Chapter 6

Saturation of the Terrestrial Carbon Sink

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6.1 Introduction

There is strong evidence that the terrestrial biosphere has acted as a net carbon (C) sink over the last two and half decades. Its strength is highly variable year-to-year ranging from 0.3 to 5.0 Pg C yr^{-1}; an amount of significant magnitude compared to the emission of about 7 Pg C yr^{-1} from fossil fuel burning (Prentice et al. 2001; Schimel et al. 2001; Sabine et al. 2004). Uncertainties associated with C emissions from land-use change are large. On average, the terrestrial C sink is responsible for removing from the atmosphere approximately one third of the CO_2 emitted from fossil fuel combustion, thereby slowing the build-up of atmospheric CO_2. The ocean sink is of similar magnitude (Sabine et al. 2004). Given the international efforts to stabilize atmospheric CO_2 concentration and climate (i.e., Kyoto Protocol, C trading markets), the terrestrial C sink can be viewed as a subsidy to our global economy worth trillions of dollars. Because many aspects of the terrestrial C sink are amenable to purposeful management, its basis and dynamics need to be well understood.

There is a sparse understanding of the multiple processes responsible for the terrestrial C sink, the relative importance of each of them, and their future dynamics. Such information is needed to predict the future strength and distribution of biospheric terrestrial sinks, and therefore critical for the design of future pathways of decarbonization.

Terrestrial biospheric models and fully carbon-coupled General Circulation Models (GCMs) are progressively incorporating more C processes thought to drive significant carbon-climate feedbacks. However, being limited by poor understanding and the slow process of building ever more complex models, they still represent a partial set of processes. This limits the ability to produce robust projections of the biosphere's role in the future atmospheric CO_2 growth and its impacts on climate change.

In this paper we demonstrate that the underlying ecology of terrestrial biospheric CO_2 sinks suggests that, despite having the potential for increased C sink owing to atmospheric and climate change over the next decades, most of the biological sinks will eventually level-off and subsequently declined to zero (hereafter referred as "sink saturation") whereby no further C will be removed from the atmosphere.

Coupled with this sink decline, global warming and deforestation have the potential to destabilize large biospheric C pools (hereafter referred as "vulnerable C pools") which would add CO_2 to the atmosphere. This C source component will further diminish the net gains of C sinks and could even diminish the sink strength beyond zero, thereby moving from being a C sink to a source during this century. Such an eventuality would put further pressure on society to select higher targets of CO_2 emission reductions from fossil fuel burning.

In this paper, we first introduce briefly the current understanding of the global distribution of C sources and sinks. Second, we discuss processes thought to be responsible for the current terrestrial C sink and their likely future dynamics highlighting their potential contribution to saturation of the terrestrial C sink. Third, we discuss the dynamics of the future terrestrial C balance based on results from stand-alone and fully coupled carbon-climate models, and their accuracy in representing various physiological and ecological sink/source processes. Recommendations for future research are made at the end of the chapter.

6.2 Location of the Current Terrestrial Carbon Sinks

The dual constraint of the measured global atmospheric CO_2 concentration growth rate and the relatively weak north-south CO_2 concentration gradient indicate a net sink of about 2 Pg C yr^{-1} in the northern extratropics (Tans et al. 1990). These observations are corroborated by additional information from $\delta^{13}C$ and O_2/N_2, as well as increasingly sophisticated inverse modeling with augmented flask data (Rayner et al. 1999; Kaminiski et al. 1999; Gurney et al. 2002).

Despite the robustness of the global and latitudinal average estimates, inverse modeling calculations are poorly constrained at the continental scale, particularly with regard to the longitudinal partition of the sink. Model intercomparisons suggest that the Northern Hemi-

sphere terrestrial sink is relatively evenly distributed across North America, Europe and Asia, with boreal North America being C neutral (Gurney et al. 2002; Yuen et al. 2005). Large uncertainties still remain for northern Africa, tropical America, temperate Asia and boreal Asia. Interpretations of global distributions of atmospheric O_2/N_2 ratio (Keeling et al. 1996) and inverse analyses (Rayner et al. 1999; Gurney et al. 2002) conclude that tropical ecosystems are not a strong net sink, and most likely they are C neutral. Given that tropical forests are well documented as current large CO_2 emitters via deforestation, this implies that there must be a strong balancing tropical sink of equivalent magnitude to that of emissions from tropical deforestation. During the 1980s–1990s this was estimated to be 0.6 to 2.2 Pg C yr^{-1} (DeFries et al. 2002; Archard et al. 2002; Houghton 2003).

Atmospheric data provides limited information about the processes driving terrestrial sinks. Such information requires the addition of ground-based observations, experiments, and modeling (Canadell et al. 2000; Global Carbon Project 2003). Processes responsible for the sinks in the tropical, mid-, and high-latitudes need to be equally understood since the saturation of any one can have large impacts on the growth rate of atmospheric CO_2.

6.3   Dynamics of Processes that Contribute to Carbon Sink Saturation

A decrease in the long term terrestrial net CO_2 sink can arise from a decrease in the C uptake component or increase in the emission component (see Box 6.1). Key sink processes are CO_2 fertilization of photosynthesis, N fertilization of net primary production, woody vegetation thickening and encroachment, forest regrowth in abandoned cropland, and afforestation/resforestation. Key C emission

> **Box 6.1. Description of GPP, NPP, NEP and NBP**
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> Plants take up carbon dioxide via photosynthesis. The sum of photosynthesis over a year is termed Gross Primary Productivity (GPP). The actual C fixed into plants, the Net Primary Productivity (NPP) of an ecosystem or growth of an individual plant, is the balance between GPP and the C lost through plant respiration (i.e., the construction and maintenance cost). It is not until C losses by microbial respiration (heterotrophic respiration, Rh) in litter and soil, and by herbivory are accounted for that we obtain the net C balance of an ecosystem. This net balance is termed Net Ecosystem Productivity (NEP). However, when considering long periods of time and large regions (or the whole terrestrial biosphere for that matter) we need to include other processes that contribute to loss of C such as, fires, harvest, erosion, and export of C in river flow. The ecosystem C balance measured over the full disturbance/recovery cycle is often termed the Net Biome Productivity (NBP) which for a system in equilibrium is zero. Continuous enrichment of the resource base on which the biome depends by CO_2 enrichment or N-deposition could lead to a positive NBP. Continuous global warming might lead to a positive NBP initially followed by a negative NBP later.

processes are soil respiration in response to warming, permafrost thawing and subsequent decomposition of organic matter, fires, deforestation, and peatland drainage. Sink saturation occurs when the increase in efflux becomes equal to the increase in uptake for a given period of time producing no increase in net uptake.

When assessing whether and when C sinks might saturate in the future, there are two types of dynamics to be considered: *(i)* Saturation of C uptake processes which refers to the diminishing sensitivity of a sink process to an external forcing or stimulus (e.g., CO_2 fertilization effect on photosynthesis as atmospheric CO_2 increases); and *(ii)* increasing C emission driven by external forcing (e.g., warming effects on heterotrophic respiration).

We refer to C pool saturation when the annual emission itself becomes equal to the annual uptake which involves two types of processes: *(i)* the "fast response" C uptake (e.g., net photosynthesis) is matched by the "fast response" C emission (e.g., heterotrophic respiration), or *(ii)* long timescale, "slow response" processes cause pools to reach maximum size (e.g., tree death rate matches new tree growth rate or, for soil, the capacity for organic matter stabilization by clay surfaces is exhausted).

6.4   Processes Contributing to Terrestrial Carbon Sink Saturation

Our current understanding of the global net C sinks requires invoking three types of processes (Table 6.1): *(i)* processes driven by changes in atmospheric composition (e.g., CO_2 fertilization on photosynthesis, nitrogen (N) deposition fertilization, pollution damage); *(ii)* processes driven by climate change (e.g., temperature and precipitation effects on gross primary production and heterotrophic respiration); and *(iii)* processes driven by land-use change or land management (e.g., deforestation, forest regrowth, woody encroachment, forest thickening due to fire suppression, afforestation and reforestation, changes in soil C pools under cultivation and grazing). The relative strengths of each of these processes contributing to the current sink and its spatial distribution remain largely to be evaluated (Houghton 2002; Smith 2005).

6.4.1   Processes Driven by Atmospheric Composition Change

6.4.1.1   *CO$_2$ Fertilization Effect*

Short term responses of leaf net photosynthetic rate to CO_2 concentration saturate at around 800–1 000 ppm. The response shows a compensation point (zero net rate) of about 50 ppm for C$_3$ species, followed by a near-linear increase up to around present ambient atmospheric CO_2

Table 6.1.
Sink and source processes driving the anthropogenic perturbation of the net C exchange between land and atmosphere

Processes	Sink, source
Processes driven by atmospheric composition change	
CO_2 fertilization of photosynthesis	Sink
Nitrogen deposition fertilization on net primary production	Sink
Air pollution effects on net primary production	Reduced sink
Processes driven by climate change	
Temperature and precipitation effects on net primary production	Sink/reduced sink
Temperature and precipitation effects on heterotrophic respiration	Source
Permafrost thawing	Source
Shifts in natural disturbances (wildfires and insect attacks)	Source/sink
Shifts in vegetation types	Sink/source
Processes driven by land-use change and land management	
Afforestation and reforestation	Sink
Forest regrowth in abandoned cropland	Sink
Vegetation thickening and encroachment	Sink
Shifts in fire regimes and other disturbances	Source/reduced source
Soil erosion and C burial	Source/sink
Crop management	Source/sink
Deforestation	Source
Peatland drainage	Source

concentration which is determined by the abundance of the CO_2-fixing enzyme RuBP carboxlyase (Rubisco) (Farquhar et al. 1980; Farquhar and Sharkey 1982). That is, when the activity of this enzyme is limiting, overall photosynthetic rate rises proportionally with CO_2 concentration. At higher concentrations, the photosynthetic rate becomes limited by the regeneration capacity of the CO_2 acceptor molecule, ribulose-1,5-biphosphate (RuBP), resulting in a curvilinear response to increasing CO_2 concentration (Fig. 6.1a). From this leaf-level saturation response, we can infer that in the absence of any other limiting factor (e.g., light, nutrients, water) net primary productivity will not increase with increasing CO_2 beyond 800–1 000 ppm (Fig. 6.1b). When other limiting factors co-occur with CO_2, the response and saturation point is different. In the short term, water limitation increases the saturation point with respect to atmospheric CO_2 concentration because water stress reduces stomatal conductance thereby decreasing the CO_2 concentration inside of the leaf. Nutrient limitations, however, can significantly decrease the saturating concentration.

Plants growing in a CO_2 richer atmosphere with suboptimal resources acclimate to the greater C availability by lowering photosynthetic capacity below that expected for a given CO_2 concentration (Long et al. 2004). This down regulation of C fixation appears to express an optimization of the deployment of limited C and nutrient resources, particularly N (Xu et al. 1994; Wolfe et al. 1998).

Herbaceous species growing in sub-optimal conditions with multiple levels of atmospheric CO_2 (includ-

ing sub-ambient CO_2 concentrations) showed that the biomass of C_3 plants is linearly related to CO_2 at CO_2 concentrations corresponding to pre-industrial (280 ppm) to present (380 ppm) (Fig. 6.1c; Polley et al. 1992; Polley et al. 1993; Polley et al. 1994; Dippery et al. 1995). At future CO_2 concentrations greater than 450–550 ppm most species enter the saturating responses zone. Woody species have similar responses with a consistent linear increase in plant biomass from pre-industrial to ambient concentrations, and often a saturating growth response in the same range as for herbaceous species (450–550 ppm) (Fig. 6.1d). Hättenschwiler and Körner (2000) found the same behavior for six European temperate tree species growing in the low light regime of a forest understory.

It is well established that N limitation prevents the full expression of the CO_2 fertilization effect on plant growth, and that a progressive N limitation is likely to be the principal factor driving the above saturation curves (Luo et al. 2004; Reich et al. 2006; Groenigen et al. 2006). That is, N becomes increasingly limiting (at least in the period up to a few years typical of elevated CO_2 experiments) as both N and C are immobilized in new plant and soil organic matter fractions. Other nutrients, particularly phosphorus, also limit plant growth responses to increasing atmospheric CO_2 (Barrett and Gifford 1995; Dukes et al. 2005; Edwards et al. 2005).

There are systems that already show saturation at present CO_2 concentrations due to nutrient or other environmental constraints (Körner et al. 2007, Chap. 3 of this volume). This is the case for: *(i)* a mature temperate trees
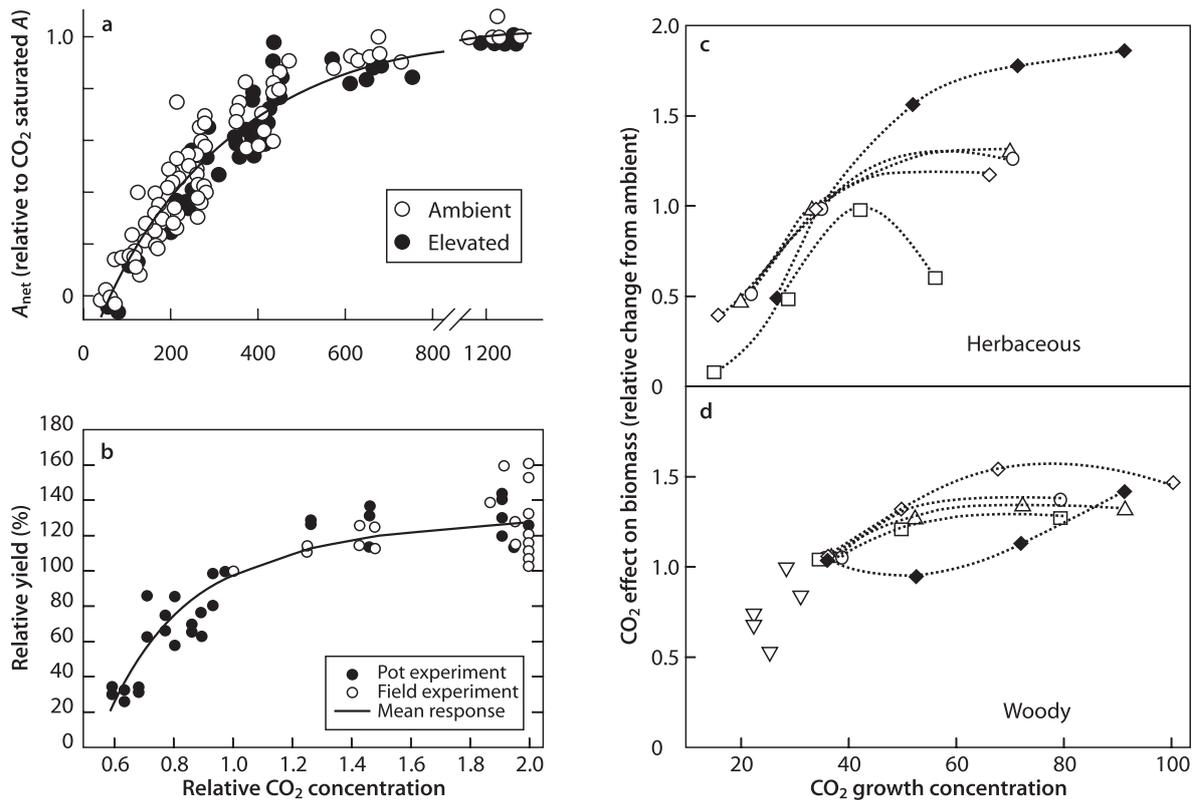
Fig. 6.1. Plant responses with saturation dynamics to increasing levels of atmospheric CO_2. **a** Typical A/C_i curve: Leaf net C assimilation at increasing leaf internal CO_2 concentrations (CO_2 concentration at the site of carboxylation in the chloroplasts). **b** Wheat responses to increasing CO_2 levels (data from field experiments (*open symbols*) and pot/glasshouse experiments (*solid symbols*, Olesen and Bindi 2002). **c** Data for the herbaceous species *Abutilon theophrasti* (Dippery et al. 1995), *Homogyne alpina* (Hättenschwiler and Körner 1996), *Arabidopsis thaliana* (Ward and Strain 1997), and *Glycine max* (*closed symbols*, Rogers et al. 1983; *open symbols*, Allen et al. 1991). **d** Data for the woody species *Quercus alba* (Norby and O'Neill 1989), *Liriodendron tulipifera* (Norby and O'Neill 1991), *Pinus taeda* (Rogers et al. 1983), and *Liquidambar styraciflua* (*closed symbols*, Rogers et al. 1983; *open symbols*, Tolley and Strain 1984). For *Prosopis glandulosa*, data were available for subambient concentrations only, and the *dashed line* is a linear regression ($R^2 = 0.55$, $P = 0.09$, Polley et al. 1994)

in Switzerland (100 years, 30–40 m tall) which after three years of CO_2 fumigation showed no growth response (Körner et al. 2005); (*ii*) trees growing nearby natural CO_2 springs (Tognetti et al. 2000); (*iii*) 30 yr old *Pinus* spp. growing at the high elevation tree-line in the Swiss Alps (Handa et al. 2005); and (*iv*) grassland in California after 5 years of elevated CO_2 exposure (Dukes et al. 2005). A number of other studies showed no responses to increased atmospheric CO_2 (Mooney et al. 1999; Luo et al. 2006).

Eleven Free-Air CO_2 Enrichment (FACE) experiments encompassing bogs, grasslands, desert, and young temperate tree stands report an average increased NPP of 12% at 550 ppm when compared to ambient CO_2 (Nowak et al. 2004). Four FACE studies on forest stands showed a 23% median increased NPP, an expectedly high response for stands made up of young trees and saplings (Norby et al. 2005). A meta-analysis of over a hundred studies shows about ⅔ of the experiments responding positively to increased CO_2 (Luo et al. 2006). This analysis also shows that ecosystems under elevated CO_2 can accumulate N, supporting the hypothesis that terrestrial ecosystems have certain capacity to gradually acquire

additional N required to continue accumulating C under increasing CO_2 concentrations (Gifford 1992, 1994; but see Reich et al. 2006; Groenigen et al. 2006).

Just as increased leaf-level photosynthesis does not necessarily translate into whole plant growth, increased NPP does not necessarily translate into increased Net Ecosystem Productivity (NEP) (i.e., net C sequestration). Ecosystem-level net C fixation is constrained by water, temperature, light, nutrients, and a host of biotic interactions between individuals and species that involve CO_2 emissions from ecosystems. There are few elevated CO_2 experiments that have been able to measure annual net C exchange owing to intrinsic methodological difficulties. Over a three-year period, NEP of a nutrient-limited tundra ecosystem exposed to a step increase in CO_2 in Alaska exhibited a pulse-increase producing a temporary C sink (Oechel 1994). The positive NEP declined over time and completely disappeared after three years. More recently a Mojave Desert shrubland that was still accumulating C (i.e., positive NEP) exposed to 8 years of elevated CO_2 using FACE technology showed in its 8th year a decreased annual net C uptake at elevated CO_2 compared

to ambient levels (Jasoni et al. 2005). These examples illustrate that a variety of responses of NEP to CO_2 can occur in different ecosystems.

In summary, ecosystem responses to large step increases of CO_2 concentration can show a strong saturation behavior driven by resource limitation. Plant growth is most stimulated under pre-industrial to present atmospheric CO_2 concentrations with some systems showing saturation at present CO_2 concentrations. For a large portion of systems studied, CO_2 exposure to higher CO_2 concentrations does stimulate plant growth. In these cases, increasing CO_2 concentrations results in a saturation response at around 500–600 ppm, a much lower level than that expected based on solely leaf-level physiological grounds. It is unclear to what extent over decades of gradual annually increasing CO_2 concentration the acquisition of atmospheric N by symbiotic and free-living N fixation (and by increased N capture from N deposition and N-cycle turnover) will be enhanced by elevated CO_2 concentration.

The CO_2 fertilization effect on C accumulation is not solely the result of photosynthesis fertilization, but also of increased water use efficiency by plant canopies owing to increased photosynthesis at lower stomatal conductances (Morison et al. 1985; see review by Mooney et al. 1999). The result can be a slower rate of depletion of soil moisture, so that higher rates of photosynthesis may be maintained for longer fractions of soil wetting/drying cycles under high CO_2, particularly in arid and semiarid ecosystems. This effect has been consistently reported for water-limited systems such as deserts (Pataki et al. 2000), Mediterranean grassland (Field et al. 1996) and savanna ecosystems (Owensby et al. 1997). The implication is that "well-watered" ecosystems, including the most productive forest regions of the tropics and boreal zone (likely to play the largest role in C sequestration), take the least advantage from the beneficial effects of reduced stomatal conductance. Therefore, they may be the first systems to approach an upper limit for the physiological effects of increasing CO_2.

6.4.1.2   *Fertilization by Nitrogen Deposition*

Given that N often limits plant growth in terrestrial ecosystems, particularly temperate regions (Vitousek and Howarth 1991), it has long been speculated that increased plant growth from N deposition is likely to be an important C sink process. Much of this reactive atmospheric N results from N fertilizer use in agriculture and the combustion of fossil fuel in industrial activities. The first process produces reactive N, predominantly as NH_3 (and NO_3); the second produces NO_x ($NO + NO_2$), which eventually deposit on vegetation and soils as dry and wet deposition (Galloway and Cowling 2002). However, it is known from fertilization trials and long term N deposition studies that there is an upper limit to which additions

of N into the system will results in increased NPP. Past this threshold, pollutants associated with N deposition have a negative effect on ecosystems particularly tropospheric ozone (Aber et al. 1998; see Sect. 6.4.1.3).

Early modeling studies showed that N deposition could have a significant impact on enhancing C sequestration in forests, due to their high C/N ratio, particularly at mid-latitude forests in the Northern Hemisphere, where N deposition is highest and N-limitation is common (Holland et al. 1997; Townsend et al. 1996). These models suggested that N deposition could result in increased C sequestration at a rate of 0.1 to 2.3 Pg C yr^{-1}. At its upper plausible limit, N fertilization could account for the entire net terrestrial C sink. Subsequent field experimentation suggests 0.25 Pg C yr^{-1} as the upper realistic limit of C stimulation by N deposition (Nadelhoffer et al. 1999). More recent analyses using the Biome-BGC model and new N deposition data reinforce this lower end value with an additional C uptake due to N deposition of 0.19 Pg C yr^{-1} and 0.26 Pg C yr^{-1} for the decades of 1980s and 1990s respectively (Galina Churkina, in preparation). That analysis attributed 8 Pg C of the terrestrial sink during the period 1950–2000 to fertilization by N deposition in Northern Hemisphere forests.

Global analyses of N deposition impacts continue to be poorly constrained by lack of high quality spatially-explicit datasets of N deposition (but see Holland et al. 2005 for the US and Europe) and an incomplete knowledge of the various pathways by which deposited N is incorporated into ecosystem components, including N absorption directly by plant canopies.

Taking into account future projections of increased N deposition over the next 100 years, largely in tropical Americas, Southern Africa, and China-India (Galloway and Cowling 2002; Fig. 6.2), it is unlikely that N deposition will create any major additional C sinks. Field experiments using N additions in phosphorus-limited
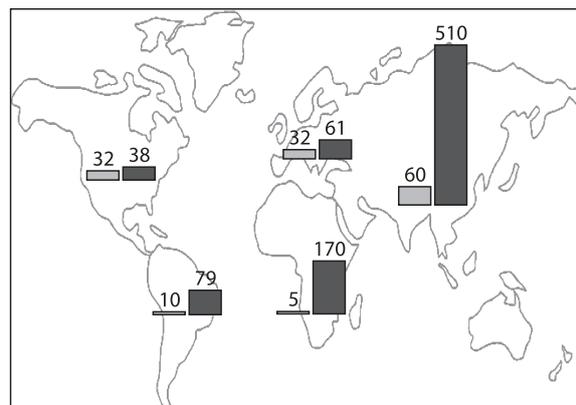

Fig. 6.2. Current and future projections of reactive nitrogen N creation rates in different regions of the world (Galloway and Cowling 2002). *Left bar* indicates the reactive N in 1996 (Tg N yr^{-1}), *right bar* indicates maximum reactive N at the time of projected peak population in 2100

tropical forest show higher emissions of nitrogen oxides than in N-limited forests, both after single pulse and chronic N additions (Hall and Matson 1999). In fact, anthropogenic N inputs may reduce productivity due to the indirect effects on acidity and the availability of phosphorus and cations (Matson et al. 1999). In China and India, most of the N deposition increases in will occur on agricultural lands with little or no sink potential. Therefore the fertilization effect of N deposition on C uptake at present is probably close to the saturation level if it has not already been reached.

There are, however, two related processes by which N may be able to stimulate further C uptake. First, N-deposition concurrent with increasing atmospheric CO_2 concentration may be acting synergistically to progressively increase the standing C pool in terrestrial ecosystems by partially reducing the limitation that its availability imposes on C uptake (Gifford et al. 1996). Second, increased soil respiration in a warmer world, provided water is not limiting, will release additional soil N which can stimulate C uptake to some degree (see Sect. 6.4.2.2).

6.4.1.3 *Air Pollution*

Air pollution, particularly tropospheric ozone (O_3) background concentrations, has increased since the beginning of the industrial revolution. Tropospheric O_3 is formed in photochemical reactions between carbon monoxide, methane (CH_4) and other hydrocarbons, and NO_x. All precursors are largely the result of fossil fuel combustion (Felzer et al. 2004 and references within). Unlike other more localized pollutants, tropospheric ozone is a major regional issue with hot spots in Eastern US, Eastern Europe and China, and with teleconnections to other regions making tropospheric O_3 a global pollutant (Sanz and Millán 1999; Derwent et al. 2004).

Plants exposed to tropospheric O_3 reduce C uptake because O_3 damage cellular tissue which in turn reduces photosynthesis and stomatal conductance. Historical analyses of the possible negative effects of O_3 damage on terrestrial carbon sequestration show a reduced sink between 0.1 and 0.3 Pg C yr^{-1} for the period 1950–1995 (Felzer et al. 2005). The suppression of the sink although small is very significant over the 45-year cumulative period, and the relative change was several times larger for hot spot regions such as US, Europe and China.

Physiologically stressed vegetation is often more susceptible to insect and fungal attacks. Karnosky et al. (2002) showed that aspen (*Populus tremuloides*) developed 3–4-fold increases in levels of rust infection (*Melampsora medusae*) following long exposures to O_3. This further exacerbated the reduction in C update.

Background O_3 concentrations and peak levels are predicted to stay high or increase in the future (Fowler et al. 1999).

6.4.2 Processes Driven by Climate Change

6.4.2.1 *Precipitation and Temperature Effects on Net Primary Production*

Increased length of the growing season, increased precipitation, increased temperature, and decreased cloud cover are factors thought to be responsible for an estimated 6% increased terrestrial NPP (3.4 Pg C) over the last two decades of the 20th century (Nemani et al. 2003); this suggests a possible increase in C sequestration. IPCC (2001) reported that the net sink in the Northern Hemisphere increased from 0.2 Pg C yr^{-1} during the decade of the 1980s to 1.4 Pg C yr^{-1} during the 1990s with associated large uncertainties.

These are the processes and trends, along with CO_2 fertilization, that support model projections suggesting a likely increase in the terrestrial C sink over the next half a century. The same model projections suggest that soil respiration will surpass levels of C uptake later in this century, and that the tropical sink will be ultimately destabilized due to drought (Cao and Woodward 1998; Cramer et al. 2001; Cox et al. 2000; Fung et al. 2005; Friedlingstein et al. 2006). This is projected to lead to an overall decline of the net terrestrial sink which becomes a source by the end of the century in some of the model projections. A key model assumption here is that soil respiration will be as highly temperature dependent in the long term as is found in short term experiments (months to a few years).

Interestingly, new observations are already showing a dampening of regional C sinks strength due to increased climate variability (particularly due to hotter droughts), and thus challenging the expectation of C sink enhancement over the next few decades. An analysis of the entire Northern Hemisphere shows that since 1994 the well known acceleration of C uptake during early spring was offset by decreased uptake during summer, most likely due to drier summers in mid and high latitudes (Angert et al. 2005). The heatwave in 2003 alone reduced the gross primary productivity of European ecosystems by 30% resulting in a net CO_2 source to the atmosphere of 0.5 Pg C yr^{-1} (Ciais et al. 2005), equivalent to about 4 years of C accumulation in these systems lost in a single extreme event of just a few weeks duration.

In the tropics, drought can enhance NEP in the short term because it suppresses heterotrophic respiration (Saleska et al. 2003). However, over the longer term, drought and associated fire (Werf et al. 2004) are likely to diminish the Net Biome Productivity (NBP) in regions that experience an increasingly dry environment.

Thus, increases in intra- and inter-annual variability (frequency and/or intensity) as predicted by some climate models under higher greenhouse gas (GHGs) concentrations (Giorgi et al. 2004; Meehl and Tebaldi 2004)

could depress the net C sink of the Northern Hemisphere and even push it to become a CO_2 source. If the increase in climate variability (e.g., frequency of extreme events) that has occurred in the past decade were to persist as part of climate change, the net terrestrial C sink could soon be declining, well before model predictions that show a possible sink saturation between the middle and end of this century (Cox et al. 2000; Cramer et al. 2001; Friedlingstein et al. 2006).

6.4.2.2 Temperature and Water Effects on Heterotrophic Respiration

The sensitivity of soil C pools to global warming is the single biggest uncertainty in the C cycle to determine future growth rates of atmospheric CO_2. Terrestrial ecosystem models and coupled C cycle GCMs suggest large positive feedbacks to climate due to an assumed high sensitivity of soil heterotrophic respiration to warming (Cramer et al. 2001; Cox et al. 2000; Friedlingstein et al. 2006). This assumption is supported by some short term soil warming field experiments and laboratory soil incubations which consistently show an increase in soil respiration when exposed to elevated temperatures. A meta-analysis of thirty-two field studies showed a 20% increase in soil respiration when soils were experimentally heated 0.5 °C to 5 °C for periods ranging from 2 to 9 years (Rustad et al. 2001; Fig. 6.3). Studies with laboratory soil incubations at various temperatures have also shown that temperature sensitivity is usually high at low temperatures (Q_{10} as high as 8 close to 0 °C) and low at higher temperatures (Q_{10} at around 2 for temperatures above 20 °C) (Kirschbaum 1995).

Other studies show either no correlation between respiration fluxes and temperature or enhanced soil respiration that disappears after some time (Giardina and Ryan 2000; Valentini et al. 2000; Jarvis and Linder 2000; Luo et al. 2001). The latter experiments suggest that once the more labile C fraction has been respired, the remaining more recalcitrant and stable pools are not sensitive to temperature.

More recently, modeling and experimental studies have shed light on this apparent paradox by showing that more recalcitrant pools of Soil Organic Carbon (SOC) do not show lower sensitivity to temperature than labile pools (Knorr et al. 2005; Fang et al. 2005; Fierer et al. 2005). In fact the studies show that slow turnover pools, the largest and therefore dominant component, are equally (Fang et al. 2005) or more (Knorr et al. 2005) sensitive to warming than fast pools.

It is important to note that concurrent with organic C mineralisation to CO_2 during heterotrophic respiration, there is N mineralisation to ammonium and nitrate that could feedback onto productivity. The meta-analysis of soil warming experiments found that associated with an average 20% increase of heterotrophic respiration with warming there was an average 46% increase in net mineralisation of soil N (Rustad et al. 2001). Given the C cycle is tightly coupled with the N cycle (Gifford 1994) some of this increased production of mineral N can be taken up by plants and converted to biomass; this would compensate to some extent for soil C lost under warmer conditions.

Thus it is not well established as to the extent global warming will decrease soil C pools. Certainly the N-cycle feedback can reduce the magnitude of the net long term decline in soil C due to increased temperature. In some ecosystems the compensation can be significant whereas in ecosystems with largest soil C pools (e.g., tundra and taiga with frozen soils and peatlands which are highly sensitive to temperature) the compensation will be smaller. A more detail analysis of C emissions from frozen soils and peatlands are described in Sect. 6.4.2.3 and 6.4.3.7.
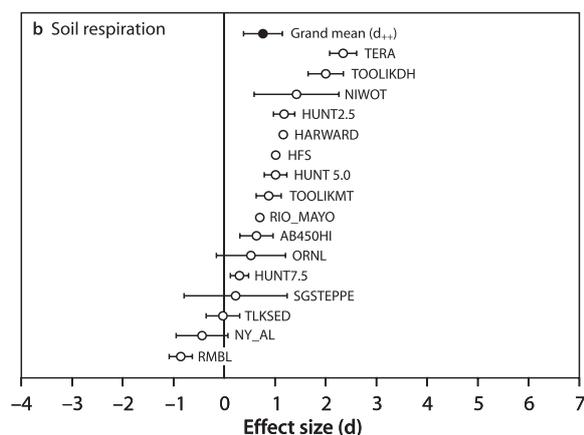
6.4.2.3 Permafrost Thawing

Frozen soils hold over 400 Pg C which have been accumulated over thousands of years (Sabine et al. 2004). Approximately 54% of frozen soils occur in Eurasia, largely in Russia, and 46% in North America, largely in Canada (Tarnocai et al. 2003). An additional 500 Pg C exist in frozen loess that have been accumulated in glacial times in the north plains of Siberia (Zimov et al. 2006). With the rapid warming of the northern latitudes well above the global averages (IPCC 2001), C in frozen ground can be exposed to warmer aerobic conditions which would result in increased decomposition of organic matter and thus C emissions. In fact, widespread observations exist of permafrost thawing leading to the development of thermokarst and lake expansion, followed by lake drainage as the permanent permafrost further degrades (Camill 2005; Smith et al. 2005).

Fig. 6.3. Effects of field experimental warming on soil respiration in 16 studies (Rustad et al. 2001). Mean effect sizes and 95% confidence interval for individual experiments. Soils exposed to elevated temperatures of 0.3 to 6.0 °C for periods of 2 to 9 years

Preliminary estimates show that permafrost area could shrink by up to 25% with a mean global warming of 2 °C (Anisimov et al. 1999). More recent estimates using a fully coupled GCM show that of the 10 million km^2 of present-day permafrost as little as 1 million km^2 near surface permafrost will remain by the end of this century (Lawrence et al. 2005). Melting permafrost will increase CO_2 and CH_4 emissions, and it is estimated for the Canadian permafrost alone that up to 48 Pg C could be sensitive to oxidation under a 4 °C warming scenario (Tarnocai 1999).

In addition to C exchange between land and atmosphere, lateral transport of DOC from thawing permafrost is another process by which C is lost from peatlands. Frozen permafrost watersheds in Western Siberia release little DOC to the rivers while permafrost-free watersheds show considerably higher amounts that are proportional to the extent of the peatland area (Frey and Smith 2005). For Western Siberia, climate models predict a doubling of the area above a mean average temperature of –2 °C (which coincides with permafrost distribution) and an associated 700% increase in DOC concentrations in streams. Part of the DOC reaching the Arctic Ocean will oxidize and return to the atmosphere.

A preliminary global estimate suggests that up to 5 Pg C could be released from permafrost over the next 20 years and up to 100 Pg C in the next 100 years if it is assumed that 25% of the C locked in frozen soils could be oxidized (Gruber et al. 2004; Raupach and Canadell 2006). This amount could increase to as much as 200 Pg C if we include the loess frozen sediments which have been recently reported. Whether this C is released initially as CO_2 or as CH_4 depends on the local hydrological conditions. Although studies suggest an increased gross C uptake by the newly established vegetation (Payette et al. 2004), long term warming and fertilization studies show increased losses of C from deep soil layers (Mack et al. 2004). These losses were sufficient to counteract increased plant biomass C, thereby contributing a net C flux to the atmosphere.

There are compensating feedbacks to consider. The northwards movement of the permafrost boundary is accompanied by northward movement of the vegetation that fixes C and thus compensating to some extent the loss of soil C (see Sect. 6.4.2.4). However, thawing will expose deep deposits of C to aerobic conditions and the losses of the integrated soil profile (including deep soil layers) will offset the possible C gains from increased productivity.

6.4.2.4 Shifts in Vegetation Types

The distribution of the world's vegetation has changed throughout glacial and interglacial periods largely forced by climate and to lesser extent atmospheric CO_2. Conse-

quently it is expected that the distribution of vegetation will change commensurately to the extent of future climate change with associated changes in the size of carbon pools and fluxes (VEMAP members 1995).

Vegetation shifts will change significantly regional C balances, but two of the most important ones at the global scale are: (i) woody encroachment in high latitudes due to warmer temperatures (e.g., boreal summer green trees moving into herbaceous systems) (see Sect. 6.4.3.3 for other types of woody encroachment), and (ii) the more controversial die-back of Amazonia and subsequent savannization driven by increased water stress (Cox et al. 2004; Schaphoff et al. 2006). In the first case, warmer temperatures are already allowing for longer growing seasons and a substantial increase of C pools in vegetation which will likely result in an additional C sink (but see Schaphoff et al. 2006 who argue otherwise due to the concomitant increased soil respiration; see Sect. 6.4.2.1). In the second case, savannization of Amazonia would result in substantial emissions to the atmosphere which is the largest contributor to a positive carbon-climate feedback projected to increase global warming by 0.1–1.5 °C by the end of this century (Friedlingstein et al. 2006).

Some of these vegetation shifts may have transient dynamics with abrupt releases of C emissions, often associated with disturbances (e.g., fire), several times bigger than the slower increased of C pools in biomass and soils in other regions of the biosphere (Smith and Shugart 1993).

6.4.3 Processes Driven by Land-Use Change and Land Management

6.4.3.1 *Afforestation and Reforestation*

Globally, the upper "biological" boundary of C sequestration by land management could be estimated as the amount of C lost historically from land-use change, about 180–200 Pg C (DeFries et al. 1999). This could decrease atmospheric CO_2 concentration at the end of this century by 40 ppm to 70 ppm (House et al. 2002). Taking account of expanding demands for food, fiber, energy and urbanization, the potential of forest C sequestration is reduced to an achievable capacity of only 10 to 50 Pg C (Cannell 2003) or 5% to 25% of the upper boundary for C sequestration estimated based only on biophysical considerations.

In addition, there is a whole fabric of interlinked environmental and socio-economic constraints and opportunities which will ultimately determine the achievable capacity to sequester C such as costs of land for sequestration and for maintaining C stores, environmental requirements for other resources, environmental constraints, social factors, economical feasibility, institutional factors, and demographics (Raupach et al. 2004).

Thus, despite a significant potential for creating C sinks through afforestation and reforestation, the ultimate impact on atmospheric CO_2 growth will be relatively modest and limited in time because globally there will be a balance between intensification (and reduced requirements of land) and an expanding population largely in less developing world with a large demand for new land to meet food and energy requirements. However, a larger potential exist for future enhancement of the C sink if there is greater adoption of biofuels, which could provide the incentive for the establishment of additional plantations.

6.4.3.2 Forest Regrowth on Abandoned Cropland

Forest regrowth on abandoned agricultural land has been identified as one of the most significant process to explain the net C sink in the Northern Hemisphere (USA: Houghton et al. 2000; Pacala et al. 2001; Europe: Janssens et al. 2005). The western movement of agricultural lands from the forests of the east to the prairies of the midwest has been the main driver of land abandonment and subsequent regrowth in eastern North America since the beginning of the last century, while agricultural intensification was the principal driver in Europe (Kauppi et al. 1992; see in Ramankutty and Foley 1999; Goldewijk 2001).

This agricultural shift has resulted in a substantial expansion of relatively young forests with fast growth rates in abandoned cropland, and therefore, with high C sink capacity. Forest cover in Europe and North America has increased over this period and is considered to be an important, if not the dominant, component of their current C sink. Caspersen et al. (2000) showed that for Eastern USA, where the largest amount of cropland abandonment has taken place, that regrowth explains 98% of the sink that is driven by the dynamics of forest demographics (i.e., age structure). Only 2% of forest growth was attributable to increased CO_2 and N deposition.

Stand NEP declines with stand age as measured in a variety of forests in the temperate and boreal regions of Europe, Siberia, and North America (Law et al. 2003; Desai et al. 2005; Fig. 6.4). As forests reach maturity (maximum C pool size) they cease to be sinks (Bond-Lamberty et al. 2004) and even later may become C sources (Goulden et al. 1998; Schulze et al. 1999; Carey et al. 2001).

From a regional and global perspective, persistence of the C sink due to regrowth can only be sustained by continuous expansion of new forests, something that is unlikely to happen in the light of increasing pressures on existing land for multiple uses (urbanization, industrialization, recreation in the industrial world) including fossil fuel substitution with energy crops. In fact, crop abandonment has largely ceased in North America and Europe, while in other parts of the world conversion to
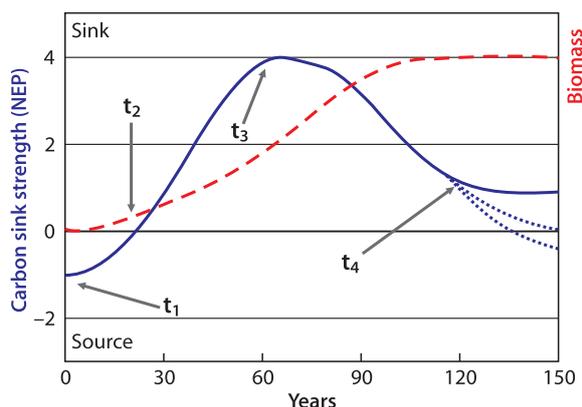

Fig. 6.4. Diagrammatic representation of C sink strength (i.e., net primary productivity, kg C yr^{-1}) and biomass (kg C) accumulation over time of an hypothetical forest stand. t_1: initial state (e.g., after clear cut); t_2: flip point from C source to sink; t_3: maximum net C uptake; t_4: slow down of the C sink, potentially becoming neutral or even flipping from C sink to source. *Dotted lines* indicate alternative trajectories at maturity

cropland is still the primary means by which to meet the food demands of an increasing human population. Forest management can maintain a sustained C sink over the years but at a much lower strength because wood extraction generally returns C to the atmosphere. Old growth forest may not be important globally for the removal of C from the atmosphere, but they hold large C pools that cannot be matched by other land uses (Knohl et al. 2003; Field and Kaduk 2004).

6.4.3.3 Vegetation Thickening and Encroachment

Vegetation thickening in forests and woodland, and woody encroachment in savannas and grasslands increase the amount of C stored on land. Many processes are responsible, but fire suppression as a means to control wildfire damage and fuel reduction by grazing may be the most important. Other factors may be also important (for a review see Gifford and Howden 2001).

Managed fire exclusion during the twentieth century has resulted in an increase of biomass in forest and woodlands in many regions (Mouillot and Field 2005; Luger and Moll 1993; Houghton et al. 2000), and the potential exists for further accumulation. For instance, for the USA, forest and woodland biomass in 1990 was 75% of what it was in 1700 and 63% of what it would be in the absence of disturbance; that is, if all forests had the biomass estimated for undisturbed forests lacking wildfire. In the unrealistic scenario of excluding fire from all areas currently lacking trees, the estimated upper limit of C sequestration by woody encroachment would be about 2 Pg C yr^{-1} (Scholes and Hall 1996), although it would not be sustainable for too long.

Woody thickening and encroachment is an important C sink process estimated to account for a substantial pro-

portion of the sink in various parts of the world although it is recognized as the single least well constrained of all sink estimates. For Australia, the introduction of live-stock grazing and fire suppression has shifted the domi-nance from natural open woodlands to tree-grass domi-nated systems with an estimated 0.035 Pg C yr^{-1} sink in northeastern Australia alone, equivalent to 25% of the total national net emissions (Burrows et al. 2002). Forest thickening and woody encroachment in the US accounts for 0.12–0.13 Pg C yr^{-1} or 22–40% of the apparent total US terrestrial sink (Pacala et al. 2001).

The suppression of wildfire has allowed C accumula-tion, a process which is not sustainable in the long term as the accumulated C can eventually return to the atmo-sphere when fire occurs (see Sect. 6.4.3.4). Forests under routine control-burn management will hold less C than the maximum possible, but over the long term they may hold more than forests experiencing occasional cata-strophic fires. The fact that the burned area in the extra tropics has increased in a number of regions (Boreal North America and Europe: Mouillot and Field 2005; Fig. 6.5) may indicate that fire exclusion has already reached, or exceeded, a maximum C storage, unless at-tempts to prevent fire are intensified. Furthermore, this increase in wildfires is also enhanced by the interactions with a changing climate.

In savannas, woody thickening also tends to saturate. As the global terms of trade for ruminant products con-tinues its long term decline, the extent of woody en-croachment onto grazed tropical grasslands may con-tinue for decades. However, C accumulation does not al-ways occur. Little or no net C accumulation was found in wet grasslands invaded by *Prosopis*, *Larrea*, and *Juniperus* (Jackson et al. 2002) and drier grasslands invaded by *Juniperus virginiana* (Smith and Johnson 2003). Thus the C sink estimates attributed to woody encroachment may not be great.

6.4.3.4 *Shifts in Fire Regimes and Other Disturbances*

Disturbances to vegetation can have dramatic effects on short-term C dynamics, although they are neutral in the long term C cycle. That is, the often quick C release dur-ing a disturbance event (e.g., wildfire) will be balanced by C uptake of regrowing vegetation during the follow-ing decades (fire as a tool to clear land is considered in Sect. 6.4.3.7 on Deforestation). However, changing cli-mate, atmospheric CO_2 concentrations, and land use and management can alter the frequency and intensity of disturbance regimes and the net contribution to the at-mospheric CO_2 growth. It is during the transient dynam-ics in moving from one fire regime to another that a C imbalance is created: a sink when reducing the fre-quency of disturbances frequency and a source when increasing the frequency of disturbances.

The annual gross C flux to the atmosphere from glo-bal savanna and forest fires (excluding biomass burning for fuel and land clearing) is estimated to be in the range of 1.7 to 4.1 Pg C (Mack et al. 1996). Thus, the potential for fire to alter the terrestrial C sink is significant. One of the most striking examples of fire sensitivity to climate (e.g., drought) was the fires in tropical regions during El Niño in 1997–1998. Fire emissions contributed 2.1 ±0.8 Pg C or the equivalent of 66% of the CO_2 growth rate anomaly during that period (Werf et al. 2004). Although most of the fires ignited by human activities, the area burned and intensity were largely the result of drought.

Despite the sensitivity of fire frequency and intensity to climate change and variability, the long term trends over the past century were driven primarily by the implemen-tation of fire suppression policies in temperate regions and increased use of fire to clear forest in tropical regions.

Fire frequency in the boreal forests of Canada and Russia decreased during most of the last century but has increased during the last two decades as a result of changes in management practices perhaps reinforced by strong climate warming (Dixon and Krankina 1993; Kasischke and Stocks 2000). As a result, modeled C emissions from fires in Canada have consistently in-creased over the last 40 years, and changed the net C balance of the Canadian forest from being a sink to a small source (Fig. 6.6; Kurtz and Apps 1999).

Future fire activity will continue to be dominated to some extent by fire suppression policies and the use of fire for forest clearing (see Sect. 6.4.3.3). However, green-house induced climate change, especially if accompanied
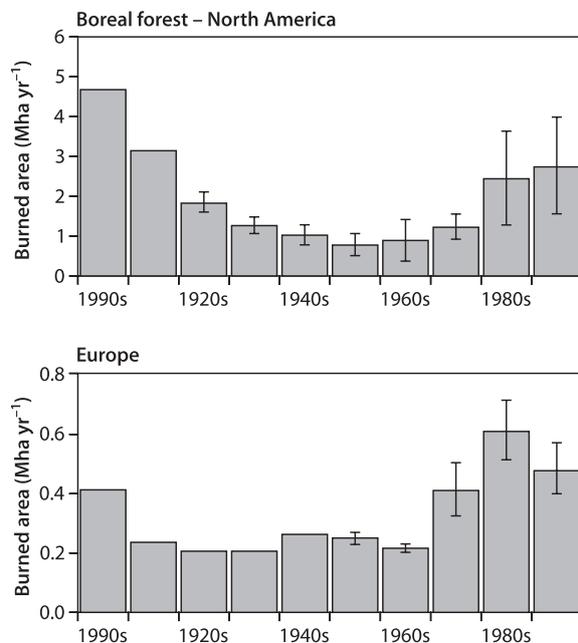

Fig. 6.5. Temporal trend of burned areas (in Mha yr^{-1}) for the 20th century for Europe and Boreal forest in North America (Mouillot and Field 2005)

by increased climate variability and particularly drought events, will affect overall fire extent and intensity. Models predict increased fire activity with a warmer climate for North America, Russia and Europe (Stocks et al. 1998; Flannigan et al. 1998; Flannigan et al. 2000).

In the tropics, El Niño-drought events have an overriding effect on fire intensity, and therefore are likely to determine the future strength of the terrestrial C sink in that part of the world to a large extent (along with deforestation). Future dynamics of El Niño/La Niña cycles are still uncertain but one view is that there will be more frequent or more severe El Niño-like conditions in a warmer climate (Timmermann et al. 1999).

In addition, the interactive nature between management practices (e.g., fire suppression in temperate forests, deforestation) and climate change largely determine the future fire frequencies. One example concerns how the unnaturally high biomass accumulation in forests due to fire exclusion may respond to warming and increased drought frequency in some parts of the world. Grissino-Mayer and Swetnam (2000) have suggested that excessive biomass accumulation in the southwest of the USA is responsible for an increase in the areas burned in recent years. A second example is the evidence that increased landscape fragmentation in tropical regions as a result of higher pressure for multiple land uses favors the rapid spread of fire (Nepstad et al. 1999; Laurance 2000).

There are other sources of disturbances linked to climate change that are important for the net carbon balance. One of the most notable examples is insect dam-age linked to warmer conditions in boreal forests which has been as extensive as fire during some years (Kurz and Apps 1999; Carroll et al. 2004). Shifts in the regimes of hurricanes, tornadoes, and windstorms are also linked to global warming although their effects on the future net carbon balance are more difficult to be determined.

6.4.3.5 Soil erosion and Carbon Burial

Changes in land cover and use are estimated to contribute 10–100 times the natural background levels of soil erosion, stimulating the increase of sediment load into the world's rivers by 2.3 Pg yr^{-1} since pre-agricultural times (Syvitski et al. 2005). Of the total C sediment in rivers, about 0.4 Pg C yr^{-1} is total organic C and 0.4 Pg C yr^{-1} is dissolved inorganic C (Richey 2004). Part of this C is intercepted by dams (estimated to be 1–3% of total river sediment (Syvitski et al. 2005)) and wetlands and the rest reaches the coastal zones. For both pathways only a fraction of that C is stored in long term pools and the rest is mineralized releasing CO_2 back to the atmosphere. Recent estimates on the CO_2 outgassing of rivers shows that a large portion of riverine C is mineralized during transport and quickly returned to the atmosphere as CO_2 (Richey et al. 2002). Although this C redistribution may have important effects on regional C balances, the fact that C previously stored in soils may end up under different sequestration characteristics (e.g., anaerobic conditions at the bottom of dams or in ocean sediments) may not necessarily result in a net global greenhouse gas sink. This is only the case if the C residence time in sediments is so much longer than that in the upper soil horizons from which soil was eroded; this would also need to overcompensate for the outgassing during river transport and mineralization at the final destination (C in sediments may be emitted as CH_4 under anaerobic conditions). Therefore, claims for net C sinks due to sediment burial as big as 1 Pg C yr^{-1} globally (Smith et al. 2001; Stallard 1998) are unlikely to be true and require further investigation. Likewise, little is know on the future of this C sink.

Fig. 6.6. Disturbances and its effects on the net carbon balance of Canada's forests (1920–1995): **a** Area in Mha yr^{-1} disturbed by fire and insect damage, and **b** net ecosystem C fluxes (after Kurz and Apps 1999)

6.4.3.6 Crop Management

Cultivation has resulted in a loss of 55 Pg C from soils worldwide (Cole et al. 1996), and improved management practices can restore some of this soil C. For example, the introduction of conservation tillage in the USA has increased soil organic matter pools by about 1.4 Pg over the past 30 years (Donigian et al. 1994), with the potential to store a further 5 Pg over the next 50 years (Kern and Johnson 1993; Lal et al. 1998). If best management practices could be implemented at the global scale, agricultural soils could mitigate up to ⅓ of the current annual atmospheric CO_2 increased. However, this would
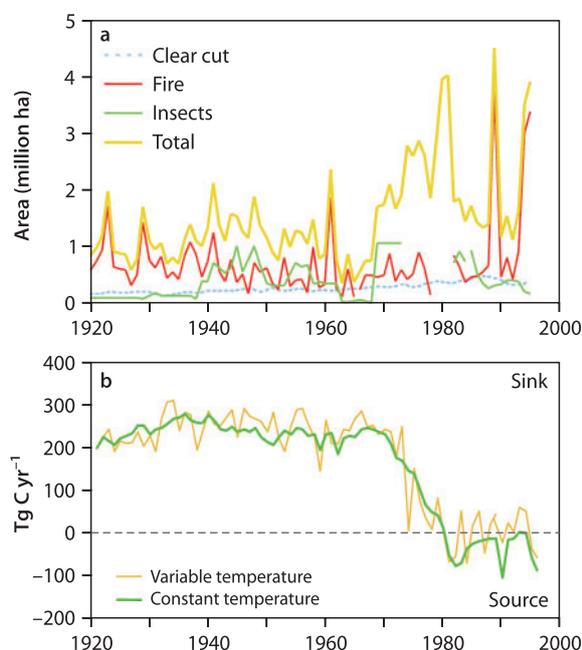
be effective for only 20–50 years (Smith 2004) because soils, as with many other biological C sinks, have an upper limit above which no additional C can be stored. This upper limit could change to a lower level from that of pre-agriculture (e.g., if major soil erosion has occurred) or to a higher level if future climate change increased productivity (e.g., at higher atmospheric CO_2 concentrations).

Over the next 50–100 years, if one includes all available management practices, the potential exists for agricultural management to sequester 0.075 to 0.208 Pg C yr^{-1} in USA arable land (Lal et al. 1998 Metting et al. 1999). In the USA, full adoption of best management practices would likely restore SOC to about 75–90% of its pre-cultivation total (Donigian et al. 1994). Similarly in Europe, it is estimated that 0.113 Pg C yr^{-1} could be offset over the next 100 years (including C offsets from bioenergy crops planted on surplus arable land) (Smith et al. 2000). Changes in management practices in Europe and the USA combined may restore ⅓ to ½ of SOC lost through agriculture globally.

In addition to the limited effectiveness of this process given the ultimate saturation of the pool, the biophysical potentials described above are probably unachievable when socio-economic and other environmental constraints are taking into account. Recent estimates for soil C sequestration potential suggest a global technical potential of about 1.9 Pg C yr^{-1}, whereas the realistically achievable potential is estimated to be 0.2–0.4 Pg C yr^{-1}, with the level of implementation of mitigation measures determined by the market price of CO_2 equivalents (Smith et al. 2006). For Europe, using projections in actual management trends suggest negligible actual sequestration between 1990 and 2000, and negligible sequestration expected by 2010 due to lack of policy incentives to encourage C sequestration (Smith et al. 2005).

6.4.3.7 Deforestation

Land-cover shifts from one type to another are responsible for large C fluxes in and out of the terrestrial biosphere. Human activities have altered land cover for thousands of years but the intensity and magnitude have accelerated during the last centuries. Historically, between 3 253 and 3 470 × 10^6 ha have been converted from natural vegetation, approximately 10% of the total land surface. This has reduced global NPP by about 5% and released 182–199 Pg C to the atmosphere (DeFries et al. 1999). Overall, changes in land use and cover since 1850 are responsible for 33% of the increased concentrations of CO_2 observed in the atmosphere (Houghton 1998), 68% of which was due to cropland establishment (Houghton 1999). Some of these values have been challenged on the basis of lower deforestation rate obtained

using remote sensing measurements (Archard et al. 2002; Defries et al. 2002) but the overall values are likely to remain high (Fearnside and Laurance 2004). Recent high resolution remote sensing analyses in the Brazilian Amazon show that unaccounted selective logging increased previously reported deforestation areas by 60% to 120% (Asner et al. 2005).

Future conversion of forest and grasslands to pastures and croplands will without doubt increase atmospheric CO_2 concentration. With increasing population, increased consumption, and apparent shifts in diets, either land must become more productive or agricultural area must be expanded. In the most developed world the former is being achieved through technological intensification, but major expansion of agricultural land must occur in many parts of the world, particularly in subtropical and tropical countries where the technology or capital for intensification may not be available. Using the UN intermediate population estimates about ⅓ of the Earth's land cover will change in the next hundred years (Alcamo et al. 1996a,b), with the largest changes expected within the next few decades (Walker et al. 1999).

In a compilation of dozens of scenarios of C emissions from anthropogenic land-use change, the IPCC-SRES (2000) predicts that the largest emissions will result from deforestation in tropical Africa, Asia and Latin America. Deforestation in all these three major regions is expected to decrease towards the end of this century to a small fraction of the levels in 1990. It is worth noting that the declines in Asia and Africa are driven by the depletion of their forests, while Latin America shows the highest uncertainty given the extent of forest resources. Carbon emissions from tropical deforestation may release between 85–130 Pg C by 2100 (Houghton 2005).

In addition to the primary effect of increased C emissions resulting from deforestation, there is a secondary (or amplifier) effect on increasing atmospheric CO_2 due to the reduced sink capacity owing to the shorter residence time of C pools in the newly established pastures or croplands (Gitz and Ciais 2004). Using the IPCC-SRES A2 future scenario, the amplifier effect is responsible for an extra 61 ppm of atmospheric CO_2 by the end of the century, after subtracting the additional C uptake by the oceans and remaining vegetation. The magnitude of this effect, which is not accounted for in climate model projections (IPCC 2001), is of a similar magnitude to the carbon-climate feedbacks reported recently (Jones et al. 2005; Friedlingstein et al. 2006).

It is difficult to know whether the current size of the deforestation source will remain steady for a number of decades before declining or whether it will accelerate C emissions due to the implementation of new land-use practices (Carvalho et al. 2001) followed by a step decline. The role of deforestation in terrestrial sink saturation is, therefore, complex but large given the size of this emission term in the global C budget.

6.4.3.8 *Peatland Drainage*

Water-logged peatlands are subject to CO_2 emissions from drying out and from fire. A drop of the water table brought about by land use and/or climate change, can expose large quantities of C to decomposition that otherwise are locked under anaerobic conditions. Interestingly, recent observations suggest that increased atmospheric CO_2 concentrations alone could also lead to a loss of peatland C through larger fluxes of Dissolved Organic Carbon, DOC (Freeman et al. 2004).

High-latitude peatlands have been a net C sink since early in the Holocene (Smith et al. 2005 for Russia's West Siberian lowland peatlands) but there has been a recent slowdown or cessation in net C accumulation (Botch et al. 1995; Peteet et al. 1998; Turunen et al. 2001; Turetsky et al. 2002; Friborg et al. 2003). There exists a balance in wetlands between increased CO_2 emissions when water tables fall or temperature goes up, and CH_4 emissions when water tables rise. In cold regions, changes in precipitation and temperature will largely determine the net balance between CO_2 and CH_4 emissions. Friborg et al. (2003) illustrates this complexity by showing that a Siberian wetland, despite being a net C sink, was an important source of radiative forcing owing to emissions of CH_4 (which has a warming potential 21 times larger than that of CO_2 over 100 years).

Lowland tropical peatlands, accounting for about 10% of global peatland extent, contain up to 70 Pg C in deposits as deep as 20 m (Page et al. 2002; Page et al. 2004 and references therein). Indonesia, Malaysia, Brunei and Thailand hold the largest tropical peatlands in the world which have been an overall net C sink since late Pleistocene (Page et al. 2004). However, over the last decade a combination of intense draining for agriculture and increasing climate variability in the form of more intense droughts (associated or not to El Niño events) have destabilized this millennia-long C sink (Page et al. 2004; Aldhous 2004; Murdiyarso and Lebel 2007, Chap. 21 of this volume).

During El Niño 1997–1998 events in Indonesia, burning of peat and vegetation resulted in an estimated loss of C between 0.81 and 2.57 Pg in 1997 equivalent to 13% to 40% of the mean annual global C emissions from fossil fuels (Page et al. 2002).

With large tracts of drained swamp forests and new peatland conversion projects to agriculture, tropical peatlands will continue contributing to increased C emissions over the coming decades. With some projections of decreased precipitation over the tropics during the dry season (Wenhong Li, in preparation) and the possibility of El Niño-like events becoming more intense or frequent under a warmer climate, C emissions from peatlands have increased and will continue to diminish the strength of the terrestrial net C sink.

Together, high latitude and tropical peatlands account for over 450 Pg C (Sabine et al. 2004). A preliminary estimate suggests that up to 100 Pg C of CO_2 equivalents could be released to the atmosphere from wetlands and peatlands over the next 100 years (Gruber et al. 2004).

6.5 Integration and Model Predictions

So far it has been difficult to integrate the above disparate knowledge under a common framework that allows the various processes and top-down measurements to constrain one another, and informs estimates of the relatively contribution and spatial distribution of the multiple processes driving the net terrestrial C sink. Suggestive of how little constrained the results are is that, added together, the sink estimates from each individual sink process described in Sect. 6.4 amount to a larger terrestrial net C sink than can be accounted for in the global C budget or the observed growth of atmospheric CO_2. This creates a paradoxical problem of an apparent missing large C source.

The limited understanding of some of the processes and pools has also prevented models from representing the necessary dynamics and underlying processes. The result is that there is no single model that includes all phenomena described above. Global climate models used to predict the effect of increasing GHGs on future climate (GCMs), and specifically the ones with coupled C cycle models (i.e., in the C4MIP family of models; Friedlingstein et al. 2006) deal just with simplified versions of photosynthesis and respiration.

Nevertheless, there have been important global and regional analyses. McGuire et al. (2001) attempted a global analysis to partition the historical global C sink from 1920 to 1992 including climate and atmospheric CO_2 change, land conversion to crops and crop abandonment, and forest harvests. They reported that from 1920 to 1957 terrestrial ecosystems were responsible for a net release of 8.8 Pg C, largely due to cropland establishment. During the period 1958 to 1992 the terrestrial biosphere stored 14.3 Pg C largely due to the physiological effects of rapidly rising atmospheric CO_2, although cropland establishment continued to release C. This is one of the best global mechanistic analyses to date but it only accounted for a portion of the sink processes believed to be important.

A sectoral approach has been used by some nations to attribute national sinks and sources to specific activities of their economy and ecosystems. Such studies do not provide information on the ultimate causes of the various sink processes but partition the individual fluxes contributing to a regional C sink. For the conterminous USA (1980–1989) eight atmosphere-land fluxes contributing to the overall C sink were reported (Pacala et al. 2001): (1) forest tree growth, (2) other forest components such as nonliving organic matter, (3) agricultural soils, (4) non-forest systems including woody encroachment, (5) wood products both

in use and in landfills, (6) sediments of reservoirs and rivers, (7) atmospheric C fixed by U.S. ecosystems and then exported by rivers, or (8) exported commercially (food and fiber). The analysis showed that only 30% of the sink was the result of sectors susceptible to the influences of CO_2 fertilization (i.e., (1), (3), and (4)), which was until recent years thought to be the primary driver of the entire terrestrial sink. A similar analysis has been also presented for Europe (Janssens et al. 2003, 2005).

Several attempts have been made to assess the future terrestrial C sink using different types of model: *(i)* physiological and biogeochemical models without dynamic vegetation (do not allow for carbon-climate feedbacks), *(ii)* Dynamic Global Vegetation Models (DGVMs) (do not allow for carbon-climate feedbacks) (Fig. 6.7a), and *(iii)* the C4MIP GCM family (provide coupled carbon-climate projections and allow for carbon-climate feedbacks, but most of them do not have yet dynamic vegetation) (Fig. 6.7b).

Despite the large differences in complexity of the various models, all projections show varying degrees of increased NPP and heterotrophic respiration during the next century due to increasing atmospheric CO_2 concentration (i.e., CO_2 fertilization effects) and associated climate changes. This results in increased global NEP during the next few decades and a subsequent decline of the terrestrial C sink strength. The models under categories *(i)* and *(ii)* that emphasised N limitation without increased N fixation showed the fastest saturation and subsequent decline of NEP. The timing and magni-

Fig. 6.7. Terrestrial net C uptake from 1850 to 2100 with atmospheric CO_2 and climate change: **a** six dynamic global vegetation models (Cramer et al. 2001); and **b** ten global circulation models with coupled carbon cycle model (Friedlingstein et al. 2006)

tude of the downturn depends critically on the sensitivity of soil respiration to warming and how the response of biological N fixation is represented (Gifford et al. 1996; Cao and Woodward 1998; Kicklighter et al. 1999; Cramer et al. 2001; Fig. 6.7a).

More recently GCMs that have been further developed to take into account carbon-climate feedbacks after Cox et al. (2000) showed that the feedback resulted in an additional 200 ppm in the atmosphere by the end of this century (model type iii). There are now 11 GCMs with such capacity and they were recently evaluated for the magnitude of the carbon-climate feedback (Friedlingstein et al. 2006). By the end of the century, the C feedback is modeled to account for an additional CO_2 of 20 to 200 ppm with 8 of the models falling between 50 and 100 ppm (Fig. 6.7b). This could lead to an additional 0.1 to 1.5 °C in surface warming.

A few of the models in categories *(i)* and *(ii)* account for additional processes such as N limitation and fire; models in category *(iii)* are driven exclusively by simple representations of photosynthesis and soil respiration as the overall controls of C exchange between the terrestrial biosphere and the atmosphere. The exclusion of some of the sink processes and vulnerable C pools (Fig. 6.8) tend to overestimate the CO_2 fertilization effect and underestimate biospheric C emissions, thus underestimating what we think is a larger magnitude of carbon-climate feedbacks than currently estimated. The omission of four of these processes/pools will illustrate this point:
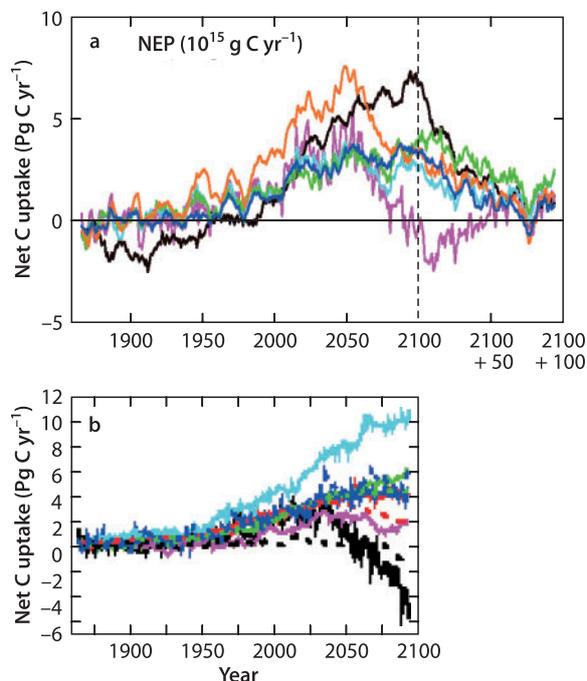
The legacy of land-use change. We discussed in Sect. 6.4.3.2 and 6.4.3.3 that there is strong evidence that a significant component of the current northern hemisphere net terrestrial C sink may be the result of past land practices, particularly from forest regrowth in abandoned cropland and vegetation thickening due to fire exclusion. If that is the case and given the saturation nature of these processes, we should expect a rather quick sink decline within a few decades as forests mature and its sink strength diminishes accordingly (Fig. 6.9). Our current estimates of the future strength of the terrestrial sink (IPCC 2001; Friedlingstein et al. 2006) falsely attribute any C sink that is due to vegetation recovery from past disturbances to the CO_2 fertilization effect with the consequent overestimation of the sensitivity of CO_2 for future sinks.

Nitrogen limitation. Hungate et al. (2003) calculated the amount of N required to accumulate between 260 and 890 Pg C over the 21st century as predicted by terrestrial ecosystem models used in the IPCC Third Assessment (IPCC 2001). Despite accounting for large increases in atmospheric N deposition and biological fixation, and decreased C/N ratios, Hungate et al. were only able to account for 1.2 to 6.1 Pg of N by the end of the 21st century, not sufficient to meet the models demands in all but two simulations for the high end N production (6.1 Pg N),

Fig. 6.8.
C pools vulnerable to global warming and land-use change. They include: *(i)* frozen ground, *(ii)* tropical and high-latitude peatlands, and *(iii)* vegetation susceptible to land use and fire
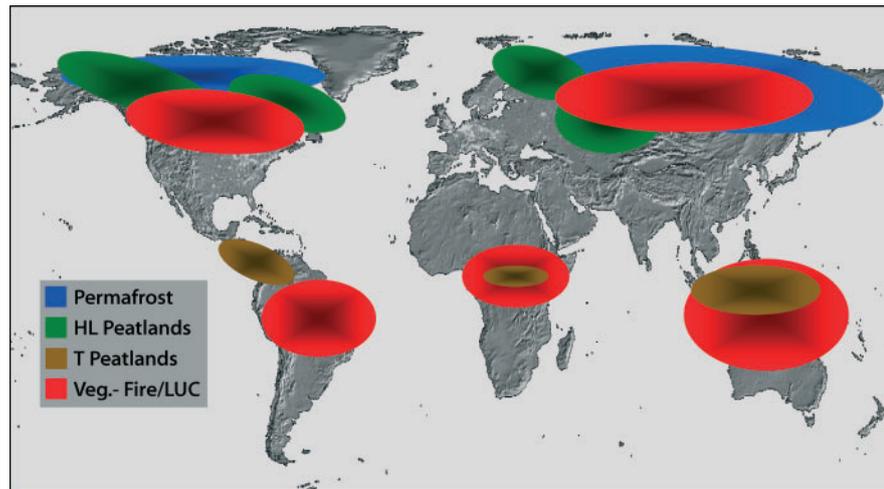
and in all simulations for the low end (1.2 Pg C). Therefore, failure to account for what is a well established constraint on terrestrial NPP (P is also important in the tropics) will result in a gross overestimation of the CO_2 fertilization effect.

Carbon oxidation by fire. Currently there are no coupled carbon-climate models that distinguish soil respiration emissions from fire emissions. Very warm years such as El Niño in 1997–1998 were used to test the sensitivity of soil respiration to temperature which was then used to forecast the soil C sensitivity to future warming (Cox et al. 2000) without accounting for the fact that those same years had high fire emissions. The unintended consequence of this false attribution is to overestimate C emissions from soil respiration. The total C emissions are not necessarily overestimated but omitting fire as a process results in a more smooth dynamics of changes in the sink capacity over time while changes brought about by high fire activity can result in major vegetation shifts which would diminish the strength of the terrestrial sink more abruptly.

Carbon in frozen ground. Soil C databases used in global modeling are still incomplete. Particularly C stored in frozen ground, both in soils and sediments, is poorly known and often estimated only for to the first 30 cm of the soil profile. Sect. 6.4.2.3 reported new findings on the depth and an extension of frozen carbon, particularly with the finding of 500 Pg C in loess sediments in Siberia not previously accounted for (Fig. 6.8). An underestimation of the size of C pools will lead to the underestimation of the potential carbon-climate feedback and therefore overestimation of the potential terrestrial C sink.

6.6 Summary and Conclusions

The current measured net C sink in the Northern Hemisphere and the inferred biosphere sink over the tropics
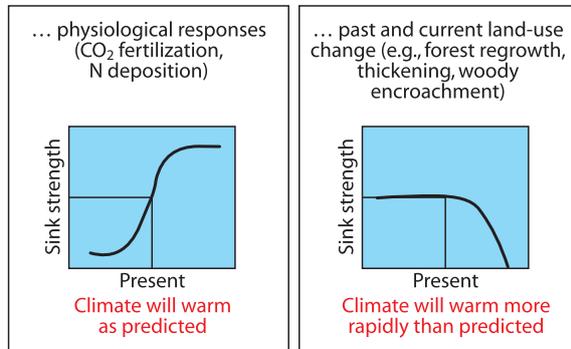
Fig. 6.9. Two alternative scenarios on future dynamics of the terrestrial net biospheric C sink

are not permanent features of the terrestrial biosphere but the result of an imbalance driven by past and present human activities. The sink strength shows high interannual variability and there is evidence that the sink cannot be sustained indefinitely. It is conceivable that by the end of this century or earlier the terrestrial sink may significantly decrease or disappear. Uncertainties in partitioning the historical and current C sinks into various possible processes limits our capacity to predict the future sink strength. Plausible processes that can explain the current sink are land-use change, including forest regrowth on abandoned croplands and fire exclusion, and responses of ecosystem C storage to the related increases in atmospheric CO_2, N deposition and climate change, among others acting individually or synergistically/antagonistically. Some of the sink processes have response curves that saturate at higher forcing levels expected to be reached during the next few decades to a century, while others produce inherently temporary sinks that disappear over time. The saturation of the sink will occur sooner if the prevalent sink processes are related to legacies of past land-use change, and slower, if the prevailing processes are driven by physiological effects of atmo-

spheric CO_2 and N fertilization which are expected to continue in the future (Fig. 6.9). High C-density regions on land such as permafrost and peatlands are vulnerable to global warming and land-use change, and are likely to add large amounts of C into the atmosphere over the coming decades (Fig. 6.8). With the present level of our understanding, it is certain that the terrestrial biosphere sink cannot be maintained indefinitely.

Future Research:

- Integrated analyses of key C sink processes for regional and global C budgets using a multiple constraints approach (bottom-up and top-down measurements and models).
- Resolution of the extent and effect of woody encroachment and forest thickening.
- Integration of the legacy effects of past land-use change (e.g., forest age structure) in biogeochemical models.
- Persistence of new sinks created by afforestation and reforestation.
- Temperature and soil water effects on heterotrophic respiration integrated with feedback constraints, and associated temperature sensitivity of soil organic matter pools.
- Accurate assessment of the extension and pool size of vulnerable C pools to climate and land-use change (e.g., C in frozen ground, peatlands, forest biomass).
- Increase sophistication of C models in coupled carbon-climate GCMs (i.e., Earth System Models) with the inclusion of key terrestrial C-cycle and land surface processes (e.g., vegetation shifts, anthropogenic and wild-fires, forest-age structure, nutrient feedbacks) and vulnerable C pools (e.g., C in frozen soils and sediments, C in cold and tropical peatlands).

Acknowledgments

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