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Review

Regionalizing global climate models

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ABSTRACT: Global climate models simulate the Earth's climate impressively at scales of continents and greater. At these scales, large-scale dynamics and physics largely define the climate. At spatial scales relevant to policy makers, and to impacts and adaptation, many other processes may affect regional and local climate and perhaps trigger teleconnections that provide significant feedbacks on the global climate. These processes include fire, irrigation, land cover change (including crops and urban landscapes), and the emissions of biogenic volatile organic compounds by vegetation. Many of these interact within the atmosphere via dynamical, physical, and chemical mechanisms that lead to boundary-layer feedbacks. It is unlikely that any of these processes have a significant global-scale impact on the Earth's climate in the sense that the amount of warming due to a doubling of well mixed greenhouse gases would change if these processes were explicitly represented in climate models. These phenomena are usually local in space (e.g. urban) or in time (e.g. fire) and probably do not provide the on-going and sustained forcing to affect the global climate. However, for most impacts and adaptation research it is the regional and local climate that defines climate risk. At these scales, processes missing in climate models can have a substantially larger local-scale impact than the additional radiative forcing due to increasing greenhouse gases. Thus, while climate models are well designed for global and continental scales they exclude a suite of important processes that are locally and/or regionally important. We review these missing processes and highlight the research required to resolve the representation of these regional-scale processes in climate models. We also discuss the experimental methodology required to rigorously determine whether these processes are restricted to a local or regional-scale role or whether they do trigger robust teleconnections that would demonstrate global-scale significance. Copyright © 2010 Royal Meteorological Society

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1. Introduction

The Earth's climate is driven by large-scale dynamics that move energy and mass around the planet. These are coupled with physical processes that cause cloud formation, drive turbulent exchanges with the surface, generate rainfall, and interact with solar and infrared radiation in the atmosphere. These dynamical and physical processes are included in coupled climate models (McGuffie and Henderson-Sellers, 2001). Climate models have evolved enormously since the 1st Assessment Report by the Intergovernmental Panel on Climate Change (IPCC, Houghton *et al.*, 1990). By the 4th Assessment Report of the IPCC, Randall *et al.* (2007) concluded that coupled climate models provide reliable projections of climate at continental scales and greater, at a variety of temporal scales.

Climate models are routinely used to model the Earth's sensitivity to increasing carbon dioxide (CO_2), methane, and other greenhouse gas concentrations. Most greenhouse gases that are emitted through human activity and contribute to radiative forcing are globally well mixed and contribute a quasi-uniform forcing on the Earth's surface. While there remains uncertainty in the global-scale response to a doubling in the effective concentration of CO_2 about half of this uncertainty is due to emission pathways and about half is related to difficulties in the modelling of the large-scale dynamics and physics. In particular, remaining uncertainties in key feedbacks affect individual model's sensitivity to increases in radiative forcing at continental scales and greater. These include physical feedbacks (water vapour, clouds, and snow Bony *et al.*, 2006), and biological feedbacks that affect the net terrestrial balance of carbon on a variety of timescales (Friedlingstein *et al.*, 2006; Arneth *et al.*, 2010a).

While some assessments of climate models point to significant skill at below continental scales (e.g. Perkins *et al.*, 2007), the challenge of providing robust and

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reliable projections of future climates at scales at which the impacts occur, for example, a large drainage basin or at the scales of a major city remains daunting. Owing to computational constraints, climate models run at spatial scales that are too coarse for many users of climate models projections. Most impacts of climate change on human activity, ecosystem functioning, production of food, water availability, vulnerability to climate extremes, adaptation to and planning for climate change occur at spatial scales that are small relative to the coarse grids of coupled climate models.

Many regional-scale drivers of regional scale climate do not feedback on the larger-scale climate or do not modify regional conditions such that changes driven by global warming are amplified or moderated significantly. Under these circumstances, the coarse resolution of global climate models can be enhanced via statistical or dynamical downscaling (e.g. Fowler *et al.*, 2007) to provide data for the impacts and adaptation community at resolutions of a few kilometres. However, if regional-scale forcing or regional-scale climate anomalies trigger changes in larger-scale climate, regional climate models cannot resolve these because the boundary conditions used to force the regional climate models are prescribed. Under these circumstances, regional-scale drivers would need to be explicitly resolved in the global climate models, not just in the regional climate models. This is not to suggest that high-resolution regional climate modelling is not fundamentally valuable (e.g. http://ukclimateprojections.defra.gov.uk/content/view/1795/519/) but strong variability among regional climate models (Takle *et al.*, 1999) and on-going challenges in respect of boundary conditions (Denis *et al.*, 2003) remain concerning.

This paper is focussed on phenomena that have a regional-scale signature in a climate sense. These phenomena or processes may feedback directly on the regional climate to affect the sensitivity of that region to large-scale processes or they may 'prepare' the surface to be more, or less, sensitive to increasing CO_2 concentrations. These regional-scale processes may, or may not, have a global-scale impact. A regional process does not need to be shown to have a global impact in order to be significant for global climate modelling. If adding a specific phenomena that only exists in a few regions does not affect the global climate sensitivity this can quite reasonably be ignored in climate simulations focussed on global climate sensitivity. However, this is not what climate models are now used for (ref. Working Group 2 of the IPCC – http://www.ipcc.ch/ipccreports/ar4-wg2.htm). Climate models are used to inform policy makers on regional-scale simulations over some regions where regionally specific processes may play a major role and not necessarily with the additional step of downscaling. Moreover, there is also a strong incentive to treat climate change and air pollution within similar policy frameworks, and the latter unquestionably operates strongly on the local to regional scales, including long-range transport patterns. We will show that there is a very strong coincidence of regionally important processes that affect regional climate with human society. We suggest that via enhancing the modelling of these processes, explicitly within the global climate models, regional prediction can be improved, by forming a stronger foundation for dynamical and statistical downscaling or by directly improving regional simulations.

This paper therefore explores processes that are highly spatially heterogeneous at a global scale but are grouped spatially at a regional scale. We focus on processes and phenomena that would very likely not affect global or continental-scale climate, but could fundamentally limit the application of climate models in climate change impact assessments in some specific regions. We highlight processes that are highly regionalised and currently omitted from the climate models used in virtually all impacts and adaptation studies. To set the context for this paper we first highlight the key regional-scale elements of climate models that are represented in climate models.

2. Included land-surface elements with regional signatures

There are a suite of regionally specific processes, mechanisms, and characteristics that control how the Earth's surface interacts with the atmosphere. Some regions are heavily forested, others grassed or cropped or urbanised. In some regions, snow is a dominant feature; in others, lakes may affect the regional climate, etc. Climate models take changes in net radiation and partition this between sensible and latent heat, and take changes in rainfall and partition this between evaporation and runoff (Pitman, 2003). To do this, the differences caused by snow, and other surface characteristics (roughness, stomatal function, etc.) are included. Some land surface models explicitly 'tile' the surface to represent heterogeneity within a grid element of vegetation, soils, and associated processes including sub-grid scale snow, rainfall interception, and runoff generation. A few land surface models do now represent an urban tile (Section 4.3). There are many challenges to represent these terrestrial quantities well, there are considerable advances required before we would claim the physics and biophysics of terrestrial systems are 'right', but to first order this regionalisation of climate models to reflect the physical and sometimes biological nature of the landscape is included.

We do not suggest that the responses of snow, canopy or surface conductance, soil carbon, vegetation dynamics and function, etc. are fully understood and fully captured by climate models. Major uncertainties in processes remain, and new approaches to evaluation (e.g. Medlyn *et al.*, 2005; Abramowitz *et al.*, 2008) highlight weaknesses in how we have approached evaluation of *components of models* in the past. While the evaluation of climate models is rigorous at continental scales and above (Randall *et al.*, 2007), the evaluation of some model components is not as rigorous across spatial and temporal scales of relevance to climate impacts and adaptation.

Sometimes this is despite extensive data existing (e.g. for forested sites; FLUXNET, Baldocchi *et al.* 2001) but only beginning to be used to its full capacity (Williams *et al.*, 2009). In other systems, rigorous evaluation is limited by lack of data on suitable timescales across a range of surface types. The wealth of remote sensing observations is expected to help refine the existing models and address remaining challenges.

A significant remaining concern is how ecosystems will be affected by climate change (Foley *et al.*, 2003) and how the terrestrial carbon balance might change, particularly the soil carbon pool (e.g. Davidson and Janssens, 2006). Some recent studies have highlighted that the response of the carbon cycle to future climate change may be affected by interactions between the carbon and nitrogen cycles (Sokolov *et al.*, 2008; Thornton *et al.*, 2009). In addition, a recent analysis by Sitch *et al.* (2007) showed that deposition of ozone (O_3) could induce a decrease of the terrestrial carbon sink. The consequences of this decline would be an increase in radiative forcing, an effect as large as the direct radiative forcing by O_3 itself. Arneth *et al.* (2010a) discuss these biogeochemical feedbacks in detail.

There is, clearly, a lot of significant research to do and this paper in no way attempts to undermine these efforts. However, overall, existing models of the terrestrial system, now including carbon exchange, capture the first-order response of increasing atmospheric CO_2 and associated increases in radiative forcing adequately (Randall *et al.*, 2007). It is a reasonable hypothesis that global and continental-scale projections of the impacts of global warming, at least over the next 30–50 years, are not significantly affected by remaining weaknesses in how these basic physical and biophysical processes are parameterised. There are, however, processes that might affect regional scales significantly that are not included in climate models. We focus on these additional areas that are of significance for some regional climates. These processes are currently not included in climate models used in IPCC assessments.

If regional climates were simply an expression of the large-scale atmospheric dynamics and physics with the physics and biophysics of terrestrial processes moderating these, then climate models could simulate regional climates simply by increasing spatial resolution. This paper highlights research that demonstrates that other key processes play a climatologically significant role that is highly regionalised. These additional processes, not represented in the IPCC climate models due largely to computational costs, are unlikely to affect the *global* scale sensitivity of the models to changes in CO_2 because each process operates only in some regions and are 'lost' in global averages. However, at the scales of climate impacts and adaptive responses these additional regional processes can sometimes play a key role in amplifying or moderating the physical and dynamically induced regional patterns of climate change. The challenge is knowing which processes are regionally important, under which circumstances, and how significant their impact is. We highlight a selection of processes, mostly confined to land–atmosphere exchange of energy, moisture, and chemicals. These processes cause changes in local and regional-scale dynamical and physical processes, e.g. boundary layer evolution, cloud processes, convection, and radiative forcing. It is conceivable they could initiate larger-scale changes remote from the specific regional phenomenon. We do not discuss direct anthropogenic emissions in this paper.

3. Missing land-surface elements with regional signatures: atmospheric chemistry and aerosols

Atmospheric chemistry plays a key role in determining the lifetime of greenhouse gases such a methane through regulation of the abundance of the main oxidant of methane; the hydroxyl radical. However, this oxidation process results in a methane lifetime on the order of about 8–10 years and does not introduce significant heterogeneity in the spatial distribution of methane such that one would anticipate a regional-scale enhancement or dampening of radiative forcing. In contrast, some of the key chemical precursors of the hydroxyl radical, which are oxidised reactive carbon and nitrogen compounds show particular large source regions, e.g. tropical forests, mid-latitude agricultural regions, which contribute to O_3 formation in the troposphere. This pollutant and greenhouse gas shows a distinct spatial distribution and imposes a regional-scale radiative forcing. The tropics and the boreal forests are also large sources of aerosols through emissions associated with biomass burning as well as production of secondary organic aerosols from BVOCs. We focus our discussion on these compounds and their sources (and sinks in form of deposition) to demonstrate the importance of atmospheric chemistry and aerosols at regional scales.

3.1. Fire and emissions of chemicals and aerosols

Fire is an inherent ecological characteristic of most terrestrial ecosystems (Bowman *et al.*, 2009). Climate models used in AR4 did not include fire in terms of how it affects emissions of aerosols. However, total radiative forcing has been shown to react sensitively even to small regional changes in fire emissions (Naik *et al.*, 2007). Fire releases large quantities of CO_2 into the atmosphere. Typical estimates of annual carbon released by vegetation biomass burning vary between 2–4 Pg C a^{-1} (Seiler and Conrad, 1987; Thonicke *et al.*, 2001) with large uncertainties, since (i) remote sensing products of area burnt are still rather unreliable, and (ii) translating burnt area into consumed biomass and fire emissions requires additional modelling of productivity as well as conversion into emitted compounds using fixed emission factors (Andreae and Merlet, 2001). A system that is in equilibrium (over annual to decadal periods) re-absorbs the emitted carbon into new biomass via regrowth. However, fire contributes also to the conversion of

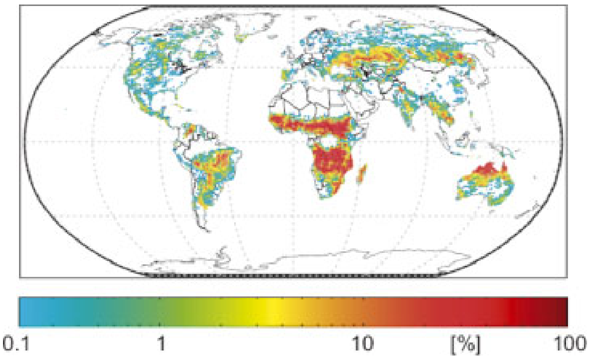

Figure 1. Mean area burned (2001–2008) derived from Terra MODIS, expressed as a fraction of each grid cell that burns each year. Data were obtained from http://modis-fire.umd.edu/MCD45A1.asp; further details are provided by Roy *et al*., 2008. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

forested into agricultural area which leads to a net release of carbon via land-use change (Bowman *et al*., 2009).

Biomass burning is also a chief source of climatically relevant non-CO_2 trace gases and their precursors (NO_x, CO, O_3, VOC, SO_2) and aerosols (BC, organic carbon) (Andreae and Merlet, 2001). The mix of chemical species released by fire depends on the physical and chemical conditions in the source region, the type of vegetation that burns, and the fuel moisture content that can affect whether the fire burns freely or smoulders. Importantly, the ratio of organic *versus* black carbon aerosols (with an opposite direct radiative forcing effect) varies greatly depending on vegetation type, from approximately 16 (boreal and temperate forest) to 3 (agricultural waste) (Andreae and Merlet, 2001; Naik *et al*., 2007). Biomass burning of forests and crop residues contributes about 40% of the total black carbon (BC) emissions (Ramanathan and Carmichael, 2008).

Key burn regions are the savannas of Africa, South America, and Australia, the boreal forests in the northern hemisphere, and Mediterranean regions (Figure 1; also ref. Ellicott *et al*., 2009). Many of these are clearly regions of high population density, including India, China, the North American west coast, Central America, and parts of eastern Europe. Table I shows the estimated annual area burned (Giglio *et al*., 2006) for various regions.

Fires probably contribute between 15 and 30% of current global total emissions of aerosol or O_3 precursors (Naik *et al*., 2007, Ramanathan and Carmichael, 2008; Denman *et al*., 2007), with African and South American fires being the chief source regions. From an atmospheric perspective, a change in biomass burning must be examined as the balance between ‘cooling’ compounds like scattering (organic and sulphate) aerosol and warming, absorbing (black carbon) aerosol, and greenhouse gases methane and ozone (Naik *et al*. 2007; Bowman *et al*., 2009). Globally, different assumptions about location of chief burn regions, anthropogenic fuel use and deforestation patterns for past and present conditions can regionally affect surface O_3 changes by up to $\pm$25 ppb, or more, and can introduce substantial uncertainty into radiative forcing estimates ($> \pm 0.3$ W m^{-2}; Ito *et al*., 2007).

The overall impact of fires in the climate system cannot be separated from long-range transport of pyrogenic emissions, especially that of particulate matter. The arctic appears particularly vulnerable to aerosols (Garrett and Zhao, 2006), transported, for instance, from boreal forest fires or agricultural burning in parts of Europe and northern Asia (Generoso *et al*., 2007; Stohl, 2006). The deposition of black carbon on ice and snow has a large effect in terms of radiative forcing and this could accelerate regional warming and trigger important biophysical (albedo) and biogeochemical (release of soil and peat carbon) feedbacks in the climate system (Law and Stohl, 2007; Quinn *et al*., 2008). A second region that shows potential sensitivity to aerosols is Australia (Rotstayn *et al*., 2007) where increasing rainfall over northwest of the continent has been linked to Asian aerosol emissions. The Australian monsoon may also be affected by seasonality and severity of the continent’s savannah fires (Lynch *et al*., 2007).

Table I. Estimated annual area burned (2001–2004) for various regions (from Giglio *et al*., 2006).

Region	Area burned ($\times 10^4$ km^2 = Mha)			
	2001	2002	2003	2004
Boreal North America	0.4	2.6	2.3	4.0
Temperate North America	1.4	1.7	1.5	1.2
Central America	1.8	2.2	2.9	1.8
NH South America	4.4	3.6	4.8	3.8
SH South America	12.4	12.7	10.8	13.4
Europe	2.9	1.6	2.6	1.9
NH Africa	153.2	135.2	125.5	129.8
SH Africa	84.0	82.4	79.6	75.3
Boreal Asia	6.3	9.3	14.5	4.9
Central Asia	16.5	26.7	17.1	18.9
Southeast Asia	10.8	10.2	8.4	16.1
Equatorial Asia	0.8	3.4	1.4	2.9
Australia	78.7	58.9	24.8	44.9

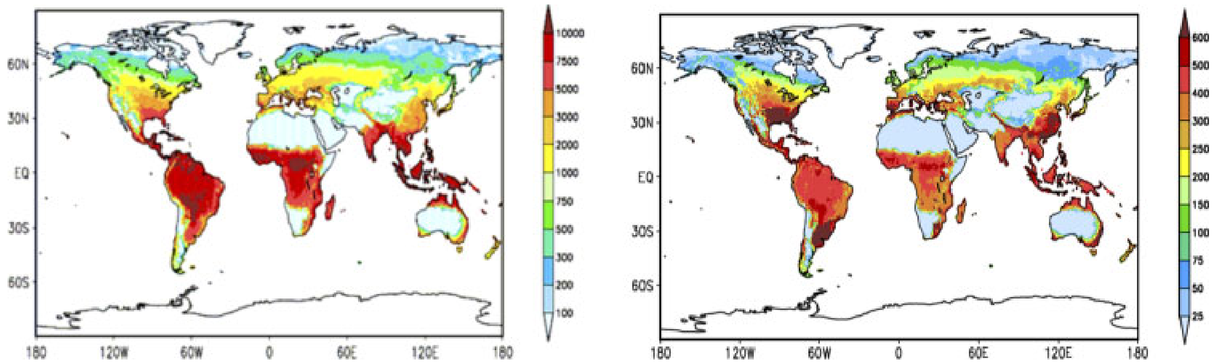

Figure 2. Global isoprene (left) and monoterpene (right) emissions (mg C m^{-2} a^{-1}; after Arneth *et al.*, 2007; Schurgers *et al.*, 2009). This figure is available in colour online at wileyonlinelibrary.com/journal/joc

Modern fire modules in terrestrial ecosystem models account for fuel combustion, fire intensity, spread, and pyrogenic emissions based on a parameterisation for ignition and simulated fuel moisture, fuel amount, and fuel type (Thonicke *et al.*, 2010). Under present conditions, ignition sources are thought to be mainly driven by human activities, although it is often forgotten that in many regions, human fire suppression and extinction also has a substantial impact on a region's fire regime. The assumption of biomass burning emissions reduced uniformly to 10% compared to present-day that is used in many calculations of the pre-industrial atmosphere, and hence pre-industrial to present radiative forcing is highly questionable (Ito *et al.*, 2007; Marlon *et al.*, 2008). A chief modelling challenge is, therefore, how to include both dynamically changing human-driven and natural fire regimes into chemistry-climate models. Human-induced fire differs from natural fire in terms of frequency, seasonal timing and impact. Removing human ignition is often thought to lead to an accumulation of fuel over several years which eventually ignites by lightning, resulting in large and high-intensity fires. By contrast, human-dominated landscapes are characterised either by more frequent and/or controlled (e.g. fire as part of land clearing or grazing management) fires of lower intensity, or by absence of fire in case of active control policies, and due to landscape fragmentation (Archibald *et al.*, 2009).

3.2. Biological emission of reactive carbon and nitrogen

The term BVOC subsumes a varied group of reactive carbon species emitted mostly from living vegetation. Of chief importance for atmospheric chemistry and climate are the terpenoids, isoprene, monoterpenes, and sesquiterpenes. Isoprene is, on a carbon mass basis, most likely the single largest emitted compound with global emission estimates of between 400 and 600 Gt C a^{-1} (Guenther *et al.*, 1995, Arneth *et al.*, 2008). Global terrestrial monoterpene emission estimates range from ∼30 to 130 Gt C a^{-1} (Arneth *et al.*, 2008). Models disagree on the total global source strength but tend to have rather similar geographic patterns (Figure 2). For sesquiterpenes, the current understanding on what controls emission and information to parameterise emission in vegetation models is insufficient to allow global estimates even within a broad margin (Duhl *et al.*, 2008).

BVOCs contribute to O_3 formation in high NO_x environments and are chief secondary organic aerosols (SOA) precursors (Tunved *et al.*, 2006). BVOCs also affect the atmospheric lifetime and concentration of methane since reduced hydrocarbons have the same major atmospheric sink reaction in their oxidation through OH (Poisson *et al.*, 2000). On the other hand, in remote locations without significant anthropogenic sources of reactive carbon, BVOC emissions regulate jointly with biogenic emissions of oxidised reactive nitrogen (NO_x) production of OH (e.g. Chameides *et al.*, 1992). Biogenic emissions greatly exceed anthropogenic VOC emissions and peak in tropical regions and during warm summer months in temperate and boreal climates when high radiation fosters O_3 production. Thus, the importance of these emissions is not merely a regional concentration, but has also a seasonal signature.

Different plant species vary greatly in the type and amount of BVOC they emit and a broad categorisation based on functional or genetic traits remains elusive (Guenther *et al.*, 1995; Kesselmeier and Staudt, 1999). Unknown totals and geographic patterns are consistently listed amongst chief uncertainties in simulations of future tropospheric O_3 or SOA levels, and associated radiative forcing (Stevenson *et al.*, 2006; Liao *et al.*, 2006; Shindell *et al.*, 2006; Tsigaridis and Kanakidou, 2007). An increasing number of field observations point also to as yet incompletely understood atmospheric oxidation pathways involved in the chemical destruction of VOCs (Lelieveld *et al.*, 2008; Li *et al.*, 2008). Amongst these is the importance of the formation of isoprene (and isoprene oxidation product) nitrates for remote tropospheric ozone burdens. These nitrates are of a sufficient lifetime to undergo transport and hence affect concentrations some distance from emission source areas (von Kuhlmann *et al.*, 2004; Ito *et al.*, 2007; Law and Stohl, 2007; Young *et al.*, 2009). They therefore have the potential to cause remote regional changes in climate but it is not known if this potential is realised.

Emission hot-spots (Figure 2) for isoprene are tropical evergreen and rain-green forests and woodlands (Guenther *et al.*, 1995; Lathière *et al.*, 2006; Arneth *et al.*, 2007). During summer months, the forests of the southeastern US, south and eastern China, parts of southern Europe and central and southeastern Asia are also important sources because these regions contain a relatively large number of high-emitting species (Guenther *et al.*, 1995; Lathière *et al.*, 2006; Arneth *et al.*, 2007). Monoterpenes tend to have an additional source in coniferous temperate and boreal forests and in Mediterranean and seasonal tropical ecosystems (Guenther *et al.*, 1995; Lathière *et al.*, 2006; Schurgers *et al.*, 2009). What is common to terpenoid BVOC emissions is their strong response to short-term variations in leaf temperature and light (Guenther *et al.*, 1995). The short-term emission response also underlies projections of strong increase in BVOC in a warmer future climate. Recent simulations argue for a more complicated picture as CO_2 appears to have an inhibitory effect in case of leaf isoprene production (Possell *et al.*, 2005).

Tropical soils are also an important source of reactive oxidised nitrogen in the form of nitric oxide (NO) which microbial production depends on biogeophysical and chemical properties of the soil, e.g. soil porosity, soil water content, temperature, nutrient status, and land management. The emitted NO is rapidly oxidised to NO_2 (within minutes), resulting in a net atmosphere–biosphere emissions flux of NO_x ($NO + NO_2$) in pristine regions. Nearby anthropogenic sources the supply of NO_x by advection results in a net NO_x deposition (Ganzeveld *et al.*, 2002) also relevant for Net Primary Production (NPP). A large input of nitrogen via wet and dry deposition can also result in forest decline as a consequence of nitrogen saturation or acidification of the soil (de Vries *et al.*, 2007). NO_x is crucial for tropospheric photochemistry in remote and rural areas involved in the production of ozone and OH. Intense cultivation and application of fertilizers and animal manure actually controls to a large extent agricultural emissions of NO but also of ammonia (NH_3) in Europe, Asia and North America. NH_3 is involved in rain- and cloud-water chemistry, the formation of N-containing aerosols, acidification of ecosystems and is essential to assess the role of sulfate aerosol in climate (Luo *et al.*, 2007). The main difficulty with assessing the role of atmosphere-biosphere exchange of NO_x and NH_3 in atmospheric chemistry-climate interactions is the limited number of observations, a large heteorogeneity in soil emissions, significance of sub-grid scale deposition but also consideration of management practices This is also reflected by the wide range in global inventories, e.g. that of soil biogenic NO_x emissions with a large range between 4 and 21 Tg N yr^{-1} (Ganzeveld *et al.*, 2004).

Deforestation associated with human agriculture and land cover change reduces regional emissions of isoprene and monoterpenes that tend to be emitted from woody rather than herbaceous vegetation (Arneth *et al.*, 2008, Lathière *et al.*, 2006). A notable exception may be where natural forest or herbaceous crop-land is replaced by woody biofuel plantations (Arneth *et al.*, 2008). Typical woody biofuel species are willow, eucalypt and oil palm, all of which have substantial rates of isoprene emissions that likely exceed emissions of local vegetation. Large-scale conversion of tropical rainforest into oil palm plantations in areas of high (or projected to increase) NO_x emissions of SE Asia therefore has the potential to be detrimental for local O_3 levels (Hewitt *et al.*, 2009). At the same time will an increase in agricultural *versus* forest area also affect emissions of oxygenated BVOC (methanol, acetaldehyde; Chung *et al.*, 2002; Goto *et al.*, 2008). The effect this may have on regional chemistry and climate is unknown as for these the oxygenated biogenics regional emission patterns and chief source areas are unknown and their atmospheric reaction pathways are incompletely understood (Kwan *et al.*, 2006; Guimbaud *et al.*, 2007; Heiden *et al.*, 2003).

Estimates of future BVOC emissions have been identified as having a substantial effect on corresponding projections of ozone formation and SOA. Due to possible compensatory effects of changes in climate and CO_2 concentration on the direction of the emissions there is as yet no agreement in terms of magnitude, sign, and geographic distribution of emissions (Heald *et al.*, 2008; Sanderson *et al.*, 2003; Young *et al.*, 2009), in some regions deforestation, rather than climate change, will likely be of overriding importance (Lathière *et al.*, 2006). For a global isoprene emission increase by 34% compared to present, in response to a near-5 °C warming, Sanderson *et al.* (2003) calculated a positive ozone response above 30 ppb compared to present in highly polluted regions of the continental northern hemisphere. A 50% reduction of isoprene in tropical areas, at otherwise unchanged conditions, caused an increase in surface O_3 by about 5–15% in the tropical source regions due to reduced isoprene ozonolysis (von Kuhlmann *et al.*, 2004), a sensitivity confirmed by Young *et al.* (2009), who compared O_3 levels in response to different future isoprene projections. Simulations with a chemistry-climate model constrained with land cover and land use change scenarios provided by an integrated assessment model showed a less drastic decrease in isoprene emissions associated with tropical deforestation (Ganzeveld *et al.*, in press).

These simulations, which considered the impact of land cover change on micrometeorology and surface exchange of chemical compounds (dry deposition and biogenic emissions), showed a ∼15% decrease in global annual isoprene emissions. Simulated changes in atmosphere–biosphere exchange of NO_x and O_3 associated with these land cover and land use changes appear to be small also pointing at the importance of compensating effects.

SOA absorb and scatter radiation and depending on size, chemical composition and particle age. These particles also act as efficient cloud condensation nuclei (Andreae and Rosenfeld, 2008). Low volatility oxidation products of BVOC condense on stable atmospheric clusters and contribute to the growth of secondary organic

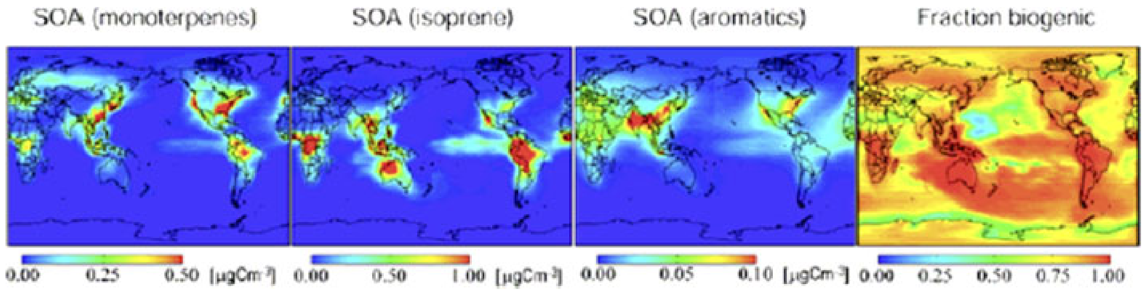

Figure 3. Annual mean simulated SOA concentration for the year 2000. SOA from each precursor is shown separately. The fourth panel shows the fraction of SOA from biogenic precursors (isoprene + monoterpenes). The colour scales are saturated (after Heald *et al.*, 2008). This figure is available in colour online at wileyonlinelibrary.com/journal/joc

aerosol (SOA) (Tsigaridis and Kanakidou, 2007; Heald *et al.*, 2008). A number of controlled chamber experiments have demonstrated considerable aerosol yield from monoterpenes and sesquiterpenes (Hoffmann *et al.*, 1997; Bonn and Moortgat, 2003; Lee *et al.*, 2006) and isoprene is also an aerosol source (Claeys *et al.*, 2004). SOA are estimated to increase substantially in response to enhanced future BVOC emissions, responding to warmer temperature. The BVOC effect on SOA can be 3- to 7-fold relative to results that ignore the climate-BVOC interactions (Liao *et al.*, 2006; Tsigaridis and Kanakidou, 2007; Heald *et al.*, 2008) and simulations indicate biogenic SOA to become one of the dominant aerosols over the 21st century (Tsigaridis and Kanakidou, 2007) in some regions. Figure 3 shows the simulated concentrations of surface SOA (Heald *et al.*, 2008) sourced from BVOCs.

Estimating how aerosols and cloud condensation nuclei above terrestrial regions might be affected by different BVOC emissions of local vegetation independently from anthropogenic emissions is difficult. Still, the link between BVOC emissions and particle formation is being quantified under field conditions (Laaksonen *et al.*, 2008; Held *et al.*, 2004, Holzinger *et al.*, 2007). For boreal forest, Tunved *et al.* (2006) provided a first quantitative analysis for the influence of regional BVOC emissions by demonstrating that the integrated mass of aerosol particles of diameter below 450 nm increases linearly with the accumulated monoterpenes an air mass containing continuously growing particles takes up while moving over land. It is as yet not possible to judge whether the slope of the observed relationship would vary with vegetation type or whether a similar slope would be obtained in a truly pristine atmosphere (Andreae, 2007; Kanakidou *et al.*, 2000). Effects on cloud albedo over boreal forests linked with BVOC-aerosol interactions have been estimated to cause a radiative cooling between -2 and -7 W m^{-2} over the forests (Spracklen *et al.*, 2008).

A number of chemistry-climate experiments have recently calculated not only radiative forcing but actual regional climate effects of changing aerosols and ozone (Shindell *et al.*, 2008; Levy *et al.*, 2008). So far, focus has been on the effects of anthropogenic pollution control and results demonstrate that changes in pollutant (and precursor) emissions can enhance or diminish north-south temperature change gradient, compared to effects of long-lived greenhouse gases alone. This clearly has the potential to affect atmospheric circulation patterns (Shindell *et al.*, 2008; Levy *et al.*, 2008; Shindell and Faluvegi 2009). A systematic study that separates effects of biogenic emissions of reactive substances from those of anthropogenic emissions is to date lacking, despite large sensitivities, in simulation experiments to changing biogenic emissions. The large role of BVOC for SOA burdens pose the question of their future role in the climate system, considering that reduced sulphate aerosol may be compensated by larger amounts of biogenic SOA as BVOC emissions as vegetation positively responds to warmer temperatures. Whether or not proposed, biogenic SOA-climate feedbacks may be dampened by the inhibition of some BVOC by CO_2 still needs to be investigated (Kulmala *et al.*, 2004; Arneth *et al.*, 2007). Clearly, untangling the complex puzzle of positive and negative feedbacks that exists due to the interacting climate, biogenic, and anthropogenic emission (and deposition) processes remains a challenge.

We do not suggest that these feedbacks could significantly change the projected global mean warming from a doubling of atmospheric CO_2. However, we do suggest that the evidence would strongly support a hypothesis that these feedbacks could significantly change the regional pattern of warming or cooling driven by the doubling of CO_2.

4. Missing land-surface elements with regional signatures: physical and biophysical characteristics

4.1. Fire and the impacts on physical and biophysical properties

In global climate models there are a suite of challenges relating to modelling fire. Fire changes the albedo and roughness of a landscape which affects the amount of available energy, and the partitioning of available energy between sensible and latent heat. The changes in interception by burned canopies, and the increased transpiration as forests re-grow affects the partitioning of rainfall between evaporation and runoff. These may, if the fire is over a large enough area, affect the boundary layer, cloud formation, and associated processes (Andreae

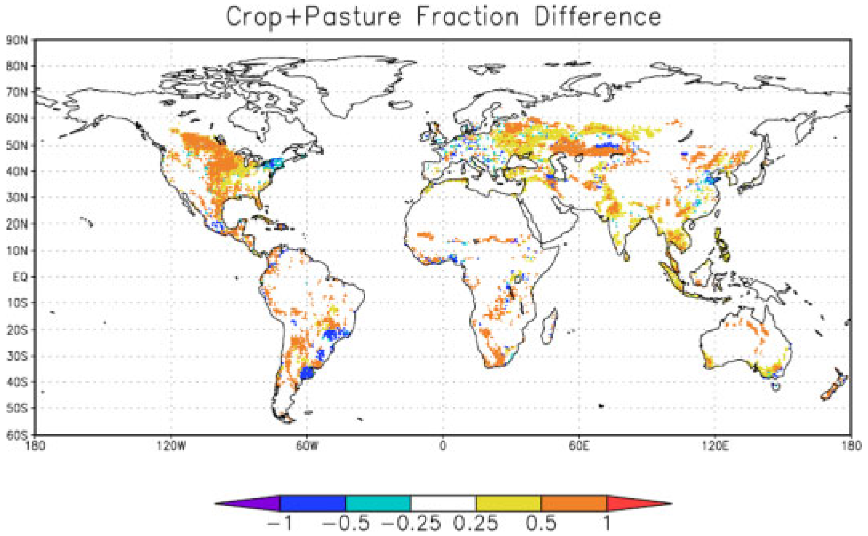

Figure 4. Extent of land cover change expressed as the difference in crop and pasture cover between 1992 and 1870. Blue colours represent changes that decrease pasture and crop cover while yellows and browns are increases (25–50% and 50–100%, respectively) (after Pitman *et al*., 2009). This figure is available in colour online at wileyonlinelibrary.com/journal/joc

et al., 2004). Altered frequency of fires in response to climate change (rather than anthropogenic land use change) may reduce the ability of forests to dominate some landscapes and lead to transient shifts in vegetation towards more savanna or grassland dominated ecosystems. Over a large enough area this could begin to be a dominant regional driver but it is likely to be on a timescale of a century or more before transitions at a scale that could be regionally significant on climate would take place.

4.2. Land cover-climate interactions

Humans have modified close to 50% of the Earth's surface (Vitousek *et al*., 1997). The location of intensive land cover change is highly regionalised (Figure 4) and most strongly seen in Europe, the eastern states of the US, and parts of Asia.

Land cover change (LCC, removal of forests for crops or grazing, replacement of crops and grasses by forests, etc.) affects regional climate through impacts on the surface albedo and radiative forcing (Forster *et al*., 2007), partitioning of available energy between sensible and latent heat, boundary layer temperature, moisture profile and depth and the partitioning of rainfall between evaporation and runoff (Betts *et al*., 1996; Pitman, 2003). The global and regional climate modelling communities have demonstrated impacts on surface temperature, rainfall, and turbulent energy fluxes if land cover is perturbed (Henderson-Sellers *et al*., 1993; Chase *et al*., 2000; Werth and Avissar, 2002; 2005; Findell *et al*., 2006).

The evidence that LCC has a direct and significant effect on climate over the regions of change is indisputable. Most recent demonstrations include Findell *et al.* (2009) who note that over areas of large-scale land cover change (US east coast, Europe, India, China) the impact on the regional scale hydrometeorology can be on par with sea surface temperature anomalies including ENSO or the observed warming trend. Pitman *et al*. (2009) performed a LCC experiment with seven different climate models and concluded that a direct and significant impact on the regions exposed to land cover changes occurred. Figure 5 shows a strong signal from LCC co-located with the perturbation on the latent heat flux, temperature, and precipitation. All models show a strong and statistically significant impact of LCC on the latent heat flux and on temperature over the regions of LCC. Four of the seven models show a statistically significant impact on precipitation. Since humans live, grow crops, source water, etc. at local-to-regional scales, Figure 5 highlights the need to represent LCC in climate model projections that are used to explore the surface impacts of climate change on regions that have undergone LCC.

There is no agreement on whether LCC directly affects climate over regions remote from the perturbation. Remote teleconnections, where LCC in one region is used to explain changes over another continent, has been addressed many times. Some authors find clear teleconnections (e.g. Gedney and Valdes, 2000), while others do not (e.g. Findell *et al*., 2007). Clarifying this issue is important because if LCC does trigger significant teleconnections, a global rather than regional-scale response could be anticipated. The first multi-model study that used statistical methods that accounted for autocorrelation and performed multiple realisations did not find common remote teleconnections from observed LCC (Pitman *et al*., 2009). This is also shown in Figure 5 which demonstrates changes remote from the LCC are below the levels expected by chance. However, the experiments by Pitman *et al.* (2009) used observed LCC and this did not include a large perturbation over tropical regions. It is possible that future LCC over the tropics could perturb the fluxes of energy and water sufficiently to trigger larger-scale responses. It is also possible that LCC affects regional emissions of reactive carbon and

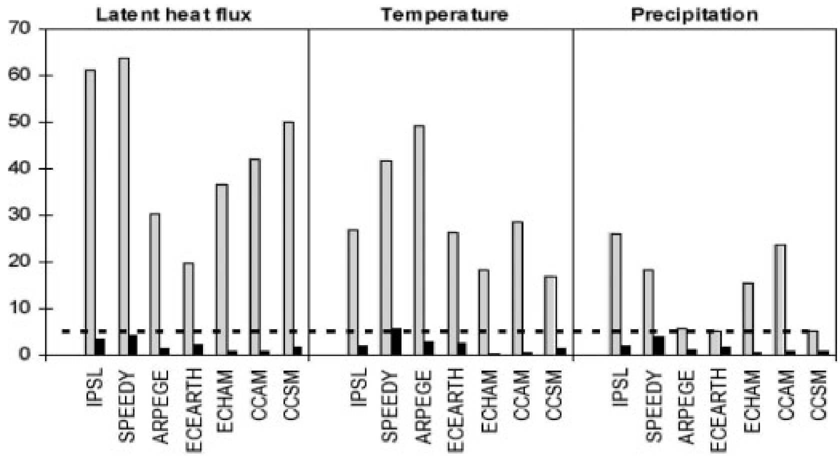

Figure 5. Percent of land area that exhibits statistically significant changes in June-July-August latent heat flux, temperature and precipitation. Stippled bar is the percent of grid points with statistically significant changes where land cover changes (change in leaf area index >0.5) within each climate model. The solid bar is the percent of grid points with statistically significant changes where land cover is not changed. The horizontal line is the 5% significance level, expected by chance (from Pitman *et al*., 2009).

nitrogen compounds enough to trigger responses that teleconnect globally (ref. Section 3.2).

LCC also affects extremes of maximum temperature and convective precipitation. Zhao and Pitman (2002) showed that the return values of the annual daily maximum temperature and the seasonal changes of frequency in daily maximum temperature and convective precipitation over Europe and China were affected by LCC – in effect, the distribution of land cover can affect the climate's sensitivity to increasing CO_2.

A very new area of research is the interaction between LCC and atmospheric chemistry. Ganzeveld and Lelieveld (2004) suggested that the understanding of the impact of LCC on the climate also required a coupling to atmospheric chemistry, but these preliminary analyses were limited in both temporal and spatial scales. Ganzeveld *et al*. (2010) used a coupled chemistry-climate model to demonstrate that the impact of LCC on atmospheric chemistry appears to be most significant in the tropics in regions where vast deforestation is expected to occur. There does not seem to be significant large-scale effects distant from the locations where LCC occurs. In addition, a consistent consideration of the impact of land cover and land use changes on emissions, deposition, canopy-interactions, and meteorology revealed the significance of compensating effects.

4.3. Urban systems

Climate models, used for the 4th assessment report of the IPCC, did not include representations of urban areas. This is extremely unlikely to have affected continental-scale simulations, or the response of the Earth's climate to increased greenhouse gases. At the spatial resolution of a typical climate model, 1–5% of an individual grid square would be urban over eastern North America, most of western Europe and isolated regions elsewhere based on data from Loveland and Belward (1997). However, urban areas are concentrated in some regions and as the spatial resolution of climate models increase some grid squares in some regions will include an increasingly high proportion of urban surfaces.

There is no doubt that urban surfaces affect regional climates (Shepherd, 2005; Arnfield, 2003). There are literally hundreds of scientific papers that highlight the impact of urban surfaces on temperature, rainfall (Shepherd and Burian, 2003), convection (Baik *et al*., 2001), storms (Gero *et al*., 2006), air pollution, boundary layer structure (Martilli *et al*., 2002), etc. Calls for adequately characterizing the urban environment in climate models including the links between the urban system and aerosols (e.g. Jin and Shepherd, 2005) have been made, and urban schemes exist (Best, 1998; Masson, 2000; Martilli, 2002; Oleson *et al*., 2008) although the links with urban aerosols and air quality remain underdeveloped.

One of the very few attempts to represent the urban system in a climate model was conducted by Betts and Best (http://www.cru.uea.ac.uk/projects/betwixt/documents/BETWIXT_TBN_6_v1.pdf). Figure 6 shows the distribution of urban areas they used in a climate model. Note that fractions are very small, but this is dependent on scale. While only up to 5% of a climate model grid square may be urban land at $3^\circ \times 3^\circ$ (a typical resolution of a coupled climate model), as resolution increases to $1^\circ \times 1^\circ$ some grid squares would become dominated by urban surfaces coincident with major cities.

They noted that the interactions between urban landscapes and radiative forcing, and the impact on regional climate from additional heat sources could not be included in regional climate projections without coupling models of urban landscapes into the climate models. Most critically, they showed that including urban effects changed the *shape* of the distribution of temperatures which they interpret as demonstrating that it is not possible to take present-day temperature distributions and simply add a change in the mean temperature to this distribution. The effects of the urban surfaces, the feedbacks and the additional heat from the urban systems need to

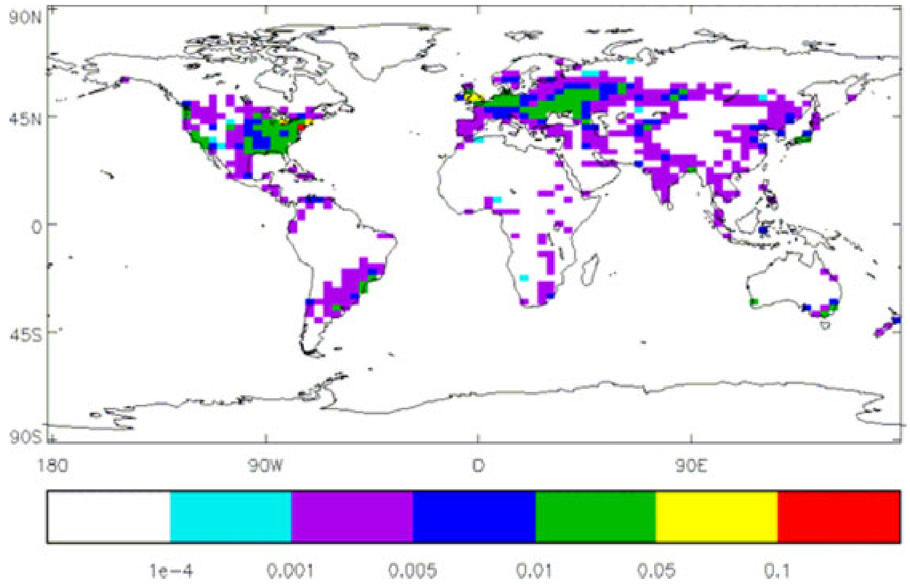

Figure 6. Fraction of climate model GCM grid squares specified as urban land (data from Loveland and Belward, 1997). This figure is available in colour online at wileyonlinelibrary.com/journal/joc

be coupled into the climate models – but only for some areas.

The issue of direct heating of the atmosphere by urban landscapes via release of heat due to energy use may also have been under-considered. Chaisson (2008) provides a global perspective on warming associated with heating resulting from energy usage and indicates that the amount of potential global warming is significant, but occurs slowly relative to the projections of warming due to increasing greenhouse gases. De Laat (2008) also note the potential role of heating from energy use, noting that this is a highly regionalised phenomenon. They note that for large energy using countries energy consumption is approximately 0.37 W m^{-2} (US), 0.24 W m^{-2} (China), 0.18 W m^{-2} (India), but increase to 1.37 W m^{-2} (UK), 2.07 W m^{-2} (Japan) and 4.19 W m^{-2} (The Netherlands) due to higher density of populations. Thus, while globally the energy released is negligible (0.03 W m^{-2}), regionally it can be of an order equivalent to the increase in radiative forcing from a doubling of the concentrations of greenhouse gases. Block *et al*. (2004) showed (in a very limited region, over only 3 months in spring) an impact on the regional atmosphere from heating of 2 W m^{-2} indicating a possibly important regional-scale effect. A remaining issue is how the urban heat island effect may also affect the dispersion of the concentrated emissions of reactive compounds which could induce a non-linear response of the chemistry-climate interactions. Current state-of-the art chemistry-climate models do not yet consider these subtle features of the urban dispersion of reactive compounds and aerosols.

4.4. Crops and phenology

The basic albedo, roughness, and stomatal function of crops have been included in many global climate models for several decades via the specification of a specific plant functional type (Dickinson *et al*., 1986; Sellers *et al*., 1986). However, the phenology of crops is highly complex and difficult to model in global or regional climate models. In some recent attempts (e.g. Krinner *et al*., 2005) the phenology of crops was treated the same way as natural vegetation, but with different maximum leaf area index values and modified parameters for critical temperature and humidity in the phenology scheme. Lawrence and Slingo (2004a, 2004b) noted a large and regionally significant impact on simulated climate resulting from how phenology was included and Pitman *et al*. (2009) attributed a part of the range of responses to land cover change to how crop phenology was represented.

The physiology and natural phenology of crops can probably be represented reasonably in models via specific parameters or with relatively simple crop models. Crops that are perennial (tea, many tree crops, etc.) probably do not function very differently in a climate sense to natural shrubs and trees. Harvesting does not affect vegetation cover, albedo, leaf area, and ability of the vegetation to access water for transpiration. Therefore, the partitioning of available energy between sensible and latent heat, and the partitioning of available water between evaporation and runoff are unlikely to be changed enough to perturb the regional climate. This can be clearly contrasted with crops including wheat, corn, and rice that are commonly spatially widespread and can be harvested by large-scale mechanisation. These can lead to sudden transformation of albedo, roughness, leaf area, etc. and have the potential to affect energy and water partition. In these circumstances, human intervention via harvesting needs to be accounted for. The timing of harvesting may be estimated via the crop model (when the crop has matured, etc.) but the details of how to do this, and potential sensitivities that might be included into the climate models, needs careful attention. A limited number

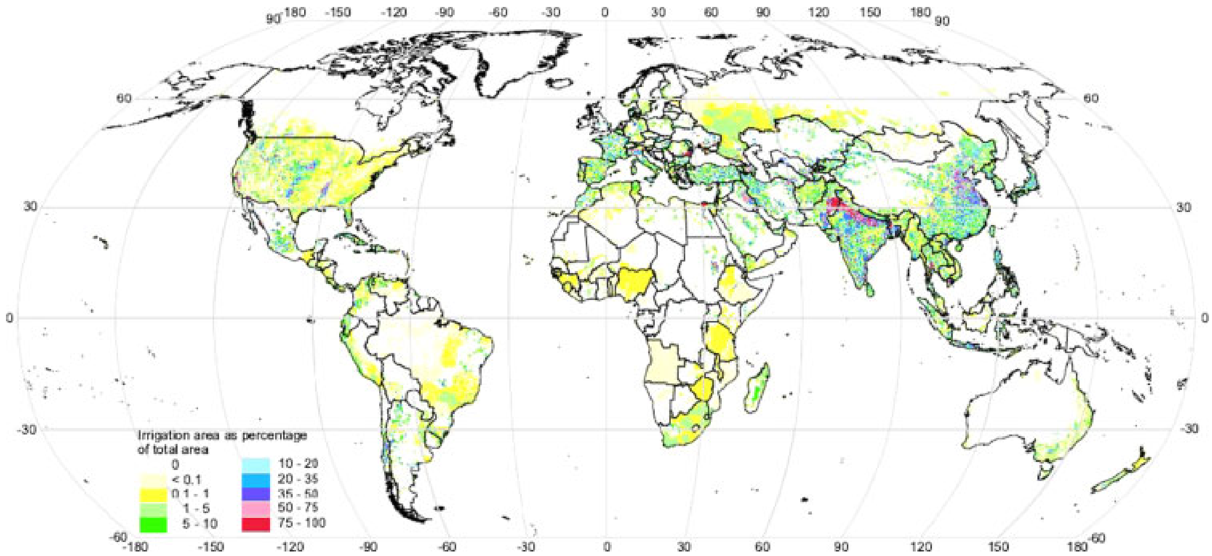

Figure 7. Global estimate of irrigation (after Siebert *et al*., 2005). This figure is available in colour online at wileyonlinelibrary.com/journal/joc

of dynamic vegetation models have started to address these issues with a variety of approaches, an overview is provided in Arneth *et al*. (2010b, and references therein).

4.5. Irrigation

Irrigation provides around 2600 km^3 of water to the land each year, around 2% of the natural flux of rainfall (Sacks *et al*., 2009). It is highly concentrated in areas of high and dense populations including China and India (Figure 7), and the estimated abstractions in those countries are as large as those in the United States and European countries. While the global signature of irrigation may be negligible, regions where irrigation may have a significant regional climate impact is coincident with high population density.

Irrigation changes the natural partitioning of available energy between sensible and latent heat. The artificial supply of water to a region must have some effect on climate at least over an irrigated region. De Rosnay *et al*. (2003) found an increase of 3.2 W m^{-2} or 9.5% in annual mean latent fluxes averaged over the Indian subcontinent resulting from irrigation. Boucher *et al*. (2004) estimated a global mean radiative forcing ranging from 0.03 to 0.1 W m^{-2} and also found surface cooling of up to 0.8 K over irrigated areas, although they used a very simple land surface scheme and changes in temperature remote from the irrigation of a scale similar to the changes over the areas of irrigation suggests problems with the statistical analysis. Haddeland *et al*. (2006) used a variable infiltration capacity macroscale hydrologic model containing an irrigation scheme to study the Colorado and Mekong River basins and showed that irrigation led to less stream flow, more evapotranspiration and decreased surface temperatures. In simulations for the Yellow River basin in China, Tang *et al*. (2007) found that irrigation also led to decreased stream flow, increased latent heat flux, and decreased ground surface temperature. In one of the few systematic analyses of the effect of irrigation on observed temperatures, Lobell and Bonfils (2008) studied stations in California to show an impact of irrigation on maximum temperatures (a cooling with a mean scale of 5.0 °C for fully irrigated sites), in agreement with Kueppers *et al*. (2007). However, they showed that minimum temperatures were not significantly affected by irrigation, leading overall to a decrease in the diurnal temperature range.

At smaller spatial scales, a study of the impact of irrigation on the surface energy budget in the US high plains was conducted by Adegoke *et al*. (2003). Their results indicated a 36% increase in surface latent heat flux coupled with a 2.6 °C increase in dewpoint temperature under irrigation. Along with these changes, they also found that both surface sensible heat flux and near-ground temperature decreased by 15% and 1.2 °C, respectively. An assessment of changes in historical near-surface temperature records for Nebraska, USA, also showed notably cooler temperatures over irrigated areas (Mahmood *et al*., 2006). Significantly, Lobell *et al*. (2008) have demonstrated that the cooling effect of irrigation masks the warming effects of increased greenhouse gases over some regions of rapid irrigation growth over the last 40 years (parts of India and China).

Kueppers *et al*. (2007) found that irrigation affected the atmosphere several kilometres beyond the area of irrigation in California due to advection. Lobell *et al*. (2008) note that it remains unclear whether irrigation affects regions remote from the source of the water. Douglas *et al*. (2009) explore the impact of agriculture and irrigation over India and conclude that regional changes need to be included in weather forecasting and multi-decadal climate variability. While this may be true, their simulations were limited to a single regional model and a 5-day period for a single year, and point to the need to significant further analysis.

Globally, Sacks *et al.* (2009) showed irrigation has a negligible impact on near-surface temperatures. However, irrigation did cool some regions by ~0.5 °C and *warmed* some regions by ~1.0 °C. The cooling was attributed to links with cloud changes rather than direct evaporative cooling. A similar result was found by Puma and Cook (2010) using ensemble simulations of the 20th century. They found some small regions of seasonal cooling of more than 2 °C coincident with intense irrigation in northern India. Overall, Puma and Cook (2010) agree with Sacks *et al.* (2009) and Boucher *et al.* (2004) that irrigation has a negligible effect on the global temperature (likely a cooling effect of <0.1 °C). However, a cooling signal due to irrigation is generally sustained through the 20th century in transitory simulations conducted by Puma and Cook (2010). This cooling is approximately 0.3 °C in the 0–30°N latitude band in northern hemisphere winter and a similar size in both 0–20°N and 30°N–60°N in northern hemisphere summer. Puma and Cook (2010) note the need to include irrigation in future simulations, particularly in regions with unsustainable irrigation resources; an important point since maintaining irrigation in these regions in future projections would tend to mask a CO_2-induced warming signal.

Overall, the impact of irrigation on climate is gradually becoming clearer. The impact on the global climate appears negligible, but irrigation has a strong regionally specific signal with a strong time dependency.

5. Conclusions

In many areas, climate models contain the appropriate dynamics and physics to provide reasonable regional projections. In these areas none of the regional drivers omitted from current climate models strongly affect the global or continental-scale patterns. In other areas where some of these regional drivers act strongly, existing regional projections may be wrong, and significantly wrong, because they do not include regionally significant processes. These weaknesses are unlikely to significantly affect global or continental-scale projections of the climate sensitivity to a doubling of CO_2 – this paper does not suggest that the IPCC assessments are flawed at the large scale. However, impacts of global warming are realised at regional scales, and we have shown that at this scale a suite of important physical, biological, and chemical processes have the potential to moderate or amplify the large-scale dynamics and physics simulated by climate models.

It is easy to show where processes that *may* contribute a significant (i.e. a statistically significant) regional driver of climate may be found. Figure 8 shows these regions, sourced from the earlier figures presented in this paper and generalised. Most continental surfaces, particularly those with large population densities, seem to be affected by one or more of the processes included in this paper – that is, one or more of the processes described in this paper affect almost the entire continental surface. The key is the definition of the word 'affected'. Are these regions significantly affected or are these additional drivers merely a minor contributor to the overall changes? We cannot answer this question because the systematic studies to determine how much of a contribution each of these regionally focussed processes have on the regional climate have not yet been undertaken.

The way forward is quite clear and the elements of the ways forward are already in place within individual science communities. The challenge is to define the circumstances whereby regional processes can dominate a regional climate. Then, the coupling of the surface,

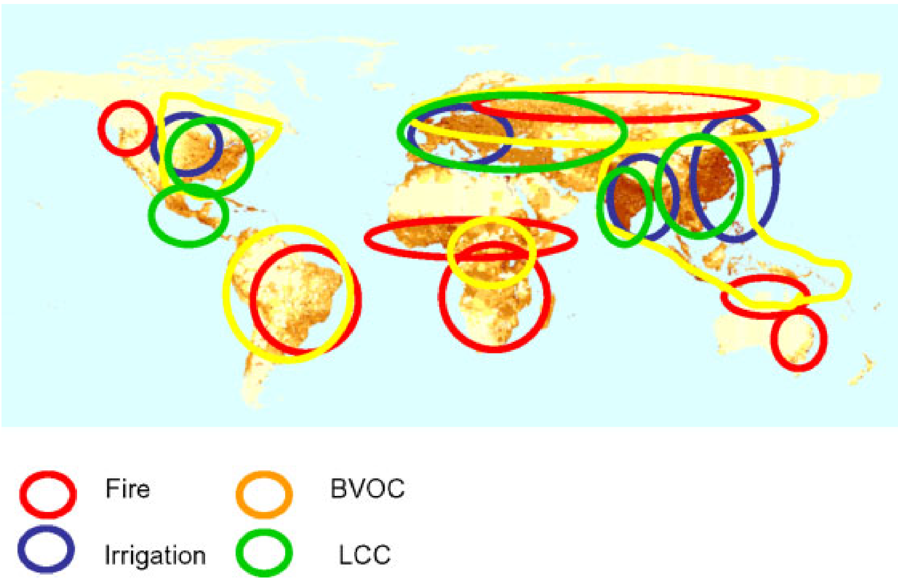

Figure 8. Synthesis of the regions that are affected by one of fire, irrigation, BVOC emissions, and LCC. These regions are superimposed on a population density map of 2000 sourced from the Population of the World, version 3 (GPWv3), Center for International Earth Science Information Network (CIESIN), Centro Internacional de Agricultura Tropical (CIAT). Palisades, NY, CIESIN, Columbia University, http://sedac.ciesin.columbia.edu/gpw/maps/gldens1km.pdf. This figure is available in colour online at wileyonlinelibrary.com/journal/joc

through the planetary boundary layer, with the clouds in terms of energy, water *and chemistry* needs to be developed. This is a new and demanding science – while the coupling of energy, water, and chemistry has been explored in coupled climate models, how these could moderate or amplify the globally imposed climate change signal at a regional scale is not known.

Our suggestion is to move from defining regional climates as a localisation of large-scale dynamics and physics to defining regional climates as a localisation of large-scale dynamics and physics, but with the inclusion of additional drivers of regional climate. A climate change of $x\,^{\circ}$C triggers different combinations of feedbacks over a natural forest to a natural grassland, to a crop or an urban surface. The nature of which feedbacks are triggered, over which surfaces, under which conditions, is not understood and offers a potentially rich vein of future research. By understanding these issues, the projections of regional climate changes, reflecting the feedbacks that are regionally dominant can be enhanced. To achieve this, the various regional feedbacks highlighted in this paper need to be explicitly resolved in the global climate models.

The problem in many earlier studies that explored climate sensitivity to specific regional processes (specifically land cover change experiments) is that computation limitations tended to mean that a single model was run once with a land cover representative of (say) 1700 and then once with a land cover representative of the present day. The standard protocol for the IPCC is to run multiple models and run each model multiple times to sample internal model variability in each model. Given this internal model variability, a statistically significant 'signal' can be looked for against the 'noise' of natural model variability. If the signal is common to many different climate models, across several independent realisations conducted with each climate model, then the level of confidence that can be attributed to the signal is significantly higher than if only one or two models display a common signal. It is also very important that appropriate statistical tests are used that ideally account for both temporal and spatial autocorrelation – to minimise the risks of 'false-positives'. The standard t-test, for example, does not account for spatial correlation within fields (Zwiers and von Storch, 1995) although it can be modified to account for time auto-correlations (Findell *et al.*, 2007). A design of this kind was used by Pitman *et al.* (2009) to show statistically significant changes due to land cover change only over the regions of land cover change.

There are two levels of proof that a process is important to a climate model. The first level of proof is that a regional process or phenomenon, like fire, irrigation, emissions of reactive compounds, etc. affects the global climate. There is little evidence that they do; there is no evidence that the probability density function of how much the global mean temperature will rise for an effective doubling of CO_2 is sensitive to these regional processes. The lack of evidence does not prove these processes are not important of course because the rigorous experiments have not been conducted systematically to explore each process. However, in the absence of any evidence to the contrary, it is reasonable to hypothesize with some considerable confidence that the regional-scale processes discussed in this paper do not affect the scale of warming projected in the IPCC 4th assessment report at continental scales and above.

The second level of proof is whether these processes, which have distinct regional signatures, affect the regional climate of some regions. This is harder to demonstrate because multiple climate models do not always respond similarly in a given region; an example is how India might be affected by irrigation in an ensemble of climate models would be partially dependent on how well the monsoon was simulated in each member of that ensemble. It is likely that a quite significant regional signature would be required to perturb the regional climate of a given region. Land cover change is an example of a perturbation that does have a regionally significant impact on the regional-scale climate. It seems, probably, that large cities also affect the regional-scale climate. It is conceivable that biogenic sources of reactive carbon and nitrogen, coupled with atmospheric chemistry and aerosols, are very important to the regional climate of some tropical regions and biomass burning may affect Arctic regional climates via aerosol deposition. However, until the systematic studies using an IPCC-style experimental design are conducted it is speculative to conclude that they would.

It is clear that the large-scale driver of future climate change, at least on policy-relevant time scales are increasing greenhouse gases in the atmosphere. A very impressive and sustained effort to understand the impacts of future increases in these gases has been underway for around 40 years. At some point in time, we must reach the point that the large-scale drivers are captured well enough, such that while incremental advances are possible, the scale of improvements in climate projections from these incremental gains will become negligible. At this point in time, a point we do not suggest we have approached, improving climate models for policy-relevant spatial scales will require processes which are regionally important to be added into the climate models. Some processes will be more important than others and the case is already clear for land cover change. For other processes, a framework is required to develop modules to represent these processes and to examine and test them in well designed international programs. This will develop an understanding of which are important, where they are important, how important they are, and how do they interact with the increasing greenhouse gases. We suggest that advances for the 5th assessment report of the IPCC will probably mean climate models have reached their limits in terms of regional scale projections *without* the inclusion of key regional scale processes. We suggest that a framework to identify key regional climate drivers, and then build, test, evaluate, and choose modules to represent key regional climate drivers is essential well before the 6th assessment report of the IPCC.

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