits are a relevant factor for the occurrence of hot extremes.

In summer 2010 evaporative fractions were especially decreasing in western Russia. In fact, the strongest temperature anomaly was also located in western Russia. In the rest of Europe these fractions remained quite average. Some regions showed a slight increase while others showed a slight decrease. Temperature anomalies were above average all over Europe, except from Scandinavia, but except from European Russia these anomalies were not high.
Figure 6: Difference of the mean evaporative fraction of summer 2003 compared to an average summer.

Figure 7: Difference of the mean evaporative fraction of summer 2010 compared to an average summer.
One explanation for the summer heat linked with soil moisture has already been made with the preceding precipitation deficit. However, another factor can also contribute to this: Regarding the fact that higher net radiation (which is also an effect of climate change due to the increase in greenhouse gases) leads to higher evapotranspiration (not the ratio) in the first place and therefore to less soil moisture, which in turn decreases the evaporative fraction. Therefore a stronger warming not only occurs because of the higher radiation but also because of the feedback effect that now a higher fraction of the radiation is actually transferred to sensible heat and thereby used for the warming of the surface.

3.2. Variability of the evaporative fraction

As explained in chapter 2.1.3, the change in the evaporative fraction is dependent on two factors: The slope of the evaporative fraction function (Figure 2) and the change in soil moisture content. If soil moisture content would never change in a region, the slope had no effect on the rate of evapotranspiration at all, as the evaporative fraction would just always remain constant. In regions with high soil moisture variability on the other hand, the evaporative fraction would vary stronger, leading to a stronger interaction between land and atmosphere.

To compute the sensitivity of the evaporative fraction to soil moisture, \( s \), the slope of the evaporative fraction function is multiplied with the soil moisture variability:

\[
s = \frac{d\text{EF}}{d\text{SM}} \times \omega\text{SM},
\]  

(6)

Where \( s \) is the sensitivity of the evaporative fraction, \( \omega\text{SM} \) is the change in soil moisture content (mm) and \( \frac{d\text{EF}}{d\text{SM}} \) is the slope of the evaporative fraction.

In contrast to the maps of the evaporative fraction in the previous section, it can be seen, that the sensitivity of EF to soil moisture shows higher homogeneity in spring all over Europe. When soil moisture content decreases from spring to summer, the southern parts generally show a stronger sensitivity of the fraction, while the northern parts and central Europe remain almost the same. The stronger change of the sensitivity in southern regions is again because of the same reasons explained in the previous section. Especially western Russia, Ukraine, Turkey, the Balkan Peninsula, Italy, France and the Iberian Peninsula show an increase of the sensitivity in summer. In autumn, when soils get wetter again, the sensitivity decreases again. Only Spain, Turkey and parts of Southern European Russia and the Balkan Peninsula have autumn sensitivities similar to the ones in summer.
Figure 8: Sensitivity of the evaporative fraction of spring 1984-2011.

Figure 9: Sensitivity of the evaporative fraction of summer 1984-2011.
Figure 11 shows the difference of the sensitivity of the evaporative fraction to soil moisture of the summer 2003 heat wave compared to an average summer. There are higher values compared to the average summer especially in the Balkan Peninsula, Italy and France. In western Russia on the other hand, they were below average. These observations again fit well with the observed temperature anomalies during this time. One exception is the Iberian Peninsula. But in general, regions that had higher sensitivities also had higher temperatures while in European Russia where the variability was decreasing, temperatures were also below average.

The 2010 heat wave showed increases especially in European Russia and France, while they were slightly below average in parts of central Europe and the Balkan Peninsula (Figure 12). The strongest temperature anomaly was also located in European Russia.
Figure 11: Difference of the sensitivity of the evaporative fraction of summer 2003 compared to an average summer.

Figure 12: Difference of the sensitivity of the evaporative fraction of summer 2010 compared to an average summer. Differences in France were up to 0.6, but in order to get a better resolution, a smaller scale was used.
The comparatively strong sensitivities of ET to soil moisture in southern Europe contribute to a stronger land-atmosphere interaction and hence an amplification of heat waves through dry soils.

### 3.3. Temperature difference after rain event

In this section the mean temperature change associated with a rain event is investigated as a proxy for land-atmosphere interaction strength. Rain increases the soil moisture content and in turn the sensitivity of EF to soil moisture decreases (see previous sections). Hence, rain events are expected to dampen the soil moisture-temperature coupling.

The observed temperatures of the day before an increase in soil moisture content were subtracted from the temperatures of the day after. Only precipitation events of at least 3mm/day were used, as weaker precipitation amounts have less impact on soil moisture. The mean of all calculated temperature changes for a grid point was taken to get the average temperature change associated with a rain event. Note that the runoff ratio also plays a role in this observation. According to formula (6), the change of the evaporative fraction is dependent on two factors: the slope of the evaporative fraction and on the change in soil moisture content. Regarding the fact that in low soil moisture regimes not only the slope of the evaporative fraction is higher, but also the fraction of precipitation taken up by the soil (see Figure 1) (resulting in a higher change of soil moisture content after a rain event in dryer regions), both factors lead to a higher temperature change in dryer regions due to evaporative cooling.

Comparing results for spring, summer and autumn, Figures 13-15 generally show temperature decreases after precipitation events which are highest during summer and autumn. In the Balkan Peninsula, Eastern Turkey and Southern western Russia changes are highest during autumn. In central Europe on the other hand, changes are highest during summer. In summer these changes can be up to -2.5 °C, while in autumn they are up to -3 °C. Comparing with the sensitivities in Figures 8-10, it can be seen that temperature changes are in fact mostly higher in regions where the sensitivities are higher. In many coastal areas this effect is not as strong (e.g. Portugal, Greece) because of the buffering effect of the sea on coastal land temperatures. In the region around Western Balkan Peninsula, Czech Republic, Southern Germany, Austria and Switzerland temperature changes are always high (even in spring, where the slopes of the evaporative fraction are low all over Europe), indicating that this is an effect independent from the slopes (further information in the next section).
Figure 13: Mean temperature difference of the day before and the day after a rain event of at least 3mm/day for: a) spring 1984-2011. (average: -0.79 \pm 0.65 °C).

Figure 14: Mean temperature difference of the day before and the day after a rain event of at least 3mm/day for summer 1984-2011. (average: -1.29 \pm 0.51 °C).
Figures 16 and 17 show the respective results for the hot summers 2003 and 2010. Considering the 2003 heat wave, temperature changes were increased mainly in the Balkan Peninsula, and central Europe up to 6°C compared to the summer mean (1984-2011), while they were decreased in western Russia and Eastern Europe. This spatial pattern fits very well with the change of the sensitivity of the EF to soil moisture in summer 2003 displayed in Figure 11.

The temperature changes in the 2010 summer were increased especially in western Russia, with differences partly exceeding 6°C. Again, the sensitivity in Figure 12 compares well with this spatial pattern. This underlines the usefulness of the employed temperature difference as a proxy for soil moisture-temperature coupling strength. Only France, where also an increase would have been expected showed a decrease in temperature change. For central and Eastern Europe the findings were often contradictory: While the slopes were decreasing, temperature variability was increasing in the same region. However, when comparing these changes with the ones in western Russia, it can be seen that they were really low. So in this case the changes of the slopes were too small to have an effect on temperature variability.
Figure 16: Difference between the temperature difference of the heat summer 2003 and average summer of 184-2011. Therefore the day before and the day after a rain event of at least 3mm/day were used. (average: -1.35 ± 1.17 °C).

Figure 17: Difference between the temperature difference of the heat summer 2010 and average summer of 184-2011. Therefore the day before and the day after a rain event of at least 3mm/day were used. (average: -1.61 ± 0.93 °C).
However, it is important to mention that for this calculation the amount of precipitation was not considered. Therefore in regions where the amount of water per rainfall is above average, leading to a stronger change in soil moisture, leading to an increased change the evaporative fraction, the calculated temperature differences are too high (see formula 6). This can be one reason why temperature changes were so high in parts of central Europe. Exactly these regions have higher amounts of precipitation.

3.4. Correlations

In the previous section hot spot regions of EF sensitivity with respect to soil moisture were identified. However, water and energy fluxes between land and atmosphere depend on actual evapotranspiration. This section investigates if the soil moisture-evapotranspiration coupling is related with the soil moisture-EF relationship.

When converting Equation (2) to

\[ E = \frac{R \cdot B}{\lambda_p} \cdot \left( \frac{\omega}{c} \right)^\gamma \]  

(7)

It can be seen that the evapotranspiration is only dependent on net radiation and soil moisture content, as the other variables are always constant for the same region. While the soil moisture content is a determinant for the evaporative fraction, ranging from 0 to 1, net radiation can vary from -100 W/m\(^2\) up to more than 200 W/m\(^2\) (however, during the three observed seasons net radiation was hardly ever below 0, as this happens mainly in winter). Till now the influence of the variability of the net radiation was not considered. However, thinking of the fact that the variability of net radiation is quite high and that the evaporative fraction changes only in small steps from day to day, it is important to examine the influence of the variability of the net radiation and its effect on net evapotranspiration. In order to consider this fact, a correlation analysis was made between soil moisture content (as an indicator for the evaporative fraction) and net evapotranspiration. Therefore all the days from 1984-2011 were used. High correlation between soil moisture content and evapotranspiration should be expected in regions, where the change of the evaporative fraction is high (meaning high variability of the evaporative fraction and therefore a stronger coupling of land-atmosphere interactions) and/or where net radiation stays more constant. Low correlations should be expected in regions where the opposite determinants can be seen (lower variability of the evaporative fraction and/or high day-to-day-variability of net radiation).

Correlations are positive in Southern regions (Iberian Peninsula, Southern Italy, Turkey, South-Eastern Balkan Peninsula) and western Russia. In these regions evapotranspiration dynamics are similar to soil moisture dynamics; they coincide with the regions where the evaporative fraction is comparatively sensitive to soil moisture. This finding shows the
pronounced soil moisture impact on land-atmosphere interactions in these regions additionally to the role of net radiation dynamics.

In contrast, in almost entire central Europe there are negative correlations. This indicates a stronger influence of net radiation and less influence of the soil moisture content because dynamics of net radiation and precipitation (and hence of soil moisture) are usually in opposite direction. Correlations especially decrease towards the coast. In these regions it is overall wetter and colder, and these conditions mitigate the soil moisture impact, in contrast to the heat-wave conditions investigated before.

As evapotranspiration is simply the transformation of radiation into sensible heat, soil moisture should then also have a bigger effect on temperature in regions with higher correlation. When comparing the correlation map with the map of temperature change after a rain event in summer, it can in fact be seen, that temperature changes are really higher in regions where correlations are higher. On the other hand, in regions where correlations were low, temperature changes were not necessarily also low. This is a representation of the fact that in these regions soil moisture does not have a big effect on local climate and is therefore more dependent on net radiation which can strongly vary. This explains why in central Europe temperature differences were often different than expected from the slopes of the evaporative fraction.

![Figure 18: Correlation of evapotranspiration with the soil moisture content.](image)
4. Conclusion

This thesis examined the sensitivity of evaporative fraction and actual evapotranspiration on soil moisture in Europe, as well as the effect of soil-atmosphere feedback on local temperature. The following conclusions can be drawn:

1) Soil moisture dynamics can influence near-surface temperature. As the evaporative fraction decreases with lower soil moisture content, below-average soil moisture content can lead to a stronger warming. Above-average soil moisture content on the other hand can lead to a stronger cooling.

2) Soil moisture anomalies impact the coupling strength between moisture content and evapotranspiration (as indicated by the slopes). While below-average soil moisture content can therefore lead to a stronger variability in temperatures, above-average moisture content can lower this variability. These changes in variability can even be experienced on a day-to-day basis.

3) These two effects get stronger, the stronger the soil moisture anomaly. Weak soil moisture anomalies often show no effect at all as then other factors overweight the effect of the evaporative cooling.

4) The lower the initial soil moisture content, the stronger the effects explained in 1) and 2). This is a representation of the fact that the slopes of the evaporative fraction increase much stronger, the lower the soil moisture content. Therefore in southern regions, where soil moisture content is lower compared to northern regions, an anomaly in the moisture content has a stronger effect on the temperature than it would have in northern regions.

5) A decrease in soil moisture content has a stronger effect on temperature than an increase, as the slope of the evaporative fraction increases much stronger, the lower the soil moisture content.

The importance of point 3) and 4) and 5) was especially demonstrated for the summer heat in 2010. In central and Eastern Europe soil moisture contents are already high and were mainly slightly above average during this summer. As a result, the impact of the change in soil moisture content was not strong enough and temperature variability was often contradictory from the expectation when comparing with the change of the slopes. In western Russia on the other hand, where conditions are already dry and soil moisture was low, the change in the temperature variability was accordingly high. When considering that Southern Europe will tend to get drier in the future (Seneviratne et al., 2012) and linking this fact with points 4) and 5), then especially southern regions that are already dry will experience a stronger land-atmosphere interaction and therefore a stronger increase in temperature and temperature variability in the future. Northern regions on the other hand are expected to be less affected by this mechanism. Therefore the above explained findings and the fact that soil moisture memory spans weeks up to months underline the importance of including this kind of land-atmosphere interactions into climate prediction models.
5. References


