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Abstract

Anthropogenic land-cover change (ALCC) is one of the few climate forcings for which the net direction of the climate response over the last two centuries is still not known. The uncertainty is due to the often counteracting temperature responses to the many biogeophysical effects and to the biogeochemical versus biogeophysical effects. Palaeoecological studies show that the major transformation of the landscape by anthropogenic activities in the southern zone of the Baltic Sea basin occurred between 6000 and 3000/2500 cal year BP. The only modelling study of the biogeophysical effects of past ALCCs on regional climate in north-western Europe suggests that deforestation between 6000 and 200 cal year BP may have caused significant change in winter and summer temperature. There is no indication that deforestation in the Baltic Sea area since AD 1850 would have been a major cause of the recent climate warming in the region through a positive biogeochemical feedback. Several model studies suggest that boreal reforestation might not be an effective climate warming mitigation tool as it might lead to increased warming through biogeophysical processes.

25.1   Introduction

This chapter addresses several major questions. Did anthropogenic land-cover change (ALCC) occur in the Baltic Sea catchment during the last two centuries, and if so, did this play a role in the recent climate warming observed in the region? If not, did it have any other effect on climate? If recent ALCC occurred, is it unique in magnitude compared to ALCC before AD 1850 and back to the Neolithic time (6000 cal year BP—calendar years before present)? Did past ALCC have an effect on past climate?

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The chapter discusses the effects of ALCC on past climate (on timescales of decades, centuries and millennia) as well as future climate. It also reviews studies on natural climate-induced (potential) land-cover change (CLCC) and its feedbacks on climate, as it may help understanding of important processes involved in land cover–climate interactions. Here, land cover relates to vegetation cover, in particular tree cover versus low herb and low shrub vegetation, as well as snow cover. ALCC may be an external climate forcing, while CLCC is part of the climate system and may cause feedbacks on climate, whatever forcing is the cause of the initial climate change (Fig. 25.1). A feedback can be either positive (enhances the climate change responsible for the vegetation/land-cover change) or negative (mitigates the climate change). Similarly, a single forcing can enhance or mitigate the effect of other forcings; that is, the effect of ALCC may mitigate the effect of greenhouse gas emissions. The effects of ALCC or CLCC on climate (as forcing and feedbacks, respectively) are due to biogeophysical and biogeochemical processes at the boundary between vegetation and the atmosphere (Findell et al. 2007). When attributing causes of regional climate change at the scale of the Baltic Sea area, the biogeophysical effects are of particular interest since they exert a direct, measurable effect on regional climate. Biogeochemical effects are more relevant in the context of global climate change since the timescale of carbon dioxide (CO_2) mixing in the atmosphere is very short. Consequently, regional changes in the carbon balance affect regional climate only indirectly by affecting the global CO_2

concentration. Therefore, this chapter focuses primarily on biogeophysical mechanisms, although biogeochemical processes are also described.

Sensitivity studies with global Earth System Models have increased our understanding of interactions between land cover and climate over the past decade (IPCC 2007). However, the mechanisms involved in biogeophysical feedbacks are mainly regional to local in scale; therefore, use of regional climate models and vegetation models should potentially provide better insights on those feedbacks. Nonetheless, few published studies have used regional climate models, and none were specifically designed to evaluate the effects of ALCC on past, present and future climate change at the scale of the Baltic Sea basin. There are also very few attribution studies using global climate models and focusing on ALCC as a possible forcing at both global and regional scales. To date (2013), there is a single study on the effect of past ALCC on climate change at 6000 and 200 cal year BP in north-western Europe using a regional climate model (Gaillard et al. 2010; Strandberg et al. 2013). Therefore, current understanding of the role of ALCC in regional climate change at the scale of the Baltic Sea basin, and in particular as a possible forcing of the warming of the last two centuries, must rely primarily on studies at the European or northern hemisphere scale using global climate models. The largest model study to date focusing on the impacts of ALCC on the climate of the northern hemisphere is that within the Land-Use and Climate, Identification of

Fig. 25.1 Schematic illustration of interactions between land cover and climate: biogeochemical and biogeophysical effects of climate-induced and human-induced changes in land cover, feedbacks and forcings. CLCC (climate-induced land-cover change, i.e. change in natural (potential) vegetation); ALCC (anthropogenic land-cover change, i.e. change in human-induced vegetation due to agricultural activities)
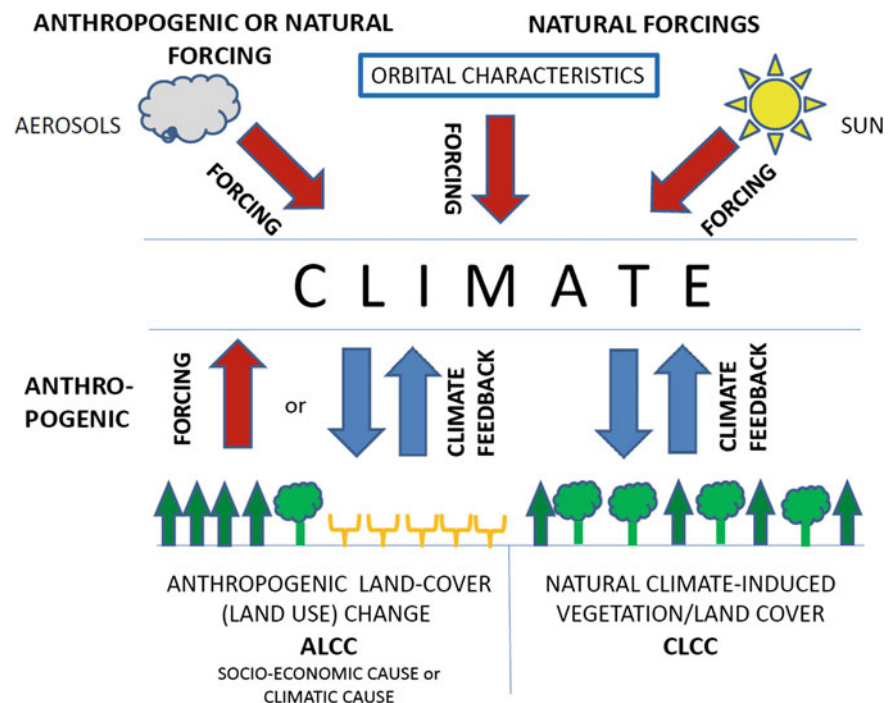
Robust Impacts (LUCID) project (Pitman et al. 2009; de Noblet-Ducoudré et al. 2012). This was set up to study the robustness of modelled biogeophysical impacts of historical ALCC on climate (roughly AD 1850 to modern time).

This chapter first explains the processes involved in land cover–climate interactions and summarises the results from modelling studies investigating these processes (Sects. 25.2.1 and 25.2.2). It then reviews current understanding of how past and recent land-cover change, both ALCC and CLCC, might have influenced regional climate in the Baltic Sea region (Sects. 25.2.3 and 25.4). As none of the chapters in the present assessment of climate change in the Baltic Sea basin deals with Holocene ALCC, this chapter also reviews studies on ALCC since Neolithic time (about the last 6000 years) (Sect. 25.3). Finally, the chapter discusses the possible effects of future resource management on land cover and, as a result, on biogeochemical and biogeophysical processes and future climate (Sect. 25.5). Holocene CLCC is presented in Chap. 2.

'Climate change' refers to systematic changes in response to external forcing, in accordance with Chaps. 23 and 24. In order to avoid any confusion of concepts, 'Land cover–climate interaction' and 'biogeophysical (or biogeochemical) effect' refer to forcing mechanisms, while the term 'feedback' is used only for the effects of CLCC on climate (i.e. as part of the climate system). See Chap. 2 for a description of the relationships between climate change and natural vegetation during the Holocene, and Chaps. 16 and 21 for a complete account of the influence of recent climate change on vegetation.

25.2 Land Cover–Climate Interactions: What Are They and What Are the Mechanisms Involved?

25.2.1 Biogeophysical Effects/Feedbacks

Biogeophysical effects (Fig. 25.2) influence physical exchange fluxes and the energy balance between the atmosphere and land surface. The major biogeophysical feedbacks are due to (i) land-surface characteristics such as albedo (referred as the *albedo effect*; albedo is the surface reflectivity with respect to short-wave radiation) and roughness (e.g. smooth snow or rough forest), and (ii) evapotranspiration (the sum of transpiration from plant stomata and evaporation from other water sources at or below ground) (Levis 2010).

Albedo is the proportion of incoming solar radiation reflected by a surface. It strongly influences the energy available for absorption by the land surface. The greatest contrast in albedo occurs between open and forested land,

especially in the presence of snow. While snow is completely exposed on open land, it is partly covered in a forested area. This is often referred to as the *snow-masking effect*. Snow masking can cause a positive feedback on climate. For example, a high-latitude northwards expansion of trees and shrubs (low albedo) due to warming will hide the snow (high albedo) on the ground and thus increase the absorption of solar radiation, which will in turn enhance the warming and lead to a further northwards expansion in tree cover, leading to further warming (Fig. 25.2). This type of positive feedback is especially strong when the forest comprises evergreen conifers that retain their needles during winter. Modelling experiments have shown that the albedo effect can be significant (e.g. Bala et al. 2007; Liang et al. 2010).

Vegetation also influences the hydrological cycle. Structural changes in vegetation, such as changes in leaf area index (LAI), roughness length, and rooting depth modify the evapotranspiration of water from the land surface. The LAI represents the amount of leaf material present in an ecosystem and is geometrically defined as the total one-sided area of photosynthetic tissue (in m^2) per unit ground surface area (in m^2). Surface roughness (often just referred to as 'roughness') is a measure of the texture of a surface. The roughness length (expressed in m) depends on the frontal area of the average element (e.g. trees in a forest) facing the wind divided by the ground width it occupies. For instance, the roughness of featureless terrain is 0.005 m (smooth), flat terrain with grass or very low vegetation 0.03 m (open), and mature forest 1.0 m (closed) (Davenport et al. 2000). While the LAI influences the amount of intercepted water and the partitioning of energy fluxes into sensible and latent heat, the roughness length affects the turbulent mixing of heat into the atmosphere. The rooting depth determines the amount of water extracted from the soils by the vegetation; a deeper and/or more extensive root system will enhance the ability of the vegetation to extract soil water. In environments where neither temperature nor water limits vegetation growth, the vegetation tends to flourish, which increases both LAI and roughness. Since vegetation transpires water through leaf stomata, an increase in LAI will be associated with increasing evapotranspiration and, as a result, an increase in latent heat at the expense of sensible heat. Sensible heat warms the atmosphere close to the vegetation surface, whereas latent heat is stored in the released water vapour and warms the atmosphere only when condensation occurs, some distance away from the vegetation and higher up. Therefore, the *hydrological cycle effect* has a dampening effect on local to regional temperature change since stronger evapotranspiration implies that more energy is required to vaporise water (Fig. 25.2). An increase in woodland cover due to climate warming or anthropogenic activities (forest planting) will also increase the land-surface roughness and, in turn, enhance the moisture convergence and lead to increased precipitation. Higher water availability triggers a

Fig. 25.2 A simplified scheme of the biogeophysical feedback system (after Ben Smith, unpublished). *LAI* leaf area index; *PAR* photosynthetically active radiation
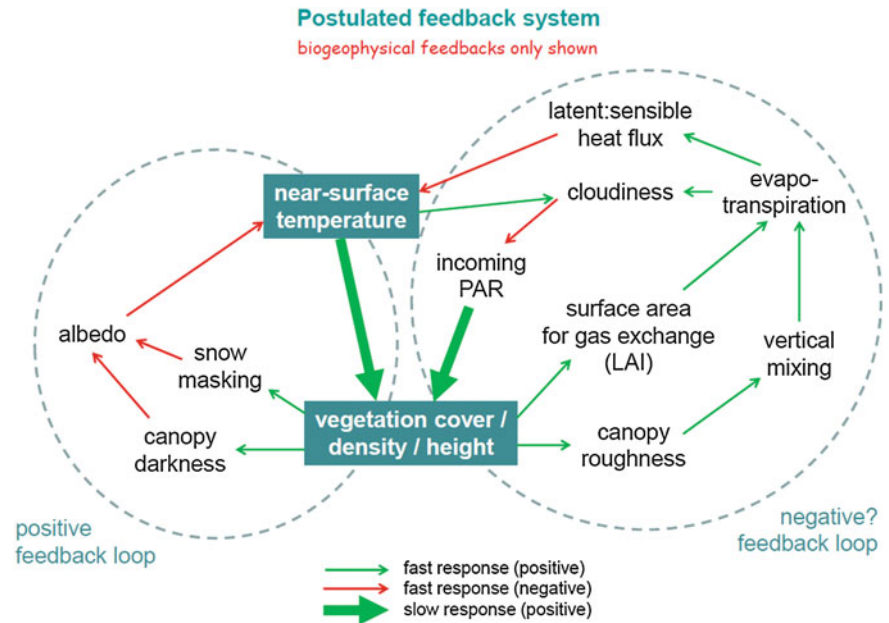
positive *precipitation effect/feedback* by producing higher vegetation density and a further increase in land-surface roughness and precipitation, etc. Higher atmospheric CO_2 concentrations may also cause an increase in vegetation density (the *fertilisation effect*, see Sect. 25.2.2) and produce similar secondary biogeophysical effects such as the *precipitation effect* through an increase in roughness. As pointed out by Levis (2010), these feedbacks (Fig. 25.2) can all be modified or eliminated by ALCC. Moreover, it is rare that a single feedback dominates or is the only one active. Several feedbacks often occur together, which increases the difficulty of interpreting the results.

The study by Zhang (2011) is one of the few investigations using observations that focus on the Baltic Sea region. The study was based on precipitation and run-off data for 1961–2003 in southern and central Sweden and showed a trend towards increased evapotranspiration. As the major cause of the increased evapotranspiration was increased winter evaporation, the author proposed this may be related to a known land-use change in the study area, namely the replacement of deciduous trees (lose their leaves in winter) by planted coniferous forest (with evapotranspiration from needles in winter, also from intercepted water, i.e. rainfall collected on the needles).

25.2.2 Biogeochemical Effects/Feedbacks

The land surface plays a major role within the global carbon cycle: vegetation takes up atmospheric CO_2 through photosynthesis and uses the carbon to build biomass, while the oxygen is released to the atmosphere; some time later, the vegetation dies and dead biomass builds up soils; the soil organic matter is then decomposed by micro-organisms and the resulting CO_2 released to the atmosphere, thus closing the cycle. However, disturbance processes such as forest and grassland fires, a climate-induced decrease in woodland and anthropogenic deforestation will also release carbon to the atmosphere. The land surface contains significant amounts of carbon in vegetation (350–550 PgC, Prentice et al. 2001) and in soils (1500–2400 PgC, Batjes 1996). Additional carbon is stored in wetlands (200–450 PgC) and in the loess soils of permafrost areas (200–400 PgC, McGuire et al. 2009).

Owing to the general character of the Baltic Sea region with its extensive forests and substantial wetland areas, the carbon storage in vegetation and soils in the region is undoubtedly significant, although no specific regional estimates appear to be available. Both humans and climate may have a significant impact on this carbon storage. The carbon balance of the land surface depends primarily on the atmospheric CO_2 concentration and temperature. In carbon cycle models, such as those used in the Coupled Climate Carbon Cycle Model Intercomparison Project (C4MIP, Friedlingstein et al. 2006), the sensitivity of the carbon cycle to climate change can be expressed by two parameters, β and γ. β describes the sensitivity to changes in atmospheric CO_2 concentration, while γ describes the sensitivity to changes in climate, especially temperature. Vegetation experiments with elevated CO_2 concentrations provide observational evidence of enhanced net primary productivity (NPP) under increased atmospheric CO_2 (Norby et al. 2005), implying a positive β. While the experiments give a direct indication of feedback between CO_2 concentration and CO_2 uptake, i.e. the *fertilisation effect*, there is still much uncertainty about

the universality of the results, especially since interactions with nutrient and water availability are likely but remain difficult to quantify (Gedalof and Berg 2010). Nevertheless, elevated CO_2 concentrations probably do enhance productivity, as long as other conditions for additional growth are met. Outputs from land-surface models (LSMs) show an increase in carbon storage under increased atmospheric CO_2 of 0.85–2.4 PgC ppm v^{-1} in early studies (Cramer et al. 2001), while later studies that consider the limitation of carbon uptake by nitrogen availability show a considerably decreased enhancement (Sokolov et al. 2008; Thornton et al. 2009; Zaehle et al. 2010). In the LSM CLM4, for example, the estimated increase in NPP when considering nitrogen availability is only 30 % of the increase without considering nitrogen dynamics (Bonan and Levis 2010).

Through the temperature sensitivity of both photosynthesis and respiration, the terrestrial carbon balance is also strongly influenced by changing temperature, although the precise response is not well known. Under water-limited conditions, an increase in temperature would lead to stronger water stress due to enhanced evapotranspiration. In contrast, an increase in temperature in cold regions would lead to a longer growing season, thereby enhancing vegetation growth. With respect to soil organic matter, an increase in temperature will lead to increased decomposition, that is enhanced carbon losses to the atmosphere (Davidson et al. 2006). Modelling studies suggest that warming will accelerate carbon losses from soils, implying a positive feedback between warming and the carbon cycle. Friedlingstein et al. (2006) found a range of −20 to −177 PgC per °C for the γ factor and Sitch et al. (2008) of −60 to −198 PgC per °C. However, the models used did not consider nutrient limitation and may have overestimated γ, since warming may increase nitrogen mineralisation and availability in soils, enhancing vegetation growth. Current climate–carbon cycle models including a nitrogen cycle show this effect (Sokolov et al. 2008; Thornton et al. 2009; Zaehle et al. 2010), but the uncertainties in γ remain very high.

25.2.3 Impact of Hypothetical Land-Cover Change on Climate: Climate Model Simulations

Ban-Weiss et al. (2011) studied climate forcing and response to idealised changes in surface latent and sensible heat. They found that globally adding a uniform 1 W m^{-2} source of latent heat flux along with a uniform 1 W m^{-2} sink of sensible heat leads to a decrease in global mean surface air temperature of 0.54 ± 0.04 °C, explained mainly by an increase in planetary albedo associated with an increase in low-elevation cloudiness caused by increased evaporation. The model results indicate that, on average, when latent heating replaces sensible heating, global and local surface temperatures decrease. Kvalevåg et al. (2010) used GCM (general circulation model) simulations to compare impacts on climate due to vegetation and albedo changes together or to albedo changes only; they concluded that effects due to changes in albedo dominate in temperate regions. The authors also claimed that divergent conclusions between similar studies are probably due to differences in specifications of albedo. Sensitivity to albedo was also shown in a model experiment where a hypothetical boreal forest expansion, decreasing the surface albedo, led to an enhancement of the summertime Arctic frontal zone and a strengthening of the jet (Liess et al. 2011). Boreal forests are characterised by lower albedo and a higher Bowen ratio (the ratio of sensible to latent heat fluxes, i.e. heat loss or gain) for similar levels of soil water availability than temperate forests (Bonan 2008). Thus, replacing boreal forests by grassland results in a cooling effect due to a decrease in both the Bowen ratio (as long as soil water is available) and net radiation (increase in albedo); the cooling effect may become higher than the warming effect of increased carbon emissions.

Eliseev (2011) showed that, at the global scale, changes in surface albedo due to the replacement of natural vegetation by agricultural land would have a greater influence on the available energy at the surface by absorbed short-wave radiation than the influence of the albedo effect. This is explained by relatively low insolation during winter at the latitudes characterised by the snow-masking effect of forest vegetation. Moreover, Cook et al. (2008) showed that feedback mechanisms including interactive vegetation and snow may be very sensitive to the parameterisation of the snow fraction. For instance, a fast-growing snow fraction produced a large-scale southward retreat of boreal vegetation and a widespread cooling. Bathiany et al. (2010) found that afforestation of all currently treeless areas north of 45°N would lead to a global mean warming of 0.26 °C due to biogeophysical effects, while the reduction in atmospheric CO_2 would only be 6.5 ppm, leading to a net warming. The albedo effect would be most significant in winter and spring when forests mask snow, causing an additional regional temperature rise. Earlier, similar studies of idealised, large-scale deforestation also found that an albedo cooling would dominate over CO_2 warming in boreal regions, indicating that boreal reforestation would probably not be an effective mitigation tool in such areas (Betts 2000; Claussen et al. 2001; Sitch et al. 2005; Bala et al. 2007).

25.3 Reconstructing Past Land-Cover Change

This section presents the various methods available to reconstruct past land-cover and their changes through time and space and describes the major ALCCs in the Baltic Sea region over about the last 6000 years, from the Neolithic (start of agriculture) until modern time. All ages are given in calibrated ^{14}C years (or calendar years) BC (Before Christ)/ AD (Anno Domini = after Christ) or BP (Before Present; present = AD1950).

25.3.1 Methodology

Attempts to reconstruct past changes in land cover have been based on two major approaches: (i) interpretation of palaeoecological data, fossil pollen in particular, and (ii) use of land use and population historical records as well as archaeological records of past settlements.

Estimates of human-induced changes in land cover based on historical records, remotely sensed images, land census and modelling (Ramankutty and Foley 1999; Olofsson and Hickler 2008; Klein Goldewijk et al. 2011) were used to provide first insights into the effects of ALCC on past climate (e.g. Brovkin et al. 2006; Olofsson and Hickler 2008). The most frequently used database in climate modelling to date is the History Database of the Global Environment (HYDE) database (Klein Goldewijk et al. 2011). However, its estimates of anthropogenic land cover during key periods of the past show large discrepancies with more recently developed scenarios of ALCC by Pongratz et al. (2008), Lemmen (2009), and Kaplan et al. (2009) (see review in Gaillard et al. 2010; Fig. 25.3).

Pongratz et al. (2008) estimated the extent of cropland and pasture since AD 800 based on published maps of agricultural areas for the past three centuries and, for earlier times, a country-based method using population data as a proxy for agricultural activity. The resulting map of agricultural land was then combined with a map of potential vegetation. One of the strengths of the study is that the uncertainties associated with the approach were quantified, in particular those relating to the estimates of technological progress in agriculture and size of human populations. These ALCC scenarios were produced at a very high time resolution and used in modelling studies (see Sect. 25.4).

Lemmen (2009) developed an independent estimate of human population density, technological change and agricultural activity during the period 9500–2000 BC based on dynamical hindcasts of socio-economic development (GLUES, Global Land Use and technological Evolution Simulator; Wirtz and Lemmen 2003). The population density estimate was combined with per capita crop intensity

from HYDE (version 3.1) to infer areal demand for cropping at an annual resolution in 685 world regions. Comparison of the simulated crop fraction estimate with the HYDE estimate showed large discrepancies attributed to missing local historical data in HYDE (Lemmen 2009; Gaillard et al. 2010).

Kaplan et al. (2009) created a high-resolution, annually resolved time series of anthropogenic deforestation in Europe over the past 6000 years (referred as KK10 scenarios) by using (i) a model of the forest cover–human population relationship based on estimates of human population for the period 1000 BC to AD 1850, and (ii) a model of land suitability to cultivation and pasture, assuming that high-quality agricultural land was cleared first and marginal land next. Alternative scenarios of deforestation were also produced by taking into account technological developments, which led to major differences in south-western, south-eastern and eastern Europe (Fig. 25.3). The Kaplan et al. (2009) KK10 scenarios are also different from the HYDE database (Fig. 25.3) and provide estimates of deforestation in Europe around AD 1800 that compare better to historical accounts than the HYDE scenarios (Gaillard et al. 2009; Krzywinski and O'Connell 2009). They are also closer to pollen-inferred land-cover change over the past 6000 years (see Sect. 25.3.3 and Figs. 25.4, 25.5, 25.6 and 25.7; Gaillard et al. 2010; Trondman et al. 2011, 2012). This implies that previous attempts to quantify anthropogenic perturbation of the Holocene carbon cycle based on the HYDE and Olofsson and Hickler's databases may have underestimated early human impact.

The second approach for reconstructing past land-use changes relies on quantifying and synthesising records of land-cover change based on palaeoecological proxy data. Such proxy-based reconstructions complement model-based scenarios of ALCC and are essential to evaluate and improve their reliability.

Objective long-term records of the inferred past changes in vegetation cover are limited. Palaeoecological data, particularly fossil pollen, have been used to approximate past vegetation changes at sub-continental to global scales (e.g. Prentice and Jolly 2000; Tarasov et al. 2007; Williams et al. 2008). However, these studies have focused on tree vegetation and are of little use for a quantitative assessment of human impacts on land cover (Anderson et al. 2006; Gaillard et al. 2010). They did not resolve problems related to (i) the non-linearity of pollen–vegetation relationships in percentages, (ii) the definition of the spatial scale of vegetation represented by pollen and (iii) the differences in pollen productivity between plant taxa (e.g. Sugita et al. 1999; Gaillard et al. 2008; Gaillard 2013). A new framework of vegetation reconstruction was recently developed that resolves these problems: the Landscape Reconstruction Algorithm (LRA) (Sugita 2007a, b). This consists of two separate models, Regional Estimates of VEgetation

Fig. 25.3 Anthropogenic land use in Europe and surrounding areas at AD 800 simulated by four modelling approaches: **a** Kaplan et al. (2009) standard scenario; **b** Kaplan et al. (2009) technology scenario; **c** the HYDE 3.1 database (Klein Goldewijk et al. 2011); **d** Pongratz et al. (2008) maximum scenario. From Gaillard et al. (2010)
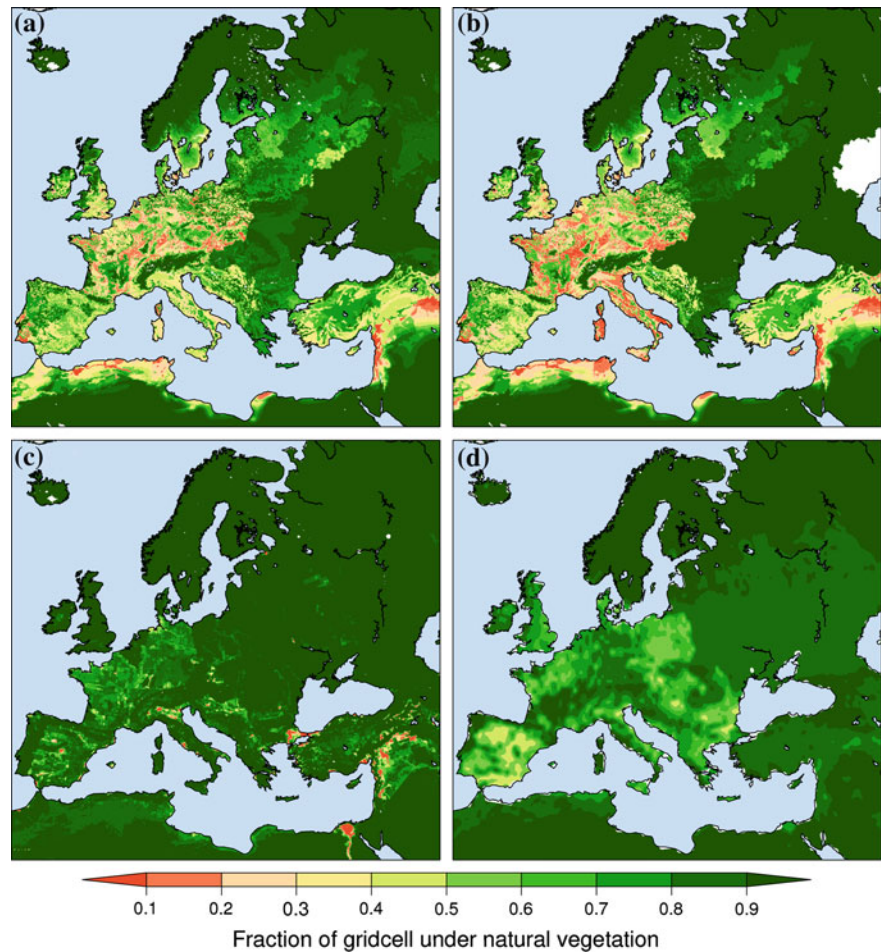
Abundance from Large Sites (REVEALS) and LOcal Vegetation Estimates (LOVE), allowing vegetation abundance to be inferred from pollen percentages at the regional (about 100 × 100 km) (REVEALS) and local spatial scales (LRA: REVEALS + LOVE), respectively. The minimum size of the area for which LRA reconstructions of local vegetation are valid is calculated by the LOVE model and varies between sites; it usually has a radius of about 0.5–3 km in southern Scandinavia (Sugita et al. 1999; Hellman et al. 2009; Fredh 2012). The LOVE model requires estimates of regional vegetation obtained using the REVEALS model. Extensive simulations support the theoretical premise of the LRA (Sugita 1994, 2007a, b). In Europe, REVEALS was empirically tested in southern Sweden (Hellman et al. 2008) and Central Europe (Soepboer et al. 2010), and the LRA (REVEALS + LOVE) in Denmark (Nielsen and Odgaard 2010; Overballe-Petersen et al. 2012) and in southern Sweden (Cui et al. 2012; Fredh 2012). The LRA approach is, to date, the best method for inferring anthropogenic land cover from pollen data (Gaillard 2013). *Human-impact pollen indicators* such as cereals, other cultivated plants, weeds and other plants favoured by human activities and cattle grazing are used

widely to interpret pollen records in terms of human-induced vegetation types (such as cultivated land, fresh/dry meadows and pastureland, ruderal land) applying the *indicator species approach*. However, these interpretations can only be qualitative. Pollen-inferred reconstruction of past human impact on vegetation is often complemented by information from plant macroremains (seeds, fruits, leaves etc.), insect remains and archaeological/historical data (e.g. Greisman and Gaillard 2009; Olsson and Lemdahl 2009, 2010).

To avoid confusion below, reconstructions of the regional vegetation are referred to as REVEALS reconstructions and reconstructions of the local vegetation as LRA reconstructions.

25.3.2 Major Past Land-Use Types in the Baltic Sea Region

The major cultural landscape/land-use types in the Baltic Sea region in the past were wood-pasture, coppices and pollards (trees cut in various ways to obtain wood for daily needs and fodder for cattle, respectively), slash-and-burn cultivation,

Fig. 25.4 Anthropogenic land-cover change in the Baltic Sea basin over the Holocene. The time trend of increasing population and agricultural land use can be seen in this series of maps covering the past 8000 years. KK10 scenarios extracted for the Baltic Sea area from Kaplan et al. (2009)
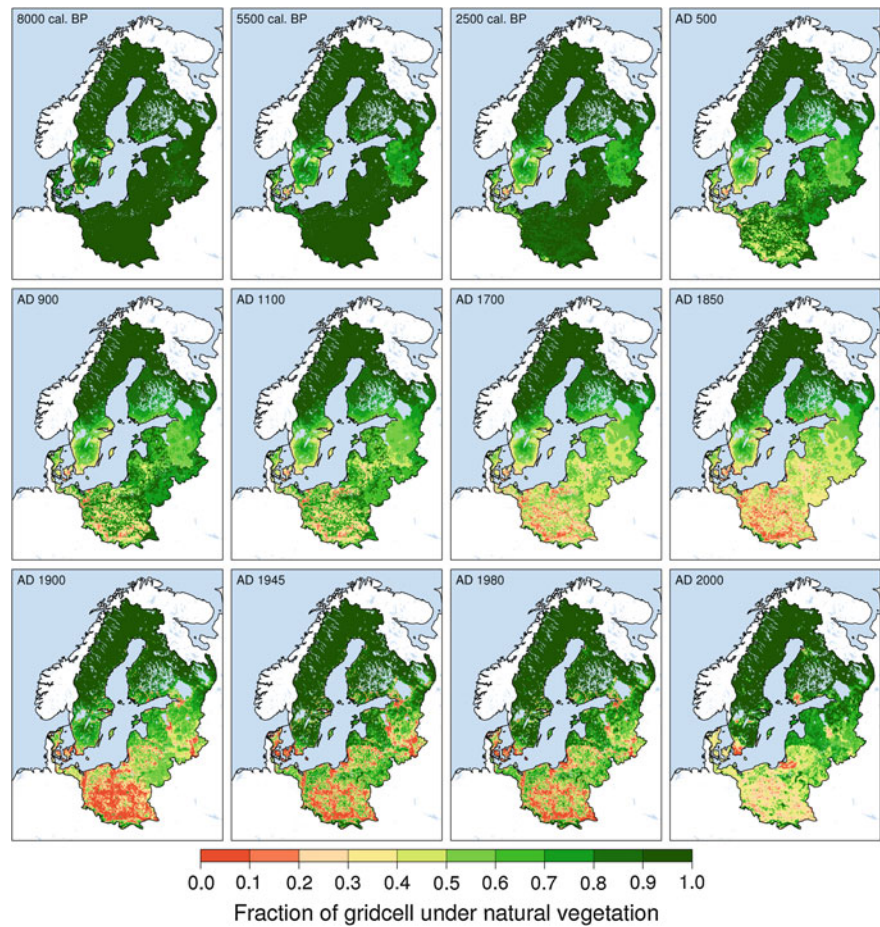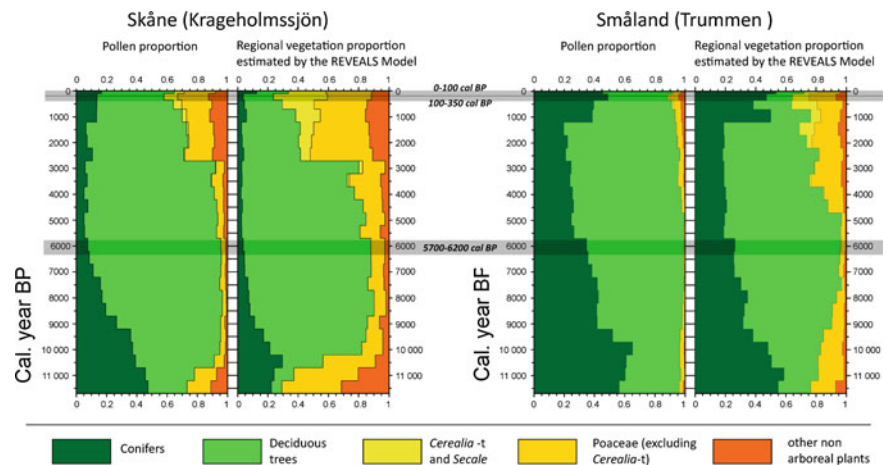

Fig. 25.5 REVEALS estimates of conifers, deciduous trees, cereals, grasses and herbs in southern Sweden, provinces of Skåne and Småland (Gaillard et al. 2010)

cultivated fields, grasslands and meadows (hay making and grazing), heathlands and summer farms (transhumance) (e.g. Gaillard et al. 2009).

Although human impact on the environment of the Baltic Sea area (and Europe in general) began in the Mesolithic (pre-Neolithic, before 6000 BP in NW Europe), it is generally accepted that farming cultures were responsible for the first

major impact on European natural environments. The pre-Neolithic lowland European landscapes are generally assumed to have been densely forested, but open land undoubtedly existed in areas where soil conditions did not allow the development of dense forests (Svenning 2002) and might also have occurred through grazing by large herbivores (e.g. Vera 2000), fire (e.g. Olsson et al. 2010; Svenning
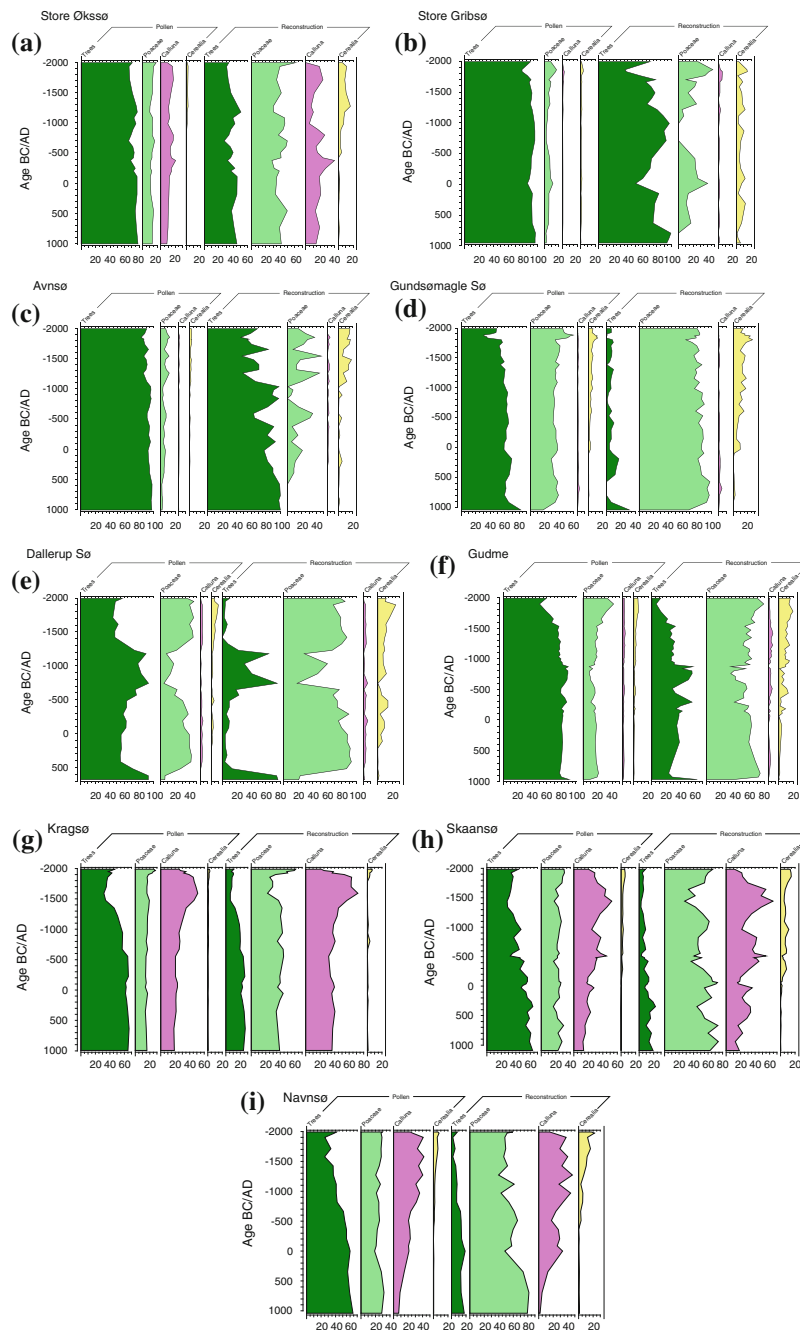
Fig. 25.6 Landscape openness in Denmark during the past 3000 years (2700 years at Dallerups Sø as modelled by the Landscape Reconstruction Algorithm (LRA)) (Sugita 2007a, b) using nine pollen records from small lakes distributed in the three major contrasting regions of the country. Pollen percentages on the *left* of each plot, LRA estimates of plant cover on the *right*. The ages are given in calibrated years BC/AD. Landscape openness is expressed by the LRA estimated cover (in percentages) of herbs (here only Gramineae (grasses, light green) and Cerealia (cereals, yellow) are shown) and low shrubs (here only Calluna (heather, violet) is shown). The LRA estimates of tree cover (*dark green*) and all herb and low shrubs (not shown here) sum up to a total of 100 %. The trees are overrepresented in the pollen percentages, while Gramineae and Cerealia (and herbs in general) are under-represented in the pollen percentages; that is, landscape openness is under-represented in the pollen percentages. The latter is true for all LRA reconstructions performed in Europe so far. In this case, heather is slightly under-represented in the pollen percentages, which is not always the case and depends on the overall vegetation composition. From Nielsen and Odgaard (2010)

2002), and the activities of Mesolithic people. However, since the beginning of the Neolithic, deforestation was a prerequisite to sow crops. Domesticated animals including cattle, sheep and pigs were introduced and also contributed to opening up the landscape. A summary of knowledge concerning the major land uses in the Baltic Sea catchment area
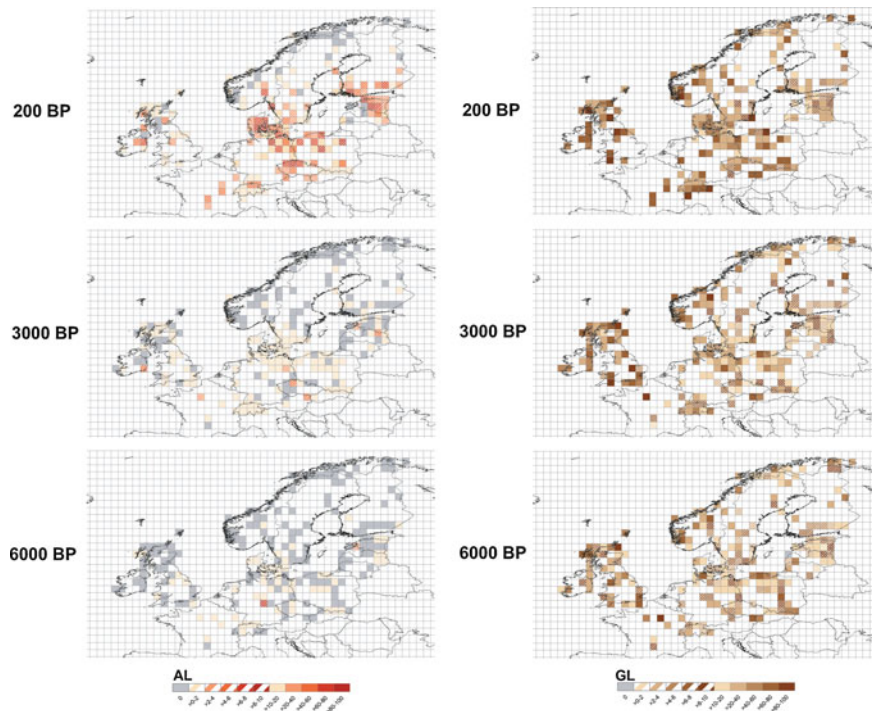
Fig. 25.7 REVEALS estimates of grassland and arable land in north-western Europe at 6000 BP, 3000 BP, and 200 BP after Trondman et al. (2011, 2012). First-generation LANDCLIM maps (produced by Anna-Kari Trondman, Florence Mazier, Anne Birgitte Nielsen, Ralph Fyfe and LANDCLIM members for the purpose of this chapter; see Gaillard et al. (2010) for a description of the LANDCLIM project). *Left* agricultural land {AL = total Cereals [Cerealia undiff., *Triticum* (wheat) type, *Avena* (oats) type, *Hordeum* (barley) type, Secale cereale (rye)]}. *Right* Grassland [GL = Cyperaceae, Filipendula, *Plantago lanceolata*, *Plantago montana*, *Plantago media*, Poaceae, *Rumex* p.p. (mainly *R. acetosa* and *R. acetosella*, i.e. *Rumex acetosa* pollen-morphological type)]. This first generation of LANDCLIM REVEALS estimates is based on all available Holocene pollen records from small and large sites (lakes and bogs) including all or part of the time windows 6000, 3000 and 200 BP (1950), and with ≥3 dates for the chronological control. These pollen records were collected from the European (EPD), Alpine (ALPADABA) and Czech (PALYCZ) pollen databases (van der Knaap et al. 2000; Fyfe et al. 2007; Kuneš et al. 2009, respectively). The pollen productivity estimates (PPEs) used are, for each taxon, the mean of all PPEs obtained within the project area (north-western Europe and western Europe north of the Alps, Broström et al. 2008). From Gaillard (2013)

over the last 6000 years follows. For more complete reviews and bibliography on the subject, see in particular Behre (1988), Berglund (1991) and Gaillard et al. (2009).

Wood-pasture was of major importance until late mediaeval times. Coppices and pollards can be traced back through written sources to the sixteenth century in north-western Europe. There is evidence from archaeological contexts for coppicing and pollarding as far back as the Mesolithic and Neolithic, respectively. Coppices and pollards were progressively abandoned during the twentieth century and replaced by open pastures and cultivated fields or developed into secondary, broadleaved forests. Slash-and-burn is also a form of woodland use. Cereals (often rye) were sown in clearings created by felling, drying and burning the woody vegetation, which enriched poor soils with ash. In Finland, slash-and-burn started with arable farming in the Neolithic and lasted until the early twentieth century. In Sweden, slash-and-burn was mainly associated with spruce forests and was common after the expansion of spruce from

the north (3000–1000 BP). It was practiced so intensively in the eighteenth and nineteenth centuries that woodlands failed to fully regenerate. For that reason and the associated fire hazard, slash-and-burn was prohibited in many parts of Scandinavia in the early twentieth century.

Pastoral activity has had a fundamental influence on the landscape and vegetation of the Baltic Sea region since the Neolithic. The major factor involved in the formation and maintenance of *pastures* and *meadows* was the need for fodder. Denser settlements, rising populations and increased demand for food resulted in an increase in livestock, which in turn demanded more grazing land and meadows. The practice of hay making brought about the development of the *infield/outland* system that optimised the available land resources thanks to an efficient regime for nutrient recycling. The infield included cultivated fields and hay meadows, while the outland was grazing land (grassland, heaths and/or forest). Livestock was also necessary to fertilise soils for cultivation, and the hay meadows had to be large enough to

sustain the amount of fodder necessary for the livestock required to fertilise the area of crop fields that would cover the food demand of the human population. In other words, the hay meadows were essential to the crop fields and their size had to be three to 20 times larger than the cultivated area depending on the soil conditions (Fogelfors 1997). Hay making is often associated with the introduction of the scythe in the Late Iron Age (ca. AD 1000). However, species-rich hay meadows may have already existed in southern Sweden at the end of the Late Bronze Age/beginning of the Early Iron Age (from ca. 2600 years BP; Gaillard et al. 1994). Mowing (hay making) probably started with the practice of stalling that may have first occurred in connection with a climate cooling in north-western Europe dated to ca. 3000 BP (e.g. Berglund 2000).

The history of heathland can also be traced back to the Neolithic. For example, in Denmark and southern Sweden, pollen analysis of soils beneath Neolithic mounds has shown that heathland arose from woodland clearance on poor sandy soils. Heathland was widespread on relatively poor soils (often in areas characterised by granitic bedrock or areas outside the maximal ice extent of the Weichselian) in large parts of southern Scandinavia and neighbouring countries around the Baltic Sea (e.g. Greisman 2009; Olsson and Lemdahl 2009). The development of wooded or treeless pastures and hay meadows in upland and northern regions of the Baltic Sea region is closely linked to upland summer grazing and collection of fodder. There is still very little known about the history of summer grazing in the region, except in the province of Värmland in northern Sweden (e.g. Regnéll and Olsson 1998) where remains of summer farms were dated to mediaeval time. During the twentieth century, traditionally managed hay meadows, pastures, heathland and upland summer farming decreased dramatically with the introduction of chemical fertilisers and feed concentrates, reclamation and afforestation.

25.3.3 Land-Use and Anthropogenic Land-Cover Change Since the Neolithic (6000 BP)

The account presented here is based on ALCC model scenarios (Figs. 25.3 and 25.4), recent REVEALS and LRA reconstructions (Sugita et al. 2008; Gaillard et al. 2010; Nielsen and Odgaard 2010; Nielsen et al. 2012; Trondman et al. 2012) (Figs. 25.5, 25.6 and 25.7), earlier syntheses of palaeoecological proxy records, of which the most important are those by Berglund (1991), Berglund et al. (1996, 2002) and Ralska-Jasiewiczowa et al. (2004), and a large number of palaeoecological studies of which only a very small fraction is cited below. Examples are provided from the major environmental zones (according to Metzger et al.

2005) of the region, that is (i) the Nemoral, Atlantic North and Continental zones in the south, and (ii) the Boreal and Alpine North zones in the north, each representing about 50 % of the total land cover of the Baltic Sea basin.

25.3.3.1 Neolithic to Iron Age (ca. 6000–1000 BP)

According to archaeological and palaeoecological data, arable farming was introduced in the loess areas of central Germany with the Linear Pottery culture around 7700–7500 BP (Kalis et al. 2003), which is reflected in pollen diagrams from the area (e.g. Beug 1992; Voigt 2006). However, along the coasts of the Baltic Sea in northern Germany, Denmark, southern Sweden, and northern Poland (north of the Elbe river), the Mesolithic Ertebølle culture persisted for a long time, possibly because of the good fishing and hunting possibilities (e.g. Regnéll et al. 1995; Kalis et al. 2003; Richards et al. 2003). Larger scale crop cultivation and animal husbandry occurred first with the Neolithic Funnel Beaker culture from around 6100 BP in north-eastern Germany and northern Poland, 5900 BP in Denmark (Richards et al. 2003) and 5900 BP in southern Sweden (e.g. Berglund 1991). The earliest cultural impact on the landscape consisted mainly of a change in forest composition towards more early-successional species (birch, hazel), but from Late Neolithic, (ca. 4300–3800 BP in southern Scandinavia) anthropogenic grassland increased in size, while the areas with arable fields were still relatively small (e.g. Berglund et al. 2002; Odgaard and Nielsen 2009). In most of the southern environmental zones, a gradual differentiation of the landscape into three more or less distinct types occurred from the Late Neolithic onwards: (i) flat areas on clay-rich moraine soils developed the most intensive agricultural impact; (ii) hilly areas were more marginal in terms of agricultural activities and, therefore, remained rich in forest; and (iii) poor sandy soils gradually became dominated by heathland (e.g. in Denmark and southern Sweden, Berglund 1991; Odgaard and Rasmussen 2000; Berglund et al. 2002; Lagerås 2007; Greisman 2009; Nielsen and Odgaard 2010). By the Mid-/Late Bronze Age (ca. 3000–2500 BP), this division was well established and the overall pattern remained in place until around AD 1800, although the composition and distribution of the landscape types varied in time and space, and the total landscape openness increased through time. In some marginal areas of Denmark and southern Sweden, and along the Baltic Sea coast, in particular in the north-eastern part of Germany, northern Poland and the Baltic States, regional forest regeneration occurred during the Migration period (ca. 1600–1450 BP or AD 400–550; Andersen and Berglund 1994).

According to the ALCC scenarios of Kaplan et al. (2009), the earliest significant deforestation in the Baltic Sea basin occurred in the earliest Neolithic period on fertile soils, and

by the Viking Age (ca. 1200–1000 BP), large areas of present-day Denmark, southern Sweden and Poland (i.e. the southern environmental zones) were ≥50 % under human use for crop and pasture land (Figs. 25.3 and 25.4). On the other hand, the scenarios by Pongratz et al. (2008) indicate that by 1200 BP, only about 3 % of the area potentially covered by vegetation on the globe was transformed to agricultural land, almost as much for cropland as for pastureland, none of the Baltic Sea basin (and almost none of Europe) was deforested by more than 50 %, and most of the region was deforested by 20 % or less except for Denmark, northern Germany, southernmost Sweden (Skåne Province) and Poland (Fig. 25.3). Thus, although both models identify the same regions as the most deforested, discrepancies are large between the estimates of the deforested land fraction for crop cultivation and pastures. Pollen-based REVEALS and LRA estimates of regional and local openness at 1000–1200 BP (≥50 % cover) agree with the Kaplan et al. (2009) scenarios for the regions of Skåne and Småland (southern Sweden) (Sugita et al. 2008; Gaillard et al. 2010; Fig. 25.5), most of Denmark (Nielsen and Odgaard 2010; Nielsen et al. 2012; Fig. 25.6), and north-western Europe in general (Trondman et al. 2011, 2012; Fig. 25.7), although between-site differences may be large at the local spatial scale (e.g. Denmark). The REVEALS-based reconstructions suggest that changes in human impact on vegetation/land cover over the past 6000 years were much more profound than suggested by earlier interpretations of pollen percentages; that is, the share of non-forested land through the Holocene is strongly underestimated by percentages of non-arboreal pollen (NAP, i.e. pollen from herbaceous plants) (Figs. 25.5 and 25.7). For instance, the REVEALS estimates of regional openness in the southernmost province of Sweden, Skåne, are 15–30 % for 6000–2750 BP and about 60 % for 2750–1000 BP, compared to 5–10 % and 30 % herb pollen, respectively, while in the province of Småland, north of Skåne, they are <10 % for 6000–4500 BP, 10–25 % for 4500–2000 BP and 25–30 % for 2000–1000 BP, compared to <2 %, <5 % and about 5 % herb pollen, respectively. These REVEALS reconstructions suggest that by the Late Bronze Age/beginning of the Iron Age, large areas of southern Sweden were under human use for crop cultivation and pastures. In Denmark, LRA estimates of local vegetation show that, on the richest soils, the openness often reached values over 80 % from ca. 3000 BP, while hilly areas were characterised by less openness (seldom over 40 %). On poor sandy soils, open heathland was dominant (70 % to over 80 %) (Nielsen and Odgaard 2010; Fig. 25.6). The maps of Fig. 25.7 clearly show the strong increase in size of the land surfaces covered by cultivated land (here exclusively cultivated with cereals) and grassland (here mainly grasses) between 6000 BP (Neolithic) and 3000 BP (Late Bronze Age), and between 3000 and 200 BP (AD 1750–1800). The

pollen-based REVEALS and LRA reconstructions indicate that the ALCC scenarios by Kaplan et al. (2009) are reasonable, except perhaps the degree of deforestation in the Neolithic time (5500 BP), which in some areas are too high compared with REVEALS reconstructions from southern Sweden (Gaillard et al. 2010; Trondman et al. 2011, 2012; Figs. 25.5 and 25.7) and from Denmark and northern Germany (Nielsen et al. 2012).

25.3.3.2 Middle Ages (AD 1050–1500)

The ALCC scenarios of Kaplan et al. (2009) showed that deforestation intensified in Poland and the Baltic countries from the mediaeval period onwards. However, the major difference is seen between AD 700 and 900 (ca. 1300–1100 BP), in particular in the southern part of the Baltic Sea basin (Denmark, northern Germany, Poland), which increases up to 10–20 % deforestation (Fig. 25.4). This is in good agreement with the REVEALS and LRA-based reconstructions of regional and local vegetation that suggest increases of deforestation in early mediaeval time up to 10–15 % in southernmost Sweden, and up to about 20–40 % on rich soils and marginal areas of Denmark.

According to pollen and other palaeoecological studies, the land under agriculture expanded in area in the Middle Ages, resulting in a significant increase in landscape openness in the southern environmental zones of the Baltic Sea catchment. This was also a period with technological advances in agriculture (Porsmose 1999) and changes in crop composition (e.g. Behre 1992; Robinson et al. 2009). In Denmark for instance, open-land areas increased especially in the period AD 1200–1400, earliest in the core agricultural areas, and about 100 years later in the more forested areas (LOVE estimates, Fig. 25.6; Odgaard and Nielsen 2009). In the heathland regions in the west, the last forests disappeared (Odgaard and Nielsen 2009). The landscape also became more open in north-eastern Germany (Nielsen et al. 2012), and in southern Sweden, regional openness reached 80 % and 35 % in Skåne and Småland, respectively (REVEALS estimates, Fig. 25.5). Nevertheless, large parts of Småland were characterised by much larger openness in areas where grazed heathland expanded (Greisman and Gaillard 2009; Marlon et al. 2010; Cui et al. 2012).

25.3.3.3 Modern Time (AD 1500–2000)

The ALCC scenarios of Kaplan et al. (2009) indicate a progressive increase in deforestation of the region, in particular its southern part, reaching a peak around AD 1900. The twentieth century in the Baltic Sea basin is characterised by a period of land abandonment that is especially marked during the period 1980–2000, however, mainly confined to Denmark, northern Germany, the Baltic States and Poland (Fig. 25.4). At a global scale, the ALCC scenarios of Pongratz et al. (2008) indicate that, around AD 1700, the

agricultural area had increased to about 9 % of the area potentially covered by vegetation on the globe (PGV), of which 3.5 % was cleared forest (85 % for cropland, 15 % for pasture) and 5.5 % was grassland and shrubland under human use (30 % for the cultivation of crops). Between AD 800 and 1700, the ALCC scenarios show that natural vegetation under agricultural use had increased by about 5 million km^2 (i.e. about 6 % of PGV). Within the next 300 years, the total agricultural area increased to about 50 % of PGV (mainly pastureland), that is roughly a 5.5 times larger area than at AD 1700. This reconstruction suggests that global ALCC was small between AD 800 and 1700 compared to the industrial time, but relatively large compared to previous millennia. During the preindustrial period of the twentieth century, the reconstruction shows clear between-region differences in histories of agriculture (Pongratz et al. 2008). However, regional reconstructions for the Baltic Sea region based on pollen records, other palaeoecological proxies, and archaeological/historical data differ significantly from the global picture proposed by Pongratz et al. (2008). According to the REVEALS and LRA model-based reconstructions, the deforested area did increase between AD 800 and 1700, however, by not more than about 50 % of the earlier deforestation. The increase in deforestation between about AD 1700 and 1850/1900 does not represent more than 50 % of the landscape openness at AD 1700, in many areas much less (10–20 %).

The pollen-based reconstructions indicate that the percentage cover of cereals was very high in the eastern parts of northern Germany and northern Poland from AD 1500 onwards, and lower in north-western Germany, Denmark and southern Sweden, where grazed grassland and heathland were the dominant human-induced vegetation types (Berglund et al. 2002; Berglund 2006; Nielsen et al. 2012). This agrees with the archaeological findings and historical sources indicating that cereals were imported to Denmark and Sweden from areas south of the Baltic Sea region (e.g. Robinson et al. 2009), while cattle were exported in large numbers from Denmark and Schleswig-Holstein to other parts of northern Germany and to the Netherlands from the fourteenth to mid-eighteenth century (Gijsbers and Koolmees 2001; Bruun and Fritzbøger 2002). Grazed heathlands had their maximum extent in the entire Baltic Sea region around AD 1500–1800. Thereafter, many heathland areas— as well as permanent grasslands and meadows—were converted into arable land or planted forests, especially with conifers (e.g. Eliasson 2002; Dahlström 2008; Frederiksen et al. 2009; Gaillard et al. 2009). Urban areas also expanded, especially after AD 1900 (e.g. Frederiksen et al. 2009; Münier 2009). In southern Sweden, southern Norway and the Baltic states, the landscape openness was at its maximum around AD 1850. Since then, urbanisation, abandonment of agrarian landscapes, land-use change and modern forestry

have led to reforestation of large areas formerly used for agriculture (see also Chap. 21). This trend is not unique to the Baltic Sea region, but is also characteristic of many other regions of Europe for which the nineteenth century was the time of most intensive land use with maximum landscape openness, while the twentieth century was characterised by reforestation after abandonment and/or through plantation, as for example in southern Norway, northern Italy, central France, the Pyrenees, central Spain and Portugal (Gaillard et al. 2009; Krzywinski and O'Connell 2009).

25.4 Effects of Land-Cover Change on Past Climate: Model Experiments

This section reviews available studies on the effect of long-term CLCC and ALCC on past climate in the northern hemisphere and Europe.

25.4.1 AD 1850 to Modern Time

The LUCID project compares responses to historical ALCCs in various climate models in a series of studies (Pitman et al. 2009; de Noblet-Ducoudré et al. 2012). Pitman et al. (2009) concluded that there was general agreement on the significant effect of vegetation patterns and land-cover change on regional climate, while their role on global climate was still under debate. In particular, the effect of teleconnections related to land-cover change was considered questionable; that is, whether a change in the climate in a given region could be related to land-cover change in other regions. Some climate modelling studies suggest that such teleconnections exist (Henderson-Sellers et al. 1993; Zhang et al. 1996; Gedney and Valdes 2000; Werth and Avissar 2002, 2005), while others indicate they do not (Findell et al. 2007, 2009; Pitman et al. 2009). The following text reviews some of the earlier studies addressing these questions.

At the global scale, Sheng et al. (2010) identified hot spots of climate-induced change in surface energy fluxes during the period 1948–2000, although these were not related to land-cover change but rather to variability in atmospheric–surface interactions. The hot spots were primarily found in northern high-latitude areas. Based on observations, Teuling et al. (2010) investigated how differences in water and energy exchange due to the differences in land cover in the temperate forest zone affected the European heatwave in August 2003. They concluded that grassland and forest areas react differently to changes in soil water availability. As long as water availability is high, woodland exerts a warming effect compared to grassland due to higher Bowen ratios over woodland; that is, more of the available

net radiation energy at the surface is used for vertical heat transfer than for evapotranspiration (see also Bonan 2008). However, woodland can also sustain its evapotranspiration rate when water availability is low, which leads to lower Bowen ratios than in grassland in dry conditions; that is, grassland becomes the source of excess heating instead of woodland.

There are still important problems in relation to how ALCCs are explored in numerical experiments using climate models. Pielke et al. (2011) concluded that most studies were based on only one or two models, which did not reflect the uncertainty between models in their responses to increased CO_2 levels or in the strength of their land–atmosphere interactions, an uncertainty that is evident in the LUCID multi-model study of Pitman et al. (2009). That study mainly focused on the Northern Hemisphere summer season, and the key result was a statistically significant impact of ALCC on the simulated latent heat flux and air temperature over the regions where anthropogenic land cover changed, but the direction of the change in summer temperature was inconsistent among the models. In terms of rainfall, four of the coupled atmosphere–land models used showed a significant impact on rainfall over regions with ALCC, while three models did not show impacts greater than the expected random variability of model outputs (Seneviratne et al. 2010). In their review, Pitman et al. (2009) did not find statistically significant impacts of ALCC on latent heat flux, temperature or rainfall remote from the actual ALCC, that is no teleconnections. The authors also suggested that robust conclusions on the effects of ALCCs on climate can only be drawn from multi-model experiments. Studies based on a single model only provide indications of possible feedback mechanisms and their implications.

The goal of the most recently published part of the LUCID project (de Noblet-Ducoudré et al. 2012) was to provide a detailed examination of why the LSMs diverge in their response to ALCC. For this purpose, the authors used seven atmosphere–land models with a common experimental design. For the vegetation distribution, each model used as a starting point the same distribution of crop and pasture, at a resolution of $0.5° \times 0.5°$, as extracted from Ramankutty and Foley (1999), combined with the pasture areas from Klein Goldewijk et al. (2011) (Fig. 25.8). However, as there are between-model differences in (i) the way land information was represented, (ii) sources of information to describe present-day and potential vegetation, and (iii) strategies to implement ALCC in the model, the resulting land-cover distribution (including natural vegetation) used in each model differed (de Noblet-Ducoudré et al. 2012; Fig. 25.8). Although the areas covered by crops increased from AD 1870 to AD 1992 in all land-cover data sets used, the increase varied; all LSMs describe temperate deforestation, but at

varying degrees. Within the Eurasian region studied in LUCID, western Europe is characterised by reforestation (i.e. land abandonment and forest planting) rather than deforestation, and the Baltic Sea catchment area by mixed deforestation and reforestation, while the entire Eurasian region exhibits overall deforestation (Figs. 25.8 and 25.9). These differences in ALCC implementations between the LUCID model runs had an influence on how ALCC affected the near-surface climate in the models' results; that is, there is no consistency in how ALCC influenced the partitioning of available energy between latent and sensible heat fluxes at a specific time (Boisier et al. 2012; de Noblet-Ducoudré et al. 2012).

These results highlight the urgent need to evaluate LSMs more thoroughly. However, there are some robust common features shared by all models: the amount of available energy used for turbulent fluxes and the almost linear relationship between the climate response to ALCC and the amount of trees removed. All models simulated a systematic increase in surface albedo in all seasons. For most models, this increase (7 % for a full transition from forest to crop/grassland) was proportional to the amount of deforestation imposed on the individual models. Moreover, the larger surface albedo was shown to cause a decrease in QA (computed as the sum of absorbed solar energy and incident atmospheric infrared radiation); QA decreased in all seasons everywhere in the temperate regions and was also proportional to the amount of deforestation imposed on a given model. In most cases, crops and grasslands were less efficient than trees in transferring energy to the atmosphere in the form of turbulent fluxes due to a lower aerodynamic roughness length. All models that underwent a change in their forest fraction that was greater than 15 % simulated cooler ambient air temperature in all seasons. These common features and their dependence on the ALCC descriptions prescribed in each model suggest that, for a specified amount of deforestation occurring over specific periods, the dispersion among the models would be significantly smaller if the ALCC descriptions had been exactly the same in all models. LUCID also compared the biogeophysical impacts of ALCC with the impact of elevated greenhouse gas concentrations on sea surface temperatures and sea-ice extent. The results show that ALCC had an impact of similar magnitude—but of opposite sign—to increased greenhouse gases and warmer oceans. However, it should be stressed that although this result is valid for the entire Eurasian region, this is not necessarily the case for individual parts of that large region, such as the Baltic Sea catchment area. Moreover, in view of the dominant reforestation of the catchment's western part and deforestation of its eastern part, it is not possible to predict the net effect of ALCC at the scale of the entire catchment area without modelling the effects of deforestation and reforestation at the regional scale using regional climate models.
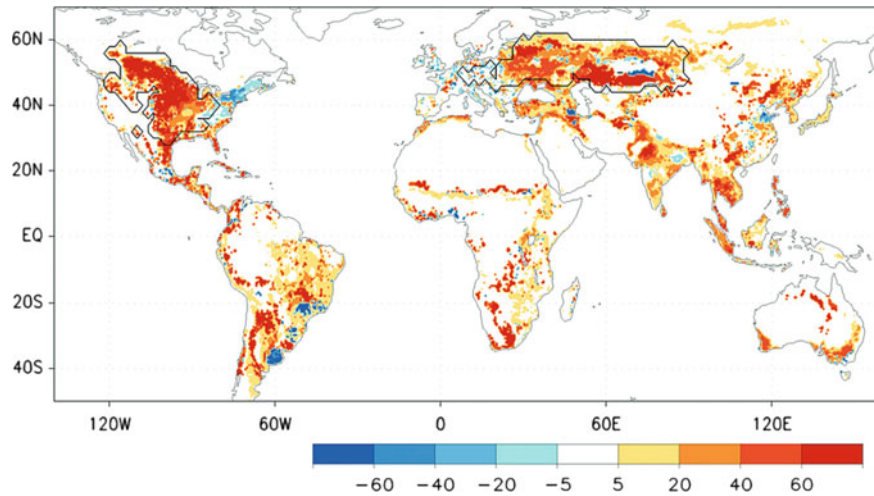
Fig. 25.8 Changes in the extent of agricultural land (crops and pastureland) between pre-industrial time (AD 1870) and present day (AD 1992). *Yellow* and *red* indicate an increase, and *blue* a decrease, in the extent of agricultural land since AD 1870. The two contours on the map indicate the regions used for specific analysis (North America and Eurasia) (de Noblet-Ducoudré et al. 2012)

Fig. 25.9 Vegetation descriptions used in the LUCID project (de Noblet-Ducoudré et al. 2012). **a** Vegetation fraction in AD 1870 in the studied Eurasian region (shown in Fig. 25.8). Surface (in fraction of total area) covered by crops (*grey*), grassland types (*orange*), evergreen trees (*green*), deciduous trees (*blue*) and desert (*white*) for all seven models used in the LUCID project (for details on the seven models, see de Noblet-Ducoudré et al. 2012); **b** Fraction difference (AD 1992– AD 1870) (in fraction of total area) for each of the vegetation types shown in (**a**). The *dashed black line* in both graphs (**a**) and (**b**) shows the crop fraction that was finally implemented in all seven models (de Noblet-Ducoudré et al. 2012)
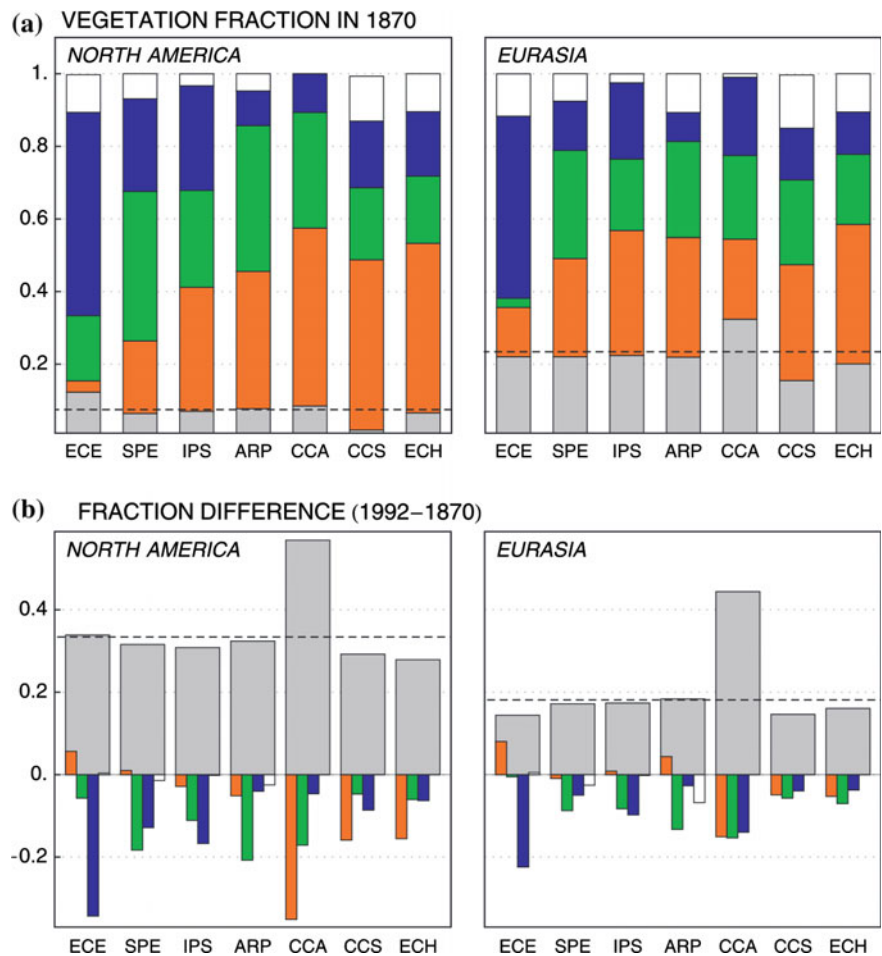
25.4.2   Before AD 1850

There are few studies of land cover–climate interactions before AD 1850, although the number has increased rapidly since 2009. To date, there are no published modelling studies on the possible feedbacks of CLCC on past natural climate warming such as the well-known Early Holocene increase in mean annual temperatures and on the effects of Late Holocene CLCC and/or ALCC on later climate changes such as the Medieval Climate Anomaly (warming) and the Little Ice Age (cooling, see also Chap. 3), except that of Pongratz et al. 2009b. However, simulations of the north-wards expansion of trees due to a warmer climate showed that such climate-induced vegetation changes produced a positive feedback on climate (e.g. Cheddadi et al. 1997).

There is only one study of the effect of past ALCC on regional climate in Europe using a regional climate model, the LANDCLIM project (Gaillard et al. 2010; Strandberg et al. 2013). The aim of the study was to evaluate the direct effects of anthropogenic deforestation on simulated climate at two contrasting times of the Holocene ∼6000 BP and ∼200 BP in Europe applying RCA3, a regional climate model with 50 km spatial resolution (Samuelsson et al. 2011). Three alternative descriptions of the past vegetation were used: (i) potential natural vegetation (V) simulated by the dynamic vegetation model LPJ-GUESS (Smith et al. 2001), (ii) potential vegetation with anthropogenic land cover (deforestation) as simulated by the HYDE model (V + H), and (iii) potential vegetation with anthropogenic land cover as simulated by the KK model (V + K). The climate model results show that the simulated effects of deforestation depend on both local/regional climate and vegetation characteristics (Strandberg et al. 2013). At ∼6000 BP, the extent of simulated deforestation in Europe is generally small, but there are areas where deforestation as simulated by Kaplan et al. (2009) (V + K) is large enough to produce significant differences in summer temperature of 0.5–1 °C, which is the case in southern Sweden and eastern Poland. However, the KK model overestimates deforestation in these areas compared to the pollen-based REVEALS reconstructions. At ∼200 BP, simulated deforestation is much more extensive than previously assumed, in particular according to the pollen-based REVEALS reconstructions (see Sect. 25.3.3.3) and the KK model. This leads to sig-nificant temperature differences in large parts of Europe. In winter, deforestation leads to lower temperatures because of the differences in albedo between forested and unforested areas, particularly in the snow-covered regions. In summer, deforestation leads to higher temperatures in Central and eastern Europe since evapotranspiration from unforested areas is lower than from forests (*hydrological cycle effect*).

Summer evaporation is already limited in the southernmost parts of Europe under potential vegetation conditions and, therefore, cannot become much lower. Accordingly, the *albedo effect* dominates also in summer, which implies that deforestation causes a decrease in temperature. Differences in summer temperature due to deforestation range from −1 °C in south-western Europe (cooling) to +1 °C in eastern Europe (warming). In the Baltic Sea area, the effects of deforestation at 200 BP are much weaker than in south-western and eastern Europe. The effect is strongest in southern Sweden where deforestation leads to lower winter temperatures by only 0.2–0.4 °C, but there is no effect in summer. The choice of anthropogenic land-cover estimate was shown to have a significant influence on the simulated climate. But the climate proxy data available for the two time windows are not precise enough to evaluate the results of the climate model runs in quantitative terms effectively.

Earlier studies also show that the *albedo effect* of his-torical deforestation was probably dominant among the effects of deforestation in the northern atmosphere. How-ever, most studies suffer from the drawback that the grid resolution is rather coarse and far coarser than the scale necessary to capture local to regional processes (Hibbard et al. 2007). Brovkin et al. (2006) used the scenarios of past deforestation produced by Ramankutty and Foley (1999) for the period AD 1800 to present-day and the HYDE database to reconstruct the effects of ALCC on climate over the past 1000 years. The outputs from six different climate models showed a cooling of 0.1–0.4 °C over the Northern Hemi-sphere due to the biogeophysical effects (mainly increased albedo) of the estimated decrease in forest cover between AD 1000 and 2000. They also found a warming of similar magnitude due to the biogeochemical effects of ALCC, therefore a net effect close to zero. Pongratz et al. (2009b) investigated the influence of historical land-use changes on radiative forcing (RF). For all of Europe, except Scandina-via, a decrease of 0.3 W m^{-2} was found between AD 800 and 1700. At the global scale, the RF was small throughout the pre-industrial period (negative with a magnitude less than 0.05 W m^{-2}) and not strong enough to explain the cooling reconstructed from climate proxies between AD 1000 and 1900 (Little Ice Age).

To date, there are few estimates of CO_2 emissions due to historical ALCC at the sub-continental scale and none using regional climate models. In the context of this review, the most interesting study so far is that of Pongratz et al. (2010) that separated the relative strength of biogeochemical versus biogeophysical effects from ALCC during the past millen-nium using a coupled atmosphere–ocean general circulation model (AOGCM) and applying the reconstruction of his-torical ALCC of Pongratz et al. (2008). They found that

biogeophysical effects had a slight cooling influence on global mean temperature (-0.03 °C in the twentieth century), while biogeochemical effects led to a strong warming (0.16–0.18 °C). During the industrial era, both effects caused significant changes in certain regions, but only a few regions experienced a biogeophysical cooling strong enough to dominate the overall temperature response. The authors concluded that the climate response to historical ALCC, both globally and in most regions, was dominated by the rise in CO_2 caused by ALCC emissions. However, the biogeophysical temperature response at the regional scale was greater than suggested by its global mean. For example, in Europe, the annual mean temperature decreased by 0.3–0.5 °C, and the cooling in northern high and mid-latitudes was found to be largely albedo-driven, leading to a winter cooling of up to 0.9 °C in north-eastern Europe, in general accordance with previous studies (e.g. Betts 2001). However, the albedo dominance over hydrological aspects in the Pongratz et al. (2010) study is only significant for the annual mean temperature, whereas transpiration effects are in some cases seasonally offsetting. The authors also concluded that strong local biogeophysical effects could substantially influence the spatial pattern of the net temperature response, as in eastern Europe for example. The global versus local effectiveness of biogeochemical versus biogeophysical effects was also demonstrated by the fact that, at the global scale, the entire land area was more strongly influenced by biogeochemical warming than the ocean, while biogeophysical cooling is particularly pronounced over agricultural areas. Pongratz et al. (2011) also quantified the contribution of local ALCC to historical global warming and showed the importance of past land-use decisions in influencing the mitigation potential of reforestation on these lands. In these simulations, they found that CO_2 warming dominated over albedo cooling at the global scale because past land-use decisions resulted in the use of the most productive land with larger carbon stocks and less snow than on average. Therefore, land-use decisions led to CO_2 warming in most agriculturally important regions of the world. This suggests that, in most places, reversion of past land-cover change may often be the most feasible step of implementing ALCC as a mitigation tool. However, because the amount of CO_2 emissions and the change in biophysical properties vary across regions and types of land-cover change, detailed analysis—that is, simulation of the regional climate response to local occurrence of ALCC—is needed for specific reforestation projects. The climate effect of past ALCC is likely to be a good indicator of the mitigation potential of reversing the area to its natural state.

The rest of this section summarises other studies of global-scale carbon emissions. Pongratz et al. (2009a) performed transient simulations over the entire last millennium with a GCM that couples the atmosphere, ocean and land surface with a closed carbon cycle. By applying the ALCC of Pongratz et al. (2008) as the only forcing to the climate system, they showed that the terrestrial biosphere experienced a net loss of 96 Gt C over the last millennium, leading to an increase in atmospheric CO_2 by 20 ppm. The biosphere–atmosphere coupling led therefore to a restoration of 37 and 48 % of the primary emissions over the industrial period (AD 1850–2000) and pre-industrial period (AD 800–1850), respectively. Atmospheric CO_2 rose above natural variability by late mediaeval times, but global mean temperatures were not significantly altered until strong population growth in the industrial period. Pongratz et al. (2009a) also found that only long-lasting epidemics or wars led to carbon sequestration because emissions from past ALCC compensate carbon uptake in 'regrowing' vegetation for several decades. Reick et al. (2010) derived the CO_2 emissions associated with ALCCs since AD 800 as reconstructed by Pongratz et al. (2008) and compared them with the pre-industrial development of atmospheric CO_2 known from Antarctic ice cores. They concluded that pre-industrial traces of CO_2 emissions from ALCC before AD 1750 was obscured by other processes of similar magnitude, while the steep increase in atmospheric CO_2 after AD 1750 and until AD 1850 (i.e. before the rise of fossil fuel emissions to significant values) was largely explained by rising emissions from ALCC. These results partly contrast with those of Kaplan et al. (2010) who found that by AD 1850, at the global scale, the cumulative CO_2 emissions due to deforestation since 6000 BC were 137–189 Pg C (using the HYDE scenarios of ALCC) and 325–357 Pg C (using the KK ALCC scenarios of Kaplan et al. (2009)). Kaplan et al. (2010) concluded that their results support the hypothesis that anthropogenic activities led to the stabilisation of atmospheric CO_2 concentrations at a level that made the world substantially warmer than it would otherwise have been. Similarly, Boyle et al. (2011) showed by using new model assumptions that the quantity of terrestrial carbon release due to early farming, even using the most conservative assumptions, greatly exceeds the net terrestrial carbon release estimated by inverse modelling of ice core data by Elsig et al. (2009). However, the conclusions of both Kaplan et al. (2010) and Boyle et al. (2011) remain an open question as the emission estimates are not compatible with current understanding of the global carbon cycle and records from ice cores.

25.5 Potential Future Trends in Land Cover and Associated Effects on Future Climate

25.5.1 Future Land-Cover Change Due to Anthropogenic Climate Change and Possible Feedbacks

Many of the existing scenarios of future trends in land-cover change are based on observations of vegetation change due to the recent climate warming, but also on socio-economic assumptions. Studies on future climate-induced (potential natural) vegetation change and related biogeophysical feedbacks to regional climate provide some indication of what to expect in the future, since the underlying mechanisms are likely to be similar for human-induced vegetation change. Such studies in Europe indicate a boreal treeline advance into the tundra regions of the northern latitudes of both the Barents Sea region (Göttel et al. 2008) and northernmost Europe (Wramneby et al. 2010; Smith et al. 2011). The most significant feedback associated with forest expansion at these latitudes is expected to be the albedo feedback (warming) that is likely to be strong enough to offset the climate gains from the increased carbon sequestration in these forests. Using the coupled regional climate–vegetation model RCA-GUESS (Smith et al. 2011) at the European scale, Wramneby et al. (2010) also showed that a future rise in the altitudinal limit of deciduous trees in the Scandinavian mountains due to increased temperature would lead to enhanced warming through the positive snow–vegetation–albedo feedback.

While the albedo feedback and its amplifying effect on climate warming is expected to be the most important biogeophysical feedback in boreal regions such as northern Europe (Strengers et al. 2010), an increase in forest cover also implies a contrasting biogeophysical feedback mechanism due to enhanced evapotranspiration. This feedback may, however, be of minor importance in boreal forests dominated by evergreen trees, since these forests have a comparatively low evapotranspiration rate (Bonan 2008). For the part of the Baltic Sea region characterised by a more temperate climate, the role of evapotranspiration might, however, be of greater importance due to the dominance of more strongly transpiring broadleaved deciduous forests, although authors disagree on the role of temperate forests in climate change (South et al. 2011). Significant feedbacks from such changes in the hydrological cycle were identified by, for example, Wramneby et al. (2010), but primarily in Central Europe. Meanwhile, there was no evidence that variations in cloudiness and precipitation over Europe could be attributed to vegetation dynamics. The lack of an established relationship between increased/reduced evapotranspiration, precipitation and cloud formation over Europe could be because these are strongly determined by the

advection of moisture from the Atlantic. This is likely to overwhelm any feedback signal from vegetation-mediated changes in evapotranspiration. In addition, the ratio between sensible and latent heat exerts a strong local control on surface temperature, but effects on cloud formation and precipitation will take place at the site of condensation, further away and higher up in the atmosphere, diffusing the signal (Wramneby et al. 2010).

Climate–vegetation feedbacks not only influence the mean climate but can also affect climate variability. Seneviratne et al. (2006) performed a suite of climate model sensitivity simulations with and without soil moisture responses to infer the role of the land surface; a substantial fraction of the future temperature variability in Europe was attributed to land-surface processes mediated by soil–moisture feedbacks. In some respect, climate variability provides a better understanding of climate change, since its concrete consequences might be extreme climate events such as floods and droughts. For the European domain, and the Baltic Sea countries, such events already have severe consequences (Della-Marta et al. 2007).

25.5.2 Resource Management and Future Land-Cover Change Scenarios: Possible Effect on Future Climate

Today, the total land area of the Baltic Sea coastal countries (Russia excluded) is roughly 160 million ha, of which 79 million ha is forest and 46 million ha is agricultural land (FAO 2009a). Sweden and Finland are mainly covered by forest and constitute approximately 63 % of the total forest area in the Baltic Sea coastal countries, while Germany and Poland are dominated by agricultural land and represent about 71 % of the total agricultural land area (FAO 2009a, see also Chap. 21). In the past 20 years, forest areas have increased in western and eastern Europe (FAO 2009b). Production of food is extremely valuable for the European Union as well as forest raw material for pulp, paper and construction material. One of the most immediate challenges facing the forest and agricultural sector in EU countries is to meet the anticipated rise in demand for raw materials resulting from the promotion of renewable energy sources (e.g. EC 2009).

The forces driving future land use in Europe include agricultural policy and international markets. Other issues include global food security/scarcity; the possible development of new, alternative agricultural products; the preservation of agricultural and forest land; and urbanisation. Moreover, future climate change may also influence land use. Globally, a number of future land-use change scenarios have been explored, and over recent decades, regional scenarios have emerged for different parts of the world (Alcamo

et al. 2008). Regional studies pinpointing future changes in the Baltic Sea region are very limited, but over the European domain, a growing number of future land-use scenarios are becoming available. The difficulty in moving focus from global to regional land-use scenarios lies in the variety of possible outcomes, since more details and locally specific questions need to be considered (Carter et al. 2007; Alcamo et al. 2008; Metzger et al. 2010). Scenarios on the future development of European land use generally rest on two assumptions (Alcamo et al. 2008): an increase in agricultural productivity and a decrease in European population. According to the United Nations projections, a population decline by 8 % is expected by 2030 (UN 2005), with a further decline likely for later years. At the same time, agricultural productivity is expected to increase by between 25 and 163 % depending on the technological developments assumed (Ewert et al. 2005). The net result of these trends is a decrease in the agricultural area required for food production. For Europe as a whole, the scenarios show a decline in cropland of 28–47 % by AD 2080 and a decline in grassland of 6–58 % (Rounsevell et al. 2006), the abandoned areas being reclaimed by either urban development or forestry, although some areas may be used to cultivate bioenergy crops. Bergh et al. (2010) argued that the expected climate conditions during the twenty-first century (according to IPCC SRES; Nakićenović and Swart 2000) are likely to imply improved growing conditions in boreal and cold temperate climates, but a decrease in production both in agriculture and forestry in Central and southern Europe (Lindner et al. 2010; Masters et al. 2010), which would result in increased pressure on both agriculture and forestry in northern Europe especially in the latter half of the century. The demand for crops and the importance of food security might imply that forests would need to be replaced by agricultural land. A few other studies have also indicated a sustained or even expanded agricultural fraction for some Baltic Sea countries (e.g. Denmark and Finland; by Audsley et al. 2006) during the twenty-first century.

The effect on climate of possible future regional land-use changes is to a large extent unexplored. Biogeochemical feedbacks from regional land-use changes have been discussed in the context of global climate change in some studies (Carter et al. 2007; Rounsevell and Reay 2009), but the direct biogeophysical feedbacks from expected land-use changes are yet to be addressed. Given that the majority of available future land-use scenarios at the European scale assume increasing fractions of forested areas in parallel with a reduction in agricultural land, it would imply a positive (warmer climate) albedo-mediated effect in winter when previously snow-covered agricultural land is replaced by snow-masking forested areas and, at least potentially, a negative (colder climate) effect from an enhanced hydrological cycle in summer due to higher LAI (Wramneby et al.

2010). It could also imply an increased CO_2 sink. Whether the net effect would be a warming or a cooling is not possible to assess without extensive modelling experiments. On the other hand, if future land-use change in the Baltic Sea region happens to be an increase in area of agricultural land as suggested by some studies, the different effects mentioned above would be of opposite sign, and so the net effect probably of different magnitude.

The International Union of Forest Research Organizations (IUFRO) has highlighted the immense potential for the forest sector to mitigate climate change at low cost, while agricultural production would have much smaller potential. Forestation/reforestation has been suggested as a tool to mitigate global warming because a growing forest takes up and stores carbon from the atmosphere (UNFCCC 2005). However, the scientific challenge is to understand how different land-use strategies can contribute to mitigation benefits (e.g. Canadell and Raupach 2008). Pongratz et al. (2011) have shown that reversion of past land-cover change may often be the most feasible step for implementing ALCC as mitigation tool, but that careful analysis of the possible effects of ALCC at the regional scale should be performed for each forestation/deforestation project (see Sect. 25.4.2). It is also debated whether carbon sequestration in forests is the most effective way of mitigating climate change and, therefore, whether forest management should be optimised to increase the carbon stock. Various forms of carbon-related land-use strategies, as well as carbon-accounting mechanisms, increasingly enter agricultural and forest politics and policy. Bioenergy, carbon sinks and raw wood products as substitutes for fossil-based materials (Sathre and O'Conner 2010) add to the traditional list of products and services that forests may provide. Ranges of energy-associated industry, in particular the bioenergy industry, together with other climate- and energy-related organisations, are now entering the forest sector with hopes of realising their strategies. However, it is important to note that many of the socioeconomic factors and assumptions controlling future land-use policies take indirect biogeochemical processes into consideration while neglecting direct biogeophysical processes (Jackson et al. 2008), presumably because the number of studies on CO_2 is substantially larger. In other words, current understanding of land-cover changes and their biogeophysical feedbacks in regional climate change is limited in comparison with the large-scale carbon cycle feedbacks.

25.6   Conclusion

1. ALCC is one of the few climate forcings for which the net direction of the climate response (warming or cooling) over the last two centuries is still not known with

certainty. The major uncertainty is due to the often counteracting temperature responses to biogeochemical versus biogeophysical effects, but also to the difficulty of quantifying the counteracting effects of changes in albedo and hydrological cycle (both biogeophysical effects), as well as obtaining precise land-cover data for the past. However, it is recognised that it is important to attempt to quantify the contribution of local/regional ALCC to past regional and global climate change. This information is necessary (i) to identify ALCC as a forcing of past climate change (over centuries and millennia), (ii) assuming that agricultural expansion continues in many regions of the world, to understand whether it will lead to similar climatic consequences as in the past, and (iii) to understand how land-use strategies may mitigate future climate warming.

2. Deforestation or reforestation by humans may have a series of contrasting effects on regional climate, of which the net result in terms of temperature and precipitation is still not possible to establish from existing modelling studies. Although the biogeochemical effect of historical ALCC on climate is theoretically well understood, the magnitude of the increase/decrease in the forcing is not accurately quantified to date. In contrast, the influence of the biogeophysical effects of ALCC is still not fully understood because it is far more complex. While the land-surface modelling community generally agrees that ALCC may affect climate through physical effects—in particular by affecting the albedo and therefore the surface–energy balance—there is no consensus on what this proposition implies. Even though the estimated global-scale impact of historical ALCC on RF through land-surface albedo changes is small relative to the CO_2-related RF (Forster et al. 2007), it does not imply that ALCC has no significant impact on regional climate (e.g. Pielke et al. 2002; Davin et al. 2007). There is common agreement that, provided ALCC is spatially coherent at a sufficiently large scale, this would affect the regional-scale climate significantly. However, to date, three factors could not be quantified yet: (i) the scale of ALCC required to produce a significant effect on climate on global and regional spatial scales (when compared to other forcings of climate change), (ii) how large the resulting change in the regional climate can be expected to be, and (iii) how much the nature of the existing climate over a region might suppress or amplify the initial impacts of ALCC. Increased concentration of greenhouse gases in the atmosphere and the subsequent changes in sea-surface temperatures and sea-ice extent are considered today to be the main drivers of climate change also over land. However, such an assumption leads to erroneous conclusions regarding the land-surface impacts on climate change in regions where ALCCs have been significant. ALCCs have the potential to mask a regional warming signal, with the resulting risk that detection and attribution studies may miss a clear greenhouse signal or misattribute a greenhouse signal if the ALCCs are poorly accounted for in the model.

3. For most of Europe, and the southern zone of the Baltic Sea basin in particular, palaeoecological studies clearly show that the major transformation of the landscape by anthropogenic activities occurred between 6000 BP and 3000/2500 BP with major deforestation during the Late Bronze Age or Early Iron Age, depending on the area. The deforested area did increase between AD 800 and 1700, however, by not more than about 50 % of the earlier deforestation. Similarly, the increase in deforestation between about AD 1700 to AD 1850/1900 did not represent more than 50 % of the landscape openness at AD 1700, in many areas much less. The effect of this long-term deforestation on past climate, at global and regional scales, is not fully understood and still much debated.

4. There do not appear to be any model studies looking specifically at both the effects of biogeophysical and biogeochemical processes related to historical (ca. AD 1850—modern) ALCC on climate change at the scale of the Baltic Sea region. The only modelling study of the biogeophysical effects of past ALCC on regional climate in north-western Europe suggests that deforestation between 6000 and 200 BP may have produced significant changes in winter and summer surface temperatures (of the order of ±0.5–1 °C) through biogeophysical processes that vary in size and direction (decrease or increase) depending on the geographical area and season. In the Baltic Sea area, the major effect is seen as a slight cooling of 0.2–04 °C during winter in southern Sweden primarily due to the *albedo effect*. The net effect of anthropogenic deforestation and reforestation through carbon sources versus sinks (warming vs. cooling) and biogeophysical cooling versus warming is not yet quantified. Reforestation has been suggested as a tool to mitigate global warming. However, several model studies of large-scale reforestation/deforestation indicate that in boreal regions, the magnitude of the positive albedo forcing following reforestation could be larger than the magnitude of the negative forcing from CO_2 uptake, therefore leading to warming. The latter would imply that boreal reforestation might not be an effective climate warming mitigation tool as it could lead to increased warming through biogeophysical processes.

5. A conclusion from available modelling studies reviewed here is that there is no indication to date that deforestation in the Baltic Sea area since AD 1850 would have been a major cause of the recent climate warming in the region through a positive biogeochemical feedback (release of CO_2). Moreover, the southern part of the

Baltic Sea area was reforested after AD 1900, which should have resulted in a different net effect on regional climate than deforestation. The ALCC scenarios do suggest a significant increase in land use at this time, but this does not agree with palaeoecological reconstructions indicating that openness was at its maximum in many parts of Europe—and in the southern part of the Baltic Sea region—around AD 1850–1900. The question is whether the change in land cover over the transition to the industrial period was large enough to make a significant contribution to the climate warming. Theoretically, the tendency towards reforestation from the end of the nineteenth century and through the twentieth century could have implied a cooling if the biogeochemical effect was the largest of all effects or a warming if the albedo effect in winter was dominant over other biogeophysical effects and the biogeochemical cooling effect. However, the respective magnitude of each potential effect and their net result are still unknown for the Baltic Sea region. Therefore, there is still an urgent need to better understand the biogeophysical effects of reforestation in this region because of the idea still prevailing that planting trees will mitigate climate warming. Hibbard et al. (2010) recognised the importance of an accurate representation of land-use and land-cover change to understand and quantify the interactions and feedbacks between climate and socio-economic systems, respectively, and highlighted recent and innovative methods that integrate observations and modelling analyses of regional to global aspects of biogeophysical and biogeochemical interactions of land-cover change with the climate system. To conclude, a quote from de Noblet-Ducoudré et al. (2012): 'the appropriate question is not whether ALCC has a globally averaged significant impact, but is rather whether ALCC has an impact on regions that have undergone intensive ALCC (such as North America, Europe, India, China, Russia, Japan, and Indonesia) that is worth accounting for when exploring the impact of other human forcings on regional climate'.

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