Relationships between forests and weather

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Contents

Executive Summary

1. The main mechanisms by which forests modify weather have been identified. They are the surface albedo, transpiration and evaporation of water vapour, aerodynamic effects, and emission of hydrocarbons whose oxidation can form aerosol particles.

2. Different mechanisms are dominant for each class of forest. Boreal forests affect local weather and climate via their low albedos, causing a local warming. Temperate forests modify weather via the albedo and transpiration of moisture, but their exact impacts on climate are the least certain. Tropical forests cool climate via their very high transpiration rates; the moisture transferred to the atmosphere forms large clouds which reflect incoming solar energy and cause a further cooling.

3. Forests have a cooling impact on global climate via the uptake of carbon dioxide. Deforestation releases the carbon back to the atmosphere causing a warming. Tropical forests absorb the largest amounts of carbon dioxide.

4. Although the main mechanisms by which forests modify climate are known, feedbacks between these mechanisms and the local climate are less well understood. These feedbacks often occur on small spatial scales which cannot be resolved by climate models, and have only been studied over limited areas with very high resolution (mesoscale) models.

5. Many modelling studies have been published in the scientific literature to assess the potential impacts of deforestation and afforestation. In boreal regions, afforestation would cause a local warming but deforestation would cause a cooling. The impacts in temperate regions are unclear, with poor agreement between the various studies, which are caused by the simulated response by different models to deforestation, the way in which physiological processes of forests and other vegetation are represented in the models, and the numbers and types of vegetation that are modelled. Other reasons include differences in modelled soil moisture amounts. Deforestation in tropical regions would produce a drier climate which may be impossible to reforest.

6. Rainfall in some regions could be enhanced if areas cleared for crops are intermingled with remaining forest. Small-scale circulations can occur which form convective clouds and rainfall above the deforested areas.

7. Feedbacks between the ocean circulation and climate appear to be important in south-east Asia following deforestation, but model results do not agree on the sign of changes in circulation and rainfall.

8. Climate change will impact on forests and climate, but the exact effects are unclear in many regions owing to poor agreement between global climate model projections of future climate, especially changes in rainfall and atmospheric circulation.

1. Introduction

This report has two main objectives. First, to review the current understanding of relationships between forests and weather worldwide, and highlight how they relate to Europe. Secondly, to synthesize current understanding of the main drivers of forest-weather relationships, the processes through which these interactions occur, and the factors that may threaten them.

1.1 What is a Forest?

There is no universal definition of what constitutes a forest. The Food and Agriculture Organisation of the United Nations (FAO) defines a forest as:

"….a minimum area of land of 0.05-1.0 hectares with tree crown cover (or equivalent stocking level) of more than 10-30 per cent with trees with the potential to reach a minimum height of 2-5 metres at maturity *in situ*. A forest may consist of closed forest formations where trees of various storeys and undergrowth cover a high proportion of the ground or open forest. Young natural stands and all plantations which have yet to reach a crown density of 10-30 per cent or tree height of 2-5 meters are included under forest, as are areas normally forming part of the forest area which are temporarily unstocked as a result of human intervention such as harvesting or natural causes but which are expected to revert to forest." (Food and Agriculture Organisation, 2006).

1.2 Types of Forest

Forests may be divided into three main groups; boreal forests, temperate forests and tropical rain forests. These three groups are described briefly below.

Boreal forests are the largest terrestrial biome and are also known as taiga. They occur between latitudes of 45°N and 70°N, with the majority (approximately 67%) located in Eurasia and the remainder in Canada and Alaska (Soja et al., 2007). The forest is predominantly coniferous and is quite dense at lower latitudes. In the north of this region the forest becomes sparser as it shifts to a tundra ecosystem. In far eastern Siberia the forests are dominated by Larch which is deciduous, and drops its needles in winter.

Temperate forests exist in many regions of the Earth, in eastern North America, northeastern Asia, western and central Europe, and parts of China. Large areas of temperate forests in the eastern United States, Europe and China have been cleared for agriculture. Temperate forests have a number of sub-classes and include both deciduous trees (which lose their leaves during winter) and evergreens.

Tropical rain forests occur in equatorial regions, and are characterised by very high annual rainfall and warm temperatures. These forests have high rates of evapotranspiration (Bonan, 2008). The climate within tropical forests is stable, and humidity levels are usually very high. These forests support a rich and diverse array of plants and animals. The mass of carbon stored in tropical forests is much higher than in temperate and boreal forests (Bonan, 2008).

2. Mechanisms by which forests influence weather and climate

2.1 Overview

Forests can influence local and regional weather and climate via a number of different mechanisms (Betts, 2006). The albedo of a surface is defined as the ratio of the radiation reflected from a surface to the total radiation falling on the surface. Forests have low albedos; for example, 0.08 (Betts and Ball, 1997), which means that only 8% of the incoming solar energy is reflected. Other surfaces, such as snow and ice, have much higher albedos and reflect almost all of the incoming solar energy. The low albedo of forests means they absorb most of the incoming solar radiation and become warmer, and then warm the air around and above them. The energy transferred from the forests to the surrounding air is called *sensible heat*, and involves a change of temperature.

Forests also absorb water from soils via their roots and release it into the atmosphere, a process called evapotranspiration, which has several effects on weather and climate (von Randow et al., 2004). Water is removed from the soil as a liquid, but is released as a vapour. Energy is required to change the phase of water, and so this process acts to cool the surface. When water changes phase, energy is either absorbed (e.g., changing from liquid to vapour) or released (changing from vapour to liquid), but the temperature does not change. This energy is referred to as *latent heat*. The water vapour released can be transported to higher altitudes where temperatures are cooler, and condense to form clouds and rain. Clouds have high albedos, and so reflect incoming solar energy and cool the surface below them.

Another mechanism by which forests can affect local weather is via aerodynamic roughness. Surfaces which are aerodynamically rough increase air turbulence above them, which causes a drag on the air flowing over them and reduces the wind speed. Surfaces which have low aerodynamic roughness include ice, grasses and crops. The surface of forests, however, is aerodynamically rough, which causes turbulence in the air and enhances the exchange of sensible heat and moisture from the forest into the air (Rotenberg and Yakir, 2010). The moisture content of air above the forest becomes larger, which could cause convection, cloud formation and an enhancement of rainfall (Millán, 2008). Trees sway in the wind which further acts to slow down the wind speed and increase turbulence (Su, 2010). Overall, momentum has been transferred from the air to the forests.

The climate within forests, especially tropical ones, tends to be more stable than non-forested areas. Removal of forests and replacing them by crops, grassland or bare soil will therefore have a large impact on the climate and hydrology of the deforested area. After deforestation, evaporation will be reduced, meaning more energy is emitted from the ground as sensible heat instead of latent heat, which acts to warm the surface. The reduced flux of moisture to the atmosphere means humidity levels and precipitation are likely to be reduced (Betts, 2006).

The mechanisms described above show how forests can influence local and regional weather and climate. However, all forests have long-term global climate impacts via interactions with carbon dioxide (CO_2) concentrations in the atmosphere. CO_2 is the most important greenhouse gas, owing to its long lifetime in the atmosphere and strong absorption of infrared radiation emitted from the Earth’s surface (IPCC, 2007). The growth and spread of forests results in an uptake of CO_2 which reduces its levels in the atmosphere and results in a net cooling. However, deforestation results in the carbon being released into the atmosphere (as CO_2) and correspondingly warmer temperatures.

The effects of a lost forest carbon sink could be long lasting, as it takes many decades for most trees to grow to maturity and absorb the CO_2.

The exchange of sensible and latent heat, moisture, momentum and carbon dioxide with the atmosphere are the key processes by which forests can modify weather and climate. In order to understand these processes fully, and identify any important feedback mechanisms, models have been developed to simulate these processes. An overview of modelling of forest-weather interactions is given in section 2.2.

All three classes of forest (boreal, temperate and tropical) interact with and influence weather and climate as described above. However, different mechanisms are dominant for each class, as discussed in the sections 2.3 to 2.5.

2.2 Boreal Forests

Boreal forests have a large impact on local temperatures, and, of the three main forest types, the largest influence on global mean temperature (Bonan, 2008). The main impact of boreal forests on climate is via their low albedo and strong absorption of incoming solar radiation. The absorbed solar energy is released primarily as sensible heat, leading to warmer temperatures all year round when compared to non-forested areas. These forests exist in and adjacent to tundra, which has a higher albedo. The tundra regions are also covered by snow and ice for large parts of the year and reflect most of the incoming radiation, whereas the much darker forests remain exposed. Even if the trees are covered with snow, multiple reflections within the canopy act to reduce the effective albedo (Betts, 2006). The uptake of CO_2 and rate of evapotranspiration by boreal forests is moderate when compared with temperate and tropical forests (Bonan, 2008).

The effect of boreal forests on climate is amplified via sea-ice feedbacks (Lee et al., 2011). As temperatures in high northern latitudes rise, the amount of snow and ice on the land and ocean is reduced, exposing the darker ocean and land surface (which have lower albedos), which in turn absorb more of the incoming solar energy. The warming is thus reinforced via a snow-ice-albedo feedback. Boreal forests act to moderate this feedback by masking the high reflectivity of snow and ice (Bonan et al., 1992). Deforestation in boreal regions would, overall, increase the surface albedo, leading to a net cooling, and an increase of snow and ice cover which would act to enhance the cooling via the snow-ice-albedo feedback. Snowmelt measurements have been obtained on forested and deforested sites, which show a strong sensitivity to the presence of the forests (Harvey, 1988; Ohta et al., 1993; Yamazaki, 1995). The forests warm the surface and cause the snow to melt earlier than if there were no forests.

2.3 Temperate Forests

The influence of temperate forests on climate is more complex and uncertain than for boreal or tropical forests (Field et al., 2007; Bala et al., 2007; Bonan, 2008). A number of climate model studies suggest that temperate forests cool the air compared to grasslands and croplands, while other studies show the opposite (see Jackson et al., 2008 and Anav et al., 2010 and references therein).

Some of these apparent contradictions can be related to season, water availability and soil moisture levels. For example, during the warm seasons, convective rainfall should change in response to modifications to land cover, since surface fluxes of moisture, heat and transpiration will be altered. When sufficient moisture in soils is available, the local cooling by forest is caused by evapotranspiration and latent heat transfer. The surface roughness may contribute to increased rainfall and surface cooling in nearby areas. Paradoxically, these forests also deliver more heat to the atmosphere because they are darker and absorb more sunlight (a transfer of sensible heat). In other regions where water availability is lower, forest plantations may warm the regional climate by absorbing more sunlight without substantially increasing evapotranspiration.

In temperate regions, it is difficult to detect the signature of forest cover changes on rainfall owing to the naturally highly variable frequency of synoptic scale meteorological systems (e.g. frontal depressions) and rainfall patterns, the regional landscape variability, the nonlinear changes in the forest cover and the related effects of urbanization, pollution loadings and regional circulation.

Tueling et al. (2010) studied the difference between the temporal responses of forests and grasslands ecosystems during heat waves. Initially, surface heating was twice as high over forest than over the grassland owing to increased evaporation from the grassland in response to increased solar radiation and temperature. However, this process accelerated depletion of soil moisture and leads to increased heating. Overall, forests contributed to increased temperatures in the short term, but mitigated the impact of heat waves in the longer term, as they can access deep reserves of water and continue transpiring even when moisture in the upper soil levels has been depleted.

In conclusion, as shown by Douville and Royer (1997), the role of both temperate and boreal forests in modifying and controlling climate should not be underestimated. These forests exert a strong influence on the surface climate in the mid and high latitudes, and also on climatic events in the tropics (Douville and Royer, 1997). Thus significant climate change could be caused merely by redistribution of the terrestrial ecosystems. This redistribution could be caused by intensive logging as already observed in some tropical areas; and in the case of the boreal forest, it could also be induced by the increase in the atmospheric concentration of CO_2 and other greenhouse gases. Furthermore, the variability of the climatic, soil and vegetation characteristics of a region, as well as the representation of land surface processes in the applied climate model, also have an influence on the simulated vegetation-atmosphere interactions.

Past deforestation in temperate regions has resulted in cooler temperatures via the change in albedo (Betts et al., 2008). Forests with low albedos have been replaced by crops which have higher albedos, and so the latter absorb less of the incoming solar radiation. Additionally, this cooling has been enhanced via the sea-ice-albedo feedback described earlier (Bonan et al., 1992; Lee et al., 2011). Without this feedback, other effects could cause deforestation to result in an overall warming. Lee et al. (2011) examined differences in measured temperatures between adjacent forested and non-forested areas. The temperature differences are partly dependent on latitude, so that the forested areas become progressively warmer relative to the open areas as the latitude increases. This dependence does not seem to hold south of 35°N.

Afforestation in temperate regions would still act to mitigate global warming through CO_2 uptake, but the warming effect of the decreased surface albedo would partly offset the cooling effect of sequestering CO_2 especially in snowy landscapes. In some parts of the boreal forests, the warming effect of decreased surface albedo could outweigh the cooling effect of CO_2 sequestration (Betts et al., 2008).

2.4 Tropical Forests

In contrast to boreal and temperate forests, a combination of high solar input, high soil moisture availability and a long photosynthetic season mean that tropical forests absorb CO_2 strongly and have high rates of evapotranspiration. The effect of the low albedo of the forests is offset by strong evaporative cooling (Bonan, 2008). The high evapotranspiration rates mean large clouds develop above these forests that cause an additional cooling by reflecting sunlight back to space (Lean and Warrilow, 1989; Shukla et al., 1990; Dickinson and Kennedy, 1992; Hoffmann and Jackson, 2000). Model simulations suggest that extensive tropical deforestation would lead to a drier, more savannah-like climate that would be difficult to reforest (Nobre et al., 1991). Overall, tropical forests cool the climate via evaporative cooling and absorption of CO_2 (Bonan, 2008).

In the Amazonian region a significant proportion of the rainfall occurs from recycled moisture released by the forests via evapotranspiration. Salati (1987) estimated that evaoptranspiration was responsible for between 50% and 75% of the measured rainfall, with the remainder of the moisture originating from evaporation from the oceans. Silvas Dias et al. (2009) also estimated that about half of the rainfall in the Amazon region originates as moisture supplied by the forests. In a brief review of isotopic data collected over the Amazon, Henderson-Sellers et al. (2002) noted that evaporation of water directly from rivers, lakes and other areas of open water contribute between 20% and 40% of the total evaporation flux. The isotopic data suggested that between about 50% and 90% of the rainfall (depending on the location) originates from recycled water vapour. Other studies (Salati et al., 1986; Costa and Foley, 1999) also estimate significant recycling of water over tropical forests.

Montane “fog” forests are common ecosystems on the interior mountain ranges in the tropics (e.g., Costa Rica) but are also found in some Mediterranean countries and small oceanic islands. These forests play a very active role in the local hydrological cycle. They capture water from rainfall, and also harvest water directly from wind- (i.e. advective) or convective-driven clouds (Bruijnzeel, 2001). Interception of cloud and fog water significantly affects the water availability for the forests itself but also in downstream locations.

2.5 Overview of the representation of forest-weather interactions and feedbacks in climate models

Several recent review papers have addressed the issue of how the state of the land surface can influence weather and climate at local, regional and global scales (Andréassian, 2004; Bonan et al., 2008; Pielke et al., 2007; Ellison et al., 2011). These reviews illustrate good theoretical reasons why the vegetation cover in particular could affect the Earth's climate and why these interactions should be included in simulations by climate models.

Feedbacks between vegetation cover and weather/climate, as represented in climate models, are summarised in the work of Pitman (2003). The effects of deforestation and subsequent feedbacks are shown in Figures 1-5 below. In all the figures, dotted lines represent a positive feedback (where the effects of deforestation are amplified) and the dashed lines represent a negative feedback (where the effects of deforestation are reduced). The effects illustrated are an increase in albedo (Figure 1); a reduction of leaf area index (LAI; Figure 2); a decrease in root depth (Figure 3) and surface roughness (Figure 4). The impacts of reduced soil moisture, which could follow deforestation or afforestation, depending on the season, are shown in Figure 5.

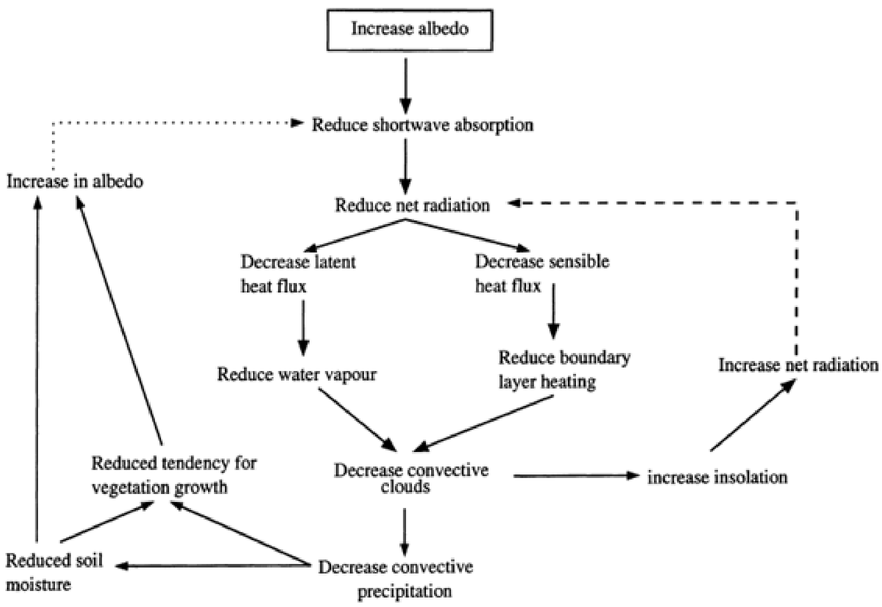

Figure 1. Effect of deforestation on surface albedo and subsequent feedbacks on radiation fluxes, clouds and precipitation.

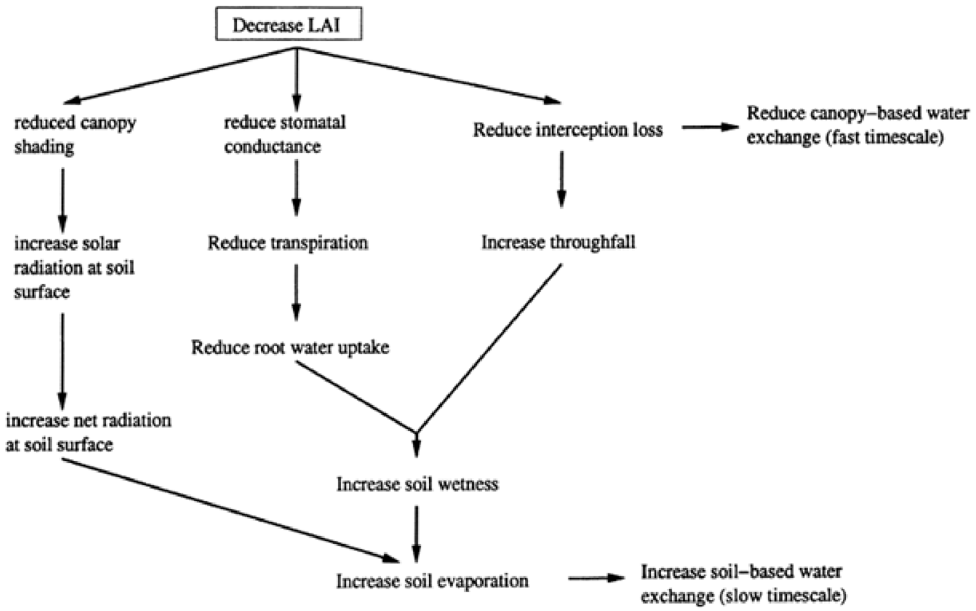

Figure 2. Effect of deforestation on leaf area index (LAI), and subsequent impacts on evapotranspiration and soil moisture.

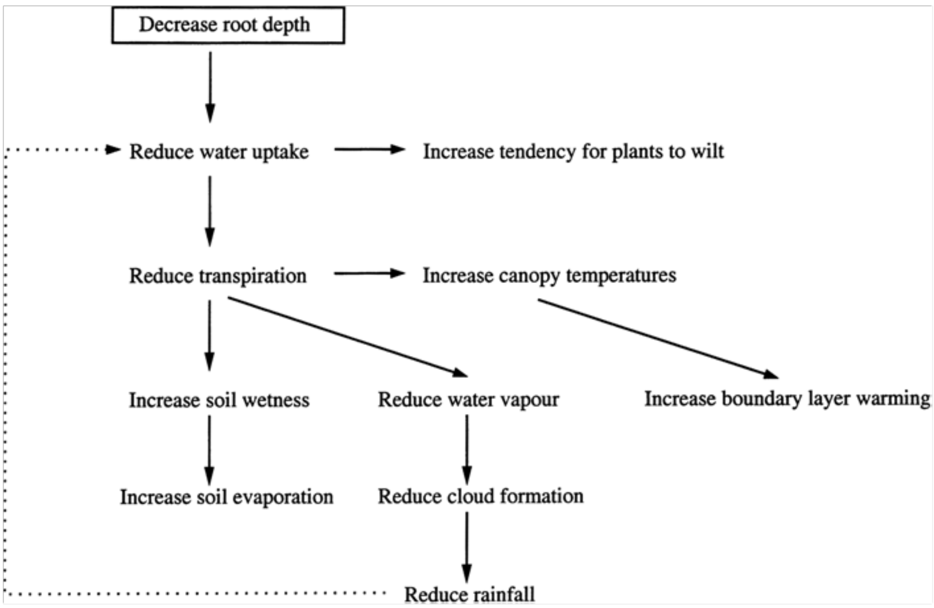

Figure 3. Effect of deforestation on the root depth of the plants and subsequent impacts on soil moisture and cloud formation.

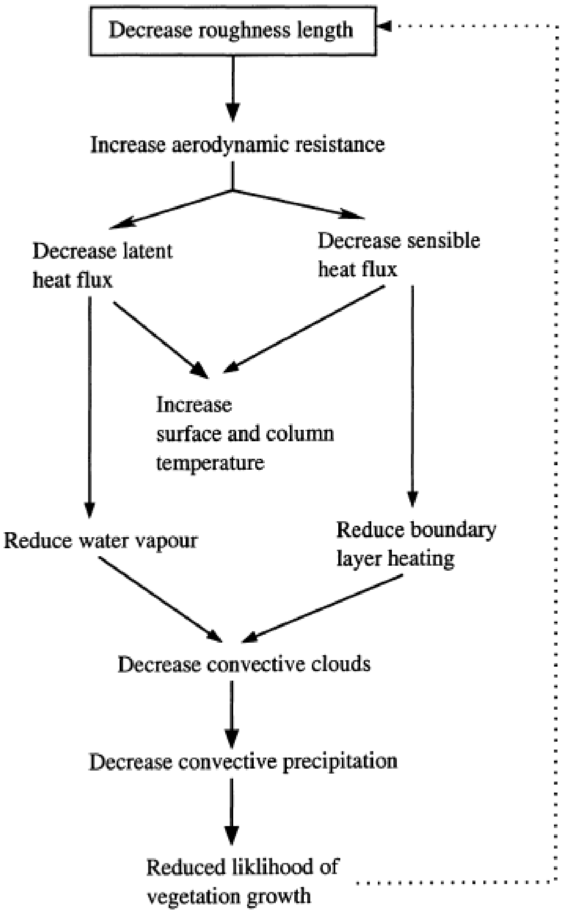

Figure 4. Effect of deforestation on surface roughness and subsequent impacts on heat fluxes and cloud formation.

As seen in Figures 1 to 5, changes in the characteristics of the land surface affect the transfer of water between the land and the atmosphere. In particular, a change in the nature of vegetation affects interception and evapotranspiration processes; then a change in the distribution of vegetation modifies the balance between fluxes of moisture originating from the soil and those derived through canopy processes. Changes in evapotranspiration, soil evaporation, re-evaporation of intercepted water etc. affect runoff and soil moisture content. These then affect a variety of other processes through the link with the surface energy balance.

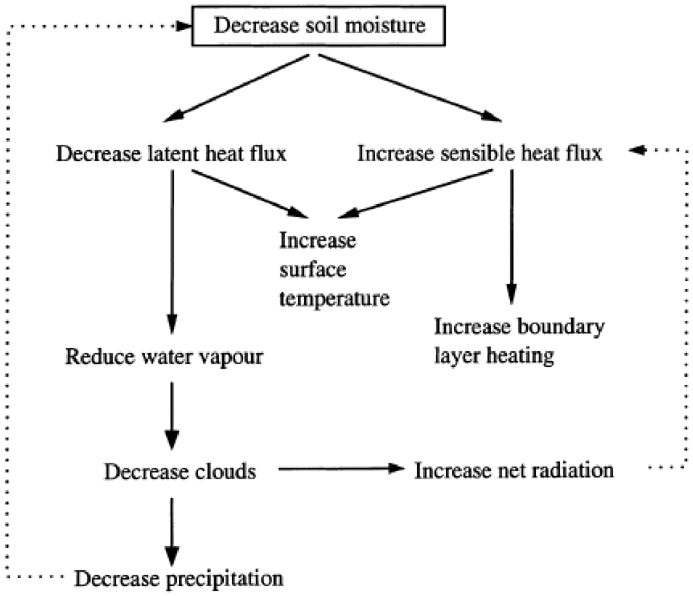

Figure 5. Effects of reduced soil moisture on heat fluxes, cloud formation and rainfall.

3. Impacts of changes in forest cover on climate

In this section, studies which use observations and models to understand impacts of changes in forest cover on weather and climate are reviewed. These studies are divided in to two groups, those which consider the physical impacts on climate discussed earlier (section 3.1), and chemical effects of forests on climate (section 3.2). In each section, the various impacts are discussed along with example studies. However, in many cases only a few studies are available, and some of the effects that have been observed or simulated may be specific to those regions.

3.1 Physical Effects

In this section, the observed physical impacts of deforestation and afforestation are discussed. These impacts occur via the change in the land cover itself, which include aerodynamic roughness, albedo and evapotranspiration rates. After human-induced deforestation, the land is usually converted to grassland or is used for growing crops. These areas tend to have a higher albedo than the forests and reflect more incoming solar energy. They have a lower surface roughness and leaf area index, and the moisture storage capacity of the ground is reduced (Maynard and Royer, 2004). The physical effects include wind speeds, cloud cover, local temperatures and rainfall.

3.1.1 Impacts of deforestation on remaining forest and climate

Laurance (2004) has reviewed the physical impacts of deforestation in tropical areas on the remaining forest. Deforestation often results in a patchy landscape of fragmented forest. The edges of the forest fragments are prone to damage by wind and fire, and enhanced tree mortality owing to the stresses of the environment caused by the harsher climate – increased temperatures and lower humidity. The climate underneath an undisturbed tropical forest is stable, and is warm, humid and dark, with little wind. The edges of the forest will be exposed to high sunlight levels and higher wind speeds. Trees which regrow around the edges of cleared forest are likely to consist of different species than those of the forest interior and are more tolerant of the drier conditions. However, a forest fragment surrounded by new growth will be protected from wind and the harsher climate. Once created, forest fragments may become less humid and more prone to fires in the future. Actual climatic data in forest fragments near their edges is lacking (Laurance, 2004).

Ghuman and Lal (1987) examined the impact of deforestation on soils, hydrology and climate in a rain forest in Nigeria. Compared with the forested area, air and surface soil temperatures were higher in the cleared area, and the diurnal cycle in temperature was larger too. Relative humidity in the forest remained around 90% throughout the day, whereas it fell to 50% during daylight hours in the cleared area.

3.1.2 Impacts of afforestation in a semiarid region

Rotenberg et al. (2008) examined the impact of afforestation in a semiarid region of Israel. They measured upward and downward fluxes of shortwave and longwave radiation, together with sensible and latent heat fluxes. Although the trees covered less than 60% of the land area, the canopy absorbed 80% of the net solar radiation. Around noon, the air temperature above the trees was warmer than the surface, but at other times the reverse was true. Rotenberg et al. (2008) concluded that, initially, afforestation would cause a local warming owing to the much lower albedo of the forest compared with the surrounding soils. After 30-50 years, the net effect of afforestation would be a cooling of global temperatures caused by the CO_2 uptake.

3.1.3 Cloud Cover

As described in section 2.4, montane “fog” forests are found on mountain ranges in the tropics but also in some Mediterranean countries and in small oceanic islands. Deforestation of montane forests can have serious impacts on the hydrological cycle of the surrounding landscape, since other vegetation types do not have the proper structure to intercept cloud and fog moisture. Ray et al. (2006) used a combination of observations and a regional climate model to study the impact of deforestation on orographic cloud formation over the Monteverde cloud forests in Costa Rica. Their results indicated that deforestation in the lowland and premontane regions has increased surface sensible heat fluxes and decreased latent heat fluxes. These changes raise the air temperature and lower the dew point temperature of air masses that blow over the lowland and premontane regions. Consequently, the orographic cloud banks form at higher elevations, and envelop smaller areas of the mountains which leads to greatly reduced harvesting of cloud water by montane vegetation. Similar results were obtained in a later study by Pielke et al. (2007). Reforestation of mountain tops and slopes can re-establish this particular hydrological interaction.

However, the impacts on clouds in many other areas is uncertain, although other studies also indicate reduced cloud cover over deforested areas compared with intact forest (Nair et al., 2011). The exact impacts on clouds are likely to be locally dependent and vary between regions.

3.1.4 Aerodynamic Effects

Aerodynamically, forests present a rough surface which acts to slow down surface winds by causing turbulence above the canopy (Rotenberg and Yakir, 2010). An analysis of wind speeds at many locations in the northern hemisphere by Vautard et al. (2010) indicated a downward trend, and that high wind speeds have declined by a greater proportion that low winds. Changes in atmospheric circulation can explain 10-50% of the surface wind slowdown. Simulations using a mesoscale model suggested that an increase in surface roughness, estimated from afforestation and biomass increases in Europe and Asia, could explain between 25% and 60% of the observed wind slowdown. Additionally, areas with pronounced stilling generally coincided with regions where biomass has increased over the past 30 years, supporting the role of forest regrowth in wind slowdown.

An example of the aerodynamic effect on local precipitation has been studied in south-west Australia, where rainfall has reduced and river flows around the city of Perth have fallen by around 40% since the mid twentieth century. This decreasing trend has been attributed to deforestation (Adams, 2010). The replacement of forests by cropland and pasture has reduced the aerodynamic roughness of the surface. When the forests were present, the rough canopy slowed the air movement down by causing drag and promoted vertical movement and rainfall. After clearance of the forests, the rainfall occurs further inland and outside of the river catchments around Perth. An analysis of observations and regional model results by Nair et al. (2011) supports these ideas. The loss of the forest has resulted in reduced wintertime rainfall over the areas cleared, which is partly caused by the reduced aerodynamic roughness after conversion of forests to crops. A more heterogeneous pattern of forest and wheat could have helped to reduce the local change in rainfall (Chapin III et al., 2008).

3.1.5 Effects on local precipitation

Depending on the scale and pattern of deforestation or afforestation, rainfall can either be enhanced or reduced. Otterman et al. (1990) proposed that an observed increase in rainfall in southern Israel has been caused by afforestation and increased cultivation in the area. They attributed the increases in rainfall to enhanced convection and advection owing to larger sensible heat fluxes from the vegetation during the day time. Other studies have reached similar conclusions (Pielke et al., 2007, and references therein).

It is evident that the chain of processes which affect local rainfall is complex and can be longer lasting than expected, as in case of regional and local droughts that may be amplified by ecosystem feedbacks by decreasing vegetation cover, energy absorption, and convective uplift. For example, climate and/or anthropogenic changes (e.g. oceanic circulation and/or overgrazing) triggered a drought in the Sahel in the 1970s that became prolonged as the reduced plant cover caused local climate and hydrological changes. As the vegetation died, the albedo increased, evapotranspiration decreased and convective uplift and associated monsoon rains were also reduced in the region. In the extreme case of full desertification, Xue and Shukla (1993) showed that this would cause a reduction in total rainfall and a delay in the beginning of the rainy season. Hence, ecosystem feedbacks apparently contributed to the magnitude and extended duration of this drought (Chapin III et al., 2008).

Meher-Homji (1991) has reviewed many studies of rainfall trends in several countries. Rainfall measurements in two adjacent areas in India were compared, one of which had experienced large scale deforestation. There was some evidence to suggest a decline in rainfall and days of rain in the deforested areas when compared with the undisturbed site. Other studies in India and Africa cited by Meher-Homji (1991) provide further empirical evidence for deforestation to be the cause of a decline in rainfall and an increase in droughts. However, these impacts may be specific to the areas studied.

It should be noted that there may be other causes for declining rainfall in some regions. Deforestation in Indo-China since the 1950s had been proposed to be responsible for the observed decrease in rainfall in Thailand of 30% (Kanae et al., 2001; Adams, 2010) during the late summer monsoon season (September). A later study by Takahashi and Yasunari (2008) found a weakening trend in tropical cyclone activity between 1951 and 2000, and concluded that this was the probable cause of the observed reduction in rainfall, although a contribution from deforestation could not be ruled out.

3.2 Chemical and Aerosol Effects

Forests have a complex but indirect effect on weather and climate via the physiological impacts of changing carbon dioxide levels, and the production and emission of hydrocarbons. Fires, which release the carbon stored in forests into the atmosphere as CO_2 together with smoke particles can also have a significant impact on climate. In this section, reported studies on these impacts are discussed.

There are also several chemical effects of forests on local and global climate. Forests absorb CO_2, which acts to cool global climate, but release a range of reactive chemical species which can alter the lifetime of methane and partly control concentrations of ozone, both of which are important greenhouse gases,. These reactive species can also produce aerosols, which have a number of different effects on climate.

3.2.1 Effects of changing levels of carbon dioxide (CO_2)

Changes in carbon dioxide concentrations can influence the climate system through its effects on plant physiology (Betts et al., 2007). Plant leaves contain small openings called stomata, through which CO_2 enters and is absorbed via photosynthesis, and water vapour is emitted into the surrounding air. The size of the stomatal openings generally decreases as carbon dioxide concentrations increase, which reduces evapotranspiration leaving more water at the land surface and increasing runoff.

Cowling et al. (2008) used a global climate model to examine the impact of climate change and increasing levels of carbon dioxide on water recycling in the Amazon region. They performed two experiments which were identical except that in one experiment a dynamic vegetation model was used, so the vegetation cover could respond to the changing climate, whereas in the other experiment the vegetation cover was held constant. Rainfall over the Amazon region reduced in both experiments, owing to the closure of stomata in response to increasing carbon dioxide levels and greater water use efficiency by the vegetation. Overall, rainfall was reduced by an additional 30% in the simulation with dynamic vegetation relative to the simulation with fixed vegetation, owing to dieback of the forest caused by a drying of the climate by the 2090s.

3.2.2 Hydrocarbon production and emission

Trees produce and emit a wide range of hydrocarbons, of which the most important are isoprene and terpenes. Isoprene (C_5H_8; 2-methyl-1,3-butadiene), has an annual emission flux of about 400 – 600 TgC yr^{-1} (Arneth et al., 2008), which is similar to the annual biogenic flux of methane. Most isoprene is emitted by trees in tropical forests, with intermediate amounts from temperate forests and low amounts from boreal forests. Trees also emit another class of hydrocarbons called terpenes, which have the generic formula $C_{10}H_{16}$. Terpene emissions are about 140 TgC yr^{-1} (Guenther et al., 1995). Larger hydrocarbons are also produced and emitted, but in much smaller quantities. These hydrocarbons are highly reactive and form a wide range of products, and affect both local and global climate indirectly.

The role of isoprene on global climate has been reviewed by Pacifico et al. (2009). Isoprene is readily oxidised by the hydroxyl radical (OH) and to a lesser extent by ozone, which is itself a greenhouse gas. Oxidation by the hydroxyl radical is the main removal process of methane, an important greenhouse gas, and so isoprene plays a role in controlling levels of ozone and methane. The exact mechanisms and products from the oxidation of isoprene are complex and not yet fully known (Pacifico et al., 2009); the same statement applies to terpene oxidation. Modelling studies indicate that, in the current state of the troposphere, isoprene emissions cause a net increase of tropospheric ozone levels, by 8-12 ppbv over mid-latitude land areas and about 4 ppbv over the oceans (Wang and Shallcross, 2000).

Products from the oxidation of isoprene and terpenes can form particles called secondary organic aerosols (SOA). These aerosols form the majority of the total aerosol mass above tropical forests (O'Donnell et al., 2011) and encourage cloud formation (Pöschl, 2010). SOA are also found in marine environments and throughout the upper troposphere. SOA, like many other aerosol species, affect climate via two main mechanisms. They can cool climate directly by scattering incoming solar radiation, which effectively increases the planetary albedo (Quaas et al., 2004); this is known as the "direct effect". SOA (and most other types of aerosol) have an additional effect on climate by modifying the properties of clouds. Water vapour can condense onto SOA, which reduces the water droplet sizes within the cloud and increases their numbers, and raises the cloud albedo (Quaas et al., 2004); this process is known as the "indirect effect".

O'Donnell et al. (2011) have implemented a simplified chemistry scheme and an aerosol size-partitioning scheme into the ECHAM5 global climate model, which was used to assess the direct and indirect effects of SOA on climate. After validation of their model, these authors estimated that the direct effect of SOA, by reflecting and scattering incoming solar radiation, was to cool the climate. The indirect effects of SOA on climate, by modification of cloud properties, were more complex. In clean areas well away from sources of pollution, SOA were found to boost the growth of small particles and increase the numbers of cloud condensation nuclei. Overall, these effects act to cool climate. However, when SOA interacted with polluted clouds, they encouraged the formation of large cloud droplets and coagulation of other aerosols, reducing both cloud lifetimes and aerosol concentrations, and so warming the climate. Globally, the overall indirect effect of SOA in the study by O'Donnell et al. (2011) was to warm the climate (a positive forcing). Further studies with different climate and aerosol models are required to test the robustness of the indirect effects of SOA found by O'Donnell et al. (2011).

Mahowald (2011) has highlighted an additional impact of *anthropogenic* aerosols on climate via modification of biogeochemical cycles and uptake of carbon dioxide by vegetation and soils. These aerosols contain sulphur (S) and nitrogen (N) which are nutrients that can promote the growth of vegetation. The effect of anthropogenic aerosols is estimated to be negative, and so would act to remove CO_2 from the atmosphere and cool the climate. It is possible that SOA originating from hydrocarbon emissions by forests could have a similar effect under certain circumstances, although the overall impact on climate is likely to be smaller. Some species, such as nitric acid vapour (HNO_3) and sulphur dioxide (SO_2) may condense or dissolve into SOA, and so be transported away from their source regions and deposited elsewhere.

3.2.3 Forest Fires

Forest fires affect local and global climate in a number of ways. Large amounts of smoke and other particles are produced which can modify both the local and global climate and potentially be transported long distances away from the fires. These smoke particles have a direct impact on local and global climate by reflecting, scattering and absorbing incoming solar energy. Smoke particles may also affect cloud properties and precipitation. The burned area of forest will take time to recover, resulting in reduced uptake of CO_2 by vegetation in the burned area. If the burned area is large, there could be other impacts on local climate resulting from reduced evapotranspiration and albedo changes. Some fires occur naturally as a result of lightning strikes. Heat waves greatly increase the risk of forest fires, as was the case in Russia in July 2010 when forest fires occurred after a month long heat wave where temperatures reached 39°C (NOAA National Climatic Data Center, 2011). Deforestation itself also increases the risk of fire in the remaining forest (Laurance, 2004; Golding and Betts, 2008; see also section 3.1.1). The edges of remaining forest tend to dry out and so will burn more readily.

Burning of forests in Indonesia during 1997/1998 coincided with a severe drought caused by an El Niño. These fires spread out of control and released large amounts of smoke particles which covered most parts of the region. Davison et al. (2004) have estimated the radiative forcing effects of these emissions. Globally, the impact was limited, but regionally the effect was substantial, causing a net cooling. Indirect effects of the smoke (acting as cloud condensation nuclei) were not included. A later study by Langmann (2007) used a global chemistry-climate model with and without aerosol-cloud interactions to study the effect of the Indonesian fires on rainfall. In approximately 67% of events, rainfall was suppressed owing to the smoke particles causing an increase in cloud droplet numbers. In the remaining 33% of events, rainfall was increased. Other possible impacts of the fires, such as heat release which could trigger convective events, were not simulated.

Fires also release carbon stored in trees to the atmosphere as CO_2, which is a potent greenhouse gas. Other gaseous species such as carbon monoxide (CO) are also produced. CO has an indirect effect on climate via its reaction with the hydroxyl radical (OH). Extra CO emitted into the atmosphere can reduce OH levels and increase the lifetime of methane. Duncan et al. (2003) used a global chemistry-transport model to estimate the impacts of the trace gas and aerosol emissions from the 1997/1998 Indonesian fires on tropospheric chemistry. CO levels were elevated and hydroxyl levels were decreased. However, the climate forcings associated with ozone, CO_2, and methane from the fires were small when compared with the forcings of the aerosols.

Golding and Betts (2008) examined the change in fire risk in the Amazon during the twenty-first century using results from an ensemble of simulations based on the HadCM3 climate model together with a deforestation scenario, although the land cover in the model simulations was not changed. The fire risk was calculated using the McArthur Forest fire danger index, which is derived from temperature, humidity, wind speed and other data. The fire risk was projected to increase throughout the 21st century, and by the 2080s large parts of the Amazon were rated as having a moderate, high or very high risk of fire. Deforestation can result in fragmented forest cover, which increases the risk of fire in nearby areas of remaining forest. In areas classed as having a high risk of fire, the risk of leakage from these areas (i.e., fires spreading out of control) is also large. In some areas the forest could recover, but would still be at risk from repeated fires. It should be noted that the HadCM3 model projects significant drying over the Amazon over the course of the twenty-first century, which is the main driver of the increased fire risk.

3.3 Climate Change

Climate change is projected to impact on all types of forest. Many different modelling studies have been used to project and understand these impacts and their possible feedbacks onto climate. However, there are still large differences between models in projections of changes in key meteorological variables such as rainfall in many regions (Christensen et al., 2007), although the spread between models is partly caused by differing emissions of black carbon aerosols (Pendergrass and Hartmann, 2012).

In all forest types, a warming climate could promote the frequency and severity of pest outbreaks, which will damage the health of the forest and may reduce the numbers of certain species, allowing others to increase in numbers. Forest fires may become more frequent and larger in extent, particularly if rising temperatures are accompanied by an increase in droughts. The species composition of the forests in boreal and temperate regions could be changed by warmer temperatures, where cold-adapted species could be replaced by those trees which occur in a warmer climate, or drought-resistant species could increase in number.

Many modelling studies have used dynamic global vegetation models (DGVMs) coupled to a climate model. DGVMs simulate physiological processes in plants (e.g., photo-synthesis and evapotranspiration) in a range of vegetation types (such as grasses, broadleaf and needleleaf trees). DGVMs can also simulate competition between different vegetation types, and so may be used to project how forested areas may change in response to a warming climate. Reu et al. (2011) have examined plant physiological processes and how they change in response to climate to understand the processes behind simulated shifts in vegetation types. These authors found that species richness decreases in the tropics but increases in mid-latitudes in response to a warming climate.

3.3.1 Boreal Forests

Observations and climate models indicate that the largest temperature increases have and are projected to occur over high northern latitudes, especially during winter and spring (Meehl et al., 2007), and so the earliest signs of climate change impacts might be expected to be seen in boreal forests. MacDonald et al. (2008) have reviewed a number of studies of tree fossils and dendrochronology of dead trees in northern Europe and Asia. These data indicate that the boreal forest treeline has moved in response to recent climate change. The trees reached a maximum northward extent during the Medieval Warm Period (approximately 600 AD to 1300 AD) followed by a retreat during the Little Ice Age (approximately 1500 AD to 1800 AD) and a second northward movement after the late nineteenth century (Ruckstuhl et al., 2008). The boreal forest treeline has not yet recovered from the retreat during the Little Ice Age, and so this natural northward movement of the forest would result in warmer temperatures in the high Arctic. There may also be effects on stream flow and fresh water input to the Arctic Ocean, which in turn may impact on the formation of sea ice during the winter months.

Soja et al. (2007) compared observed changes in boreal forests during the twentieth century with previous predictions. They noted that there has been an upward movement of the lower and upper treelines in the southern mountains across Russia. There has also been an increase in the areas burned by fires in Russia, Canada and Alaska together with an increase in the number of severe fires. There is also evidence for increased pest outbreaks in Alaska. All of these changes are consistent with those expected under a warming climate.

The northward movement and expansion of forests will increase local warming owing to the albedo effect already described. However, the rate of CO_2 uptake by boreal forests will also increase, causing a cooling effect on global temperatures. Soja et al. (2007) also note that if the climate becomes warmer and drier the forested areas may die back, but could expand if the climate becomes warmer and wetter. Some evidence for this effect exists in parts of Alaska. The exact effect of climate change on treeline expansion and tree growth is complex and is modified by water availability. Changes in permafrost melting can modify soil moisture availability and increase drought stress during the summer months (Soja et al., 2007), which would limit or reduce tree growth.

Tchebakova et al. (2011) noted that winter temperatures have increased by 1-3°C in Siberia between 1991 and 2010. Rainfall has increased in some areas, but reduced in southern regions, promoting drying in regions which are already dry. Using global climate model data for 2020 and vegetation models, boreal forests were projected to move northward, but areas of steppe are projected to increase, especially south of 56°N (Tchebakova et al., 2011).

3.3.2 Temperate Forests

Lindner et al. (2007) have reviewed the likely impacts of climate change on European forests. Many of these effects will be applicable to temperate forests in other regions of the Earth. Rising CO_2 levels could promote growth of the trees by acting as a fertilizer, as CO_2 is a key substrate used in photosynthesis. However, the exact effects vary between species, and will also depend on the availability of soil moisture and nutrients. Enhanced CO_2 levels could promote the growth of deeper roots, allowing trees to cope with drier conditions as they will be able to access deeper water reserves. The stomata (small pores on the surface of leaves through which CO_2 enters the plant and water vapour is transpired) need not open as widely to allow sufficient CO_2 to enter, further increasing the trees' water efficiency. However, evaporation of water is a key cooling mechanism used by trees, and the stomata may need to remain open to prevent overheating.

Temperature increases have different impacts in the north of Europe when compared to the south. Northern Europe currently has cold winters, and so an increase in temperature would lengthen the growing season, provided sufficient rainfall still occurred. Otherwise, water availability would limit growth of the trees. Rising temperatures in southern Europe are projected to be accompanied by a decrease in rainfall (Christensen et al., 2007), which will reduce growth and increase heat stress in trees. There are likely to be changes in cloudiness which could also affect the growth of forests, although the exact effects are unknown.

3.3.3 Tropical Forests

The impact of climate change on the Amazonian rain forest has been the subject of many studies. Changes in large scale circulation could reduce rainfall over the Amazon and cause dieback of the forest. However, agreement between climate model projections of rainfall over the Amazon region is poor (Christensen et al., 2007). Dieback of the tropical forests would release large amounts of carbon (as CO_2) into the atmosphere, amplifying the warming (Betts, 2006). The evaporation from the surface would also be smaller, resulting in reduced cloud amounts and rainfall and increased surface warming.

Salazar et al. (2007) studied the impacts of climate change on the Amazonian forest using a single DGVM driven by meteorological data from fifteen different climate models under two scenarios (with low and high emissions). Despite the large range of projected precipitation changes over the Amazon from the models (which included both increases and decreases in rainfall), forest dieback was projected in all cases. Salazar et al. (2007) highlight the fact that elevated temperatures are sufficient to cause the loss of forest and its conversion to savannah, even if rainfall increases. However, only one DGVM was used in this study, and other DGVMs could give different responses.

Huntingford et al. (2008) used a perturbed physics ensemble of models based on the HadCM3 global climate model to investigate the robustness of the projection of future Amazonian forest dieback as a consequence of climate change. The projected loss of the rain forest was robust across the model ensemble, which covered a wide range of global climate sensitivity. A parallel study cited by Huntingford et al. (2008) used five different dynamic global vegetation models (DGVMs) driven by climate data generated under four different future emission scenarios. Dieback of the Amazonian forest was projected in all experiments, although the amount of dieback did vary between the vegetation models.

Many climate modelling studies of the impacts of climate change on forests have focused on the modelled changes in forest cover during the simulation. However, a recent study by Jones et al. (2009) used a climate model coupled to an interactive vegetation model to highlight inertia in the global terrestrial biosphere as it responds to a changing climate. The biosphere can continue to change for decades after climate stabilization, and that ecosystems can be committed to long-term change well before any climate change impacts are observed. In the study of Jones et al. (2009), the risk of significant loss of forest in Amazonia rises rapidly if global temperatures rise by more than 2°C above preindustrial levels.

3.3.4 Summary

The availability of moisture, both as rainfall and soil moisture from melting permafrost and ice is the key factor controlling the spread of boreal forests. The studies above suggest that climate change will result in a northward movement of boreal forests, but their southern boundary could also move northward owing to a drying of the climate and an increase in areas of steppe.

Climate change could result in increased tree growth in northern Europe, owing to an increased growing season and higher CO_2 levels, but in southern Europe water and heat stress could increase which will reduce tree growth. Similar arguments apply to other temperate forests.

Most modelling studies suggest that some dieback of the Amazonian rain forest would occur during the twenty-first century as a result of climate change. The extent of the dieback is dependent on both the projected changes in temperature and rainfall from climate models, and the DGVM used. The reduction in the forest could ultimately be more severe than most projections owing to inertia in the response of the forest to climate change.

4. Regional Impacts of Deforestation and Afforestation

In this section, the potential impacts of deforestation and afforestation in boreal, tropical and temperate regions are discussed. Some of the results have been summarised earlier, but are repeated here for clarity, and are grouped by forest type. There are a large number of studies of tropical deforestation, but far fewer examining the impacts of boreal and temperate afforestation and deforestation.

The exact effects of deforestation are still uncertain in some areas of the Earth. Pitman et al. (2009) used seven different models to simulate the impacts of land cover change (which include deforestation) between 1870 and the present day. Five models simulated a near-surface cooling during summer over regions where the land cover had changed significantly (mostly central and eastern USA, and west central Asia adjacent to Europe), but another model simulated a warming. There were few significant changes in rainfall, but four models did simulate reductions in rainfall over deforested areas. No teleconnections, where deforestation induces a change in climate in a remote region were found.

4.1 Boreal Deforestation and Afforestation

Although the recent rate of deforestation in Europe has been less than in the tropics, local deforestation has been documented in Eastern Europe, as well as degradation of boreal forests by acid rain (Douville and Royer, 1997). The potential cooling effects of deforestation via the increase in albedo are pronounced in the boreal zone, owing to the high contrast in albedo between forests and snow surfaces, tundra and bare ground, or substitution of these surfaces and possibly the taiga by broadleaf tree species such as aspen, alder, and birch (Euskirchen et al. 2007). Logging in the boreal forest reduces the quantity of carbon remaining in the forest, with the magnitude depending on the rate of decomposition, regrowth and thawing of permafrost, lakes or wetlands.

Early experiments on boreal deforestation were conducted with energy balance models. Otterman et al. (1984) suggested that the surface air temperature of a Northern Hemisphere devoid of trees would be 1.9°C lower than that of a hemisphere fully forested south of the taiga/tundra boundary. Thomas and Rowntree (1992) and Chalita and Le Treut (1994), both using a GCM, showed that the removal of boreal forests reduces the springtime surface temperature up to 2.8°C and delays the timing of snowmelt by a month.

By replacing all forest vegetation north of 45°N with bare ground or with tundra, Bonan et al. (1995) simulated boreal deforestation by a change in the surface albedo. Their results confirmed that the forest increases surface air temperature and atmospheric moisture at all times of the year, also suggesting that the climatic change induced by deforestation could be sufficient to prevent a regeneration of the forest.

Bathiany et al. (2010) modelled tropical and extratropical deforestation and afforestation using an AOGCM coupled to a land surface scheme and an ocean biochemistry model, so allowing sources and sinks of CO_2 and the CO_2 levels in the atmosphere to respond to land-surface and climatic changes. In this study, Bathiany et al. (2010) simulated deforestation and afforestation of land areas north of 45°N, and so did not distinguish between temperate and boreal regions. Extratropical deforestation produced a global cooling and warming of similar magnitudes (0.25°C). Locally, the warming and cooling ranged between 1 and 3°C over the boreal forests; the temperature changes over the temperate forests was much smaller.The increased carbon sink in the boreal afforestation experiment was cancelled out by higher carbon emissions from tropical soils and the oceans owing to the increased global temperatures. In the results of Bathiany et al. (2010), biogeophysical responses dominate the global temperature changes in the boreal experiments.

The biogeochemical effect of a release of the large amount of soil carbon accumulated in boreal regions may be less important than other biophysical effects of forests (Bala et al. 2007). Swann et al. (2010) have examined the impacts of broadleaf forests (which have higher albedos than evergreen taiga) moving northward and replacing the tundra and other non-forested areas. Broadleaved forests have a higher transpiration rate than the taiga. The results of Swann et al. (2010) showed that the increased water vapour transpired into the lower atmosphere resulted in an additional warming.

Consequently, the change of vegetation from tundra to taiga under future climate as well as reforestation to offset carbon losses amplifies global warming through reduced albedo at spatial scales of hundreds or even thousands of kilometres (Betts, 2000), especially in winter and spring (Göttel et al., 2008).

4.2 Tropical Deforestation and Afforestation

The effects of tropical deforestation on local and global climate occur as a result of two competing processes. Forests have low albedos, and so conversion of forests to grass or crops increases the albedo and would produce a net cooling if this was the only effect. However, loss of forest also results in decreases in both evapotranspiration efficiency and surface roughness, and these changes act to warm the climate. Moreover, warm temperatures in the tropics probably cause a more rapid loss of soil carbon after logging than at high latitudes, strengthening the warming impacts of tropical deforestation (Bala et al., 2007). Conversely, reforestation is progressively more effective in reducing the potential for climate warming as the climate moves from a colder one to a warmer and wetter one. Avoided deforestation and forest establishment in the tropics cools the climate through evapotranspiration, cloud feedbacks, and sequestration of carbon from the atmosphere and so slowing the buildup of CO_2 in the atmosphere.

There have been a large number of studies which have used global climate models to project the impacts of tropical deforestation on climate in the deforested areas (Henderson-Sellers et al., 1993; McGuffie et al., 1995; Zhang et al., 1996a,b; Chase et al., 2000; Gedney and Valdes, 2000; Berbet and Costa, 2003; Avissar and Werth, 2004; Findell et al., 2006; Hasler et al., 2009; Bathiany et al., 2010; Costa and Pires, 2010; Davin and de Noblet-Ducoudré, 2010; Snyder, 2010). There are many other such studies referenced in these papers. Almost all of these studies used scenarios where tropical forests were replaced by grass and shrubs. This approach was used to ensure any response of the atmosphere to these changes would be detected; the climate signal produced by realistic land use changes would probably be small and hard to distinguish from modelled natural variability of climate. However, the assumption of such large-scale deforestation throughout the entire tropics may be unrealistic and produce misleading results. A wide range of models with various resolutions and complexity have been used which makes drawing a consensus difficult. Despite this, all these studies agree that rainfall over the deforested regions would be reduced and surface temperatures increased. Costa and Pires (2010) simulated the effects of partial or full deforestation in an area of Brazil. Their results indicated that not only did rainfall decrease during the wet season following deforestation, but the length of the dry season could be extended by up to 1 month.

The effects of the pattern of deforestation on local and regional rainfall are more complex. D'Almeida et al. (2007) have reviewed observations and modelling studies of the impact of deforestation on the hydrological cycle in Amazonia. Large scale deforestation results in a weakened hydrological cycle, with reduced rainfall and evaporation and an increased proportion of runoff. In contrast, smaller disturbed areas, where a "patchy" landscape exists, with areas of forest and croplands intermingled, could result in enhanced convection and increased rainfall (Pielke, 2001; Laurance, 2005). Roy (2009) used a mesoscale model to show that the herring bone pattern of deforestation in Amazonia triggers small-scale circulation patterns leading to more clouds and rain over the deforested patches. In a later study, Garcia-Carreras and Parker (2011) used a large eddy model to study the effects of heterogeneous land surfaces on rainfall over a part of West Africa. These authors found that rainfall over the cropland was enhanced but reduced over the forests. Net rainfall over the whole area studied was enhanced compared with a homogenous land surface.

Jonko et al. (2009) investigated the sensitivity of the tropical atmospheric circulation to a global land cover change scenario together with projected climate change under the SRES A2 scenario (Nakicenovic and Swart, 2000). The results are in accordance with previous studies and suggest that tropical deforestation, particularly in the Amazon basin, leads to reduced evaporation and decreased ascending motion over continental areas and therefore to a weakening of circulation patterns driven by this ascending motion.

A study of the radiative impacts of deforestation on global climate with a fully coupled AOGCM by Davin and de Noblet-Ducoudré (2010) concluded that the albedo effect was dominant. The response of the oceans was important, in that the temperature change over land was itself strongly influenced by the ocean response. Davin and de Noblet-Ducoudré (2010) showed that the magnitude of the effects of albedo versus evapotranspiration and roughness changes varies with latitude. The cooling effect due to albedo change is stronger at high latitudes and affects both land and ocean. Conversely, the warming effect from change in evapotranspiration efficiency and surface roughness is stronger at low latitudes and does not affect the oceans. However, the study of Davin and de Noblet-Ducoudré (2010) did not consider the release of CO_2 after deforestation and subsequent impacts on climate.

Bathiany et al. (2010) have performed a similar series of experiments to that of Davin and de Noblet-Ducoudré (2010), but they used an AOGCM coupled to a land surface scheme and an ocean biochemistry model, so allowing sources and sinks of CO_2 and the CO_2 levels in the atmosphere to respond to land-surface and climatic changes. In line with other studies, tropical deforestation resulted in increased surface temperatures and reduced rainfall in the deforested areas. Global temperatures increased by 0.4°C after tropical deforestation owing to an increase in CO_2 levels by 60 ppm and reduced evapotranspiration in the deforested areas.

A number of modelling studies have highlighted the potentially important role of the oceans in modifying the impacts of tropical deforestation on regional climate. Tropical forests in south-east Asia are located on relatively small land areas and islands in the western Pacific, and so feedbacks of deforestation on ocean temperatures and circulation are potentially important in modifying the local and regional responses to deforestation. Delire et al. (2001) and Voldoire and Royer (2005) used a dynamic ocean model coupled to a global climate model to study the local and regional impacts of deforestation in south-east Asia. In these two studies, rainfall and evapotranspiration decreased and the sensible heat flux increased over the deforested areas. Overall, the deforested areas became warmer and drier. Easterly winds were intensified leading to a cooling of the ocean surface, reduced convection and rainfall, and enhanced upwelling of deep water. These changes to the ocean had the effect of reducing rainfall in the wider south-east Asian region.

A later study by Schneck and Mosbrugger (2011) also used a fully coupled ocean and atmopshere model to study local and remote effects of deforestation in south-east Asia. In the deforested areas, surface temperatures increased but rainfall decreased, in agreement with the previous studies by Delire et al. (2001) and Voldoire and Royer

(2005). The ocean circulation was once again simulated to play an important role in modifying some of these effects. However, in the study by Schneck and Mosbrugger (2011) the upwelling of cool water was reduced, leading to higher sea surface temperatures and increased evaporation from the ocean which partly compensated for the reduced evapotranspiration over deforested areas. Rainfall in the wider Asian region was enhanced after deforestation, owing to reduced easterly winds and increased convection over the oceans.

The study by Schneck and Mosbrugger (2011) has identified the same mechanisms as the two earlier studies, but the response of the ocean following deforestation has occurred in the opposite direction. These three studies show a strong influence of the ocean circulation on the projected climate effects of deforestation in south-east Asia, but do not agree on the sign of the changes and subsequent impacts on rainfall in the wider Asian area.

The modelling studies of tropical deforestation which include an interactive ocean model (Delire et al., 2001; Voldoire and Royer, 2005; Bathiany et al., 2010; Davin and de Noblet-Ducoudré, 2010; Schneck and Mosbrugger, 2011) have shown that potential changes to the ocean surface temperatures and circulation in response to climate forcing by deforestation are important. These changes act to modify the land temperature and rainfall responses, but are often model-dependent.

Teleconnections, where the change in land surface characteristics in a deforested area in the tropics has an impact on the climate of mid- and high latitude northern hemisphere land areas have been simulated by some models (Gedney and Valdes, 2000; Avissar and Werth, 2004; Hasler et al., 2009; Snyder, 2010; Schneck and Mosbrugger, 2011). However, the strength of these teleconnections and the areas affected are model-dependent. Other model results suggest that the impacts of tropical deforestation are largely confined to the tropics and that no teleconnections exist (Findell et al., 2006; Findell et al., 2007).

Zheng et al. (2009) have reviewed several modelling studies of the impact of land cover changes (replacement of forests by cropland) on climate over China since the 1700s. All model results indicated that these land use changes have caused an enhancement of the East Asian winter monsoon as well as cooling in winter and warming in summer.

However, the impacts on annual mean temperatures and precipitation varied between the models.

There are a large number of modelling studies which have examined the potential impacts of tropical deforestation on local, regional and global climate. Many different models with varying resolutions and complexity have been used which makes drawing a consensus difficult. However, all these studies agree that temperatures would rise and rainfall would decrease in deforested areas. If the deforestation results in a patchy landscape, with croplands and forest fragments adjacent to each other, rainfall over the croplands could be enhanced by increased convection, but rainfall over the forest fragments would be reduced. Interactions with the ocean surface temperatures and circulation appear to be important in south-east Asia where the tropical forests occur on islands and small land areas in the west Pacific. Some studies have reported teleconnections, where, for example, deforestation in the Amazon impacts on the climate of Europe. However, these teleconnections are model-dependent. Other studies suggest that no teleconnections occur following deforestation, so no definitive statement can be made. Overall, more studies of tropical deforestation with full earth system models which can simulate all the potential feedbacks between local and global climate and the oceans are required.

4.3 Temperate Deforestation and Afforestation

The forest cover, particularly over Europe, has been reduced significantly over the past centuries as a consequence of human activities. Forests have been cleared and the land has been used for arable crops and pasture for grazing by animals. This land use change is likely to have considerably influenced European climate (Reale and Shukla 2000). However, the overall climate forcing of temperate forests is uncertain (Bonan, 2008). Annual mean temperatures over temperate forests are controlled by the effect of the low albedo during winter (which acts to warm the climate) and evapotranspiration during summer (which cools the climate). Carbon emissions from deforestation could be approximately balanced by the higher albedo of the crops and grass which would replace the forests (Betts et al., 2008), so that the net climatic effect of temperate deforestation would be negligible. Alternatively, reduced evapotranspiration as a consequence of the loss of the trees could amplify the warming (Bonan, 2008), although reduced canopy cover can increase soil evaporation (Pitman et al., 2009).

This uncertainty in the climatic impacts of deforestation and afforestation in Europe is illustrated by the following modelling studies. Gates and Leiß (2001) simulated a slight cooling following deforestation owing to the increased surface albedo, together with reduced summer rainfall as a result of lower evapotranspiration. Heck et al. (2001) studied the climatic effects of afforestation for the period May to August only. The increased tree cover led to a cooler and moister spring climate but warmer and drier summers as a result of changes in soil moisture and evapotranspiration when compared with present-day forest cover. Rainfall followed a similar pattern with increased rainfall between May and mid-July followed by a decrease. The warmer temperatures and reduced rainfall after mid-July were caused by soil moisture values falling below critical values.

Sanchez et al. (2007) simulated the potential impacts of tree cover change on European climate. They found that rainfall over most of Europe was larger when tree cover was increased, but, contrary to the results from other studies, surface temperatures were between 1 and 3 K cooler. Anav et al. (2010) examined the impacts of both afforestation and deforestation on European climate, and found no significant change in mean surface temperature or rainfall in either scenario when compared with the climate simulated using present-day vegetation. However, they did simulate an increased number of hot days in summer in the afforestation scenario, and a decrease in the deforestation scenario. No significant changes were simulated during winter in either scenario.

Wattenbach et al. (2007) examined the impacts of afforestation in a part of former East Germany caused by the abandonment of agricultural land. Overall, afforestation had large impacts on the local hydrology. The biggest impacts were seen in ground water recharge and runoff, which were both reduced after afforestation. This lack of recharge meant summer droughts would be worsened. Additionally, the conversion of the forest from Scots Pine (an evergreen) to native deciduous and mixed forest reduced the overall loss of water via evapotranspiration. This study shows that the particular species used for afforestation is important. Other research (Ellison et al., 2011) suggests that, overall, afforestation would enhance precipitation.

Temperate forests also partly mitigate the effects of droughts, as they can access deep reservoirs of water with their roots. Crops, however, have shallow roots, and temperatures in croplands during droughts can be much higher than in forests. For example, Zaitchik et al. (2006) report that, during the 2003 heat wave, temperatures over temperate forests were 9°C cooler than croplands in France.

Vautard et al. (2010), as reported in section 3.1.4, have found a downward trend in wind speeds in many European and Asian areas. These areas coincide with regions where forests have regrown after land formerly cleared for agriculture has been abandoned.

Results of mesoscale model studies showed that land use change in Hungary during the twentieth century had already altered weather and climate (Drüszler et al., 2010). These changes have caused a +0.15 °C temperature rise and +0.18 °C increase in the dew point depression. The maximum warming and drying also affected the urban areas. It was also shown that the change in Hungarian land cover does not have a significant nation-wide impact on average precipitation.

Climatic effects of potential afforestation in Hungary to mitigate the effects of climate change have been investigated on a regional scale by Gálos et al. (2011). Their results suggest that afforestation would support the increase of both evapotranspiration and rainfall by 10-15%, as well as a decrease of surface temperature (by up to 1°C), even during summer, which is in contrast with results of Heck et al. (2001). For rainfall, the largest effects were simulated in the north-eastern part of Hungary, where half of the projected rainfall decrease could be offset and the number of summer droughts could be reduced.

4.4 Focus on Mediterranean Deforestation and Afforestation

Millán (2008) has proposed that deforestation in the Mediterranean has had two important consequences. First, moisture from the sea, typically in the form of fog, is no longer trapped along the coast. Secondly, the increased land temperatures due to vegetation loss result in rising air columns that carry storms up and over surrounding mountains, reducing orographic precipitation. Vegetation loss leads to lower evapotranspiration and subsequently lower precipitation. Any moisture transpired is now advected away to form precipitation at other locations, instead of falling back to the ground locally (Ellison et al., 2011). A similar impact of deforestation on rainfall over south-east Australia near the city of Perth was found by Nair et al. (2011).

A study by Gangoiti et al. (2011) examined the sources of moisture responsible for the severe flooding event in central Europe in August 2002 using a combination of a mesoscale model and a trajectory model. These authors found that preceding rainfall events had saturated the ground, and the majority of the moisture associated with the August 2002 event came from subsequent evaporation of this moisture. This mechanism of rainfall and subsequent evaporation plays an important role in the inland movement and propagation of this type of extreme rainfall event. This mechanism is only applicable to the summer months, when almost all rainfall occurs via convective storms. The deforestation impacts discussed by Millán (2008) could have impaired the propagation and frequency of this type of event.

In the Mediterranean, recovery of forests led to an increase in evapotranspiration, which caused cooler and moister conditions in the period from April until mid-July. In mid-July, soil moisture dropped below the critical value and transpiration was almost completely inhibited, which resulted in drier and warmer summers, contrarily to what happens in Middle and Northern Europe (Heck et al., 1999; Heck et al., 2001). Overall, these studies and others (e.g., Anav et al., 2010) indicate that the largest impacts of deforestation or afforestation are realised during the summer months. Little impact is seen during winter as conditions are controlled by synoptic weather systems that are advected over the land. The exact impacts on surface temperatures are controlled by the balance between evaporative cooling and albedo; afforestation results in a reduced albedo (and hence greater sensible heat transfer from the forests to the atmosphere), but additionally increased cooling via evapotranspiration (latent heat transfer) (Heck et al., 2001; Sanchez et al., 2007; Gálos et al., 2011). Deforestation in the Mediterranean could be responsible for storms penetrating further inland and increased flooding in parts of Europe (Millán, 2008; Ellison et al., 2011). Other studies support the mechanisms behind this theory, but modelling studies to test the theory itself have not yet been performed.

Recent work by Matteucci et al. (2011) focused on changes in forest resources in Mediterranean Europe. The outcome of this historical assessment can be a useful basis to understand recent dynamics of similar ecosystems:

1. Mediterranean forests cover an area of 80 Mha, most of which are located in the European side of the Mediterranean (76%), while Mediterranean Africa and Asia contribute 10% and 14% respectively. Forest area in the region has increased by 10% between 1990 and 2000 due to wood encroachment, natural invasion of

abandoned lands and deliberate reforestation and afforestation plans (sources: Forest Resources Assessment 1990, 2000 and 2005 by FAO).

2. Roundwood nowadays represents 60% and woodfuel 40% of total wood products (125 M m3). Industrial roundwood is harvested mostly in Mediterranean Europe (80-90% in the last 50 years). The role of Asia is increasing and roundwood production reached recently 16% of the Mediterranean total production, while the role of Africa is negligible (less than 2%).

3. The average total economic value of the Mediterranean area was calculated to be about 133 € ha^{-1}, with Northern countries having much higher value (173 € ha^{-1}) than Southern (70 € ha^{-1}) and Eastern countries (48 € ha^{-1}) (Croitoru 2007). Non-wood forest products and services are more important as they represent 60-70% of total economic value.

Currently, the planning situation in the Mediterranean basin is very diverse. Periodical planning, when applied, is static but challenges and constraints impose a change toward a dynamic process which allows up-dating or re-planning (Palahi et al., 2008).

5. The “biotic pump” hypothesis

It is widely recognized that the presence of forests favours the recycling of water through evapotranspiration, cloud formation and rainfall, at least in the humid tropics (Silva Dias et al., 2009). Tinker et al. (1996) recalled how the removal of forest is worse in inland than coastal regions where water vapour from the nearby ocean is also available.

A new hypothesis, called the “biotic pump”, suggests that forest cover plays a much greater role in determining rainfall than previously recognized (Makarieva et al., 2006, 2009; Makarieva and Gorshkov, 2007). Pressure gradients caused by temperature and convection drive circulation in conventional meteorology. However, Makarieva and Gorshkov (2007) have proposed that evaporation and condensation play a greater role in atmospheric circulation than has been previously thought. If conditions are unstable, water vapour can rise and condense once the temperature is sufficiently low. Makarieva and Gorshkov (2007) then argue that a reduction in air pressure occurs as a result of condensation, because liquid water has a smaller volume than water vapour.

Atmospheric pressure is thus reduced at a higher rate over areas with intensive evaporation, and then the low pressure acts to draw in additional moist air from areas with weaker evaporation. Overall, a transfer of moisture occurs towards areas with high evaporation rates.

Should this hypothesis be correct, the implications are substantial (Sheil and Murdiyarso, 2009). Conventional models typically predict a “moderate” 20 to 30% decline in rainfall after continental-scale deforestation (Bonan, 2008). In contrast, Makarieva and Gorshkov (2009) suggest that even relatively localized clearing might ultimately switch entire continental climates from wet to arid, with rainfall declining by more than 95% in the interior.

However, the “evaporative force” on which the “biotic pump” theory is based has been criticized by micrometeorologists and hydrologists as not being supported by basic physical principles (Meesters et al., 2009). Although there is a reduction in volume following condensation of water vapour, latent heat is released which warms the air and causes an increase in volume and pressure. Overall, there is a net increase in pressure following condensation of water vapour. The release of latent heat appears to have been neglected by Makarieva and Gorshkov (2007) when constructing their hypothesis (Meesters et al., 2009). The release of latent heat is a key process in convection and the buoyancy of air which is well understood.

6. Summary

Before the discussion of the ways in which forests influence weather and climate, it is worth remembering that forests play a much bigger role than influencing weather. Forests, particularly tropical ones, are an important source of biodiversity, providing habitats for many unique plant and animal species, and are home to indigenous peoples. Forests provide an environment for many recreational pursuits, and are an important source of raw materials and medicinal products. Forests also act to maintain biodiverse soils and protect these soils by stabilising them against heavy rainfall and being washed away.

This review has described the mechanisms by which forests influence climate, and which particular components are dominant for each forest type. The mechanisms by which forests interact with climate are summarised in section 7. The mechanisms include albedo, evapotranspiration, aerodynamic effects, carbon sequestration, and indirect chemical and aerosol effects. The most important effects for boreal, temperate and tropical forests are albedo, albedo and evapotranspiration, and evapotranspiration respectively. The potential effects of climate change on forests were also described, followed by a discussion of the regional effects of deforestation and afforestation on local and global climate.

Significant climate change could be caused merely by redistribution of the forested areas. This redistribution could be due to intensive logging as already observed in some tropical areas; and in the case of the boreal forest, it could also be induced by the increase in the atmospheric concentration in greenhouse gases. Furthermore the variability of the climatic, soil and vegetation characteristics of a region, as well as the representation of land surface processes in the applied climate model, also have an influence on the simulated vegetation-atmosphere interactions.

Over the Sahel (Xue and Shukla, 1996) modelling exercises indicate that current rainfall increases in an afforestation/reforestation scenario compared to current land cover, owing to changes in vegetation structure, moisture patterns and evapotranspiration. In south-western Australia, land-cover changes from trees to grasslands/croplands in the last 250 years could partially explain the decreases in winter precipitation. Extensive reforestation may cause, in the long term, an increase in rainfall. On the other hand, an exercise coupling an afforestation scenario and a climate model over the U.S. found that changes in summer precipitation were marginal and depended on site location.

7. Main drivers of and processes involved in forest – weather relationships

1. **Albedo**

Forests have low albedos compared with most soils and other types of vegetation. Consequently, they absorb the majority of incoming solar energy which is released as sensible heat and warms the surface. However, forests also emit water vapour via evapotranspiration which acts to cool the surface. Overall, boreal and temperate forests warm their environment via this process. For tropical forests, cooling by evapotranspiration is the dominant effect (see 3 below).

2. **Evapotranspiration**

Forests transpire moisture from the ground to the surrounding air via small openings in their leaves called stomata, and moisten the boundary layer. This process cools the surface via the transfer of latent heat, and is especially important in the tropics. Evapotranspiration also enhances the formation of clouds, which cool the climate by reflecting incoming solar energy. Rainfall above the forests may also be enhanced, especially over tropical forests, where a large proportion of moisture is constantly recycled between the ground and the atmosphere via the forests.

3. **Aerodynamic Effects**

Forests are aerodynamically rough, which means that they induce turbulence in the air above which causes drag and slows the wind speeds. This turbulence can also enhance the transfer of heat and moisture between the forest and the air, and increase convection and rainfall. The swaying of trees also reduces wind speeds.

4. **Interactions with carbon dioxide (CO_2)**

Forests absorb CO_2 during photosynthesis, and so help to reduce its levels in the atmosphere. Rising CO_2 levels may cause a reduction in the size of the stomata which would reduce the amount of water transpired and subsequently affect cloud formation and rainfall (see (2) above).

5. **Chemical and Aerosol Effects**

Forests emit reactive hydrocarbons whose oxidation can form aerosol particles. These aerosols can reflect and scatter incoming solar energy, and also enhance cloud formation by acting as condensation nuclei for water vapour. Over tropical forests, these aerosols make up the majority of the total aerosol mass; they may also make a significant contribution over the boreal forests. Convective clouds over the Amazon have characteristics more like those found in marine environments, which is thought to be due to the aerosols. The chemical reactions of these hydrocarbons partly control levels of methane and ozone in the troposphere, which are important greenhouse gases.

8. Barriers and Knowledge Gaps

There have been a large number of studies of the local, regional and global impacts of forests on climate. Many of the early studies used atmosphere-only climate models with (by current standards) low resolutions, whereas later studies have used more complex models, and a few earth system models have recently been employed (earth system models include couplings between the atmosphere, oceans, vegetation and ocean carbon cycles, and simulate any impacts on CO_2 levels and subsequent feedbacks). For tropical regions, many of these studies agree on the impacts of deforestation on the deforested area, although changes to temperature and rainfall in other regions are model dependent. Some modelling studies have suggested that teleconnections exist, where deforestation in the Amazon would also impact on the climate over Europe, for example (e.g., Gedney and Valdes, 2000). However, other modelling studies suggest that teleconnections do not occur following deforestation (Findell et al., 2006; Findell et al., 2007). Clarifying this issue is important since significant teleconnections would imply both a regional and a global scale response to deforestation. Similarly, changes in ocean circulation and surface temperatures following deforestation seem to be important in south-east Asia but the exact effects are also model-dependent.

Studies of deforestation and afforestation in temperate regions do not agree on the impacts. Some studies reported little or no change in surface temperature and rainfall following deforestation, whereas others simulated either a warming or a cooling (Pitman et al., 2009). Some of these differences will be caused by the different models used and the representations of plant physiological processes. Simulations of soil moisture changes are also likely to be important. Models which simulate a drying of the soils during summer project increases in surface temperatures (e.g., Heck et al., 2001) whereas other models may simulate wetter soils which allow the forests to continue to transpire during the summer months and maintain cooler surface temperatures.

More recent studies include dynamic vegetation models which simulate the physiological processes in plants, and competition between different types, and so can be used to project changes to forested areas as a consequence of climate change or other external forcings. There are important differences in the way these models represent trees and other types of vegetation, and the number of different vegetation types included. Further studies using a number of different earth system models coupled to vegetation models are required to test the robustness of these studies, but earth system models are computationally demanding and costly to run. They also generate vast amounts of data which need to be stored and analysed. Many of the climate model simulations being executed for the forthcoming IPCC Fifth Assessment Report (due to be published in 2013) include land-use change scenarios, and analysis of these results should yield increased understanding of the impacts of forests on weather and climate.

Many of the studies of the impacts of forests on weather and climate have only examined changes in long-term average or seasonal average climate over a region. Very few have considered the impacts on extreme temperatures and rainfall, variability of climate over a region, and interannual variability of climate. For example, the magnitude of heat waves could be amplified by land surface preconditioning (Vautard et al., 2007; Ferranti and Viterbo, 2006). In general, the effects of land cover changes on extreme events have been poorly investigated, an exception being the study by Anav et al. (2010).

The different mechanisms by which forests interact with the atmosphere and modify weather and climate are now fairly well understood. However, feedbacks exist between these processes and the atmosphere which are not well understood. Many of these feedbacks occur at small spatial scales which cannot be resolved by global climate models, and can only be simulated by higher resolution models for limited areas.

The importance of the ecosystem feedbacks like those on the hydrological cycle suggests that there is space for management options that could potentially reduce the likelihood of extended droughts. Several factors influencing water use by tree plantations can be controlled by management and there is scope to design and manage forest plantations for increased water use efficiency. Plantation design (edges, firebreaks, streamlines, use of mixed species) has the potential to modify atmospheric coupling of forest plantations with impact on their water use. Furthermore, proper management of natural and planted forests has a relevant impact on the water available downstream for agricultural and civil uses (Vanclay et al, 2009).

Bruijzeel (1990) observed how the type and level of management influences how the soil damage caused by deforestation can favour, via lower infiltration rates, runoff and floods. Vanclay (2009) through his review stimulated the possibility to manage plantations to reduce water loss, highlighting several research needs:

1. Where does rain come from: what is the relative importance of moisture transpired from vegetation and transported from other regions?
2. How does vegetation and its management affect the distribution of rainfall?
3. How is transpiration affected by air turbulence near plantation edges, firebreaks and streamlines?
4. Can atmospheric coupling of forest plantations be modified through plantation design, for example by using mixed species plantations to alter roughness?
5. Can existing plantations be made more water-efficient by softening hard edges by thinning and pruning edges, by planting hedges along external plantation edges to control turbulence development, and by re-designing the shape of planting blocks during harvesting?

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Task A2_D1: Literature Review of datasets describing the Historical Evolution of Forest cover and Weather across the EU region

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Contents

Prefatory Note

A literature review summarising data sets which describe the historical evolution of forest cover and weather patterns across the European Union over the last 150 years is presented in this report, as part of Task A2_D1. The report is divided into two main sections, with the first focusing on forests, and the second on weather patterns and observed changes in climate. In each section the main findings and knowledge gaps are described, and extensive list of supporting references is also provided. For clarity, there is also a detailed summary of the main findings for each section. For completeness, Annex 1 and 2 provide more details on the available climate data sets.

A second report has been written which focuses on the actual changes in weather patterns across Europe (Task A2_D1), presents criteria to be used to assess the influence of EU forests on various weather-related effects (Task A2_D2) and then uses these criteria to assess the influence of forests on EU weather (Task A2_D3).

Acronyms and Abbreviations

AVHRR – Advanced Very High Resolution Radiometry
CORINE – Coordination of Information on the Environment
DISCover – Data and Information System global land Cover
DGVM – dynamic global vegetation model
EC – European Commission
ECE – Economic Commission for Europe
EEA – European Environment Agency
EFI – European Forests Institute
EFT – European Forest Types
ESA – European Space Agency
EU – European Union
FAO – Food and Agriculture Organisation of the United Nations
FCCW4 – Forests and Climate Change Working Paper 4 (2006)
FEW - Forestry in the EU and in the World
FRA – Forest Resource Assessment
FROW – Forest Resources of the World
FS – Forestry Statistics

GCM – Global Circulation Model

GLC2000 – Global Land Cover 2000

GOFC-GOLD – Global Observation of forest and land cover dynamics

HYDE - History Database of the Global Environment

IGBP – International Geosphere Biosphere Project

IRS – Indian Remote Sensing (Satellite)

IRSO – Indian Space Research Organisation

JRC – Joint Research Centre (European Commission)

MODIS – Moderate resolution Imaging Spectroradiometry

NOAA – National Oceanic and Atmospheric Administration

NPP – Net Primary Production

PELCOM - Pan-European Land Use and Land Cover Monitoring

PFT – Plant Functional Type

SAGE – centre for Sustainability and the Global Environment (University of Wisconsin - Madison)

SEF - State of Europe's Forests

SWF – State of the World's Forests

UMd – University of Maryland

UN – United Nations

UNEP – United Nations Environment Project

UNFCCC - United Nations Framework Convention on Climate Change

Units

ha – hectare = 10000 m^2

m – metre

km - kilometre

µm – micron = 10^{-6} m

Pg C/year – Peta grams of carbon per year = 10^{15} grams of carbon per year, or billions of tonnes per year.

Part A: Literature Review of the Historical Evolution of Forest Cover and Appropriate Datasets Spanning the past 150 years

Executive Summary

The first part of this report describes the literature and data sets that are available for assessing the historical evolution of forest cover across the EU over the last 150 years. The literature sources are separated into three categories: i) historical data; ii) satellite data; iii) model reconstructions. We provide references to the original sources of historical data, and describe the features of the satellites used to generate land cover data sets. In addition, we describe the techniques and assumptions that are used to generate the reconstructions of historical forest cover.

European forest cover has changed significantly throughout history due to land use changes caused by population expansion and the changes in exploitation of wood products. Historical national statistics from several European countries suggest that forest area was at a minimum in the 19th century and early 20th century. However, in recent times European forest cover has undergone a transition in which the forest area stabilised and then expanded to cover approximately 42% of the European Union land area, an increase of 5% in the past 20 years. This makes Europe the only world region to have experienced net forest expansion during that period. However, the changes in forest cover across the EU are highly heterogeneous with forest expansion occurring in Ireland, Bulgaria, Latvia, France, Italy and Sweden, but decreasing forest area in Denmark, Slovenia and Finland.

Measuring the precise geographical distribution of EU forests has only been possible over the last 20 years due to the advent of global and European land cover data sets generated from satellites. In this report, we describe the most useful data sets of this type: DISCover (Loveland and Belward, 1997; Loveland et al., 2000), GLC2000 (Belward, 1996; Fritz et al., 2003; Bartholomé and Belward, 2005; Hartley et al., 2006), MODIS (Friedl et al., 2002), CORINE (Heymann et al., 1994; Büttner et al., 2002), PELCOM (Mücher et al., 2001), GLOBCOVER (Bicheron et al., 2008; Bontemps et al., 2011), and LANMAP2 (Mücher et al., 2010).

Estimating geographically explicit land cover and forest cover maps prior to 1980 is much more difficult and subject to significant uncertainties. For example, between 1960 and 1985, several global maps of natural vegetation were compiled from ground-based measurements. However, comparison of these maps demonstrates inconsistencies, with the original sources very hard to obtain. More recently, the HILDA (Historic Land Dynamics Assessment) database has been constructed from a range of historical data. This dataset covers the period 1900 – 2010 and provides changes in land cover at decadal intervals on a 1 km $\times$ 1 km scale.

Prior to the HILDA database, estimating historical forest cover required the use of numerical models that hindcast land cover changes by combining ecosystem models and historical cropland inventories to estimate the total area of land in use at any point in time. The geographically explicit land cover maps are then produced by making assumptions about how the historical croplands are distributed. In particular, the SAGE database down-scaled the current cropland distribution, while the HYDE database assumed that croplands follow population density. The areas formerly occupied by crops are then replaced with vegetation types predicted by the ecosystem model to be most suited to the climate and soil type at that location. Since the papers describing the HYDE database are more clearly written, provide comparisons with other databases and are more open about uncertainties, we consider the HYDE database to provide a better realisation of the historical distribution of forest cover than SAGE.

1. Introduction

Since the late 1940s the Food and Agriculture Organisation (FAO) of the United Nations (UN) has produced a multiplicity of reports and inventories that describe forest resources worldwide in great detail. As a result of these reviews, we have estimates of the forested area, type and usage for each country involved in the studies. Indeed, owing to the proliferation of data during this time, the problem is not the availability of data, but rather the huge number of sources and their variability. In contrast, original forest data prior to the 1940s are extremely difficult to obtain, even if they are available at all. Not only this, but the data are often of undocumented accuracy. Furthermore, in order to determine the extent to which European forest cover has change since 1850, from both contemporary and historical sources, it is necessary to define the vegetation types and cover that constitutes a ‘forest’, as described below.

1.1 Definition of a forest

This report adopts the standard definition of a forest as given by the FAO (e.g. Forests and Climate Change Working Paper 4, 2006[1 2]). Thus, a forest is taken to be:

"….a minimum area of land of 0.05-1.0 hectares with tree crown cover (or equivalent stocking level) of more than 10-30 per cent with trees with the potential to reach a minimum height of 2-5 metres at maturity *in situ*. A forest may consist either of closed forest formations where trees of various storeys and undergrowth cover a high proportion of the ground or open forest. Young natural stands and all plantations which have yet to reach a crown density of 10-30 per cent or tree height of 2-5 meters are included under forest, as are areas normally forming part of the forest area which are temporarily unstocked as a result of human intervention such as harvesting or natural causes but which are expected to revert to forest; [FCCC/CP/2001/13/Add.1]."

In addition, the FCCW4 also states that:

"'Forest' is not defined by land use, but rather by the current or expected physical properties of vegetation cover (Verchot et al., 2005). Even land cover systems that are not intuitively perceived as forests may qualify, e.g. orchards, oil palms, trees planted as shelter belts in fields, along riversides or highways, as well as urban tree plantings (Dutschke, 2002)."

As a further inclusion, reports such as Forestry Statistics 2009[3] produced by EUROSTAT (2010)[4] define 'other wooded land' to be:

"…..land of more than 0.5 hectares not classified as a forest. It has a canopy cover of 5-10%, comprising trees able to reach a height of 5 m at maturity in situ. Alternatively, it

[1] http://www.fao.org/forestry/climatechange/53622/en/
[2] Henceforth, Forests and Climate Change Working Paper 4 (2006) will be referred to as FCCW4.
[3] Henceforth Forestry Statistics (2009) will be referred to as FS (2009).
[4] http://epp.eurostat.ec.europa.eu/portal/page/portal/forestry/introduction

has a canopy cover of more than 10% comprising trees that will not reach a height of 5 m at maturity in situ (e.g. dwarf or stunted trees) and shrub or bush cover. It does not include land that is predominantly under agricultural or urban use."

For reference, the current definition of a forest should be contrasted to that which was used in compiling that 1980 Global Forest Resource Assessment (FRA) - the first assessment to use a technical definition of forests in which measurable parameters where indicated. The 1980 FRA assessment employed the following minimum criteria for defining a forest: "10 % canopy cover density, tree height of 7 m and area of 10 hectares". As a result of the broadened definition introduced between the 1990 and 2000 FRAs, the area of vegetation world-wide classified by the FAO as forest increased by approximately 10% (~300 million ha).

1.2 European forest types

Owing to its wide range of climate and geography, Europe comprises a number of biogeographical regions[5], with different tree and vegetation types suited to particular regions. As shown in the European Environment Agency's report (EEA, 2006) on European Forest Types[6] the forests form three main groups: the warm climate of southern Europe favours mixed deciduous broad-leaved trees and scrubland; further north, in the cool temperate regions, deciduous broad-leaved and coniferous broad-leaved forests are prevalent, while the boreal forests of Scandinavia consist mostly of mixed needle-leaf and broad-leaved coniferous trees.

2. Overview of Literature of European forest cover

Recent assessments have shown that forests and wooded land currently occupy approximately 42% (177 million hectares) of the European Union's land area (Forestry in the EU and the World 2011 (EUROSTAT, 2011)[7]; State of Europe's Forests (UN-ECE/FAO, 2011b)[8]; Forestry Statistics 2009 (EUROSTAT, 2010)). These forests are distributed heterogeneously, with northern Europe being the most heavily forested region, while the climate of southern Europe favours more scattered vegetation (SEF

[5] http://www.eea.europa.eu/data-and-maps/figures/biogeographical-regions-in-europe
[6] Henceforth European Forest Types (EEA, 2006) will be referred to as EFT (2006).
[7] Forestry in the EU and the World 2011 (EUROSTAT, 2011) will be referred to as FEW (2011).
[8] State of Europe's Forests 2011 (UN-ECE/FAO, 2011b) will be referred to as SEF (2011).

2011). In particular, the Member States with the largest proportions of wooded area in 2010 were Finland and Sweden, where approximately three quarters of the land area was covered with forests or woods. Indeed, Sweden alone accounted for 17.6% of all the wooded land in the EU in that year. More generally, the five largest wooded areas (in Sweden, Spain, Finland, France and Germany) collectively accounted for well over three fifths (62.4%) of the wooded land in the EU (FEW 2011). In contrast, the least densely wooded Member States were Malta, the Netherlands, Ireland and the United Kingdom.

Recent forest surveys have found that Europe is the only region in the world to have experienced a positive net change in forest area over the last 20 years (SEF 2011). This increase is set against a global decrease in forest area of roughly 3% (UN-ECE/FAO, 2007; The State of the World's Forests 2007[9]). However, it is important to consider that, while all the forest assessments agree that the European forest area has increased, estimates of the actual level of the increase vary considerably. For example, according to the EUROSTAT website[10], EU forest area has increased by 5% over the last 20 years. That is, the present forest area has increased by 5% on the area estimated in 1990. Somewhat confusingly, statistics given in FEW (2011) imply an increase of 8.4% since 1990. Similarly, SWF (2011) puts the figure at closer to 8.5% for Europe, excluding the Russian Federation. In another example, SEF (2011) gives a figure of 7.4%. One reason for the discrepancy could be that part of the increase occurred as a result of changes in the assessment methodology – that is, more recent assessments include vegetation that was not previously considered to be forest. If this is correct, the available data suggest that roughly 3.5% of the 8.5% increase of forest area should be attributed to changes in assessment, leaving a 5% actual forest area increase. However, this has not been confirmed.

Assessments of European forests also reveal that the rate of forest area change varies substantially between countries. Again, the estimates vary considerably. For example, FEW (2011) states that since the year 2000 the largest relative expansions in wooded area were recorded in Ireland (21.4%), while Bulgaria and Latvia both recorded increases in excess of 10%. Furthermore, in absolute terms, four Member States recorded an expansion in excess of 400,000 ha, namely France, Bulgaria, Italy and Sweden, with the latter recording the highest increase (594,000 ha). However, four of the EU Member States recorded a fall in their areas of wooded land, with Denmark

[9] The State of the World's Forests reports for 2007 and 2011 (UN-ECE/FAO, 2007; UN-ECE/FAO, 2011) will be referred to as SWF (2007) and SWF (2011) respectively.
[10] http://epp.eurostat.ec.europa.eu/portal/page/portal/forestry/introduction

recording the largest reduction (-5.0%) followed by Portugal, Slovenia and Finland (FEW 2011). The findings given in SWF (2011) display many similarities to those in FEW (2011), but are also surprisingly different. In particular, SWF (2011) makes a point of noting the large increase in area of Spain's forest and states that, within Europe, only Estonia and Finland reported a decrease in forest area between 2000 and 2010, failing to mention Denmark, Portugal and Slovenia. In addition, SWF (2007) also noted that Portugal's forest area had increased since the year 2000. If both SWF (2007) and FEW (2011) are correct, it must mean that Portugal has experienced a dramatic loss of forest cover over the past 4 years. However, the 2010 FAO Global Forest Resources Assessment for Portugal suggests that forest area has continued to increase up to and including 2010.

The source of the disagreements outlined above is far from clear since much of the data appear to originate from the FAO. Presumably, the discrepancies are a result of different classification systems, but this is not abundantly clear from the reports. As an example, SWF (2007) and SWF (2011) present the different values for the European forest area in 1990; the existence of such differences is briefly mentioned in FEW (2011), but not precisely explained.

As demonstrated by the above, the EUROSTAT and FAO reports provide a wealth of relevant forestry information often on a country by country basis; this includes data on forest area, forest biomass, the area used for commercial wood production, the area assigned for the protection of soils and water, the area protected for the conservation of biodiversity. However, the majority of the data ultimately comes from UN-ECE and the FAO. As a result, they cannot be considered to be independent, or corroborating, data sets. Furthermore, such detailed statistics have only been available since 1990, which are the FRAs for 2010 (FAO, 2010), 2005 (FAO, 2006b), 2000 (FAO, 2000). Prior to this period, it is necessary to rely on, in reverse chronological order, the FRAs for 1990 (FAO, 1995), 1985 (UN-ECE/FAO, 1985), the Regional FRA for Europe (FAO, 1976), World Forest Inventories[11] for 1963 (FAO, 1966), 1958 (FAO, 1960) and1953 (FAO, 1955) and finally the report 'Forest Resources of the World' (FAO, 1948)[12].

[11] The World Forest Inventories will subsequently be referred to as WFI.
[12] Henceforth referred to as FROW (1948). A full list of FAO publications can be found at http://www.fao.org/docrep/v6585e/V6585e11a.htm, with FAO statistics being available at http://faostat.fao.org/site/377/default.aspx#ancor.

It is important to note again that care must be taken when comparing forest statistics before and after the year 2000, because of the change in definition that was introduced between 1990 and 2000 (see section 1). In addition, some reports present statistics for the whole of Europe, as opposed to the EU or on a country by country basis. For example, the 1976 European FRA states that roughly 154 million ha (~33%) of Europe is forested, i.e. covered with closed forest and other wooded land. However, FS (2009) states that 177 million ha (~42%) of the EU is covered by forest. The 1976 FRA implies a total land area of 467 million ha, while FS (2009) implies a land area of 421 million ha. The difference occurs partly because the 1976 report includes non-EU countries such as Norway, Iceland, Switzerland, and the former Yugoslavia, but does not include Latvia, Lithuania and Estonia which were formerly part of the U.S.S.R. In addition, since the 1976 European FRA does not provide data on a country by country basis, it is not possible to modify their calculation to better estimate EU forest cover in 1976. Nevertheless, a 9% increase in forest area since 1976 would be consistent with the estimated ~8% increase in area since 1990. As mentioned previously, a certain fraction of the increase must be due to changes in the assessment methodology previously described. Assuming this effect contributes an apparent area increase of 3.5%, the true increase in forest area since 1976 could be ~5.5%. If correct, this number would be similar to the estimated increase since 1990, from which we could infer that the EU forest area was roughly constant between 1976 and 1990. However, given the inherent uncertainties in the calculation, a conclusion of this sort cannot be stated with confidence.

In another example of the wide range of forest area estimates, the European Forest Institute (EFI[13]) used FAO statistics from 1960, 1970, 1980 and 1990 to show that the areas of forested and wooded land in 9 European regions increased from 145 million ha in 1960/70 to 149.3 million ha in 1990 - an increase of ~3% (Kuusela, 1994). However, when including ‘other wooded land’, the increase is closer to 10%. As mentioned previously, the disagreement with previous statistics is likely to be due to inconsistent selection of countries and differences in assessment techniques. It is also worth noting that the EFI database does provide detailed forest information at a national level[14], including a 1 km resolution forest map of Europe generated using a combination of IRS-P6, SPOT4 and AVHRR satellite data (Kempeneers et al., 2011; Païvinen et al., 2001; Schuck et al., 2002).

[13] http://www.efi.int/portal/
[14] http://www.efi.int/portal/virtual_library/databases/efiscen/inventory_database/

Consequently, it is perhaps safer to only make direct comparisons between surveys that use the same definition of forests, and which also use the same land area for Europe. As an example, earlier FAO publications (e.g. FROW 1948; WFI 1953, 1958 and 1963) provide forest area estimates that broadly agree with the 1976 European FRA. For example, FROW (1948) estimated the land area (including the former U.S.S.R) covered by forests to be ~30-38%, depending on which countries are included. While this agrees broadly with the 1976 European FRA, the 1948 FROW provides less detailed data overall, meaning that the validity of comparisons with later data is uncertain. Unfortunately, the 8% uncertainty in the FROW (1948) forest area estimate makes it difficult to identify statistically significant forest area changes in subsequent surveys.

The FAO/EFI data have also been used to calculate evolving carbon uptake rates due to the expansion of European forests. In particular, Goodale et al. (2002) described forest carbon sinks in the Northern Hemisphere and showed that, in 1990, the net change in forest-sector carbon pools was about +0.14 Pg C/year. Similarly, Nabuurs et al. (2003) described the evolution of the European forest carbon sink from (1950-1999) and showed that carbon uptake in European forests has increased from approximately 0.03 Pg C/year up to 0.17 Pg C/year – a factor of 5.67 increase. Their analyses were based on European aggregated statistics on forest area and net annual increment from the FRAs for 1950-2000 (i.e. FAO: 1948, 1955, 1960, 1976; UN-ECE/FAO, 1985, 1992, 2000).

Prior to the 1940s, general information on forests and land cover is known to be sparse with the original papers and books frequently being difficult to obtain and of undocumented accuracy. However, FROW (FAO, 1948) does provide a useful review of the early publications on forestry, mentioning that the first comprehensive estimate of world forest resources was made by Zon and Sparhawk (1923) immediately after the First World War. Subsequently, Ilvesallo and Jalava (1930) and Streyffert (1931) published important studies of the coniferous resources of the world. Following on from this, the International Institute of Agriculture at Rome published estimates supplied by national governments in a series of yearbooks which appeared from 1933 to 1938. In 1946, the information available on forests was summarised in a report submitted to the Second Session of the Conference of FAO under the title Forestry and Forest Products - World Situation, 1937-1946. In the same year Sir Hugh Watson made an independent study which appeared in the (British) Empire Forestry Handbook. However, these original sources are difficult to obtain. Furthermore, it is important to note that while they do make valuable additions to our knowledge of forests in the first half of the 20th century, they all suffered from fundamental limitations - particularly the inconsistent use of forestry definitions.

General trends of historical forest loss and gain across Europe are discussed extensively by Mather (1992), Mather and Needle (1998), Mather, et al. (1998), Mather and Needle (2000), Mather and Fairbairn (2000) and Mather (2004). However, the difficulty with investigations of this sort is that many European countries were deforested hundreds of years ago, meaning that records are not available or were never made. Nevertheless, using national records from Denmark, France and Switzerland, Mather (2001) wrote:

"During the 19th and 20th centuries, many European countries underwent a 'forest transition'. Net national forest cover stopped declining and began to increase (Mather 1992). This has led some to speculate that developing countries currently experiencing deforestation may eventually undergo a similar transition."

In addition, Mather (2001) noted that Denmark currently has nearly 3 times more forest cover than in 1800, though the country has witnessed a recent decrease of forest cover. The transition is attributed to the influence of technology, scientific forest management, and the increased use of fossil fuels.

One aspect in which all of the data sets described above lack detail is the absence of explicit information on the geographical distribution of forests. That is, forest areas and trends are given for entire countries or regions, but there is no information about the precise distribution of forests within a country. However, since the 1980s, information of the required type can be obtained from land cover maps generated from data collected by satellites. For example, Loveland and Belward (1997) constructed the International Geosphere Biosphere Project (IGBP) Data Information System land cover (DISCover) global vegetation map with 1 km spatial resolution from Advanced Very High Resolution Radiometry (AVHRR) data. At a European level, the European Environment Agency initiated the Coordination of Information on the Environment (CORINE[15]) project to study European land cover change. To date, the CORINE project has produced 0.1 km resolution maps for 1990, 2000 and 2006 which were generated from data taken by the

[15] http://www.eea.europa.eu/publications/COR0-landcover

LANDSAT-4,5,7, SPOT-4 and IRS-P6 satellites. In addition, the ALTERRA institute at the University of Wageningen has produced 1 km resolution European land cover LANMAP-1,2 data sets by combining the CORINE, GLOBAL LAND COVER 2000 (GLC2000; Fritz et al., 2003[16]) and PELCOM[17] databases. There is also a version of GLC2000[18] specifically focusing on Europe. To complement and update GLC2000, the GLOBCOVER[19] provides a global land cover data set at a resolution of 0.3 km for the year 2005. The satellites and original data sets are described in greater detail in the appendix.

Prior to the advent of remote sensing instruments, vegetation maps were constructed by combining maps produced using ground-based observations. The earliest example of a global natural vegetation map was produced by Küchler (1964), followed by Matthews (1983) and Olson et al. (1983). These data sets were used in early ecosystem models to estimate carbon uptake rates (e.g. Melillo et al. 1993) and estimates of historical land use (e.g. Houghton et al. 1983) which have heavily influenced recent simulations of historical land cover change (e.g. Ramankutty and Foley 1999; Klein Goldewijk, 2001). However, the origin of the data used to construct the maps is unclear, meaning that the associated uncertainties continue to be propagated through current research and preconceived assumptions are hard to dislodge. Nevertheless, in the absence of other information these maps do provide a useful estimate of the historical forest cover and its spatial distribution.

In the absence of historical data which explicitly tracks changing land use/forest areas over time, several studies (e.g.; Ramankutty and Foley, 1999; Klein Goldewijk, 2001; Pongratz et al., 2008; Kaplan et al., 2009; Klein Goldewijk et al., 2010) have attempted generate historical land cover data using sophisticated modelling techniques, but applying simplistic relationships to extrapolate 20th century patterns of land use to earlier times. In addition, Hurtt et al. (2011) have included the land use data produced by Klein Goldewijk et al. (2010) combined with predictions of future land use scenarios to estimate the influence of human activities out to the year 2100. At this point it is important to note that Klein Goldewijk et al. (2010) only explicitly calculated the changing distribution of crops and pasture over time. However, this information can be combined with ecosystem models to calculate the historical forest coverage.

[16] http://bioval.jrc.ec.europa.eu/products/glc2000/glc2000.php
[17] http://www.geo-informatie.nl/projects/pelcom/
[18] http://www.eea.europa.eu/data-and-maps/data/global-land-cover-2000-europe
[19] http://www.gofc-gold.uni-jena.de/sites/globcover.php

Both Klein Goldewijk (2001) and Ramankutty and Foley (1999) constructed a contemporary cropland map using satellite images correlated with current cropland inventories as a starting point for their model, and then ran a land cover change model backwards in time to the year 1700. Specifically, Ramankutty and Foley (1999) opted to estimate the historical cropland distribution by downsizing the current distribution such that, at each point in time, the total cropland area matched estimates based on historical cropland inventories. In a slightly different approach, Klein Goldewijk (2001) assumed that the distribution of croplands follows population density. Under this assumption, the total cropland area was matched to estimates from historical inventories at each point in time. Thus, the historical forest distribution may be estimated by combining population density maps with historical cropland inventories, and assuming that where cropland has expanded forested areas have reduced and vice versa. In both studies, the land cover change model is run backwards using the present land cover as a starting point. As the area of croplands decreases, both models replace the area formerly occupied by crops with natural vegetation of the type and quantity predicted by ecosystem models (e.g. BIOME, see Prentice et al., 1992; Leemans and van den Born 1994; BIOME3, see Haxeltine and Prentice 1996) to be most suited to the climate at that location. Thus, these studies make use of much of the available data: historical population and cropland data, while the ecosystem models are driven by temperature, precipitation, insolation and soil data.

Pongratz et al. (2008) used a similar approach to model land use from 800 to 1992. They started with the potential vegetation map used by Ramankutty and Foley (1999). Next, the projected spatial distribution of croplands was calculated by assuming that it followed the historical population distribution. To do this, they estimated the area of cropland used per person in the year 1700 and assumed that this per capita land usage applied back to 800. They also assumed that the ratio of cropland to pasture remained constant during the period of interest.

Kaplan et al. (2009) modelled deforestation in pre-industrial Europe using a slightly different approach. Even though their study focuses on an earlier period this approach is worth discussing since they claim their model to be more accurate than either Klein Goldewijk (2001) or Ramankutty and Foley (1999). Notably, Kaplan et al. (2009) calculated the natural distribution of forest vegetation using a related ecosystem model (FORSKA, see Prentice et al., 1993) to that employed by Klein Goldewijk (2001) and

Ramankutty and Foley (1999). However, their prescription for clearing forests was built on the correlation between population density and forest cover developed by Mather (1999). Thus, using historical population maps, they estimated the geographical distribution of croplands and forest cover from 1000 BC to the present day. This prescription predicts much greater levels of deforestation at earlier times than Klein Goldewijk (2001) or Ramankutty and Foley (1999). However, Kaplan et al. (2009) claim that their forest cover maps are in better agreement with archaeological studies (Rackham and Rackham, 1994; Gaillard et al., 2000; Gaillard, 2007; Hellman et al., 2008).

2.1 Land cover classification schemes

Before describing the available datasets further, another area of confusion surrounding forests is the different land cover classification schemes employed throughout the published literature. While this is not necessarily relevant to the definition of forests, it is of importance when interpreting comparisons between data sets and potential vegetation maps produced by simulations. For example, both GLC2000 and GLOBCOVER data sets use the FAO land cover classification scheme (LCCS[20]) which consists of 22 land cover types. The HYDE[21] database (Klein Goldewijk 2001; Klein Goldewijk et al., 2010) uses the DISCover International Geosphere Biosphere Project (IGBP) legend, which consists of 17 land cover types (FRA 2000) with 13 vegetation types. In contrast, the SAGE[22] database (Ramankutty and Foley 1999) uses an aggregation of the 94 Olson Global Ecosystem classes (Olson et al., 1983) into 18 land cover types with 15 vegetation types.

2.2 Human population datasets

Several of the land cover change models described previously employed historical population data. However, the population data are not always taken from the same sources. In an influential early paper, Houghton et al. (1983) based some of their historical estimates of land usage (back to 1700) on a hypothesised correlation between cropland and population distribution. Using the historical population distribution compiled by McEvedy and Jones (1978), they generated maps of historical natural vegetation by

[20] http://bioval.jrc.ec.europa.eu/products/glc2000/legend.php
[21] http://themasites.pbl.nl/en/themasites/hyde/index.html
[22] http://www.sage.wisc.edu/mapsdatamodels.html

replacing former cropland with the appropriate vegetation types taken from Küchler's (1964) map of natural vegetation. This is significant, because Ramankutty and Foley (1999), Pongratz et al. (2008) and Klein Goldewijk et al. (2010) cite a variety of references (e.g. Houghton and Hackler, 1995; Richards, 1990) that are closely based on Houghton et al. (1983). Hence, the SAGE and HYDE databases have an implicit dependence on the population data given by McEvedy and Jones (1978), and the unclear assumptions about land clearance used by Houghton et al. (1983). The original HYDE historical land cover model (Klein Goldewijk, 2001) took population data from International historical statistics (Mitchell, 1993; Mitchell, 1998a,b) for the period 1750-1993. However, the most recent version of the HYDE database (Klein Goldewijk et al., 2010), uses historical population numbers from three books: McEvedy and Jones (1978), Maddison (2001) and Livi-Bacci (2007). The spatial patterns were obtained based on population density map patterns of LANDSCAN[23, 24] (Landscan, 2006). Specifically, the LANDSCAN data set is a gridded worldwide population database derived using census counts (at sub-national level) from around the year 2000 that are apportioned to each grid cell based on its proximity to roads, slope, land cover, night time lights, and other data sets.

Kaplan et al. (2009) used population data from the world population atlas compiled by McEvedy and Jones (1978), as did Pongratz et al. (2008). However, as noted by Klein Goldewijk et al. (2010), the historical population estimates should be treated with care, particularly since the references on which they are based are not easily accessible.

2.3 Ecosystem models, climate and soil data

The HYDE and SAGE historical land cover data bases both use potential vegetation maps generated by slightly different versions of the equilibrium vegetation model, BIOME (Prentice et al. 1992). Specifically, for the HYDE data base Klein Goldewijk (2001) used a modified version of BIOME (see Leemans and van den Born 1994), while Ramankutty and Foley (1999) used BIOME 3 (Haxeltine and Prentice 1996). In practice, these versions are almost identical.

The BIOME model simulates the equilibrium distribution of vegetation at a location using soil texture class, monthly mean temperature, precipitation and insolation. BIOME then

[23] http://www.ornl.gov/sci/landscan/
[24] http://gcmd.nasa.gov/records/GCMD_Landscan.html

determines the dominant and secondary plant functional types (PFTs) in each grid cell that maximise net primary production (NPP). In this way, it produces a map of the vegetation expected in the absence of human intervention. Essentially, BIOME determines the vegetation distribution that maximises NPP for a given climate scenario, and assumes the vegetation obtains that configuration instantaneously. This approach differs from dynamic global vegetation models (DGVMs) used in general circulation models (GCMs) which simulate competition between different plant functional types throughout the model run at high temporal resolution (e.g., every 10 – 30 minutes), meaning that DGVMs simulate the evolution of the vegetation towards equilibrium. For example, TRIFFID (Cox et al., 2001) models competition between PFTs using the Lotka-Volterra equations[25] with competition indices given by the relative plant heights at each point in time. Both Ramankutty and Foley (1999) and Klein Goldewijk (2001) used temperature, precipitation and sunshine statistics from the Leemans and Cramer (1991) database, which contains long term monthly climate averages for the period 1931-60. The soil types are taken from the IIASA harmonized world soil database[26].

Pongratz et al. (2008) used a similar approach to model land use from 800 to 1992. However, they did not explicitly calculate any potential vegetation maps, instead using the vegetation map used by Ramankutty and Foley (1999), but reclassified according to the PFT used in the Jena Scheme for Biosphere – Energy Transfer Coupling in Hamburg (Raddatz et al., 2007).

Kaplan et al. (2009) used a model similar to BIOME known as FORSKA (Prentice et al., 1993) in order to simulate the natural forest cover across Europe. This was driven by temperature and precipitation data from Hijmans et al. (2005), sunshine hours from New et al. (2002), and soil data from the IIASA 2008 Harmonized World Soil Database.

3. Forest data sets derived from written sources and historical inventories

One of the earliest global vegetation maps was produced by Küchler (1964) for the United States Geological Survey. This map provided an estimate of the potential natural vegetation – that is, the vegetation expected in today's climate in the absence of humans. It can still be seen in Goode's world atlas. As such, it subsequently proved to be useful as a starting point in land use change models, prior to the development and

[25] The Lotka-Volterra equations are a system of coupled differential equations that model
[26] http://www.iiasa.ac.at/Research/LUC/External-World-soil-database/HTML/

use of ecosystem models in simulating potential vegetation. As described in the previous section, Houghton et al. (1983) used Küchler's map along with population data from McEvedy and Jones (1978) to hindcast land use change back to 1700. Their approach followed earlier discussions of land use change, notably Robertson (1956) and Grigg (1974). However, since these books and the original data on which these publications are based are not readily accessible, it is not clear how much emphasis should be placed on their findings.

For comparison, three other well-known land cover maps derived from ground based data were compiled in the 1980s: "Global vegetation and land use: new high resolution data bases for climate studies" (Matthews, 1983); "Carbon in Live Vegetation of Major World Ecosystems" (Olson et al., 1983), and "A Global Archive of Land Cover and Soils Data for Use in General Circulation Climate Models" by Wilson and Henderson-Sellers (1985). Thus, in principle, it is possible to estimate forest cover change in Europe from differences between the Küchler and Matthews/Olson/Wilson maps.

More recently, a new land use database for Europe, HILDA[27] (Historic Land Dynamics Assessment), has been constructed from a range of historical data. This dataset covers the period 1900 – 2010 and provides changes in land cover at decadal intervals on a 1 km × 1 km scale. A map showing areas where forest cover has increased or decreased using data from the HILDA database is provided in the next report.

4. Summary of land cover data derived from satellite monitoring

The primary reasons for developing land cover maps derived from remotely sensed data are to improve the consistency and reproducibility of land cover datasets. These maps represent significant progress compared to traditional ground-based mapping methods, such as those employed by Küchler (1964), Matthews (1983), and Olson et al. (1983) described previously. In particular, Hansen and Reed (2000) (see also DeFries and Townshend 1994) compared the ground-based maps generated by Matthews (1983) and Olson et al. (1983) and revealed that the level of disagreement present among traditional ground-based land cover maps is significantly larger than for vegetation maps generated using remote sensing instruments, as demonstrated in Table 1.

[27] http://www.grs.wur.nl/UK/Models/HILDA

Consequently, they advocated the use of satellite maps to generate consistent land cover maps, which will provide better measurements of land cover change.

Hansen and Reed (2000) also compared two land cover maps derived from the same Advanced Very High Resolution Radiometry (AVHRR) data set, but which used different algorithms for predicting land cover type from the data. These are the DISCover dataset, and another produced by the University of Maryland (UMd). The different algorithms arise for the following reasons: 1) different classification approaches can be fully/partially automated; 2) 'mixed pixels', containing several vegetation types, are common at global scales (i.e. 1 km resolution). As a result, there will be some disagreement between different classification algorithms, and it is important to compare the performance of different algorithms for the same dataset (Table 2).

Table 1: Comparing thematic agreement between Matthews (1983) and Olson et al. (1983) ground based vegetation maps (%). The average class agreement = 66.30 %, overall agreement = 68.35%.

Olson → Matthews ↓	Forest/woodland	Grass/shrubs	Crops	Bare ground
Forest/woodland	70.2	22.0	7.1	0.7
Grass/shrubs	14.9	60.0	6.7	18.5
Crops	16.1	29.5	51.1	3.2
Bare ground	0.5	14.5	1.0	83.0

Table 2. Comparing thematic agreement of DISCover and University of Maryland (UMd) vegetation dataset using satellite remote sensing (%). The average class agreement = 81.87%, overall agreement = 80.32%.

DISCover → UMd ↓	Forest/woodland	Grass/shrubs	Crops	Bare ground
Forest/woodland	88.6	9.0	2.4	0.0
Grass/shrubs	15.8	69.0	10.0	5.2

Crops	8.6	12.2	79.2	0.0
Bare ground	0.0	9.3	0.0	90.7

Since the publication of Hansen and Reed (2000), numerous other datasets have been generated from a variety of satellite data and using different classification algorithms. There are both global and European satellite-derived datasets which can be used to estimate European forest cover change over recent years. These datasets are listed below, and subsequently described in more detail. In addition, in order to interpret the literature it is also necessary to have some knowledge of the different satellites; details are given in Appendix 1.

Global land cover datasets

- **DISCover**
- **GLC2000** (http://bioval.jrc.ec.europa.eu/products/glc2000/products.php)
- **MODIS** (http://modis.gsfc.nasa.gov/)
- **GLOBCOVER** (http://www.gofc-gold.uni-jena.de/sites/globcover.php)

European land cover datasets

- **CORINE** (http://www.eea.europa.eu/themes/landuse/interactive/clc-download)
- **PELCOM** (http://www.geo-informatie.nl/projects/pelcom/public/index.htm)
- **LANMAP** (http://www.alterra.wur.nl/UK/research/Specialisation+Geo-information/Projects/LANMAP2/)
- **GLC2000-Europe** (http://www.eea.europa.eu/data-and-maps/data/global-land-cover-2000-europe)
- **HILDA** (http://www.grs.wur.nl/UK/Models/HILDA)

4.1 Global land cover data sets

DISCover (Loveland and Belward 1997)

The International Geosphere Biosphere Project (IGBP) Data Information System landcover (DISCover) data set (Loveland and Belward 1997, 2000) is a 1 km resolution land cover dataset created from AVHRR data taken during the period April 1992 to

March 1993 and was produced by the U.S. Geological Survey. The land cover is classified as one of 17 IGBP classes defined in Belward (1996). Over 250 maps and atlases of ecoregions, soils, vegetation, land use and land cover were used in the interpretation phase of the study and served as reference data to guide class labelling (Hansen and Reed, 2000).

Global Land Cover 2000 (Batholomé and Belward, 2005)

The Global Land Cover 2000 (GLC 2000) Project dataset was produced by the European Commission, Joint Research Centre, Global Environment Monitoring Unit (http://bioval.jrc.ec.europa.eu/products/glc2000/glc2000.php).
The objective was to provide a harmonised land cover database over the whole globe, for the year 2000 (1st November 1999 – 31st December 2000). To achieve this objective, GLC 2000 makes use of the VEGA 2000 dataset, which consists of 14 months of daily 1 km resolution global data acquired by the VEGETATION instrument on board the SPOT 4 satellite (http://www.spot-vegetation.com/pages/vega2000.htm), and delivered as multi-channel daily mosaics. The GLC2000 database legend consists of 22 land cover types (http://bioval.jrc.ec.europa.eu/products/glc2000/legend.php) defined by the FAO land cover classification scheme (http://www.fao.org/docrep/003/x0596e/x0596e00.htm).

Moderate-resolution Imaging Spectroradiometry (Friedl, et al. 2002)

The NASA MODIS data set provides land cover information with a resolution of 1km classified according to the IGBP scheme with 17 land cover types. Land Products are available through the Land Processes DAAC at the U. S. Geological Survey EROS Data Center (EDC) (https://lpdaac.usgs.gov/).

GLOBCOVER

The GLOBCOVER project was started in 2004 with the objective of producing a global land cover map for the year 2005, using 0.3 km resolution data from the MERIS sensor on-board ENVISAT. This land cover map was intended to complement and update existing comparable products, such as GLC2000. The project now involves international collaboration between ESA (http://www.esa.int/esaCP/index.html), FAO, UNEP (http://www.unep.org/), JRC (http://ec.europa.eu/dgs/jrc/index.cfm), IGBP and GOFC-GOLD (http://www.fao.org/gtos/gofc-gold/index.html).

4.2 European land cover data sets

CORINE (Coordination of Information on the Environment) land cover dataset (Heymann et al., 1994)

European Commission implemented the CORINE programme in 1985, with the first dataset making use of observations taken by the Landsat-4/5 satellites for 1990 (http://www.eea.europa.eu/publications/COR0-landcover). The procedures used to generate this data set are described by Heymann et al. (1994). Subsequent updates were produced for the years 2000 (Büttner et al., 2002,;Hazeu, 2003; http://www.eea.europa.eu/data-and-maps/data/corine-land-cover-2000-clc2000-100-m-version-9-2007) and 2006 (http://www.eea.europa.eu/data-and-maps/data/corine-land-cover-2006-clc2006-100-m-version-12-2009). For completeness, it is worth noting that the CLC2000 used data from the Landsat-7 satellite, while CLC2006 used data from the SPOT-4 and IRS-P6 satellites. In each data set, the resolution is 100 m, with 44 land cover types.

PELCOM (Pan-European Land Cover Monitoring) data (Mücher et al., 2001)

The 1 km Pan-European Land Cover Monitoring (PELCOM) Database (http://www.geo-informatie.nl/projects/pelcom/public/index.htm) was set up by a partnership of European research institutions for the "Development of a consistent methodology to derive land cover information on a European scale from remote sensing for environmental monitoring". It is based on the integrative use of multi-spectral and multi-temporal AVHRR satellite imagery and ancillary data. In particular, PELCOM combines AVHRR data with the 1990 CORINE land cover database as ancillary/reference data, but uses 16 major land classifications as described by Mücher et al. (2001).

GLC2000-Europe (Hartley et al., 2006)

This data set was created by extracting a window from the GLC2000 global database with the objective of providing a consistent and improved classification for the whole of the European continent, with respect to the data derived from the VEGETATION instrument. This is essential not only for management and monitoring of European ecosystems, but also, in being directly related to a global land cover product, is valuable for putting the land cover of Europe in a global context. The land is classified to be compatible with the FAO land cover classification system, thereby ensuring pan-European consistency. CLC2000 was used to assist with the classification.

LANMAP2 (Mücher et al., 2006)

LANMAP is a program initiated in 2002 by Alterra at Wageningen University to produce a pan-European landscape classification that provides a practical and easy tool for European policy implementation (http://www.alterra.wur.nl/NR/exeres/6DA551F0-D7A6-45C6-A499-AC6B80FB82EB). To do this, LANMAP combines and integrates the 1990 CORINE, PELCOM and GLC2000 datasets, but with 10 land cover types and a spatial resolution of 1 km.

HILDA (Fuchs, 2012)

HILDA (Historic Land Dynamics Assessment) is a high-resolution (1 km) harmonized model approach for reconstructing historic land changes in Europe for the period 1900 – 2010.

5. Summary of historical land cover data derived from numerical models

The latest version of the HYDE database provides publicly available land cover files that can be obtained from: http://themasites.pbl.nl/en/themasites/hyde/download/index.html. As noted in the first section of this report, these files are maps of land use from 10000 BC until 2005 AD, with files produced at 10 year intervals for the period 1700-2000 AD. Unlike the original version of the HYDE database, they do not contain data on the distribution of natural vegetation and forests. However, the evolving forest cover can be obtained by combining the land use data with an ecosystem model of the type previously described (e.g. BIOME, BIOME3 or FORSKA). The same is true for land use data available from the SAGE database which is available at:

(http://www.geog.mcgill.ca/~nramankutty/Datasets/Datasets.html). Notably, the Pongratz et al. data (which is generated by combining the HYDE and SAGE databases) does provide vegetation and forest data. It is available at: (http://cera-www.dkrz.de/WDCC/ui/Compact.jsp?acronym=RECON_LAND_COVER_800-1992). In contrast, the forest coverage calculated by Kaplan et al. does not appear to be publicly available.

6. Barriers and knowledge gaps

There is a huge quantity of literature and data that are relevant to European forest cover and its historical evolution. The data take the form of ground based surveys (1920s – 1960), published maps (~1960-1980), remote sensing data sets generated from satellites observations (~1980-present), and numerical models (~1980-present) which combine ecosystem models and historical data. However, given the lack of geographically explicit historical data, there remain definite limitations in our estimates of the spatial distribution of Europe's forests back to 1850.

At the present time, land cover maps produced from satellite monitoring instruments are the best methods for estimating forest cover over the past 20-30 years. In general, the data sets derived from several different sensors agree relatively well. Discrepancies occur because there is no universally agreed mapping between vegetation type and the spectral signatures detected by different satellites. As a result, different research groups employ slightly different algorithms to predict land cover type from the satellite observations. It is also important to note that the official FAO definition of a forest changed between 1990 and 2000. This presents a significant obstacle to making accurate comparisons of forest areas and their rates of change before and after the year 2000.

For the period 1960-1980, natural vegetation maps compiled from ground-based surveys provide a useful source of information for estimating European forest coverage. Unfortunately, there is often significant disagreement between different compilations, while the accuracy of forestry literature published prior to 1948 is impossible to verify. However, the recently constructed HILDA database (1900-2010, 1 km^2 resolution) has provided a significant improvement on previous land cover datasets generated from historical records. Nevertheless, numerical simulations which combine ecosystem models with historical data for population and cropland areas have proved to be a useful tool in estimating the evolution of forest coverage globally and in Europe. However, as shown above, only the cropland and pasture data are publicly available. This approach also has other inherent limitations: i) the estimates of historical populations and their geographical distribution are of undocumented accuracy; ii) assumptions about how land use follows population distribution strongly influence the distribution of forests, but are difficult to verify with historical data; iii) some ecosystem models estimate equilibrium vegetation distributions and so do not influence the climate via feedbacks; additionally, the vegetation in reality may not have reached equilibrium with the climate; iv) the ecosystem models are driven with climate data which are representative of the period 1931-1960 or 1950-2000, which are not likely to be appropriate for the 19th century (or earlier periods). Consequently, these models could likely benefit from the inclusion of Dynamic Global Vegetation Models (DGVMs) such as TRIFFID or ORCHIDEE[28], and the use of climate data from the 20th century re-analysis project (Compo et al., 2011) which generates global circulation data from historical measurements of surface pressure to provide more accurate historical time-series of temperature and precipitation.

7. Summary

European forest cover has changed significantly throughout history due to land use changes caused by population expansion and changes in the exploitation of wood products. Historical national statistics from several European countries suggest that forest area was at a minimum in the 19th century and early 20th century. However, in recent times European forest cover has undergone a transition in which the forest area stabilised and then expanded to cover approximately 42% of the European land area, an increase of ~5% in the past 20 years. This makes Europe, the only world region to have experienced net forest expansion during that period. At a national scale, the changes in forest cover across Europe are highly heterogeneous with forest expansion occurring in Ireland, Bulgaria, Latvia, France, Italy and Sweden, but decreasing forest area in Denmark, Portugal, Slovenia and Finland.

Measuring precise changes in the geographical distribution of European forests has only been possible over the last 20 years due to the advent of global and European land cover data sets generated from satellites. In this report, we have described the most

[28] http://orchidee.ipsl.jussieu.fr/index.php

useful data sets of this type: DISCover, GLC2000, MODIS, CORINE, PELCOM, GLOBCOVER, LANMAP2 and HILDA.

Estimating geographically explicit land cover and forest cover maps prior to 1980 is much more difficult and subject to significant uncertainties. For example, between 1960 and 1985, several global maps of natural vegetation were compiled from ground-based measurements. However, comparison of these maps demonstrates inconsistencies, and the original data used are not publicly available and hard to obtain. Estimating forest cover further back in time requires the use of numerical models that hindcast land cover changes by combining ecosystem models and historical cropland inventories to estimate the total area of land in use at any point in time. The geographically explicit land cover maps are then produced by making assumptions about how the historical croplands are distributed. In particular, the SAGE database down-scaled the current cropland distribution, while the HYDE database assumed that croplands follow population density. The areas formerly occupied by crops are then replaced with vegetation types predicted by the ecosystem model to be most suited to the climate and soil type at that location.

From the information compiled here, it is evident that the geographical distribution of Europe's forests will be conspicuous in the CORINE, GLC2000-Europe, LANMAP2 and HILDA data sets. Estimating the specific locations in which forest cover has recently changed is more difficult, however the magnitude of the area change can be obtained from FAO forestry statistics for each country. To evaluate changes in the geographical distribution of forests prior to 1990, it is necessary to rely on numerical models that predict land use and forest cover. At the present time, the HYDE database could be the most useful since the relevant literature provides clear and transparent explanations of the methods and assumptions. However, the HILDA database, although still being refined, has high spatial resolution and provides changes in forest cover once per decade.

Appendix 1

AVHRR on board NOAA series satellites

The latest National Oceanic Atmospheric Administration (NOAA) Advanced High Resolution Radiometry (AVHRR-3) sensor is an instrument launched in 1998 that detects radiation across 5 bands in the wavelength range 0.58µm -12.5µm, i.e. from the yellow part of the visible spectrum to the mid infra-red (http://edc2.usgs.gov/1KM/avhrr_sensor.php). The instrument records data with a spatial resolution of 1.1km, twice daily (http://nsidc.org/daac/projects/visible_infrared/avhrr.html).

ENHANCED THEMATIC MAPPER PLUS (ETM+) On board LANDSAT 7

The LANDSAT Program is a series of Earth-observing satellite missions jointly managed by NASA and the U.S. Geological Survey. Since 1972, LANDSAT satellites have collected information about Earth from space (http://landsat.gsfc.nasa.gov/). The ETM+ mapper on board LANDSAT 7 was launched in 1999 (http://landsat.gsfc.nasa.gov/about/L7_td.html), has 8 detectors which cover a wavelengths range 0.45µm -2.35µm with a spatial resolution of 30m, and a temporal resolution of 16 days (http://landsat.gsfc.nasa.gov/about/etm+.html). While the instrument is still in operation, a fault with the scan line corrector (SLC) means that images from ETM+ are missing data (http://landsat.usgs.gov/products_slcoffbackground.php).

LISS III on board IRS-P6/RESOURCESAT 1

The Indian Remote Sensing (IRS)-P6 satellite, otherwise known as Resourcesat 1, is a remote imaging satellite built by the Indian Space Research Organisation (ISRO), and launched in October 2003 (http://www.isro.org/satellites/irs-p6resourcesat-1.aspx). The LISS III instrument comprises 4 detectors which cover a wavelength range of 0.5µm – 1.7µm, with a spatial resolution of 23.5m (http://earth.esa.int/object/index.cfm?fobjectid=1660) and a temporal resolution of 24 days (http://www.isro.org/satellites/irs-p6resourcesat-1.aspx).

MERIS on board ENVISAT

ENVISAT was launched by ESA in 2002 to provide measurements of the atmosphere, ocean, land, ice (http://envisat.esa.int/earth/www/category/index.cfm?fcategoryid=87). The MERIS instrument on board ENVISAT comprises 15 spectral bands across the wavelength range: 0.39 µm to 1.04 µm (with a programmable bandwidth of between 2.5 and 30 nm). It has a spatial resolution of approximately 1 km over the ocean, and ~ 300 m over land (http://envisat.esa.int/instruments/meris/).

MODIS

MODIS (or Moderate Resolution Imaging Spectroradiometer) is an instrument aboard NASA's Terra (EOS AM) and Aqua (EOS PM) satellites (http://modis.gsfc.nasa.gov/about/), launched in 1999 and 2002 respectively. MODIS comprises 36 detectors which cover a wavelength range of 0.4µm-14.4µm, with a band dependent spatial resolution of 250m-1000m and a temporal resolution of 8 days (https://lpdaac.usgs.gov/products/modis_overview).

VEGETATION on board SPOT-4

The VEGETATION instrument onboard the SPOT-4 satellite was launched in 1998 and is similar to AVHRR, but has 4 detectors which cover a wavelength range of 0.43µm - 1.75µm, i.e. blue to the near infra-red (http://spot4.cnes.fr/spot4_gb/index.htm). In addition, while SPOT-4 produces data sets with a very high spatial resolution of 20m, its temporal resolution is only 26 days (http://www.spot-vegetation.com/pages/VegetationProgramme/spot4.htm). This should be contrasted to the AVHRR instrument, which has a temporal resolution of ½ a day (http://phenology.cr.usgs.gov/ndvi_avhrr.php). The VEGETATION 2 instrument on board SPOT-5 (http://spot5.cnes.fr/gb/satellite/satellite.htm) was launched in 2002.

Part B: Literature Review of the Historical Evolution of European Weather and Appropriate Datasets Spanning the Past 150 Years

Executive Summary

By making use of the available and most referenced climate datasets (e.g. ECA[29], HISTALP[30]), the evolution of climate in Europe has been analyzed at two main spatial scales: continental and regional, along with several examples for the (sub-)national level when data covering at least the medium-term time frame were available.

The first major point of note is the evidence that the warming observed in the last century mainly accelerated during the last few decades. The effect is clearer in winter than in summer, and more noticeable for southern Europe rather than the north. Notably, this recent warming appears to affect minimum temperatures in particular, thereby leading to a general reduction of diurnal temperature range. In addition, there is evidence of more moderate warming periods that occurred throughout the last five centuries.

As well as an increase in the mean temperature, studies have also revealed changes in temperature extremes, i.e. a strong increase in the number of warm-days and a decrease in the number of frost-days. However, the trend involving cold-days is comparatively more uncertain, even if the number has appeared to decrease during the winter months.

Trends in precipitation across Europe tend to be much more variable than for temperature. For example, the frequency of wet days seems to be generally increasing in the north and decreasing in the south of the continent, while the rainfall intensity of wet days is increasing more uniformly.

Several studies are described that identify mechanisms by which large scale circulation (air pressure difference) can influence extremes in both temperature and precipitation. In

[29] http://eca.knmi.nl/
[30] http://www.zamg.ac.at/histalp/

particular, the Arctic and North Atlantic Oscillations and East Atlantic variability are shown to be regionally important.

Focusing on a regional level, the Greater Alpine Region (GAR) has warmed nearly twice as much as the global average trend, with warming affecting winters more than summers. It is also important to note that the Alpine chain acts as a barrier between warmer-wetter conditions in the north-west of Europe and warmer-drier conditions in the south east, which are favoured by the effect of the NAO. Warming is also manifested as an upward shift (i.e., increasing altitude) of the zero degree line. As for the whole of Europe, the spatial variability of precipitation is greater than for temperature. However, temporal variability of the temperature is rather high. In addition, the rate at which glaciers have retreated during recent decades is found to be accelerating. The same trends were also found in Scandinavia, except that an increase in the frequency of hot days has not been observed.

The trends for central-southern Europe generally reflect those observed for the whole of Europe, i.e. more warming in winter particularly during the second half of the 20th century. Moreover, the residence time of the circulation type is positively correlated with the increase of warm extremes and negatively with cold extremes.

Drawing conclusions from national and local studies is more challenging. The heterogeneity of findings reinforces the general observation that results are highly affected by the spatial scale of evaluation, as well as by the number and distribution of weather stations, and how accurately they capture the data. Moreover, the time-scale (e.g. annual, seasonal) of analyses and the time frame duration (very long-, long-, medium-term) yield different responses.

More details on the data sets are given in Annexes 1 and 2.

1. Evolution of climate in Europe

1.1 Pan-European analysis

Europe is among the world regions where the temperature rose slightly more than the global average (Hiebl, 2006), namely by 0.076°C per decade during the 20th Century; notably, the ongoing warming (0.425°C per decade for the 1977–2001 period) has been especially strong there (Jones and Moberg, 2003). Indeed, Luterbacher et al. (2004) suggest that this late 20th century warmth is unprecedented in Europe for at least the past 500 years. Otterman et al. (2002) attributed a large part of the observed warming in Europe during 1948-1995 to more predominant south-westerly winds in winter.

Before the recent advances in climate data availability and improvements, Heino et al. (1999) found evidence using data for the period 1901-1995 in northern and central Europe (i.e. considering stations in Norway, Sweden, Finland, Poland, Czech Republic, Germany, Switzerland) of decreasing numbers of frost-days since the 1930s, which appear to be associated with strong increases in winter minimum temperatures. No major change in the precipitation extremes[31], strong wind speed and the number of thunderstorms was observed.

The following sections describe the evolution of European climate, in terms of averages and extremes. This information is drawn from datasets previously described for Europe, such as the ECA, which are well-evaluated in Wijngaard et al. (2003), Klein Tank et al. (2005), Hofstra et al. (2009) and Klok and Klein Tank (2009). The daily resolution, the large number of stations and its nine climate variables mean ECA is an excellent dataset to be used in studies on changes in climate and its extremes over the whole European region.

On a medium-term historical scale, Klein Tank et al. (2002a,b) and Klein Tank (2004) showed that the daily ECA dataset for the period 1946-1999 is in good agreement with two other datasets at lower temporal resolution for temperature (Jones et al. 2001; Jones, 1994) and precipitation (Hulme, 1992; Hulme et al., 1998) respectively (see Table 3).

[31] Maximum daily precipitation and number of days with precipitation ≥ 10 mm.

	Temperature trend (°C/decade); $n = 41$		Precipitation trend (mm/decade); $n = 97$	
	ECA stations	Land grid boxes (Jones)	ECA stations	Land grid boxes (Hulme)
1946–99	0.04 (−0.04–0.08)	0.03 (−0.05–0.11)	11.1 (5.5–16.8)	6.4 (1.3–11.4)
1946–75	−0.04 (−0.24–0.16)	−0.03 (−0.21–0.15)	16.1 (2.1–30.1)	4.7 (−8.4–17.8)
1976–99	0.42 (0.10–0.74)	0.38 (0.08–0.68)	−2.8 (−20.0–14.4)	−0.2 (−15.0–14.6)

Table 3. Comparison of ECA dataset with previously developed gridded datasets.

The trends in mean temperature of winter (defined as October-March) and the number of cold/warm-spell days[32] for stations in the ECA dataset between 1976 and 1999 are shown Figure 1(a). Less than 10% of the stations recorded a winter warming accompanied by a decrease in the number of cold-spell days (Figure 1(b)). In Figure 1(c) it is shown that the winter warming is accompanied by a diffuse increase in the number of warm-spell days.

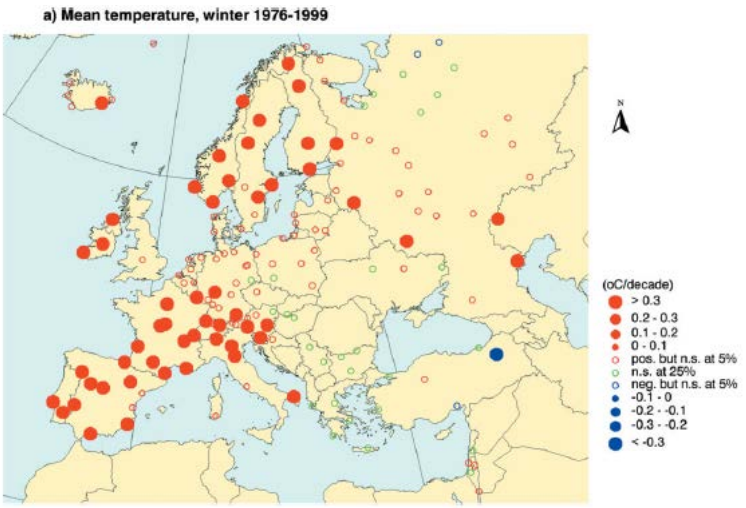

Figure 1. (a) Trends in mean temperature for winter between 1976 and 1999 derived from the ECA data set.

When averaged over all of the 168 ECA stations used, the increase in the number of warm-spell and cold-spell days during winter is 3.0 and 0.2 days per decade, respectively. In summer (April-September), the increase of warm-spell days is five times larger in magnitude than the decrease of cold-spell days.

[32] Cold/warm spells at a given site are periods of at least six consecutive days with daily mean temperatures below/above the lower/upper 10th percentile of the temperature distribution for each calendar day in the 1961–90 standard normal period. These calendar-day specific percentiles were calculated from 5 day windows centered on each calendar day. This gives a total sample size of 30 years × 5 days = 150 for each calendar day. The length of each spell is expressed in a single index that comprises the total number of days in that spell (from Klein Tank, et al. 2002a).

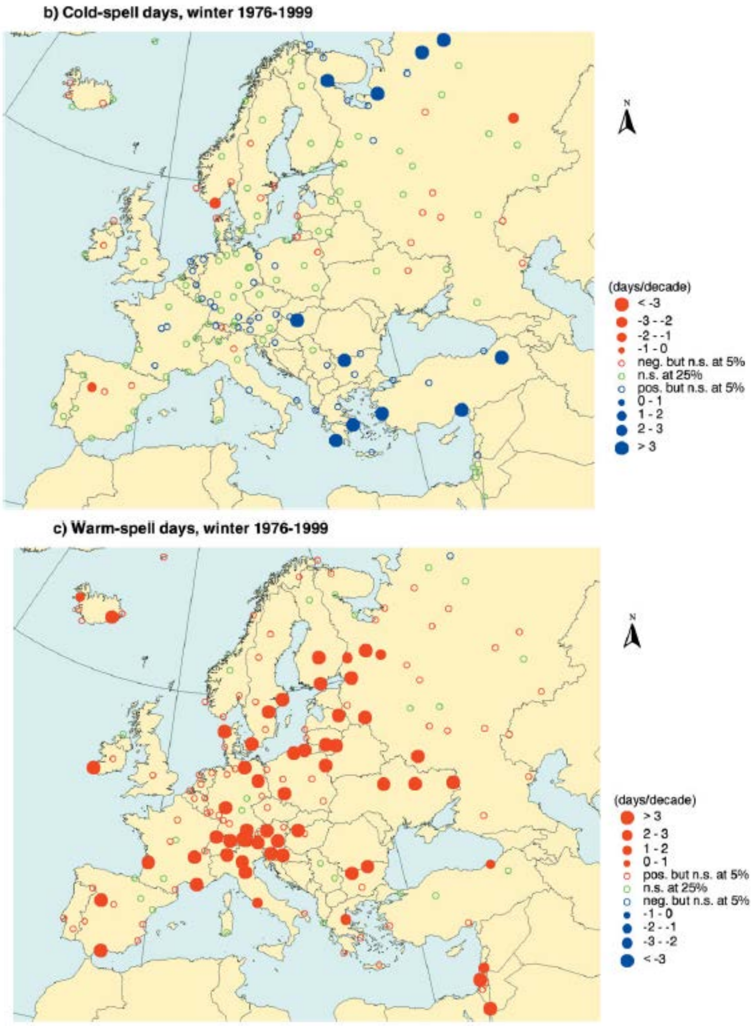

Figure 1.(b) Trends in cold-spell days and (c) trends in warm spell days during winter derived from the ECA data set.

A further application in Klein Tank et al. (2002a) deals with precipitation changes. The trends in winter precipitation metrics between 1946 and 1999 are shown in Figure 2. There are clear increases in the winter precipitation total, the number of wet days and the mean precipitation amount per wet day over most of Europe. The exceptions are the Iberian Peninsular and other parts of southern Europe, where decreases in winter precipitation amount and the number of wet days can be seen.

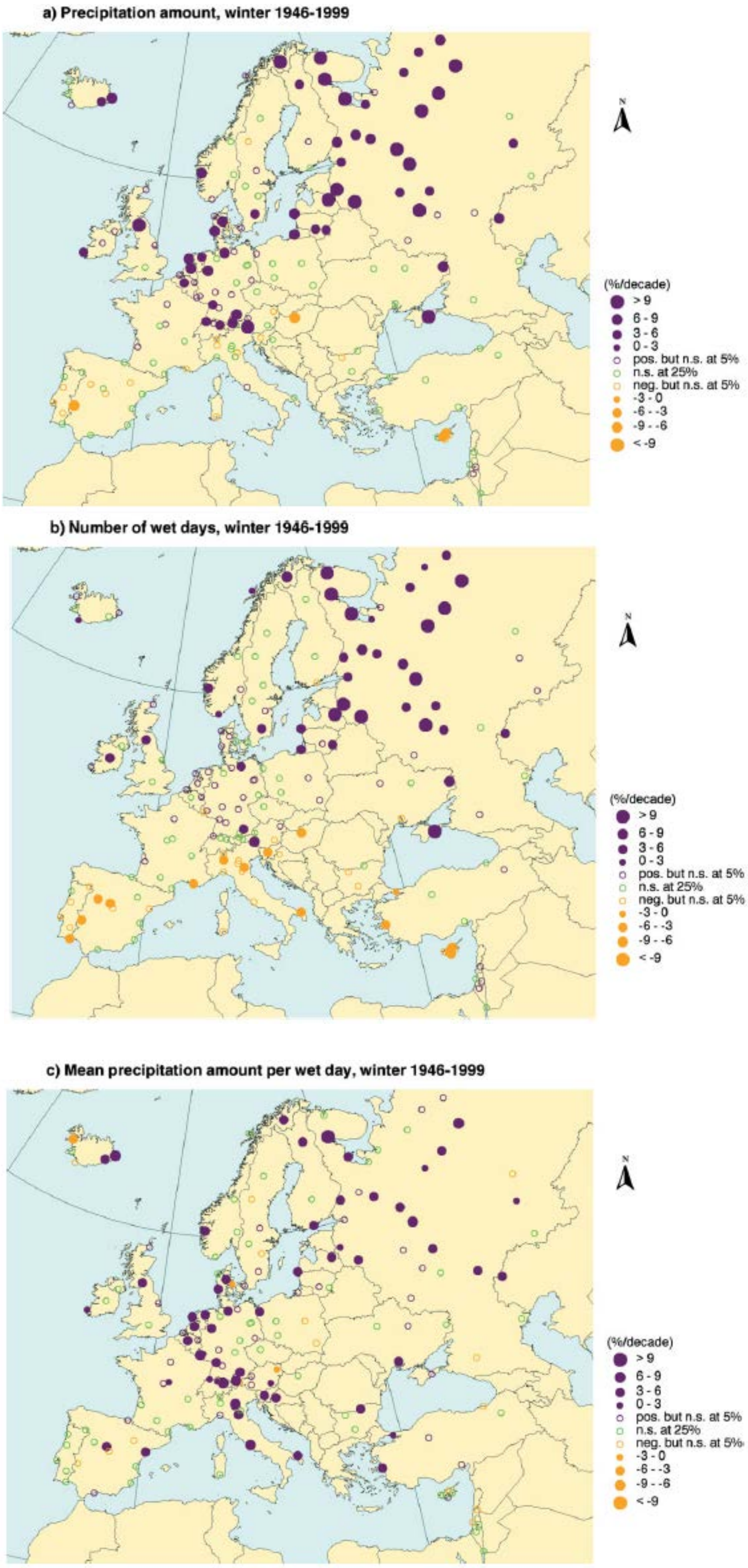

Figure 2. Trends in precipitation metrics for winter during the period 1946-1999 (shown as % change per decade). (a) precipitation amount, (b) number of wet days and (c) mean precipitation amount per wet day.

Still in terms of extremes (considering percentile-based, absolute and threshold indices)[33], and relying on the same ECA dataset from 1951 to 2003, the work of Alexander et al. (2006) reports how central-eastern Europe and the Iberian peninsula are experiencing a significant[34] increasing trend in the number of warm-nights and days (4-8 per decade), while central Europe and, in particular, France are experiencing a decrease in cold-nights and cold days (up to -6 per decade). Cold and warm conditions are evaluated using 10th and 90th percentile of both minimum (for nights) and maximum (for days) temperatures.

Seasonally, warming in Europe (in terms of decreasing cold-nights) appeared significant mainly in spring (for central north-eastern Europe) and summer (for central western Europe), while it is less significant and more uncertain in winter and autumn, respectively. A general increase in warm-nights is observed across all seasons. These findings are in line with the increase in the frequency of warm-spells across Europe already highlighted in Klein Tank (2002a). In addition, cold-spells are reported to have increased in frequency along the Mediterranean coasts, but decreased in the rest of Europe. Also, significant decreases in the annual occurrence of frost-days over parts of central western Europe are reported (up to -8 per decade).

Analysis of precipitation records has shown that Mediterranean countries have experienced an increase in consecutive dry-days and, seasonally, a decrease of maximum 5 day precipitation levels that is more discernable in summer and winter. In addition, daily precipitation intensity and the number of heavy precipitation days have increased in particular for central-eastern Europe during the whole year. The main decreases are evident in the Iberian Peninsula (Alexander et al., 2006). These results are in line with the work of Frich et al. (2002) who studied changes in extremes during the second half of the twentieth century.

[33] *Percentile-based* indices include: occurrence of cold nights, warm nights, cold days, warm days, very wet days and extremely wet days. *Absolute* indices include: maximum daily maximum and minimum temperature, minimum daily maximum and minimum temperature, maximum 1-day precipitation amount and maximum 5-day precipitation amount. *Threshold* indices include annual occurrence of frost-days, of ice-days, of summer-days, annual occurrence of tropical nights, number of heavy precipitation days > 10 mm and number of very heavy precipitation days > 20 mm. *Duration* indices include: cold and warm spell duration indicators, growing season length, consecutive dry and wet days. Other indices comprise: annual precipitation total, diurnal temperature range, simple daily intensity index, extreme temperature range and annual contribution from very wet day.

[34] Assumed at 5% level

Extending the period of study back to 1901 confirms trends which identify an increasing number of warm days and warm nights and a decreasing number of cold days and cold nights. For precipitation, the increasing trend in extreme events (amount and frequency) for central-eastern Europe, for which long term station data are available, is observed (Alexander et al. 2006).

Moberg et al. (2006) used the ECA data set plus additional national/local data sets for the same ~100-year period to study climate mean, extremes[35] and their correlations. Averaged across all the stations, all the calculated temperature indices confirm a warming in the 20th Century which is more evident in winter than summer. On average, winter has warmed more (~1.0°C/100 yr) than summer (~0.8°C), considering both daily maximum and daily minimum temperatures. The increase in daily maximum temperatures in winter was stronger at the warm end of the distribution (~1.5°C) than at the cold end (~1.0°C).

Using the 98th percentile of daily maximum temperature to represent warm days in winter, a large warming trend (often 2.5 – 4.0°C) south of 45-50°N is seen, whereas trends in the northern parts of Europe are generally smaller (1 - 2°C). The 98th percentiles of daily minimum and maximum temperatures during summer, which are used to represent warm nights and warm days respectively, reveal a significant[36] warming trend over central and western Europe. Over the Iberian Peninsula, only the trend in daily maximum temperatures (i.e. warm days) is significant.

Significant wetting trends in winter are common in central and western Europe for all indices. There is also slight evidence for larger wetting trends in total precipitation compared to the average precipitation per wet day, but unusually heavy precipitation events do not show an obvious increase. A few stations show significant drying trends, some of which are very large (40–80%), but as they are located in regions with scarce data it is not easy to assess if such drying trends are real or a result of inhomogeneous data.

[35] These include 14 temperature indices: mean daily minimum and maximum temperature, 2nd, 5th, 10th, 90th, 95th, 98th percentile of daily minimum and maximum temperatures; and 5 precipitation indices: total precipitation, simple daily intensity index, and 90th, 95th, 98th percentile of daily precipitation.
[36] Assumed at the 5% level.

All the precipitation indices have a high degree of variability. A noteworthy point for summer precipitation data is that while north-eastern and south-eastern Europe have positive trends (+10 to 15% increases), data for Scandinavia, Germany and the Greater Alpine region display no or slight negative trends. In Iberia, trends in the highest percentiles of precipitation in the summer are too small to give a useful measure in percentage terms. In the winter, positive trends are seen for all the regions except north-eastern Europe, where trends are close to zero or slightly negative. Some of these results are different to those of Alexander et al. (2006), and are likely to be due to both different spatial domains (global versus Europe) and the different periods analysed, as discussed by Klein Tank and Können (2003).

Influence of large scale circulation

van der Besselaar et al. (2011) used the ECA dataset to assess how trends in temperature extremes[37] are related to large-scale circulation types. Perhaps the best known example is the influence of the North Atlantic Oscillation (NAO), which is defined using the sea-level pressure difference between Iceland and the Azores, on European temperatures. The effects of the NAO have been studied by Castro-Diez et al. (2002), Beniston and Jungo (2002) and Scaife et al. (2008).

Wibig and Glowicki (2002) examined the variability of minimum and maximum temperature, their daily range and their relationships with NAO in Poland; Saenz et al. (2001) did the same for south-eastern Europe and found that temperature variability is more influenced by east Atlantic variability than by the NAO. Rebetez (2001) analyzed the variability of two century-long daily minimum and maximum temperature series in Switzerland; from this, they found that higher NAO index values tend to be associated not only with warmer temperatures, but also with lower day-to-day variability, concluding that the temperature warming during the 20th Century has happened mainly through the loss of the coldest part of the series.

Pozo-Vazquez et al. (2001) studied the relationship between spatial and temporal modes of European winter temperature variability and the NAO during the period 1852-1997. The results indicate that quasi-periodic oscillations in the NAO do not lead to more extreme episodes, but rather that an extreme value of the oscillation is more likely to

[37] Namely the 10th and 90th percentiles of minimum, maximum and mean temperature.

persist for a few years. They found that winter temperatures in a large part of Europe do not vary in a linear manner with respect to phase and intensity of the NAO.

Turkes and Erlat (2003) examined the relationship between the NAO and Turkish rainfall, finding strong correlations during winter. Uvo (2003) did the same for the north European winter precipitation, Tomozeiu et al. (2002) for Italy, and Fowler and Kilsby (2002) for the UK. Munoz-Diaz and Rodrigo (2003) and Goodess and Jones (2002) investigated relationships between the NAO and monthly and daily rainfall, respectively, over the Iberian Peninsula.

Further to this, Lucero and Rodriguez (2002) identified the spatial patterns of decadal and bi-decadal fluctuations in annual rainfall in Europe: decadal relationships with the NAO were identified. Martin et al. (2004) studied the spring precipitation patterns in the western Mediterranean area concluding that the NAO and the east Atlantic jet explain about 50% of the total spring precipitation variability in the area.

The summertime Palmer Drought Severity Index was examined across Europe by Mares et al. (2002) who emphasised the NAO signal in interannual variability. Relying on the ECA data, Haylock and Goodess (2004) examined daily winter rainfall at 347 European stations for 1958-2000. They found that a large part of the observed trends and interannual variability in the maximum number of consecutive dry days and the number of days above the 90th percentile of wet-day amounts (defined over the period 1961-1990) could be explained by changes in large-scale mean atmospheric circulation, particularly the NAO.

Trigo et al. (2002) presented an analysis of European temperatures and precipitation and their relationship with the NAO using NCEP reanalyses. They showed that NAO-related temperature patterns are mainly controlled by the advection of heat by the anomalous mean flow. Precipitable water is shown to be strongly related to the corresponding anomaly fields of temperature, while precipitation rate appears to be controlled by the surface vorticity field and associated strength of the troposphere synoptic activity.

Finally, the relationship between the Arctic Oscillation (AO), NAO and Eurasian snow cover has been examined by Saito et al. (2004), Bednorz (2004) and Hori and Yasunari (2003). On the seasonal time scale, the winter AO is found to be significantly correlated with the preceding autumn Eurasian snow cover throughout the period observed. Consistent with this finding, snow cover variability led the AO variability on the sub-decadal time scale in the early half of the record. However, starting in the mid 1980s, the AO and snow cover vary in phase.

Results presented by van der Besselaar et al. (2011) show that variations in the European snow cover extent are likely to be affecting the cold and warm day indices in winter. There is a correlation between the decreasing trend of the snow cover extent in Europe and the increasing (decreasing) trend of the number of warm (cold) days for stations throughout Europe.

1.2 Regional analyses

Regional meteorological monitoring programmes and long term observational networks allowed the production of a wide number of works, e.g. for characteristic regions as Alps and Scandinavia, as well discussed in Hiebl et al. (2006). In the next sections, the main findings for different European regions are summarised.

Alps

According to the first complete temperature trend analyses from the HISTALP project data pool (Auer et al., 2007; Matulla et al., 2005) the Greater Alpine Region (GAR) has warmed nearly twice as much as the global average trend. Studies for Austria (Auer and Böhm 1994; Auer, et al. 2001b), Switzerland (Begert et al., 2005) and Italy (Brunetti et al., 2006) confirm this result.

Long term observations show, in particular, that after an initial warming trend in the end of 18th Century, the 19th Century showed a cooling trend. The globally observed warming during the 20th Century is also evident in the GAR data and captured by the sub-periods 1901-40 and 1971-2004, with an intermediate cooling during 1941-70. As previously noted, warming particularly affected winters (Matulla et al., 2005).

Auer and Böhm (1994) showed that the western part of Austria has experienced an increasingly warm and wet climate over the last 150 years, while in the eastern parts warm and dry conditions have become more common. Seasonally, the spatial variability is even larger.

Weber et al. (1997) showed that several mountain stations for central Europe located in Switzerland, Austria, Eastern France, Germany, Czech Republic and Slovakia present a small change in the diurnal temperature range during the 1901-1990 period, while the low-lying stations in the western part of the Alps showed a strong decrease (stronger than in the eastern part), driven by an increase in the minimum temperature that is also confirmed in the shorter period 1951-1990.

Analysing extremes, Frei and Schär (2001) examined seasonal trends of heavy daily precipitation at 113 rain gauge stations in Switzerland from 1901 to 1994. A statistically significant increase in frequency was found in winter and autumn for a high number of stations.

In the Alps and their surroundings, there are a further 90 years of instrumental temperature information prior to 1850. This extra data changes the results of trend analysis because the early period is not characterized by low temperatures, but shows high temperatures especially in spring and summer. Auer et al. (2001a) and Böhm et al. (2001) have used this data series, known as the ALP-CLIM series, to compare the climate of the "pre-industrial" period (prior to 1850) with the warmer (anthropogenically influenced) 20th century climate.

Annual mean temperature data recorded in different parts of the Alps are shown in Figure 3. These data show that there were two relative maxima in the 1790s and the 1820s, interrupted by a sudden cold event in the 1810s. After the 1820s there is a gradual trend towards lower temperatures in the 1840s and 1850s and the 1880s and 1890s with a relative maximum in the 1860s and early 1870s. The whole 20th Century is characterized by rising temperature towards a first maximum near 1950 and a second in the 1990s, which is the main maximum of the 240-year ALP-CLIM series. When compared to the evolution of global mean temperature (black curve), the temperature increase in the Alps since the late 19th Century appears significantly stronger.

The altitude of the zero degree level has increased by approximately 250 m since the late 19th Century. The increase has not been steady due to a number of superimposed decadal scale variations. The earlier parts of the series show again higher altitudes - making the early 19th Century comparable to the recent values especially in the warm season, less for autumn and the annual mean.

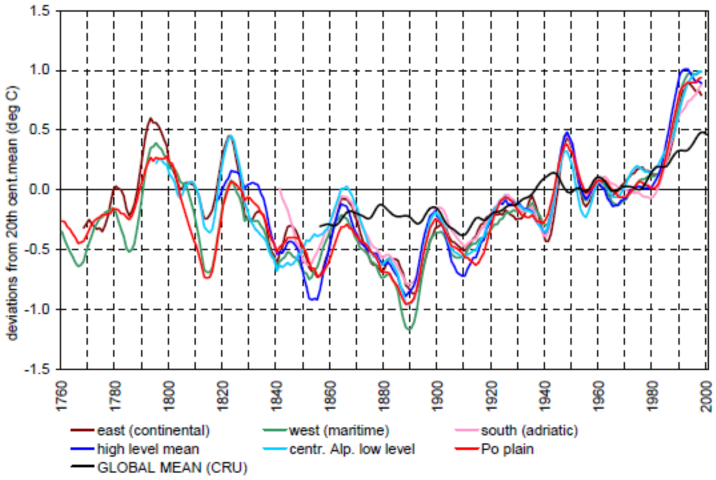

Figure 3. Temperature trends for the Alps from the ALP-CLIM series.

According to the same study of Auer et al. (2001a), precipitation variability shows a number of different evolutions across season and sub-regions, as shown in Figure 4. The Alpine chain acts as a sharp climatic divide between long-term generally increasing precipitation west and north of the Alps versus a general decreasing trend south and east of the Alps at least during the last 120 years.

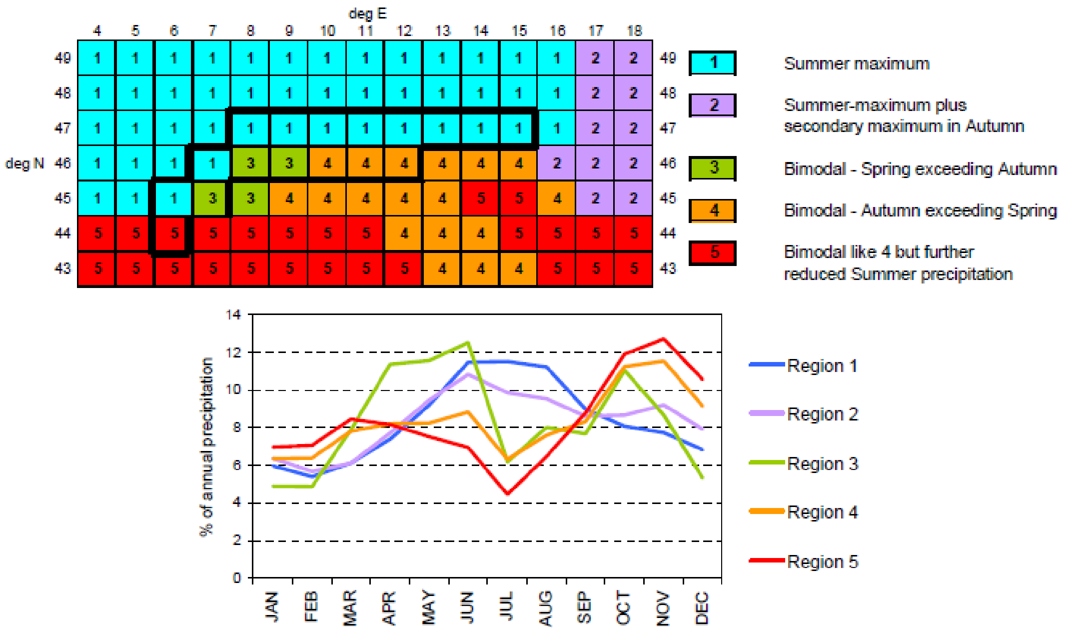

Figure 4. Monthly and regional temperature trends for the Alps from the ALP-CLIM series. Top panel – spatial patterns of monthly precipitation maxima. Bottom panel – monthly cycle of precipitation in each region of the top panel.

The stronger sub-regional differences in precipitation than in temperature are also confirmed by Auer et al. (2001b). In terms of temperature, the summer maximum curves have slightly larger amplitudes on the decadal scale than summer minimum curves and a similar but weaker effect is given for the winter minimum vs. the winter maximum curves.

The western region of the Alps shows a long-term increase of precipitation in summer and in winter. In the north there is a high precipitation level prior to the 1860s in summer, a weaker maximum in winter and a strong drying towards the 1860s, which are the driest summer years in all four sub-regions. After the mid-19th Century minimum, the western and northern regions are characterized by a long-term increase of annual precipitation totals which is mainly caused by winter precipitation. The more continental sub-regions, eastern and southern, also start with low precipitation amounts in the 1860s, and quickly increase in the subsequent year, and remain at a high level for more than five decades followed by a drying trend until the 1980s.

Concerning other climate variables, an increasing trend of bright sunshine hours was observed for high-elevation stations, while the second half of the last century is especially characterized by a reduction of sunshine at low elevation sites. As for cloudiness, the observations show a long-term increasing trend of the annual means in the west, but with no trend in the north, and decreasing cloudiness in the continental sub-regions of the east and south.

Matulla et al. (2005) reported analysis from the 250-year series for regionalizing trends in temperature in the greater alpine region, finding that:

- temperature variability is very similar over the entire study region, which is likely to be representative of an even larger region, for example Central Europe;
- temperature variability shows considerable differences on the monthly and seasonal scale;

Precipitation analysis also presented in Matulla et al. (2005) permits the reconstruction of more complex patterns

- 1801-1850 was remarkably wet; this is particularly true for summer and autumn, but not for spring;

- during 1831-1900, extraordinarily dry winters occurred;
- the decade 1861-1870 was very dry all year round;
- the decade 1881-1890 was the driest winter decade;
- the period 1910-1924 was wet in winter and all year round;
- the period 1941-1950 was exceptionally dry in spring (in all regions);
- autumn during 1991-2000 was remarkably wet (general trend within GAR).

van der Schrier et al. (2007) noted the periods 1850-1870 and 1940-1950 as persistent and exceptionally dry periods, whereas the first two decades of the 19th Century and the 1910s were exceptionally wet. The combination of temperature and precipitation also plays a key role for glaciers. The same work of Matulla et al. (2005) and literature therein provide a means for delineating a series of periods:

- the cool summers of the 1810s, when many alpine glaciers strongly advanced;
- the all year round warm period 1856-1873, when alpine glaciers strongly retreated;
- the period of 1887-1895, which shows cool winters and summers. This period contained a comparable large fraction of stationary and advancing glaciers;
- the maritime period 1910–1924 (mild winters cool summers) coincides with the 1920s glaciers moraines;
- the combination of cool summers and warm winters caused up to 70% of the alpine glaciers to advance;
- the continental 1940s coincide with a period of glacier retreat. High summer temperatures cause negative mass balances during this period;
- the all year cool period 1956–1985 most significant in summer: even somewhat more than 70% of the Alpine glaciers advanced;
- the year round warm period that started 1990, with a very pronounced glacial retreat.

In summary, from 1850-1975 the alpine glaciers lost about half of their volume; between 1975 and 2000 further 25% of ice volume disappeared and 10-15% only since then (Haeberli et al., 2006): the mean loss rate of alpine glaciers since 1980 has doubled compared to the whole 20th Century's mean (Haeberli and Holzhauser, 2003). Zemp et al. (2006) reported how Alpine glaciers lost 35% of their total area from 1850 until the 1970s, and almost 50% by 2000. Total glacier volume around 1850 is estimated at some

200 km^3, but is now close to one-third of this value. Model simulations suggest that a 3°C and 5°C warming of summer air temperature would reduce the currently existing alpine glacier cover by 80 and 100%, respectively. Annual precipitation changes of ± 20% would modify such estimated percentages of remaining ice by a factor of less than two.

Very long (500 years) time series of precipitation and temperatures for the Alps were investigated by Casty et al. (2005). Annual, winter and summer Alpine temperatures indicate a transition from cold conditions prior to 1900 to present day warmth, especially after 1970. Very harsh winters occurred at the turn of the 17th Century. Warm summers were recorded around 1550, during the second half of the 18th Century and towards the end of the 20th Century. The years 1994, 2000, 2002, and particularly 2003 were the warmest since 1500 (Luterbacher et al., 2004). Unlike temperature, precipitation variation over the Alps showed no significant low-frequency trend and increased uncertainty back to 1500. The years 1540, 1921 and 2003 were very likely the driest in the context of the last 500 years.

Recalling what was written in section 1.1, the winter NAO correlates positively with Alpine temperatures and negatively with precipitation. These correlations, however, are temporally unstable. One can conclude that the Alps are situated in a band of varying influence of the NAO (e.g. Schmidli et al., 2002; Quadrelli et al., 2001), and that other atmospheric circulation modes controlled Alpine temperature and precipitation variability during the recent past (e.g. Efthymiadis et al., 2007).

Scandinavia

The ALP-IMP project also enabled an examination of changes in the climate of the Scandinavia region (Hiebl, 2006); that is, stations in Sweden, Norway, Denmark, Lithuania, Estonia, Finland, Poland, and the Russian Federation. An investigation of 68 temperature series distributed across Scandinavia (Tuomenvirta et al., 2001) revealed a positive trend in temperature during 20th century at most stations. Moberg et al. (2005) quantified this warming to be 0.64°C between the two periods 1891-1920 and 1971-2000, based on time series from six selected station. In terms of temporal development, a first warming from the 1850s to the 1930s was followed by a slight 30 year cooling, after which the current warming phase started.

Seasonally, springs have warmed the most (0.94°C) while autumns have warmed the least (0.44°C); the high temperatures of the 1990s arose primarily from warm winters in this decade. Moberg and Alexandersson (1997) homogenised temperature data recorded in Sweden into six regions and found great spatial similarity. The relatively small temperature difference between 1861-1890 and 1965-1994 was obtained for the north-eastern region (0.51°C), while the largest difference was found for the south-western region (0.74°C) of Sweden. The long time series recorded at Uppsala and Stockholm in Southern Sweden provide data with which to investigate temperature variability as far back as 1722 and 1756, respectively. From these data, Moberg and Bergström (1997) examined temperature variability on different time scales. Looking at ten-year periods in the Uppsala series, 1862–1871 showed the largest downward anomaly, while 1730–1739 showed the largest upward anomaly. In addition, the harshest winter and coolest summer seasons occurred in the nineteenth century, while the mildest winters and the hottest summers occurred at the beginning of the series. In contrast to the greater Alpine region, no large change in the annual number of hot days in Uppsala was seen (Moberg et al., 2005). Nothing exceptional was found for the 30 years of warming until 1995 that did not significantly differ from the climate two centuries earlier (Moberg and Bergström, 1997).

Concerning precipitation variability in Scandinavia, Tuomenvirta et al. (2001) detected a positive trend in annual precipitation sums that is at least significant for the western parts of Scandinavia. Although the behaviour of individual glaciers depends on summer temperature, exposition or slope etc., it is this precipitation intensification that accounts for the general equilibrium state or even advance that has been typical for maritime western Scandinavian glaciers during the 1980s and 1990s. Even there, a trend reversal has taken place during the last years (Haeberli et al., 2005). On a somewhat longer time scale, maximum glacial extension was reached centuries ago even in Scandinavia (Nesje and Dahl, 2000).

Central/Southern Europe

Monthly temperature reconstructions for central Europe since AD 1500 have been developed using documentary index series from Germany, Switzerland and the Czech Republic covering 1500-1854 and instrumental temperature records for 1760–2007. Limited data from Poland and the Carpathian basin were also included (Dobrovolný et al., 2010).

The recent warming expressed in the new reconstruction, is most pronounced in annual mean temperatures, but is also very clear in winter, spring and summer, but somewhat less so in autumn. For annual mean temperatures, the last 20 years (1988–2007) stand out as the warmest 20-year period. The new reconstruction also displays a previously unobserved long-term decrease in winter, spring and summer temperature variability over last five centuries.

Changes in maximum and minimum daily temperatures in nine regions of Central Europe (Germany, Poland, Czech Republic, Switzerland, Austria, Slovakia, Slovenia, Hungary and Croatia) and Bulgaria during 1951-1990 were investigated by Brázdil et al. (1996). During the period 1951-1990, the increase in the annual maximum daily temperatures in Central Europe were slightly lower than that of minimum daily temperatures (0.52°C and 0.60°C, respectively). This results in a small decrease in the daily temperature range by -0.08°C.

Domonkos et al. (2003) studied the variability in frequency of winter extreme low-temperature events and summer extreme high-temperature events using daily temperature series (1901–1998) from 11 sites in Central (Poland, Czech Republic, Hungary) and Southern Europe (Italy, Serbia, Croatia).

A slight warming is detected in almost all of the seasonal mean temperature series investigated. However, the rates of the warming differ considerably. For example, there is a mean rate of change of more than 1.5°C / century for the whole 98 year period and more than 4°C / 100 years for the last 50 years in the winter mean temperature series at the northern sites. In contrast, a slight decrease is detected in the summer mean temperatures at the Hungarian station. The warming during the 20th century was more pronounced in the northern sites than in other parts of the study area; it was slightly higher in winter than in summer and usually higher in the last 50 years than during the first half of the century.

The relatively high warming rates detected in Krakow and Prague are caused partly by the increasing urban heat island effect, although real climatic differences are likely to play a substantial role in the observed spatial variability as well. Several strong statistical connections were found between the frequencies of extreme temperature events and those of certain circulation classes. Such connections are usually stronger for extreme cold temperatures than extreme warm temperatures. The mean residence time of individual circulation types has a positive correlation with the frequency of extreme warm temperatures and a negative correlation with the frequency of extreme cold temperatures.

Moving to the southernmost part of Europe, Xoplaki et al. (2003a) examined the inter-annual and decadal variability of summer air temperature over the Mediterranean area for the period 1950 to 1999. They show that more than 50% of such variability can be explained by three large-scale predictor fields: 300 hPa geopotential height, 700-1000 hPa thickness and sea surface temperatures. Similar results are reported from 24 stations in the north-eastern Mediterranean (Xoplaki et al., 2003b).

Alpert et al. (2002) analyzed daily rainfall over the Mediterranean to determine the changes in rainfall intensity categories for 1951-1995. They found increases in extreme rainfall in Italy and Spain but no change in Israel and Cyprus.

1.3 (Sub-)National level

Nationally and locally, results presented by Brunetti et al. (2006) highlight a positive trend for mean temperature of about 1°C/century over Italy; generally higher for the minimum temperature than for the maximum temperature. However, the progressive application of trend analysis shows that, in the last 50 years, the behaviour is the opposite, with the maximum temperature trend being stronger than the minimum temperature trend.

Kysely (2002), Jungo and Beniston (2001), Brabson and Palutikof (2002) and Garcia et al. (2002) examined daily temperatures at Prague-Klementinum, Switzerland, central England and Madrid, respectively, to determine relationships between extremes in temperatures and circulation patterns.

In terms of precipitation extremes, Fowler and Kilsby (2003a,b) examined annual maxima of 1, 2, 5 and 10-day rainfall amounts for 1961-2000 in the UK. Significant decadal-level changes are seen in 5 and 10-day events in many regions. Osborn and Hulme (2002) examined daily precipitation in the UK over the same period. Their findings showed that precipitation has become generally more intense in winter and less intense in summer. Recent increases in total winter precipitation are shown to be mainly due to an increase in the amount of precipitation on wet days, with a smaller contribution in the western UK from a trend towards more wet days.

Brunetti et al. (2002; 2004; 2006) studied daily rainfall over Italy to determine changes in the longest dry period, the proportion of dry days and the greatest 5-day rainfall totals. They found that precipitation shows a decreasing tendency, even if low and rarely significant, the negative trend being only 5%/century. There has been a large increase in longer dry spells and summer droughts accompanying warming over all of Italy but there is no significant trend in the extreme rainfall intensity.

Crisci et al. (2002) examined daily rainfall in Tuscany, concluding that extreme episodes increased for many different durations (1, 3, 6, 12 and 24 hours). A 100-year daily rainfall record for Uccle (Belgium) was examined by Vaes et al. (2002), who found no significant trend in extreme rainfall.

2. Barriers and knowledge gaps

In general, meteorological data are available at suitable temporal resolutions to permit reconstructions of climate trends over the preceding centuries. However, despite the spatial network of stations becoming denser over time, there remain gaps to be filled; for example, when stations do not function for given time intervals. Indeed, one of the main limitations found in analysing the available literature was the heterogeneity of source data both within a given study and across studies, making the results very difficult to compare. In addition, the number of representative stations for the different landscape types (e.g. mountains vs. lowlands, coastal vs. inland area, forestland vs. agricultural lands) is often not comparable. Thus, when the data recorded by these stations are interpolated onto a regular grid, there are necessarily different degrees of uncertainty in different regions. As a result, meteorological data are highly affected by the spatial scale of evaluation, as well as by the number, distribution of stations, and how accurately they capture the data. It is also important to note that the time scale (e.g. annual, seasonal) of analyses and the time frame duration (very long-, long-, medium-term) also yield different responses.

3. Summary

Analyzing the evolution of climate in Europe at continental to sub-regional scale has highlighted that, although prior periods of warming occurred in the last five centuries, the warming observed in the last century mainly accelerated during the last few decades. This effect has been more noticeable for minimum rather than maximum temperatures, in winter rather than in summer, and for southern Europe rather than in the north. The warming of the mean climate is also reflected in the frequency of extremes in temperatures.

Trends in precipitation across Europe tend to be much more variable, in time and space, than for temperature, resulting in a wetting in the north and a drying in the south. This is illustrated by the increased frequency of wet and dry extremes, respectively, even if the rainfall intensity of rainy days is uniformly increasing.

Continental trends are generally reflected at regional scales. However, at these scales, climate variability is influenced by the interactions between large scale processes such as atmospheric circulation and small scale features, such as landforms and topography. This explains why the Alps act as a barrier between warmer-wetter conditions in the north-west of Europe and warmer-drier conditions in the south east.

Differences between the studies analysed in this report suggest that the results are highly affected by the spatial scale of evaluation, as well as by the number and distribution of stations used, *and* how accurately they capture the weather data. Moreover, the time-scale (e.g. annual, seasonal) of analyses and the time frame duration (very long-, long-, medium-term) also yield different responses.

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ANNEX 1: Demography data

Demographic and economic tables can be compiled from data provided directly by the national statistical offices to EUROSTAT[38], in particular for the following variables: (i) total population (ii) population by gender and age, and (iii) population projections.

Specifically for demographic data, they are available at NUTS3 level as already used in the ESPON-CLIMATE[39] project. Population data in each country are available at municipal level (NUTS4 - NUTS5), while more detailed data (e.g. rural vs. urban areas) are available for a few countries only. The others among above listed socio-economic variables, at the European level are only available by NUTS0 (EU Member State national level); while NUTS2 (Regions) or NUTS3 (Province) administrative units can be extracted from national statistics and national censuses. However, not all the EU Member States follow the same methods for collecting the statistical information, especially those countries that have joined the EU more recently. Concerning long term demographic projections (e.g. up to 2100), comparable with the time frame of climate scenarios, useful information are expected from the DEMIFER[40] project.

The Joint Research Centre (JRC) in collaboration with the EEA calculated the population density disaggregated in connection with the CORINE land cover classes for the year 2000 (while the future goal is to make the same for CORINE 2006). This methodology provides approaches to combine municipal population with land cover data to produce an EU-wide population density grid, where each 100 m x 100 m pixel value is the estimated density of inhabitant per km^2 (Gallego, 2010). Furthermore, 2.5 arc-minute (a resolution of about 5 km) population and population density data (including projections up to 2015) are available at the Socioeconomic Data and Applications Center (SEDAC)[41], where also preliminary (alpha release) data from the Global Rural Urban Population Mapping project (GRUMP) (CIESIN *et al.*, 2004) are available at 30 arc-second (about 1 km) resolution.

[38] http://epp.eurostat.ec.europa.eu/portal/page/portal/statistics/search_database (last access: 29.12.2011)
[39] http://www.espon.eu/export/sites/default/Documents/Projects/AppliedResearch/CLIMATE/ESPON_CLIMATE_revised_interim_report_22-03-2010.pdf (last access: 29.12.2011)
[40] http://www.espon.eu/main/Menu_Projects/Menu_AppliedResearch/demifer.html (last access: 29.12.2011)
[41] http://sedac.ciesin.columbia.edu/gpw/global.jsp (last access: 29.12.2011)

ANNEX 2: Climate-related data sets

Besides (sub)national networks of meteorological stations (e.g. aiming at urban, river basin, sea monitoring etc.), whose full coverage is not so easy to assess, several large to medium scale databases have been built that merged, and in some case harmonized, the format of data from different sources. A good inventory of available datasets and access rules for essential climate variables is stored at:

http://gosic.org/ios/MATRICES/ECV/ecv-matrix.htm

The most significant databases are below recalled.

Focus on temperature and precipitation

Site level

The most comprehensive example are data from the U.S. National Climatic Data Center (NCDC[42]) which, among other activities, collect surface data daily from meteorological stations around the world (Surface Data - Global Summary Of The Day) making these freely available in user-friendly formats. Primary variables are temperature, pressure, precipitation, visibility and wind.

Additional download via FTP is possible for hourly data from 1901 to today in function of required location (ftp://ftp3.ncdc.noaa.gov/pub/data/noaa/), as well as GHCN (Global Historical Climatology Network)-Daily is an integrated database of daily climate summaries from land surface stations across the globe. GHCN-Daily now contains records on maximum and minimum temperature, total daily precipitation, snowfall, and snow depth; however, about two thirds of the stations report precipitation only. Both the record length and period of record vary by station and cover intervals ranging from less than year to more than 175 years (http://www.ncdc.noaa.gov/oa/climate/ghcn-daily/index.php?name=data).

Also, GHCN-Monthly[43] database contains historical temperature, precipitation, and pressure data for thousands of land stations worldwide. The period of record varies from station to station, with several thousand records extending back to 1950 and several

[42] http://www.ncdc.noaa.gov/oa/mpp/freedata.html
[43] http://www.ncdc.noaa.gov/ghcnm/v2.php

hundred being updated monthly. First in May 2011, then in November 2011, versions 3 and 3.1 respectively replaced version 2 as the dataset for operational climate monitoring activities (http://www.ncdc.noaa.gov/ghcnm/v3.php), including calculation of global land surface temperature anomalies and trends. GHCN-Monthly is used operationally by NCDC to monitor long-term trends in temperature and precipitation, and it has also been employed in several international climate assessments, including the IPCC-AR4.

Also starting from above sources, the World Monthly Surface Station Climatology from 1738[44] to present (Quayle, 1989) has data for over 4,700 different stations (2,600 in more recent years). Among land, marine and roving ship data, Figure A1 shows the land station spatial coverage, where each dot represents a 3° box containing one or more observations. Among others, standard variables for land points are geopotential height, humidity, surface pressure, precipitation and temperature.

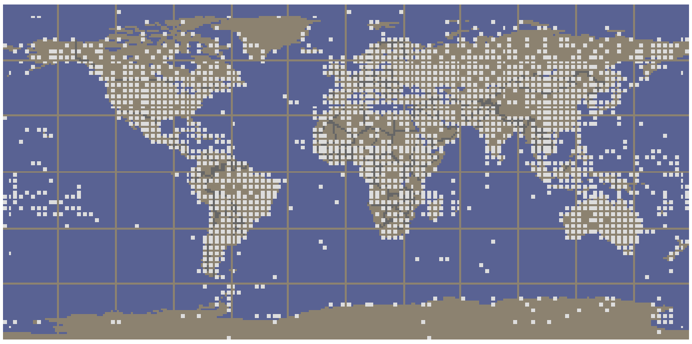

Figure A1: Spatial coverage of world Monthly Surface station Climatology

Representative of a more complete work on temperature, the Berkeley Earth Surface Temperature Study has created a preliminary merged data set by combining 1.6 billion temperature reports from 15 preexisting data archives. More information may be found at: http://berkeleyearth.org/data/

After eliminating duplicate records, the current archive contains 39,390 unique stations. This is more than five times the 7,280 stations found in the GHCN-Monthly that has served as the focus of many climate studies.

[44] U.S. National Climatic Data Center, Harvard University [Harvard College Observatory], Florida State University [Meteorology Department], and U.S. National Center for Atmospheric Research [CGD Climate Analysis Section], updated yearly: World Monthly Surface Station Climatology, 1738-cont. *Dataset ds570.0 published by the CISL Data Support Section at the National Center for Atmospheric Research, Boulder, CO, available online at* http://dss.ucar.edu/datasets/ds570.0/.

Still in terms of air temperature, multi-layer radiosonde network-based data are available at University of Wyoming[45], NOAA (RATPAC[46]), and University of New South Wales[47]. In some cases, data have been averaged into global regions (Sterin, 2007; Angell, 2011).

At the European Level, the European Climate Assessment and Dataset project (ECAandD; http://eca.knmi.nl/) presents information on changes in weather and climate extremes, as well as the daily dataset needed to monitor and analyze these extremes. Today, ECAandD receives data from 58 participants for 62 countries and the dataset contains 26109 series of observations for 12 elements at 4824 meteorological stations throughout Europe and the Mediterranean. About 48% of these series is public, which means downloadable from this website for non-commercial research.

Such a dataset is being integrated by/into other projects or programmes: e.g. climatology of severe weather events (METEOALARM, ETCCDI); global temperature dataset (International Surface Temperature Initiative); integration with satellite/reanalysis data (EURO4M, ACRE); climate predictions (ENSEMBLES); various temporal aggregation of precipitation and temperature (MILLENNIUM, EEA); records for the Mediterranean Region (MEDARE). This dataset has been then interpolated into various gridded formats as explained in the following pages.

Regionally inside Europe, the HISTALP project (http://www.zamg.ac.at/histalp) provides a database consisting of monthly homogenized temperature, pressure, precipitation, sunshine and cloudiness records for the Greater Alpine Region (Figure A2). The longest temperature and air pressure series' extend back to 1760, precipitation to 1800, cloudiness to the 1840s and sunshine to the 1880s.

Spatial averages for 4 variables (temperature, sunshine, precipitation and cloudiness) are also available for 5 sub-regions as shown in Figure A2, as well as interpolated datasets at different grid resolutions, described in the following section. This database, jointly with other records in the same region, is also included in the ALP-IMP[48] project.

[45] http://weather.uwyo.edu/upperair/sounding.html
[46] http://www.ncdc.noaa.gov/oa/climate/ratpac/index.php
[47] http://www.ccrc.unsw.edu.au/staff/profiles/sherwood/radproj/index.html
[48] http://www.zamg.ac.at/ALP-IMP/index.htm

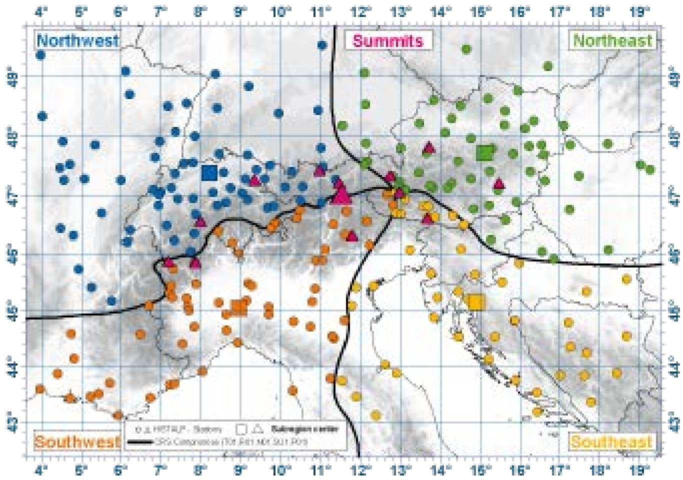

Figure A2: Meteorological stations used in the HISTALP project

Finally, a WMO database on world weather and climate extremes[49] aims at archiving and verifying extreme record events, such as the highest/lowest recorded temperatures and pressures on the Earth, the strongest winds, the greatest precipitation, as well as records involving the world's most destructive storms, hurricanes and tornadoes.

Grid level

Interpolated long-term historical global datasets on climate often have the drawback of a very coarse spatial resolution in order to reduce computation time and storage requirements.

Looking for the highest resolutions, past/recent global climate variability can be examined using coarse scale (0.5° resolution) monthly dataset resolution produce at Climate Research Unit (CRU[50]; Mitchell and Jones, 2005), whose version TS3.1 store data from 1901 to 2009. Monthly variables available are:

- cloud cover percentage;
- diurnal temperature range;
- frost day frequency;
- precipitation;
- daily mean temperature;

[49] http://wmo.asu.edu/#continental
[50] http://badc.nerc.ac.uk/data/cru/

- monthly average daily minimum;
- monthly average daily maximum;
- vapour pressure;
- wet day frequency.

CRU TS3.1 also served as input for producing a set of further analyses.

To drive land surface and vegetation models, several studies created historical decadal time series of forcing data (mainly precipitation, temperature, humidity, and radiation, at daily to several-hourly timescales). Most of these products have been based on reanalysis data corrected against globally available observations. For example, Ngo-Duc et al. (2005) created 53 years (1948-2000) of 6-hourly forcing data from NCEP/NCAR reanalysis data with correction of precipitation and radiation using the CRU data, and scaled the monthly mean long-wave and short-wave radiations to fit those of the Surface Radiation Budget (SRB) project. Sheffield et al. (2006) and Qian et al. (2006) followed similar frameworks. The datasets created by Sheffield et al. (2006) can be obtained from: (http://hydrology.princeton.edu/data.pgf.php). These authors used the Global Precipitation Climatology Project (GPCP) and TRMM mission (see below) for statistical downscaling to 1°, while other meteorological variables (downward short- and long-wave, specific humidity, surface air pressure and wind speed) were downscaled in space with account for changes in elevation. Berg et al. (2005) obtained 15-year (1979-1993) 6-hourly forcing data from ECMWF reanalysis data by scaling temperature, dew point temperature, precipitation, and long- and short-wave radiations to the monthly observations of those variables.

Although the above studies scaled the variables based on monthly observations, atmospheric forcings based on reanalysis products still contain some specific biases at shorter timescales such as daily precipitation intensity and number of precipitation days. Therefore, creating forcing data sets in which the daily statistics are similar to those of observations is important.

Hirabayashi et al. (2005) estimated first 1° × 1° global 100-year (1901-2000) atmospheric forcing data extended the data period with respect to Nijssen et al. (2001) by using a stochastic weather generator to statistically create daily atmospheric forcing from monthly precipitation and temperature observations in CRU, applying statistical parameters derived from available daily or 3-hourly observations. More recently, Hirabayashi et al. (2008) created a 59-year (1948-2006) near-surface meteorological data set with daily to 3-hourly timescales and at 0.5° spatial resolution. Valuable global datasets used (all at 0.5°) were:

- Monthly precipitation: PREC/L (Chen et al., 2002);
- Daily precipitation and monthly rainy days: GTS grid product by NOAA and CRU;
- Monthly temperature: GHCN/CAMS, Fan and van den Dool, 2008;
- Monthly min-max temperatures: GTS grid product by NOAA and CRU;
- Daily max. and min. temperature: GTS grid product by NOAA;
- Daily shortwave radiation: NASA Langley SRB (Gupta et al. 2006).

GHCN-M, jointly with the FAOCLIM (FAO, 2001[51]) and regional databases, was also the support dataset for the WORLDCLIM[52] (Hijmans et al., 2005) database at 30' (~1km) gridded resolution for the 1950-2000 period.

In case of precipitation based datasets, the Global Precipitation Climatology Centre (GPCC[53]) operated by the German Meteorological Institute under the auspices of the WMO, released gridded products (at 2.5, 1, 0.5 and 0.25°) on precipitations, focusing on the period 1951-2000 and consisting of data from ca. 64'400 stations, and related re-analyses, anomalies, etc.

Temperature and precipitation interpolated fields are also available in terms of anomalies. The former e.g. in the HadCRUT3[54] (land and sea) and CRUTEM3[55] (land) datasets, each since 1850, or in the GSTA[56] and GISTEMP[57] datasets, each since 1880. Furthermore, the Japan Meteorological Agency (JMA) processed monthly and annual anomalies, from 1891 to the present day, averaged in 5° x 5° grid boxes, for the global surface temperature[58]. The land part of the combined data for the period before 2000 consists of GHCN-M above illustrated, while that for the period after 2001 consists of CLIMAT reports archived at JMA from National Meteorological and Hydrological Systems through the Global Telecommunication System (GTS) of the WMO.

[51] FAOCLIM2 is now available at http://www.fao.org/nr/climpag/pub/EN1102_en.asp
[52] http://www.worldclim.org/
[53] http://www.dwd.de/bvbw/appmanager/bvbw/dwdwwwDesktop?_nfpb=trueand_pageLabel=_dwdwww_klima_umwelt_datenzentren_wznandactivePage=and_nfls=false
[54] http://www.metoffice.gov.uk/hadobs/hadcrut3/
[55] http://www.metoffice.gov.uk/hadobs/crutem3/
[56] http://www.ncdc.noaa.gov/cmb-faq/anomalies.php
[57] http://data.giss.nasa.gov/gistemp/
[58] http://ds.data.jma.go.jp/tcc/tcc/products/gwp/temp/ann_wld.html

In the case of precipitation anomalies, the period 1951-1979 was taken as a reference point in the work of Dai et al. (1997), recently extended to cover the period 1850-2008 at 2.5° resolution, and relying on data compiled by Escheid et al. (1991).

Satellite acquisitions are a good source of spatialized weather data, even if limited to the time period specific satellite missions became available. The historical precipitation data could be completed by the recent information gathered by the NASA satellite Tropical Rainfall Measurement Mission (TRMM[59]), providing spatially distributed and accurate rainfall information for the portion of the globe between -50 and 50° N lat, at 0.25° resolution and every 3 hours from 1998 to today.

Specifically for air temperature, many radiosonde network-based gridded products are recently available, both in terms of absolute values or anomalies: RAOBCORE[60], HadAT[61].

MSU (Microwave Sounding Unit; started on 1978) and its successor AMSU (Advanced MSU; started on 1998), using microwave spectrum, allowed reconstructing originally temperature only and then precipitation (amount and rate) fields, gradually improving both spatial (from 110 to 16 km) and spectral (from 4 to 20 channels) resolutions (http://amsu.cira.colostate.edu/; http://www.ssmi.com/msu/msu_data_description.html).

The Second Global Soil Wetness Project (GSWP-2[62]) produced a complete set of 3-hourly near-surface meteorological data sets based on the re-analyses and gridded observational data, for the period 1982-1995 at 1° resolution, useful to land surface schemes.

At Pan-European Level, E-OBS version 5.0, based on ECAandD information above described, has been released (http://eca.knmi.nl/download/ensembles/ensembles.php) in form of daily gridded observational dataset for precipitation, temperature and sea level pressure in Europe (Haylock et al., 2008) at 25 km resolution. Actually, the full dataset covers the period 1950 until mid 2011. It has originally been developed and is maintained as part of the ENSEMBLES and EURO4M projects, respectively.

[59] http://trmm.gsfc.nasa.gov/data_dir/data.html
[60] http://www.univie.ac.at/theoret-met/research/raobcore/
[61] http://www.metoffice.gov.uk/hadobs/hadat/
[62] http://www.iges.org/gswp/

Additionally for European regions, one of the main outputs of the Crop Growth Monitoring System (CGMS) developed by the MARS Project are the meteorological interpolated data (http://mars.jrc.ec.europa.eu/mars/About-us/AGRI4CAST/Data-distribution) from 1975 to today and at 25 km spatial resolution, covering the domain illustrated in Figure A3. Treated variables are:

- maximum temperature and minimum temperature;
- mean daily vapour pressure;
- mean daily wind speed at 10m;
- mean daily rainfall;
- Penman potential evaporation from a free water surface;
- Penman potential evaporation from a moist bare soil surface;
- Penman potential transpiration from a crop canopy;
- daily global radiation;
- snow depth (data with no quality check).

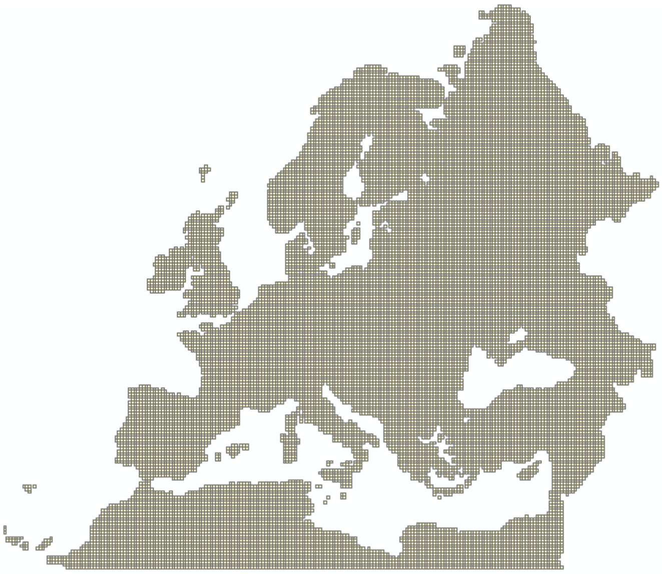

Figure A3: Domain of the AGRI4CAST meteorological dataset

The same HISTALP project previously mentioned produces gridded datasets from station data at 1°, 5' and 1 km (in progress) grid resolution. Data are from about year 1780 for temperature and 1801 for precipitation. (http://www.zamg.ac.at/histalp/content/view/18/36/index.html).

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Task A2_D2 and A2_D3: Historical evolution of forest cover and weather in Europe

European Commission (DG Environment)
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Contents

Executive Summary

1. Forest cover has increased over most of Europe since 1900. However, the increases are scattered over the entire continent and are unlikely to have had a significant effect on weather and climate in Europe except at the local scale.

2. The climate of Europe has warmed over the past 100 years, although the trends vary considerably between different locations. Rainfall has increased in the north, but decreased in the south. At least in north-western France, seasonal rainfall patterns have changed over the last 150 years, where summers have become drier. Forests only play a minor role in the regulation of large-scale floods, but can reduce the impact of small scale localised floods through intercepting rainfall, slowing surface runoff and enhancing the permeability of the land surface.

3. Temperature patterns are modified by forest cover – daily maximum temperatures under the forest canopy are reduced, whereas daily minimum temperatures are increased compared with open areas. The altitude, tree type and direction of the slope are also important. Modelling studies suggest that increased vegetation cover in and around urban areas can reduce the magnitude of the urban heat island effect, but a large increase in vegetation cover (at least 10% - 20%) would be needed before an effect would be seen. On a smaller scale urban trees provide shade, evaporative cooling, reduce wind speeds and increase the permeability of the urban land surface.

4. Droughts appear to have increased in severity in the Mediterranean and central Europe since the middle of the twentieth century, but any trends are dependent on the drought index used and the data analysed. It is not clear what the role of forests or any changes in forest cover are. However, modelling studies suggest that deforestation in historical times has contributed to reduced rainfall and increased aridity in the Mediterranean region. Regions experiencing decreased precipitation may also manifest an increased fire risk, with many studies projecting a lengthening of the fire season.

5. The effect of forests and changes in forest cover on the above points is unclear, as these changes in climate are caused by changes in large-scale circulation

patterns and anthropogenic warming caused by increased greenhouse gas emissions.

6. Forests are important for supplying clean water, as they can filter out some harmful chemicals and stabilise soils, reducing sediment levels. Any links between forest cover and water quantity are less clear. However, at larger scales, forests are an important component of the hydrological cycle and act to increase rainfall.

7. Forests also act to stabilise soils on sloping areas, provided the forest is well-maintained and in good health. Open areas within forests, or locations where the trees are in poor health, are more prone to land slips. On very steep slopes, the forest cover has a much smaller effect on the stability of the soils.

Task A2: Historical evolution of forest cover and weather in the EU

A separate report has been written which reviews and compares different land use and climate datasets for Europe (E. Pope, M. Sanderson, M. Santini and R. Valentini, “Literature review of the historical evolution of forest cover and weather across the EU region, and appropriate datasets”). The objectives of Task A2 presented here are:

(1) Summarise current literature that describe the historical evolution of forest cover and weather across the EU region during the last 150 years (A2_D1)
(2) Use this information to devise and apply objective criteria to assess the influence of EU forests on precipitation patterns, Storm regimes, Temperature patterns, Water quality and quantity, Soil stabilisation and Flood regulation (A2_D2 and A2_D3).

Regarding the first objective, a new land use database (HILDA) has been used to assess forest cover changes since 1900. The advantage of this dataset is that it makes use of historical records to identify the specific locations at which forest cover has changed. This is in marked contrast to other datasets (e.g. HYDE and SAGE) which make assumptions about the geographical distribution of historical human land use in order to derive changes in forest cover. For the second objective, we have used published studies to assess the influence of forests on the six weather-related phenomena listed in item (2); however, when reading this report, it is important to bear in mind the following. European forest cover has generally increased over the past 150 years. During this time, global temperatures have also increased. This does not mean that forest cover is the cause of the recent global temperature increase, nor, indeed, that global mean temperature and current European forest cover are related in any precise or fundamental way – direct human interference is likely to be a far more important consideration. Therefore, any correlation between current forest cover and other variables must be considered carefully and examined for other potential links and common causes. For example, care must be taken when investigating links between forests and precipitation patterns, and frequency and intensity of storm regimes, because these phenomena are strongly influenced by large-scale, synoptic processes. However, relationships between forests and water quality and quantity, soil stabilisation and flood regulation are generally much more straightforward to identify because they occur on local scales.

This part of the report is structured as follows. In Section 1, the land use database (HILDA) is briefly described, and the changes in forest cover from this database are presented. Next, the results of literature reviews on the changes in European weather and climate over the past 150 years are given in Section 2. Criteria to assess the influence of forests on the six items are summarised in Section 3. Section 4 illustrates the influence of forests on the six items listed in (2) above, plus crop yields and forest fires. A summary is presented in Section 5.

1. Task A2_D1: Evolution of forest cover in the European region.

There are many datasets describing land use change (including forest cover). For this report, the land use database HILDA (Historic Land Dynamics Assessment) has been used to identify when, where and by how much forest cover has changed. This dataset covers the period 1900 – 2010 and provides changes in land cover at decadal intervals on a 1 km × 1 km scale. A map showing areas where forest cover has increased or decreased using data from the HILDA is shown in Figure 1.

The green plus symbols "+" indicate where forest cover has increased, and the red circles where forest cover has decreased. Each symbol indicates where and when forest cover over an area of approximately 1 km^2 has changed. It can be seen that the forest cover changes are scattered throughout Europe, with very few instances of contiguous changes - the coastal area of Sweden being one of the few exceptions. Consequently, many of the changes in forest cover are localised and may be too small to have exerted an identifiable influence on regional weather patterns. As a result, the effects could be limited to the area in which forest cover has changed, and the immediate surroundings.

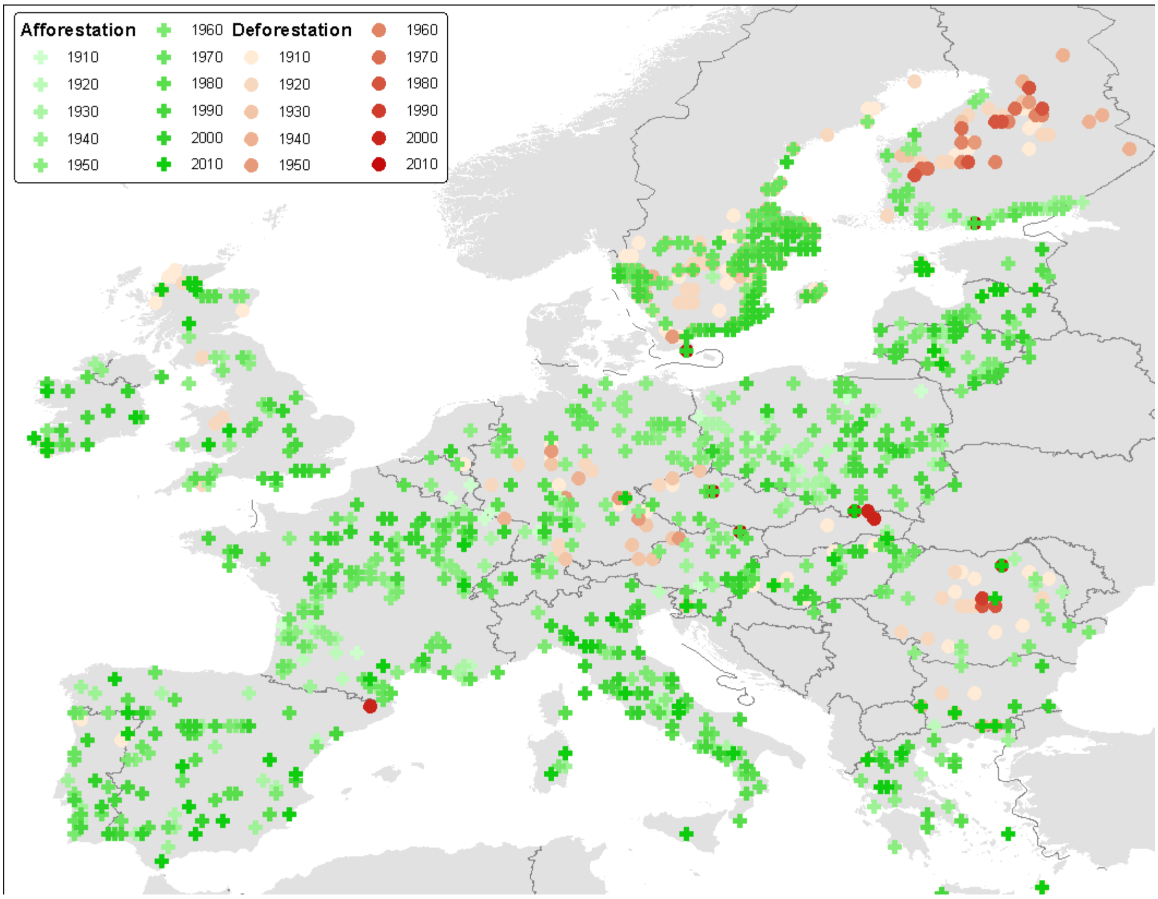

Figure 1. Afforestation and deforestation in Europe between 1910 and 2010 using the HILDA database. The symbols show the change in forest cover and when it occurred over an area of approximately 1 km^2.

2. Task A2_D1: Evolution of weather and climate in the European region.

2.1 Storminess

Possible changes in storminess in Europe since 1871 have been investigated by Donat et al. (2011). This study used the newly available Twentieth Century Reanalysis (20CR; Compo et al., 2011). The 20CR assimilates observations of surface pressure only into an atmospheric model and uses observed monthly mean sea-surface temperatures and sea-ice distributions as boundary conditions. An ensemble of 56 realizations of the reanalysis have been produced, allowing the uncertainty in any analysis to be produced (Compo et al., 2011). Donat et al. (2011) used annual 95th percentile of daily maximum wind speeds as an indicator of storminess, and found significant upward trends in the wind speeds and the number of days with gales over Scandinavia, the British Isles and North Sea and central Europe. Trends in the remaining parts of Europe were mostly not significant. These results are consistent with previous studies which analysed surface pressure data from weather stations and also diagnosed an increase in storminess (Wang et al., 2008). All these studies have noted the high levels of storminess that occurred during the late nineteenth and early twentieth centuries, and around 1990. With this in mind, it is important to note that climate model simulations often show an increase in storminess in the future, suggesting that the trends identified by Donat et al. (2011) could be caused by rising levels of greenhouse gases, although increased variability of the climate cannot be ruled out.

In contrast, a previous study by Matulla et al. (2008) which also used long-term surface pressure data from weather stations did not identify any increase in storminess over the past 130 years. However, these authors analysed annual mean changes in pressure which may have masked increases in winter storminess and decreases in summer identified by Wang et al. (2008) and Donat et al. (2011). In agreement with all the previous studies, Matulla et al. (2008) also noted considerable variability in storminess at decadal and longer time scales.

As a final example, Vautard et al. (2010) analysed observed wind speeds at many locations in Europe and found that mean wind speeds had declined since 1979, or possibly earlier at some locations. Notably, a slowing of winds has occurred in eastern Europe where forests have re-grown owing to the abandonment of agricultural land. In general, large-scale circulation changes are thought to be responsible for some of the reduced wind speeds, but Vautard et al. (2010) estimated that the increased forest growth could explain 25% – 60% of the observed wind speed reductions.

2.2 Daily mean temperature

Barbosa et al. (2011) have examined trends in daily mean air temperature from 1901 over Central Europe using quantile regression. This technique allows the estimation of trends in the mean and in all other parts of the data distribution. They used long term daily mean temperature measurements from weather stations (beginning in 1901 or earlier) which were located in and around Germany, southern Scandinavia, north-eastern France, Slovenia and Croatia. The seasonal cycle in the data was removed before analysis. The results show considerable spatial diversity in temperature trends over the central European region. Higher trends of about 0.15 °C decade^{-1} are seen at the 5% quantile and around 0.20 °C $decade^{-1}$ at the 95% quantile. The largest trends (>0.2 °C per decade) occurred in the Alps. Parker and Horton (2005) calculated a trend of 0.077 °C $decade^{-1}$ in annual mean temperatures from the central England temperature record since 1901. This trend is significant at the 1% level.

2.3 Rainfall: Trends and wet periods

Zolina et al. (2010) have analysed the duration of wet spells (consecutive days with significant precipitation) in Europe over the period 1950–2008 using daily rain gauge data. Wet periods were found to have increased in length by about 15–20% over most of Europe. However, this is not because of an increase in the total number of wet days. Furthermore, the contribution of short duration events to periods of heavy rainfall has decreased, while the contribution of longer events has increased. As such, heavy precipitation events during the last two decades have become much more frequently associated with longer wet spells and have intensified in comparison with rainfall in the1950s and 1960s.

Masson-Delmotte et al. (2005) have reconstructed temperature variations and rainfall anomalies for summer and winter for north-western France using a 400-year tree ring record. A comparison with summer temperature variations in the instrumental records highlights decades of consistent fluctuations. Warmer and drier summers occurred in the late eighteenth century and twentieth century, in contrast to cooler and wetter summers in the nineteenth century. The reconstructions highlight the summers of the period 1800 – 1850 as particularly wet and cool, while cool winter and annual mean temperatures were more typical until the end of the nineteenth century. An analysis of a long rainfall record (1680 – 2001) from Paris indicates that in the current climate rainfall occurs all year round, but during the second half of the nineteenth century, more rainfall occurred during the summer months (Slonosky, 2002).

2.4 Droughts

There have been several analyses of trends in droughts in Europe. However, different indices of drought have been used by these studies which produce different conclusions. Indeed, drought may be defined in three different ways (Dai, 2011):

(1) Meteorological drought is a period (lasting from months to years) with below average rainfall, and is often associated with above average temperatures.
(2) Agricultural (or soil moisture) drought is caused by low soil moisture levels, which can be caused by a preceding meteorological drought, intense but infrequent rainfall, or increased evaporation.
(3) Hydrological drought occurs when stream flows and groundwater levels fall below a long-term average.

There are many published indices which measure one of these types of drought, and analyses of trends in these indices can sometimes produce different results. For example, Dai (2011) measured trends in droughts using three different indices pertaining to agricultural drought, and found a consistent drying signal over southern Europe between 1950 and 2008. Similarly, van der Schrier et al. (2011) calculated an annual average drought index for the period 1901 to 2006 using two different estimates of potential evapotranspiration, and found a drying trend over southern Europe in qualitative agreement with the results of Dai (2011). Using a different approach, Stahl et al. (2010) analysed annual mean streamflow data for the period 1932-2004 and found only a weak drying signal. However, when the analysis incorporated data collected from 1952 (and 1962), a drying signal was seen across northern Spain, southern France and central Europe. An analysis of monthly streamflow data collected between 1962 and 2004 also showed a strong drying trend over most of Europe between April and August.

However, it is worth noting that strong drying trends have not been found by other studies. Alexander et al. (2006) calculated trends in consecutive dry days between 1951 and 2003 using a global precipitation database, and found only a weak drying signal over Europe. Sheffield and Wood (2008) analysed soil moisture data produced by a land surface hydrological model which was driven by meteorological forcings based on observed data, and also found only a weak drying trend over Europe.

Overall, these results indicate a drying trend over the Mediterranean and central Europe since the 1950s, but it is important to remember that some of the conclusions are dependent on the choice of drought index and data analysed.

2.5 Summary

Published studies on the evolution of forest cover, weather and climate over Europe have been summarised. Many studies only explore changes and trends from the beginning of the twentieth century or later owing to a lack of quality controlled long term records. Forest cover in Europe has changed since 1900 in many locations, but these changes are scattered piecemeal throughout the continent, an exception being southern Sweden. Consequently, any influences of European forests on weather and climate are likely to be relatively small-scale and confined to the forested areas and immediate surroundings only.

The climate of Europe has changed over the past 100 – 150 years, but these changes are due to a combination of increasing greenhouse gas emissions, causing the climate to warm, and changes in large-scale circulation patterns. The influence of forests on these changes is likely to be very small. The numbers of storms per year are highly variable, which makes an assessment of trends difficult. Nevertheless, storminess appears to have increased during the winter months in parts of northern Europe, but decreased in summer over the last 100 years. Wind speeds have also decreased in some areas where forests have grown back on land formerly used for agriculture which has subsequently been abandoned.

Daily mean temperatures and rainfall amounts have increased over Europe during the twentieth and early twenty-first centuries. Temperatures have risen over central England and central Europe, but there is considerable spatial variation in the trends over continental Europe. Rainfall has occurred in longer-lasting events over most of Europe since 1950, and the contribution of heavy rainfall events has also increased. A reconstruction of the climate of north-western France using tree ring records indicates shifts in summer and winter temperature and rainfall patterns. Warmer and drier summers occurred in the late eighteenth century and twentieth century, in contrast to cooler and wetter summers in the first half of the nineteenth century. Droughts, as measured using a variety of indices, appear to have increased in severity in the Mediterranean and central Europe since the 1950s, although some of these trends are dependent on the choice of drought index.

It is difficult to understand how changes in forest cover may have affected these trends in climate, as they have mostly been caused by changes in large-scale weather patterns. The re-growth of forests in some areas may have increased local rainfall amounts by slowing the wind speeds and increasing turbulence above the forest canopy; this effect has been shown to be important in south-western Australia where a decrease in rainfall and river flows to the city of Perth has been caused by loss of forest (Nair et al., 2011). However, the areas where forests have re-grown in Europe are small and are scattered throughout the continent. Only an increase in forest cover over a very large area could produce a detectable change in European climate.

3. Task A2_D2: Objective criteria for assessing influence of EU Forests on weather.

In this section, the criteria used to assess the influence of EU forests on the topics which will be discussed in section 3 (results of Task A2_D3) are given.

1. Wind speeds. Owing to their high aerodynamic roughness, forests cause a reduction in wind speeds in comparison to similar open areas. They can also increase turbulence in the air and increase local rainfall.
2. Forests can influence temperature patterns in two main ways. Firstly, they provide insulation, meaning that daily maximum temperatures within forested areas tend to be lower than non-forested areas, whereas daily minimum temperatures tend to be higher. Consequently, the diurnal temperature range in forested areas is smaller than non-forested areas. Secondly, forests can alter heat wave intensities and patterns. If sufficient soil moisture is available, forests can cool the surface via evapotranspiration. The aerodynamic roughness of the forest canopy also acts to increase evapotranspiration and low-level cloudiness which also acts to reduce the intensity of heat waves.
3. Forests may affect the severity of storms by changing the moisture flux from the surface; however, this link is currently unproven.

Forests act to stabilise soils on slopes and reduce the number of land slides, by mechanical stabilisation via the roots, and the forest litter which reduces the splash effect of raindrops. They are also important sources of clean water. However, the exact effects of forests in these latter cases is much harder to prove, and long term data sets are scarce or simply not available.

4. Task A2_D3: Influence of EU forests on weather, flooding and soils

4.1 Precipitation patterns

Studies of precipitation patterns in eastern Spain indicate that, during the last 30 years, there has been a significant reduction in the occurrence of summer storms over the mountains which face the western Mediterranean basin, thereby leading to inland drought (e.g. Millán, et al. 2004). It has been claimed (Millán, 2008) that the drainage of coastal marshes, increasing urbanisation and vegetation loss on the slopes of mountains in eastern Spain could be responsible for this change, since the reduced evapotranspiration rate from the land surface may no longer be sufficient to ensure that the lifted condensation level (LCL) of the Mediterranean sea breeze falls below the mountain tops. The reduction in storm occurrence has been necessarily accompanied by an accumulation of water vapour above the western Mediterranean, which can potentially influence flooding events along the Spanish coast and in central Europe. In particular, the greenhouse effect due to increased water vapour above the western Mediterranean has been posited as an explanation for increased cyclogenesis which has contributed to flooding along Spain's eastern coast. In addition, the same accumulated moisture lies on the tracks of storms that are well-known for bringing severe floods to central Europe (e.g. July 1997, August 2002). More specifically, the central European floods of 1997 and 2002 were associated with a particular storm track known as Vb[1] (Kreienkamp et al., 2010). These storms begin over the eastern Atlantic Ocean, and travel over the Iberian peninsular and the north-western Mediterranean before moving north-eastwards into central Europe. Therefore, if there is an increase in accumulated moisture above the western Mediterranean and this is incorporated into a Vb storm, there could be an increased Mediterranean contribution to floods in central Europe.

However, it is important to note that the August 2002 flood was an extreme event on a scale that has historically occurred, on average, once in 100 years. In contrast, the loss of storms over mountains in eastern Spain has been observed during the last 30 years. While the quantity of accumulated moisture above the western Mediterranean may have increased during this period and contributed to the 2002 flood, a time series of 30 years

[1] Storm tracks across Europe were first classified by van Bebber (1891) into five main groups. The Vb storm track has remained in use today owing to its links with flooding events in central Europe.

is not long enough to determine a change in the frequency of extreme events which have historically exhibited a return period greater than 30 years. Furthermore, even if there has been a statistically significant increase in the frequency and intensity of Vb weather patterns, as suggested by Frick and Kaminski (2002) (although a separate analysis by Mudelsee et al. (2004) disputes this claim), the cause is most likely to be changes in large scale atmospheric flows generated by disturbances above the Pacific, as described in the next section. Land use changes might modify the properties of the storms and possibly even the intensity, but are unlikely to affect the actual frequency. In addition, it is also important to note that the central European flood of August 2002 followed a wetter than average July and a heavy rainfall event earlier in August, meaning that the ground was saturated and river levels were already high (Grazzini and van der Grijn, 2003).

In principle, the influence of vegetation on orographic rainfall applies in other regions, not just around the Mediterranean and Black sea basins. For example, it is clear from observations that precipitation in the UK is strongly influenced by orography. Due to the prevailing wind from the Atlantic, areas east of the high ground tend to experience much lower precipitation rates than those in the west of the country. This is because the LCL of the incoming moist air moving inland from the Atlantic frequently lies below the height of the hills near the west coast. However, with a warming climate and deforestation, it is possible that the LCL of this air might exceed the hill tops. This is most likely to occur in the south west of the UK where land surface temperatures could rise by several degrees due to climate change, and the hills are no higher than several hundred metres. In particular, Dartmoor has a maximum height of ~600 metres and experiences annual precipitation of up to ~2000 mm per year. Consequently, the runoff provides a locally and regionally important source of fresh water. However, simple calculations suggest that this supply could be vulnerable to increases in both land and sea surface temperatures which could dramatically reduce local rainfall and lead to water shortages. Given the lessons learned from Spain's Mediterranean coast, it could be prudent to pre-empt the impact of changing weather patterns by investigating whether reforesting such regions can protect water supplies. However, it is important to note that detailed studies would be required to determine the likely impacts of planting forests, and whether as well as where they would be of greatest benefit.

4.2 Storm regimes

It is well known that Vb storm tracks are associated with heavy rain events and floods in central Europe. These tracks form part of the TrM Grosswetterlage (Hess and Brezowsky, 1999) which occurs when a low pressure region forms over central Europe, while being situated between regions of high pressure. Fundamentally, the characteristic sequence of high and low pressure regions that signify a Vb track are the remnants of Rossby waves generated by atmospheric disturbances above the Pacific Ocean that have travelled across North America and the Atlantic (Ambrizzi and Hoskins, 1996; Grazzini and van der Grijn, 2003; Richardson, 2005). The arrangement of high pressure to the west of Europe, and in the east, with low pressure over central Europe causes the jet stream to pass from the Atlantic across northern Spain/southern France over the western Mediterranean, then across northern Italy, before passing over the Adriatic and then back into central Europe. Such weather patterns typically occur for a few days per month, though the exact figure is difficult to determine with accuracy. As an illustration of this, Hofstätter and Chimani (2012) identified 146 Vb storm tracks between 1961 and 2002 using ERA-40 and ERA-Interim reanalysis data, but found that only 3 were reported in the literature. In their explanation, these authors point out the highly subjective nature of storm track identification. Attempts to achieve a more objective identification have been made (e.g. James, 2007) which also find more occurrences of Vb-type events than are identified by more subjective studies. Interestingly, James (2007) lists the longest unbroken sequence of TrM patterns as July 12th - 26th 1981, which is coincident with a 1 in 50 year flood across parts of central Europe (e.g. Kaspar and Müller 2008; Böhm and Wetzel, 2006)

It is also important to note that there is, as yet, no consensus on whether the frequency and intensity of Vb/TrM events is changing. As mentioned in the previous section, the findings presented by Fricke and Kaminski (2002) suggest that Vb/TrM patterns have become more frequent and intense over the last 150 years. In contrast, Mudelsee et al. (2004) were unable to find a significant increase, though they used monthly mean data, making it difficult to detect events which occur infrequently. Similarly, Hofstätter and Chimani (2012) did not detect any trend in Vb numbers, although the interannual variability in the numbers of Vb storms was very large.

While some details of this weather pattern may be influenced by local processes, such as the European land surface, it is now well known that the skill of weather forecasts for Europe (on the times scale of days to weeks) is sensitive to the accuracy of weather data from the western Pacific that is supplied to numerical weather prediction (NWP) models. Indeed, Rossby waves generated by cyclogenesis off the east coast of Japan can propagate across North America and may lead to Vb storm tracks in Europe (Grazzini and van der Grijn, 2003) when combined with cyclones that frequently form in the gulfs of Lyon and Genoa. It is also important to note that cyclogenesis east of Japan is/can be related to convection in the western Pacific warm pool off the east coast of Indonesia (e.g. Ambrizzi and Hoskins, 1997). Therefore, an increase in the severity and frequency of TrM/Vb events may be influenced more by rising temperatures in the western Pacific and conditions in the gulfs of Lyon and Genoa, than by deforestation in Spain, for example. However, local vegetation patterns may play a significant role in contributing to the moisture to storms or slightly diverting the Vb track, although this possibility is unsubstantiated.

It is also important to note that many studies of climate change impacts on European storm tracks using global climate models suggest that the North Atlantic and Scandinavia will be subjected to an increase in cyclone activity as the climate warms, while the storm tracks in the Mediterranean area (the V-group of van Bebber (1891) ecrease in importance. Consequently, any role of forests on the V-group of storm tracks may also become less important. There are many studies of the impacts of storms on forests, but very few (if any) on the influence of forests on storms. As outlined above, the literature reviewed for this task suggests that storms are formed and controlled by synoptic scale weather patterns and so at most forests could modify the severity of the storms or possibly change the direction of a storm, but not by much. Nevertheless, while forests may not significantly alter the paths of storms, they may be effective in urban environments at reducing turbulent winds that can affect city streets, and potentially reducing energy usage for heating requirements (see Stewart et al., 2011 and references therein).

4.3 Temperature patterns

Two recent studies (Lee et al., 2011; Ferrez et al., 2011) have compared temperatures inside a forested area with those measured in a nearby open area in mid-latitude regions. The results of these studies are consistent, showing that forests moderate maximum and minimum temperatures, such that maximum temperatures in forests are cooler than those in open ground, whereas minimum temperatures are warmer. More specifically, Lee, et al. (2011) calculated monthly and annual mean temperature differences between open and forested sites in USA and Canada (as open site minus forested site) at a number of sites with latitudes between 28°N and 56°N; these latitudes are similar to those which enclose Europe. Monthly maximum and minimum temperature data (denoted as T_{max} and T_{min} respectively) showed that T_{max} in the forested sites were cooler than in the open sites, but the T_{min} values in the forested sites were warmer than the open sites. That is, the range of temperatures measured in forested sites was narrower than, and within, the range displayed by open sites. For sites located between 45°N and 56°N, the effect on T_{max} was noticeably smaller than the effect on T_{min}. Between 28°N and 45°N, the effects of forest cover on T_{max} and T_{min} were similar. Overall, the diurnal temperature range was smaller in forested areas than open areas.

Lee et al. (2011) also found some dependence of the annual mean temperature differences on latitude, such that the difference in temperature (calculated as open land minus forest) increased with increasing latitude. The effect on annual mean temperatures is 0.85 ± 0.44 K (mean ± one standard deviation) northwards of 45°N and 0.21 ± 0.53 K to the south.

In a similar approach, Ferrez et al. (2011) analysed air temperature data from 14 sites in Switzerland, each with two weather stations in close proximity, one under a forest canopy and the other in the open. Their analysis method was more complex than that of Lee et al. (2011). Ferrez et al. (2011) fitted splines to their data to de-trend it and remove the seasonal cycle. Using the fitted splines, they found that the T_{max} values in the open site were 1.24 – 3.58°C warmer than the forested site, with an average difference of 2.71°C. The difference between the T_{min} values in the open and forested sites ranged from 0.74 to -2.60°C, indicating that minimum temperatures were colder in the open site. Most T_{min} differences were smaller than the T_{max} differences, with an average of -0.36°C.

The altitude of the site was also found to have an important effect on the maximum temperatures such that the sheltering provided by the forest was more effective at high altitudes. The orientation of the site affects the minimum temperatures: south-facing forests have extreme minimum temperatures similar to those for open ground.

The influence of the tree type on the observed temperature range is more complex. However, overall, conifers were found to insulate (for both maximum and minimum temperature extremes) more than beech and other deciduous trees, meaning that extreme maximum and minimum temperatures occur less commonly under coniferous forests than other forest types.

In and around cities, urban and peri-urban forests (UPFs) have been shown to play a role in moderating local climates. This is important because heat island effects can make cities uncomfortable places to live and work (e.g. Stewart et al., 2011). Indeed, a recent study by Bowler et al. (2010) found that an urban non-green site would be, on average, around 1°C warmer than an urban park, and that larger parks tend to be cooler. Furthermore, by providing shade and cooling buildings, trees can help to reduce energy usage. As an example, a study in Athens concluded that the cooling effects of trees could reduce summer time consumption of air conditioning during the day by 2.6-8.6% (Tsiros, 2010). A study by Lion et al. (2009) found that a 30% increase in forests in a radius of 50 km around Paris could reduce the urban heat island in the centre of Paris by 2-3°C during heat waves.

Bohnenstengel et al. (2011) used a high resolution model to simulate the urban heat island of London. They compared two simulations of May 2008, one which included the effects of the buildings and a second where the urban areas had been replaced by grass. Although forests were not considered, these results are still useful in illustrating the effect of vegetation cover on the urban heat island. The simulated urban heat island was strongly dependent on the urban surface fraction, and small vegetated areas within densely built areas had almost no effect on the simulated surface temperatures. The non-urban land fraction had to be between 10% and 20% before a reduction in the urban heat island occurred. Advection of heat across the city by winds had a large effect on the heat island at downwind locations, such that even small urbanised areas lead to strong urban heat islands, suggesting that large areas of vegetation cover might not reduce the urban heat island by very much. McCarthy et al. (2011) used a regional climate model which had a horizontal resolution of 25 km to study the urban heat island of London under a warming climate. Their results indicated the vegetation cover of London would have to be increased considerably to mitigate the effects of the warming climate on the urban heat island. However, they note that increased vegetation cover could have larger effects at the street level.

It is important to note that the characteristics of simulated heat waves are sensitive to the parameterisation of forests in numerical models. Notably, Clark et al. (2010) examined projected changes in heat waves using a large number of physically plausible future climate projections, and selecting those models in which the global average temperature increase is 2.0 ± 0.5°C warmer than present. They performed tests to identify the processes that had the greatest impacts on heat wave intensities. Their results showed that vegetation root depths had the greatest impact on heat waves in southern Europe. If the root depths were assumed to be shallower, the heat wave intensity was increased because the vegetation could not access deeper reserves of water. For Scandinavia and eastern Europe, the forest roughness had the largest influence on heat waves; if the roughness was reduced, evapotranspiration (which acts to cool the surface) was also reduced. Clark et al. (2010) also found that if soils remained moist, the severity of heat waves was reduced owing to evaporative cooling of the surface via increased evapotranspiration rates.

4.4 Water quality and quantity

Despite significant advances in scientific understanding of forest and water interactions, the role of forests in relation to the sustainable management of water resources, in terms of both quantity and quality, remains a contentious issue. Uncertainty persists because of difficulties in transferring research findings to different countries and regions, watershed scales, forest types and species, and management regimes. In addition, while there are strong links between the presence of forests and purity of water, any links between forest cover and the constancy of water flow and water supply are less clear (Stolton and Dudley, 2007). These are described in more detail below.

4.4.1 Water quantity

A broad description of the influence of forests on water availability is as follows. Considerable quantities (up to 35 %) of rain falling on forested land is intercepted by the foliage of the canopy and evaporated back into the atmosphere without contributing to soil water reserves. Much of the water that does soak into the soil is used by the trees themselves. Rainwater that penetrates to the land surface and is not used by the trees flows into rivers, lakes and eventually the oceans. A fraction of this rainwater penetrates the soil before resurfacing to join the surface flow, while some water becomes part of underground reserves of water or aquifers. For this reason, extensive afforestation will not increase the low stream flow rates in the dry season (Hamilton and Pearce, 1986). Therefore, replacing forest cover with other land uses almost always results in increased runoff and stream flow, but this is to the detriment of soil stability and water quality. Despite this, it remains theoretically possible that in degraded agricultural catchments the extra infiltration associated with afforested land might outweigh the extra evaporation loss from forests, resulting in increased rather than reduced dry-season flows, but this effect has rarely been seen.

The quantity of water used by forests depends on factors such as the local climate, the age and type of forest, the mix of trees and the soil type. For example, cloud forests (as found in Costa Rica, for example) are likely to increase water availability, some old natural forests may also increase water flows, but young forests and plantations are likely to decrease flows (Stolton and Dudley, 2007). Forests also use more water than shorter types of vegetation because of higher rates of evapotranspiration. As described previously, forested land affords lower surface runoff, groundwater recharge and water yield. Consequently, forest management practices can have a marked impact on forest water use by influencing the mix of tree species and ages, the forest structure and the size of the area harvested and left open.

Global climate models project large changes in snowfall, rainfall and evaporation in many parts of the world (IPCC, 2007). The influence of forests on water quantity and quality may be negative or positive in conjunction with these changes. Where large-scale planting of trees is proposed for mitigation of climate change, it is essential to ensure that water shortages will not be worsened. It is important to remember that the effects of forests on rainfall and water supplies in tropical and sub-tropical regions are likely to be very different to those in temperate regions. Although some links between forests and water resources are understood, there are still many unanswered questions.

The spatio-temporal scale used when assessing the impacts of forests on water quantity is crucial. At the micro-scale (e.g. catchment level) forests are often considered to be "consumers" (attractors) of water, removing it from the surface via interception and evapotranspiration. At the macro-scale, forests transfer water to the atmosphere by acting as recyclers. Evapotranspiration of moisture by forests favours cloud formation and precipitation (Ellison et al., 2011).

The discrepancies among scales are also shown in D'Almeida et al. (2007). At the micro-scale, deforestation generally results in decreased evapotranspiration and increased runoff and discharge. At the macro-scale (using model predictions), atmospheric feedbacks may significantly reduce precipitation regionally and, if larger than the local evapotranspiration changes, may decrease water yield, runoff and discharge. Meso-scale observations (i.e. a few km – a few 1000 km) are more ambiguous (Ellison et al. 2011) with deforestation increasing precipitation in some locations, but with a regional decrease (Silva Dias et al., 2009). Increased surface water flow following deforestation is larger in coastal areas than inland areas (Shuttleworth, 1989; Durbidge and Henderson-Sellers, 1993).

Other studies have suggested that afforestation, which increases water demand from plants by evapotranspiration, reduces runoff and leads to declining water availability, particularly in areas not forested for long periods of time (Bosch and Hewlett, 1982; Zhang et al., 2001; Andréassian ,2004; Brown et al., 2005; Farley et al., 2005). Similar conclusions were drawn by a number of small-scale studies in Europe, e.g. the Netherlands (van der Salm et al., 2006) and Norway, Denmark and Sweden (Rosenqvist et al., 2010).

The tree species is also important when considering impacts on water quantity. Deciduous trees typically transpire less moisture than evergreen species, potentially mediating forest impacts on the local water balance. Wattenbach et al. (2007) found evidence that planting deciduous species such as oak in temperate zones diminishes negative effects on the water balance resulting from the increased evapotranspiration associated with evergreen species. Another study by Zhang et al. (2001) showed that conifers tend to use more water than hardwoods or eucalypts. Overall, coniferous trees seem to have greater impacts on water availability than deciduous species (Sahin and Hall, 1996). While afforestation results in a decrease of water yield, clear-cutting and deforestation, on the other hand, lead to a period of increased water availability, at least until the forested state returns.

Recent reviews of forests and their impacts on water availability (Vertessy et al., 2001; Brown et al., 2005; Farley et al., 2005; van Dijk and Keenan, 2007) have postulated that water yield from afforested catchments declines with increasing plantation age, specifically through the relationship with leaf area index (LAI). Observations reported by Vertessy et al. (2001) for mountain ash suggest that decline of sapwood area index (SAI) with age produces a concomitant decrease in stand transpiration. The decline in overstory SAI is accompanied by a decline in overstory LAI, while understory LAI increases, but this layer transpires at only about 63% of the mountain ash rate on a per unit leaf area basis. Hence, while total stand LAI decreases are modest over time, the trend is for a significant decline in total stand transpiration as the forest ages. Rainfall interception also declined over time. Such reductions can be explained by lesser turbulent mixing and elevated humidity around the bulk of the leaf area in the mature forest. However, like the consequences of the aging of forest stands and of forest type, the impacts of the densification of forest cover at the watershed-scale are still poorly understood, as well as effects on the seasonal water cycle.

Micro-scale studies are limited by their spatial coverage (generally of the order of 2 km^2; Andréassian, 2004). Catchment basins of this size can easily receive precipitation from other locations and the impacts of change in a catchment (e.g., deforestation, reduced evapotranspiration and declining precipitation) may be not evident, not observable, or become evident only further away. This raises another scale issue, the one of time: while impacts of deforestation occur at shorter time period, most studies of the impact of afforestation consider relatively short periods (1–10 years; Brown et al., 2005). As other authors have concluded, ecosystem regeneration (and subsequent impacts on soil surface and ground water) can take longer periods of time (tens of years) to be realised. Then, although deforestation is always immediately followed by a period of water yield increase, the subsequent period of recovery when the forest grows back may or may not be characterized by a decrease in water yield (relative to pre-treatment conditions; Andréassian 2004).

Modelling experiments are invaluable for overcoming the spatio-temporal limits of observational studies. The results from regional and global models suggest that the overall impact of forests is to improve water availability.

4.4.2 Water quality

Forests make their most significant contribution to the hydrological characteristics of watershed ecosystems by maintaining high water quality. The presence of forests acts to minimize soil erosion and stabilise soils on sloping land areas (discussed in more detail in Section 5), reduce sediment in water bodies (wetlands, ponds, lakes, streams, rivers) and trapping or filtering of other water pollutants in the forest litter. The protection afforded by forests is recognised as being crucial for maintaining drinking water supplies for many of the world's cities - approximately 1/3 of cities receive drinking water from protected forested areas (FAO, 2007).

As indicated above, many questions regarding the links between forests and water supplies remain. What area of forest is needed to ensure a high quality water supply? What are the best management practices? The answer depends on which pollutant is of greatest concern. For drinking water agrochemicals are of greatest concern, whereas for hydropower sediment in the water is of more importance.

Regarding drinking water supplies, forests provide protective ground cover because forestry activities (with the exception of intensively managed plantations) generally use no fertilizers or pesticides and avoid pollution from domestic sewage or industrial processes. In addition, pollution from domestic, industrial and agricultural use can be greatly reduced or even eliminated by maintaining sufficient forested areas along watercourses, particularly in urban areas where concentrations of pollutants can be very high. Similarly, forests planted in agricultural areas can help to reduce pollutant levels, especially when planted in riparian zones. However, these forested areas will not prevent groundwater contamination. For example, atmospheric pollutants, such as sulphur and nitrogen-containing compounds, can be captured and deposited onto trees owing to their height and aerodynamic resistance. Subsequently, these pollutants can then enter the watercourse, in which case they may cause acidification.

Sources of physical and chemical constituents in natural forest catchments include hill slopes, channels and floodplains, with additional inputs from disturbances such as access roads, forestry activities, prescribed burns, and wildfire. Specifically for wildfires, Smith et al. (2011) reviewed those water quality constituents from natural forest catchment sources that may occur in streams and reservoirs after burning.

A modelling study in Northern Sweden by Erikson et al. (2011) indicates that the requirements to maintain or improve water quality stipulated by the EU Water Framework Directive goals may influence forest management, both in how the forest is managed and the economic consequences. Limiting maximum increases in concentrations of these substances to 10% above reference values resulted in an economic loss of circa 20% or 35%, depending on whether the limits were applied to the whole area or to each sub-catchment individually.

4.5 Soil stabilisation

When considering the role of forests on the stabilisation of soils, it is important to make the distinction between soil erosion and land slides. The effects of forest cover (and its removal, or replanting) will be discussed first. While forest cover does tend to reduce soil erosion, it is not the tree canopy that is directly responsible (FAO, 2005). On an area of sloping land, the soil will move downhill because of gravity and displacement by the splash action of raindrops. Experiments indicate that the erosive power of raindrops under trees tends to be very high because the raindrops merge before dripping off the leaves and therefore hit the ground with greater force (Wiersum, 1985; Hamilton, 1987; Brandt, 1988). Natural forest cover provides an effective barrier to splash-induced soil erosion, because the ground litter and leaves of the lower canopy reduce the force of the raindrop impact and their erosive power. In plantations where the soil has been cleared of vegetation and litter, serious erosion problems can develop.

The network of tree roots acts to anchor soil and rocks and improves soil stability. The degree of protection afforded by forests depends on the gradient of the slope, the root systems of the forest and undergrowth, as well as other land uses such as grazing, the stability of the underlying soils and rock surfaces, and the intensity of winds, rainstorms and snowfall. While forest cover is important for environmental protection and stabilisation of soils, the presence of forests alone does not guarantee that that there will be no erosion or other degradation of the environment.

The removal of forest cover does not necessarily result in erosion of soils. Poor land use practices following forest removal, such as overgrazing and removal of litter can result in degradation of the land (Bruijnzeel, 1990, 2004; Hamilton and King, 1983). The reverse procedure, planting forests on erosion-prone soils and runoff pathways, can reduce erosion and intercept sediment. Furthermore, much of the erosion that occurs after harvesting of timber is caused by the movement of soils during logging operations (e.g., road construction, movement of vehicles, etc.). Compaction of the soils results in increased surface runoff and lower water storage capacity. As a result, Reduced Impact Logging (RIL) techniques can significantly diminish these negative effects on soil stability.

As mentioned earlier, landslides are a quantitatively different phenomenon to soil erosion, although both may occur in the same place. Landslides are a downward mass movement of rock and/or soil controlled by gravity and other factors (Muzira et al., 2010). They may occur following loss of forest cover because the mechanical support provided by the tree roots has been removed. Furthermore, it is now known that shallow mass movements of soils (less than 1 m in depth) are quickly stabilised by forests, and do not usually result in high amounts of sediment entering the surrounding rivers (Bruijnzeel, 1990, 2004; O'Loughlin, 1974). In contrast, deep-seated landslides (over 3 m deep) are not noticeably influenced by the presence or absence of forest cover (Bruijnzeel, 1990, 2004). Instead, these events are most strongly influenced by geological, topographical and climatic factors (Ramsay, 1987).

As an illustrative example, Rickli and Graf (2009) studied the effects of forests on shallow landslides in Switzerland. They found that, while such landslides commonly occur on both forested and open areas, the number of landslides in forested areas was smaller than those in open land. These results are in agreement with previous work on the positive effects of forests and tree roots on slope stability (Greenway 1987; Abe and Ziemer 1990; Ekanyake et al. 1997; Muzira et al., 2010), and the negative consequences of intensive harvesting activity and deforestation which are generally considered to lead to an increase in landslide frequency (Ziemer, 1981; Tang et al., 1997; Sidle and Wu, 1999; Jakob, 2000; Sidle et al., 2006; Claessens et al., 2007). This increase in landslide frequency often occurs a few years after the removal of the forests, when the roots of the cut trees begin to decay.

Landslide frequencies are generally larger in forests which have many gaps (Temperli, 2006) or where the forest is in poor health as a result of damage by storms or pests (Rickli, 2001). In the Sachseln area of Switzerland, the number of landslides per unit area in forest stands identified as being in poor condition was even higher than in open land (Rickli, 2001), which was also observed in Prättigau (Temperli, 2006). The effect of a forest's condition on the frequency of landslides is confirmed by the observations of Schmidt et al. (2001) and Markart et al. (2007).

Overall, landslide density within forests is generally smaller than in open land, but is strongly related to the condition of the forest and the size and number of gaps. It is also important to note that, while forests do exert a stabilising influence on shallow slopes, the effect is negligible for very steep slopes (Boll, 2002).

Land degradation resulting in desertification was investigated in a recent study in Sardinia (Italy), which assessed how the increase in desertification risk can be related to the loss of vegetation (forests and shrubs) that negatively impacts soil resistance to erosion, overgrazing and land productivity (Santini, 2008).

Forest fires can result in increased degradation of soils; this effect is discussed in section 4.8.

4.6 Flood regulation

Forests play a role in regulating some aspects of local hydrology, but cannot be used for large-scale reduction of flooding risks or control of floods. The largest effects of forests are seen following short duration and low intensity rainfall events, which are often the most frequent in temperate regions. Under these conditions, the partial or complete removal of tree cover may accelerate water discharge and increase flood risk. Forest cover can mitigate the effects of small-scale rainfall events and reduce the likelihood of downstream flooding due to enhanced infiltration and storage capacities.

However, as rainfall duration or intensity increases, or as the distance of the rainfall area from the watershed increases, the influence of tree cover on flow regulation decreases. Consequently, natural processes in the upper watershed are more important than land management practices in the development of large floods. One possible exception is reduction of downstream flooding by floodplain forest, where the simple hydraulic roughness of the forest, such as forest litter, dead wood, twigs and tree trunks, may slow down the flood flows. For example, it is unlikely that the effects of floods which arise from extremely high rainfall events like the one caused by a cyclone in the Paznaun

Valley, Austria, in August 2005 could have been mitigated by the presence of forest. Although there are many good reasons for reforesting watersheds, as discussed in the previous sections of this report, reduction or even control of flooding is not one of them. However, the complex relationships between forests and water in large river basins continue to be a matter of debate, and it is clear that more work is needed to develop a full understanding of these relationships (FAO, 2007). On smaller scales, the impermeable surfaces and limited vegetation coverage of urban environments can make cities vulnerable to floods due to storm-water runoff. Trees can reduce the risk of flooding by intercepting and re-evaporating water from their canopies, and also by allowing water to percolate into the soil beneath (Stewart et al., 2011).

In terms of flooding extremes, in general the impact of forest removal results in increased flood peaks, while reforestation causes only very limited reductions of flood peaks. Both deforestation and reforestation appear to have little or no impact on severe floods (e.g. Mudelsee et al., 2003). In the short term, reforestation decreases low flows and deforestation increases them (Andréassian, 2004).

Linking back to forest fires and vegetation removal, Soulis et al. (2010) studied a watershed affected by two wildfires in August 1995 and in August 2009 in Greece. Before the first wildfire the watershed was mainly covered by a dense pine forest, with a small upstream part covered by shrubs or bare rock. The pine forest was almost totally destroyed by the wildfire in 1995. Before the second fire the watershed had a mixed vegetation cover consisting of pasture, shrublands, and pine forest at the first stages of development. The forest fire in August 2009 destroyed most of this vegetation cover.

Soulis et al. (2009) observed peak runoff values as much as 10 times higher in the period soon after the first wildfire compared to the period soon before the second wildfire. A longer term study of the Bisagno River in the Liguria region of Italy was made by Rosso and Rulli (2002). Land use data for 1800 to 2000 showed increasing urbanization in the downstream area and agricultural expansion in the upstream area, and both occurred at the expense of wooded areas. Results of hydrological analysis showed that forest removal was a strong component of the observed increase both in peak flood and peak flow frequencies with respect to the (more highly vegetated) period.

4.7 Crop yields

In addition to the six criteria discussed above, it also worth noting the effect small areas of trees can have on crop yields. Trees have been planted on and around farms to form shelterbelts and provide protection for buildings, animals and crops from strong winds, sunlight, rain and snow (Donnison, 2012). These shelterbelts are commonly linear in nature, but small wooded areas with more rounded shapes can also provide shelter. A suitably dense shelterbelt can protect a large area of cropland via many interacting factors. Lower wind speeds means the crop yield may improve, and the damage caused by strong winds will be reduced. The humidity levels in the cropped area will be higher, reducing evapotranspiration and increasing both the water use efficiency of the crop and the overall yield. Yields can be reduced in crops close to the trees owing to shading and competition for water and nutrients; however, these effects can be reduced by careful selection of native tree species and the use of deciduous trees in place of conifers (Donnison, 2012). Shelterbelts in drier areas of Europe can increase rain-fed crop yields by 16% by increasing the water use efficiency of the crop.

4.8 Fire risk

In this section the impacts of fires on forests are assessed. First, the changing risk of forest fires in Europe is briefly assessed, and then the effect of the fires on soil stability is discussed. The latter part includes a discussion on methods used to encourage the growth of forests following fire and limit soil erosion.

Fire frequency is expected to increase with human-induced climate change, especially where precipitation remains the same or is reduced (Stocks et al., 1998). This can result in a change in vegetation structure that, in turn, exacerbates this risk (Cramer and Steffen, 1997). Indeed, Miranda et al. (1994) suggested that has been an increase in risk, severity, and frequency of forest fires in Europe. In addition, several authors have suggested that climate change is likely to increase the number of days with severe burning conditions, prolong the fire season, and increase lightning activity, all of which lead to probable increases in fire frequency and areas burned (Price and Rind, 1994; Goldammer and Price, 1998; Stocks et al., 1998).

Forest fires play a key-role in soil degradation and removal via water erosion. For instance, soil destabilization is mostly expected to occur in the short term, i.e. a few months after the fire, when the recovery of the vegetation cover is still low (Pausas and Vallejo, 1999). Re-establishment of pre-fire plant communities may span several years to decades, depending on the pre-existing vegetation types, fire severity and the environmental conditions (Keeley, 2009).

While the vegetation response capacity is a function of climate, water availability, aspect, and reproductive strategy, vegetation density and height influence the burnt vegetation protective capacity, which, combined with rain erosivity, soil erodibility and slope, determine soil erosive susceptibility (Duguy et al., 2012). Soil is a non-renewable primary resource which may be exposed to the risk of degradation and erosion after fire. In Mediterranean ecosystems affected by wildfires, one of the main objectives of restoration programmes could be to conserve the soil (Vallejo and Alloza, 1998). Post-fire rehabilitation measures are short-term actions designed to mitigate soil degradation until natural vegetation regeneration covers the burned area. Regulation of the hydrological cycle is mainly aimed at controlling soil erosion and runoff and preventing off-site impacts of sediments and floods. The most common post-fire rehabilitation measures are grass seeding, mulching, and contour-felled logs, as hill slope measures, and check dams (straw bales, log, and rock dams) as channel measures (Napper, 2006; Cerdà and Robichaud, 2009).

As reported by Duguy et al. (2012), a study in eastern Spain (Gimeno et al. 2005; Bautista et al., 2009) tested the effectiveness of new rehabilitation treatments (seeding, mulch, and seeding plus mulch) to mitigate soil degradation and enhance vegetation recovery in the short and medium term in burned, highly degraded woodlands. The combined seeding plus mulching treatment enhanced total plant cover throughout the two post-fire years studied, with plant cover values being around 50% higher than for the untreated plots. However, neither the seeding nor the mulch alone influenced the vegetation recovery.

With respect to soil protection, the treatments with mulch (seeding + mulch and mulch alone) greatly reduced soil surface compaction and enhanced water infiltration. Nearly two years after fire and treatment application, these effects were still significant. The mulch layer also greatly reduced post-fire soil loss: the non-mulched sites lost about 20

Mg ha^{-1} year^{-1} of soil owing to erosion during the first post-fire year while the mulched sites had negligible losses (Bautista el al., 2009).

A further study in Liguria, northern Italy (Rosso et al., 2007) coupled experimental and modelling activities to assess peak runoff and erosion rates under different rainfall intensities and durations for both unburned and burnt plots, showing the high vulnerability of the latter. Moreover, sediment removal and transport of sediment along slope and channels, as well as debris flows (including pieces of wood) were simulated confirming the complexity of processes remodelling the landscape after changing land cover.

5. Summary

This report has summarised relevant literature on the historical evolution of forest cover and weather across the European Union. The influence of forests on precipitation patterns, storm regimes, temperature patterns, water quality and quantity, soil stabilisation and flood regulation have been assessed. The forest influences are summarised below.

5.1 Precipitation Patterns

Precipitation in Europe is predominantly controlled by large scale weather systems which propagate across the continent. As a result, it is extremely difficult to disentangle the role played by forests, and changes in forest cover, in modifying precipitation patterns. In addition, although there have been changes in forest cover over Europe since 1900, these areas have tended be scattered throughout the continent, so any changes are too small in area to have exerted a strong influence over large areas. However, local precipitation levels could be altered by the forests. An analysis of a long (approximately 300 year) rainfall record from Paris shows that there has been a change in seasonal rainfall amounts over the past 150 years. During the second half of the nineteenth century, summers tended to wetter than at present. This shift in rainfall seasonality is likely to be due to the end of the Little Ice Age, and so is controlled by large-scale circulation patterns.

5.2 Storm Regimes

We have focused on the Vb storm track, since these types of storms are responsible for severe flooding events in central Europe (e.g., August, 2002). It is not possible to state whether the presence or absence of forests has affect the frequency of these storms, since their numbers are highly variable, and the methods used to calculate their numbers. The presence or absence of forests could alter the severity of the storms and the total rainfall from the storms; this aspect will be investigated as part of Task A3.

5.3 Temperature Patterns

Forests can influence surface temperatures by insulating the ground from extreme air temperatures. As a result, temperatures under forest cover tend to show lower maximum temperatures and higher minimum temperatures than comparable areas of open ground. This effect is a function of tree type, with conifers seeming to provide the greatest insulation. However, forests are generally darker than open ground, meaning that forests absorb a greater fraction of the sun's incident radiation.

Evapotranspiration from the forest acts to reduce surface temperatures. In particular, models show that forests with characteristically deeper roots can access greater quantities of water, leading to higher evapotranspiration rates which enhance evaporative cooling. Increased areas of vegetation cover within cities could reduce the urban heat island, but current model results suggest that at least 10% - 20% of the urban area would have to be replaced by vegetation before any effect would be seen. In areas immediately downwind of an urban area, even large vegetated areas would reduce the urban heat island by only a small amount.

5.4 Water Quality and Quantity

Due to their higher evapotranspiration rates, forests use more water than other forms of vegetation, such a cereal crops. Consequently, increased forest cover is unlikely to increase river flow rates, especially during dry spells. However, forests reduce soil erosion and absorb atmospheric pollutants, thereby protecting water supplies. Looking at larger scales, the presence of forests enhances rainfall, especially during the summer months in areas such as eastern France and south-west Germany, where the influence of weather fronts is small and most rainfall is produced by from convective clouds.

5.5 Soil Stablisation

Forests improve soil stability in two main ways. Firstly, the network of roots provides mechanical stability, which can reduce number of landslides on shallow slopes. Secondly, the canopy and ground litter reduce the impact of raindrops, which is largely responsible for soil erosion. However, it is important to note that the benefits are greatest when the forest is in good condition, i.e. it comprises a wide variety of species, with trees of different ages, and with the forest litter intact. In addition, while the severity of shallow (<1 m deep) landslides may be reduced by the presence of a well-maintained forest, deeper landslides (> 3 m deep) are largely unaffected by the presence or absence forest. Similarly, forests do not seem to dramatically decrease the rates of landslides on steep ground.

5.6 Flood Regulation

Forests can play a role in reducing flooding risks for short duration rainfall events which occur on local scales by re-evaporating water intercepted by the canopy, and providing increased hydraulic roughness via tree trunks and forest litter (including dead leaves, twigs) that slow down floods. However, as the rainfall duration increases and the size of the watershed increases, the influence of tree cover on flow regulation decreases.

5.7 Crop Yields

By selecting the appropriate tree species and location, forests can increase crop yields by providing shelter from strong winds, and increasing humidity levels which can have the effect of improving the efficiency of water usage by the crops.

5.8 Fire Risk

Fire risk is projected to increase in regions where precipitation remains the same or is reduced. In addition, studies indicate that climate change is likely to increase the number of days with severe burning conditions, prolong the fire season, and increase lightning activity, all of which lead to probable increases in fire frequency and areas burned. The soils on burned areas are susceptible to erosion, and protection of those soils is important for reforestation to succeed.

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