

An overview of regional land-use and land-cover impacts on rainfall

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ABSTRACT

This paper documents the diverse role of land-use/land-cover change on precipitation. Since land conversion continues at a rapid pace, this type of human disturbance of the climate system will continue and become even more significant in the coming decades.

1. Introduction

The role of landscape change in altering convective rainfall has been well documented (e.g. Pielke, 2001; Pitman, 2003). This paper summarizes the subject by landscape conversion type with a particular focus on how regional change results in changes in rainfall, in the same area. The teleconnection effect, where regions remote from the landscape conversion have altered rainfall (e.g. as discussed in Chase et al., 2000 and Avissar and Werth, 2005) is not the focus of this paper, as this is discussed elsewhere (e.g. Pielke et al., 2002; Marland et al., 2003).

Cotton and Pielke (2007; Table 6.2) list papers on the issue as to how regional weather patterns are affected by land-use and land-cover change. Warm season rainfall should be expected to change whenever deep cumulus convection is common in a region, since the surface fluxes of moisture, sensible, and latent heat change. This is the fuel for thunderstorms both in terms of moisture and in altering the convective available potential energy (Stull, 1988). The effect on cold season rain and snowfall, if any, is much less clear, and is not discussed in this paper.

The structure of the paper is to present examples of the role of human land-cover/land-use change for several landscape types. The main goals are to update earlier review papers as

well as to further demonstrate the important role of landscape change as a first-order climate forcing. Land-use/land-cover change, while highlighted as a major climate forcing in National Research Council (2005), is still not generally recognized in international climate assessments as having a role on precipitation that is at least as large as caused by the radiative effect of the human addition of added well-mixed greenhouse gases.

We categorize the human landscape conversions with respect to several biome classes. We chose the specific categorization framework in Sections 2 to 9 since the different responses can be more effectively presented. While not inclusive of all landscapes, the examples that we present clearly document the important (and diverse) role of land-use/land-cover changes on climate.

2. Short-grass conversion to dryland agriculture and irrigated agriculture

Prior to agricultural settlement in the mid-19th-Century, grasslands comprised 300 million hectares of central North America, 21% of which (61.5 million ha) was short-grass steppe (Küchler, 1964; Sims and Risser, 2000). Short-grass steppe occupies a region that stretches from Western Nebraska to Western Texas, adjacent to the eastern front of the Rocky Mountains. It is dominated by low stature (5–30 cm), drought-tolerant, warm-season grasses such as *Bouteloua gracilis* and *Buchloë dactyloides* (Archibald, 1995; Sims and Risser, 2000).

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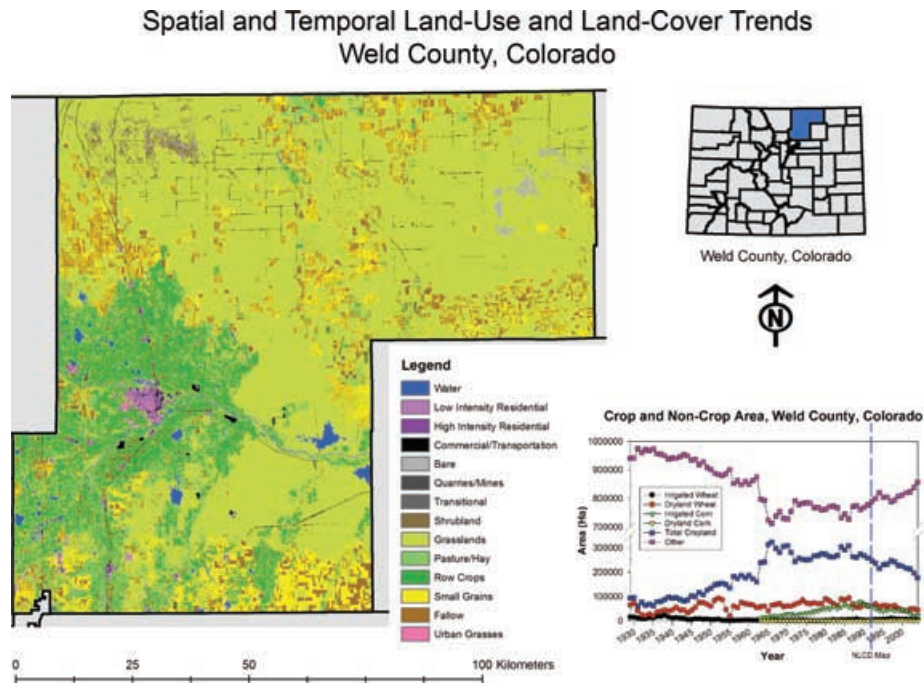


Fig. 1. National Land Cover Data (Vogelmann et al., 2001) illustrate the 1992 distribution of land-cover classes in Weld County, Colorado, a county within the short-grass steppe that has undergone extensive modification since the mid-19th Century. In general, areas closer to rivers are irrigated, while upland areas support dryland farming and short-grass steppe rangelands. The graph illustrates changes in agricultural area in Weld County from 1929 to 2004 due to precipitation changes, economic conditions, and implementation of irrigation (statistics from National Agricultural Statistics Service, <http://www.nass.usda.gov/>). With the exception of wheat, most crop statistics start in the late 1950s and early 1960s, hence the apparent jump in total cropland during 1963 (corn reported) and 1965 (hay included). The “other” class is cropland area subtracted from the county total, and it includes short-grass steppe, fallow, water, and residential classes.

Over the course of the next 150 years following settlement, somewhat less than half of the short-grass steppe was converted to agricultural land covers, principally dryland and irrigated croplands, to create a mosaic of native vegetation and croplands (Lauenroth and Milchunas, 1991; see Fig. 1 as a county scale example). Overall, vast areas of short-grass steppe were converted to dryland farming, where a wheat-fallow rotation is used for periodic soil moisture recharge. With the development and implementation of irrigation from the 1940s to 1980 (Chase et al., 1999; Parton et al., 2005), dryland farming area declined slightly, yields increased (Burke et al., 1994), and crops with higher water requirements (e.g. *Zea mays*) were planted. Further, ongoing technological advances (e.g. tillage practices, crop rotations, genetic characteristics) and changes in economic conditions and government programs (e.g. Conservation Reserve Program) produce chronic shifts in land cover and occasional conversion back to grasslands.

The conversion from native short-grass grassland to cropland is manifested in biophysical effects that influence energy and water cycling. Seasonality, albedo, leaf area index, surface roughness, and moisture fluxes were altered with conversion to cropland. Cropland and grassland albedos are similar during the crop growing season (Oke, 1987), but croplands have bare soil for much of the year and native vegetation has a higher albedo than

bare soil (Bonan, 2002). Dryland and irrigated crops are taller, and possess more leaf area than native short-grass steppe (Paruelo et al., 2001). Moreover, surface roughness is higher with the taller agricultural plants (Chase et al., 1999; Bonan, 2002). Lastly, moisture fluxes are higher in the agricultural systems, especially irrigated croplands (Baron et al., 1998; Chase et al., 1999; Stohlgren et al., 1998).

Growing season dynamics are changed with conversion to agriculture. With short-grass steppe featuring a mixture of cool- and warm-season grasses, photosynthesis occurs during the entire growing season and peak biomass occurs in early summer (Paruelo et al., 2001). In contrast, croplands have one dominant plant with dramatically different growing seasons and peak biomass, with dryland and irrigated crops peaking earlier and later than short-grass, respectively (Paruelo et al., 2001).

The effect on air temperature and precipitation produced with short-grass conversion varies with the conversion, spatially and temporally. The magnitude of change from short-grass steppe to irrigated agriculture is much more dramatic than the shift to dryland agriculture (Baron et al., 1998). At relatively fine scales during short periods of the growing season, lower temperatures and higher atmospheric moisture levels are closely tied with irrigated croplands (Segal et al., 1989; Chase et al., 1999). At larger scales, given the same brief temporal scale, this difference

produces a regional cooling effect and precipitation increase in the immediate lee of adjacent Rocky Mountains (Stohlgren et al., 1998). In a coarse resolution (50 km grid increment) regional modeling comparison of natural and current vegetation change over a much larger area of the Great Plain, Eastman et al. (2001) showed a 0 to 1°C increase in maximum temperatures over unconverted short-grass steppe, while a 2 to 3°C increase occurred in areas converted to dryland crops. Precipitation changes were heterogeneous. Irrigation has recently been shown in a global model to be an important climate forcing (e.g. Lobell et al., 2006).

Future work on the short-grass steppe must address scaling issues, future scenarios and linking altered energy and water dynamics to ecosystem functions. A gap in our knowledge exists with respect to spatially fine-scale resolution over longer time periods (e.g. seasonally) and detailed knowledge of annual precipitation patterns are lacking. Future scenarios of relevant economic and environmental forcings on land cover should also be considered. For example, how does the short-grass steppe change with reductions in available irrigation water? How would changes in tillage practices and crop rotations change weather impacts? Last, how do altered land-cover types and their associated physical changes impact ecosystem functions such as decomposition, nitrogen cycling and soil carbon (Epstein et al., 2002; McCulley et al., 2005)?

3. Tall grass conversion to dryland and irrigated agriculture

The tall grass conversion is similar to the short-grass conversion except the loss of aboveground biomass (leaf area index- LAI) is greater when the tall grass prairie was removed, and almost 100% of the original tall grass region is gone. Further studies on the role of grassland conversion include the initial evidence of the role of irrigation in modifying surface climate trends which came from observational studies (Marotz et al., 1975; Barnston and Schickendanz, 1984; Alpert and Mandel, 1986; Pielke and Zeng, 1989).

Barnston and Schickendanz (1984), for example, found that irrigation increased precipitation in the Texas Panhandle when the synoptic condition provided low-level convergence and uplift such that the additional moisture produced by irrigation was allowed to ascend to cloud base. These studies were followed by regional-scale climate model investigations of the effect of irrigation on various planetary boundary layer (PBL) properties (De Ridder and Gallée, 1998; Segal et al., 1998; Adegoke et al., 2003). Segal et al. (1998) used the fifth-generation Pennsylvania State University–National Center for Atmospheric Research (PSU–NCAR) Mesoscale Model (MM5) (Grell et al., 1993) in their study of irrigated areas in North America. Their model results suggest an increase in the continental average rainfall for the present irrigation conditions compared with those of past irrigation. De Ridder and Galleé (1998) used an European re-

gional numerical model (Modèle Atmosphérique Régional) and reported a reduction in the diurnal amplitude of temperature and wind speed when a semiarid surface is replaced by a partly irrigated one. The potential for moist convection also increased with surface moisture availability in their simulations. Lohar and Pal (1995) used the Regional Atmospheric Modeling System (RAMS) to show that irrigation can reduce the intensity of sea-breeze convection during the pre-summer monsoon season in eastern India and lead to the observed reduction in regional rainfall. The primary thermodynamic impact of irrigation is the repartitioning of the sensible and latent heat fluxes at the affected sites. Thus, an increase in irrigation or surface wetness reduces sensible heat flux while increasing physical evaporation and transpiration (Pielke, 2001). The resulting additional moisture flux can enhance the moist static energy within the convective boundary layer (CBL) and consequently become thermodynamically more conducive to an increase in rainfall (Betts et al., 1994; Segal et al., 1998).

In Nebraska, as in much of the U.S. High Plains, corn is the dominant crop cultivated during the warm season months (Williams and Murfield, 1977). Much of the corn area replaced the tall grass prairie. Irrigated corn, which represented about 10% of total corn-producing areas during the early 1950s, now comprises nearly 60% of the total corn-producing areas in Nebraska (National Agricultural Statistics Service, 1998). This rapid land-use change was achieved largely by converting rain-fed corn areas to irrigated areas.

To investigate the likely impacts of this agriculture-related land-use change on surface energy partitioning and summer climate, as reported in Adegoke et al. (2006) a modeling study consisting of four land-use scenarios over the 15-day period from 1–15 July 1997 was conducted for this region. The first scenario (control run) represented current farmland acreage under irrigation in Nebraska as estimated from 1997 LANDSAT satellite and ancillary data. The second and third scenarios (OGE wet and dry runs) represented the land-use conditions from the Olson Global Ecosystem (OGE) vegetation data set, and the fourth scenario (natural vegetation run) represented the potential (i.e., pre-European settlement) land cover from the Küchler vegetation data set. In the control and OGE wet run simulations, the topsoil of the areas under irrigation, up to a depth of 0.2 m, was saturated at 0000 universal time coordinate (UTC) each day for the duration of the experiment (1–15 July 1997). In both the OGE dry and natural runs, the soil was allowed to dry out, except when replenished naturally by rainfall. The ‘soil wetting’ procedure for the control and OGE wet runs was constructed to imitate the center-pivot irrigation scheduling under dry synoptic atmospheric conditions, as observed in Nebraska during the first half of July 2000 (i.e., when little or no rainfall was recorded throughout the state). The observed atmospheric conditions from NCEP re-analysis data (Kalnay et al., 1996) were used to create identical lateral boundary conditions in the four cases (see Adegoke et al., 2006, for additional details on the experimental design).

A key finding of this study is that midsummer 2 m temperature over Nebraska might be cooler by as much as 3.4°C under current conditions. The domain-average difference between the control and OGE dry runs computed for the 6–15 July 2000 period was 1.2°C. The cooling effect and the surface energy budget differences identified above intensified in magnitude when the control run results were compared to the potential natural vegetation scenario. For example, the near-ground domain-average temperature was 3.3°C cooler, the surface latent heat flux was 42% higher, and the water vapour flux (at 500 m) 38% greater in the control run compared to the natural landscape run. Important physical changes between the natural prairie of this region and the current land-use patterns include alterations in the surface albedo, roughness length, and soil moisture in the irrigated areas. These changes are capable of generating complex changes in the lower atmosphere (PBL) energy budget. For example, the simulated increase in the portion of the total available energy being partitioned into latent heat rather than sensible heat resulted directly from the enhanced transpiration and soil evaporation in the control run. Although not examined in detail in this study, elevated dewpoint temperature and moisture fluxes within the PBL can increase convectively available potential energy (CAPE), promote atmospheric instability, and enhance daytime cloud cover (Alapaty et al., 1997; Stohlgren et al., 1998; Pielke, 2001; Holt et al., 2006).

4. Mid-latitude deciduous forest conversion to agriculture

Similar to the grasslands, mid-latitude forests have a discernible impact on the regional weather patterns and climate. However, the forests are a significant terrestrial sink of global carbon. Forest covers have declined through the earlier 20th Century and there has been regrowth of newer forests (Fig. 2). This could be one reason that long-term observational impacts are not well doc-

umented for mid-latitude deciduous forests. Lambin et al. (2003) synthesized several cases to present the following pathway for the midlatitudinal landscape conversion as a complex function of: (i) population and resource availability stresses; (ii) opportunities due to market process, production and technology; (iii) policies related to subsidies, taxes, property rights, infrastructure and governance; (iv) vulnerability to external perturbations and coping capacity; and (vi) social organization related to resource access, income distribution, household features and urban-rural interactions which has been further highlighted in Pielke (2004) and Douglas et al. (2006).

Asner et al. (2004) document that agricultural activities associated with grazing operations is another important driver for the deciduous forest to be converted to grasslands and pastures particularly under poor soil conditions. This conversion is summarized to have hydrological impacts through a reduction in the rate of the spring snow melt, a reduction in the cloud condensation level and hence moisture availability, a reduction in the moisture interception due to a reduction in the LAI, and increased runoff and soil evaporation and decreased transpiration. This latter effects leads to higher fluxes of soil moisture and discharge over the landscape, which results in erosion and poor soil conditions over the region. Typically, the change in landscape from forest to grasslands leads to a significant reduction in moisture flux, while the change to cropland can have variable influences on the regional moisture flux (because of crop photosynthetic pathways and transpiration rates, cropping patterns, irrigation, etc.).

Pinty et al. (1989) showed that a nonlinear feedback between soil moisture availability and the forest cover can lead to higher regional precipitation. Anthes (1984), Pielke (2001) and National Research Council (2005) reviewed observational and modeling studies to conclude that landscape changes can modify large-scale atmospheric conditions, and result in increases of convective precipitation. The landscape change impacts are

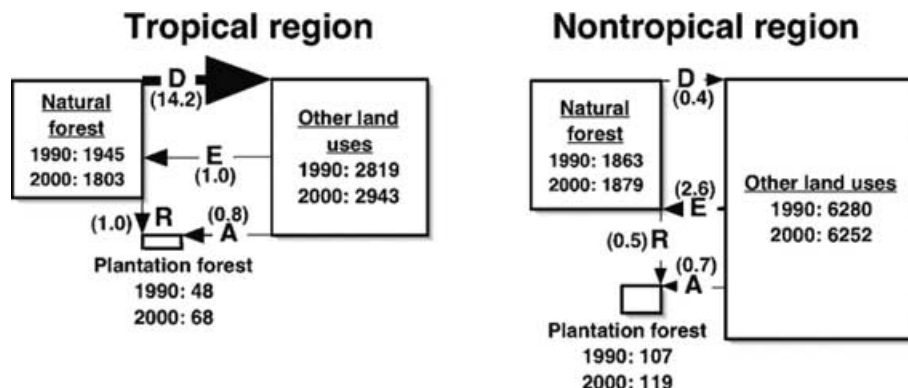


Fig. 2. Forest land-use change dynamics for tropical and nontropical regions. All data are averages for 1990–2000, and the units are million (10^6) hectares. Pools (squares) and fluxes (arrows) are drawn approximately to scale. Abbreviations for fluxes are D, deforestation; E, expansion of natural forests; R, reforestation; and A, afforestation (from Gower, 2003).

manifested through modified moist convection and low-level moisture and heat, changes in albedo, net radiation, and evaporation/transpiration leading to altered circulation and convection, and thus a modification of the regional water recycling. Otterman et al. (1990) provide observational evidence that suggests landscape changes have caused significant precipitation variability in southern Israel. The increase in precipitation is specifically attributable to an intensification of the convection and advection processes due to afforestation and increased cultivation-induced enhancement of the daytime sensible heat flux from the generally dry surface; the enhancement is from both the reduced surface albedo and the reduced soil heat flux, when insolation is strong. Greater daytime convection can lead to penetration of inversions capping the planetary boundary layer, while strengthened advection can furnish moist air leading to circulation changes, and higher precipitation.

Hogg et al. (2000) used field measurements over deciduous forest to show that the distinctive climate of interior Western Canada is the feedback associated with the leaf phenology of the aspen forest. Latent heat fluxes are largest under high LAI conditions leading to cooling but increased moisture availability and precipitation. A number of studies such as Pielke et al. (1991), Xue (1996), Chase et al. (1999), Bounoua et al. (2000), Zhao et al. (2001), Feddema et al. (2005), Niyogi and Xue (2006) have each asserted with different models and independent experiments (and ranging from global to landscape levels) that landscape changes impact surface radiative and biogeochemical fluxes, which in turn affect regional surface temperatures and precipitation either directly or as a feedback to changes in the regional circulation pattern.

For instance, results from Xue et al. (1996) indicate that the LAI changes of deciduous vegetation cause regional changes and propagate high uncertainty into general circulation model (GCM) simulations. Bounoua et al. (2000) concluded that for the midlatitudinal forest region, the resulting impact of LAI changes was a decreased albedo; cooling of about 1.8 K during the growing season and slight warming during winter due to snow albedo masking; and decreased effective precipitation and an increase in the low-frequency variability of weather in the northern latitudes.

Baidya Roy et al. (2002) simulated a 300 year (1700–1990) time series of U.S. land use/land cover using the Ecosystem Dynamics (ED) model constrained by available data. The RAMS model was used to simulate 3 case scenarios with 3 different surface vegetation distributions – 1700 (pristine), 1910 (maximum deforestation) and 1990 (current conditions). They found that changing the land-use/land-cover pattern can lead to several degrees of warming/cooling at the surface accompanied by significant changes in precipitation patterns.

Unlike the tropics, the impact of deforestation and midlatitudinal precipitation changes are difficult to estimate because of frontal systems and the variety of air mass source regions. Therefore, surrogate hydrological data fields need to be further

considered to assess the precipitation changes. Swank and Vose (1994) reviewed four decades of research on changes in water yield and timing of streamflow, following landscape changes from two mixed deciduous hardwood forests to plantations of eastern white pine. Within 10 years after the change, annual streamflow in watersheds was less than expected from mixed hardwoods due to greater transpiration from pines as a result of a higher LAI and corresponding higher interception loss in the dormant season and more transpiration loss in early spring and late fall. Flow duration analysis showed that the conversion to pine reduced the frequency of both high and low flows by 33 to 60 percent both due to precipitation changes and ensuing water loss due to transpiration/interception.

This summary is not exhaustive but yet provides significant clues that the midlatitudinal deciduous land-use changes have also resulted in regional precipitation changes. The complexity of detecting the signature within a naturally high frequency meteorological system and associated changes in the rainfall patterns, and the nonlinear changes in the forest covers (initial decline followed by an asynchronous regrowth in several regions), and the variable changes in the urbanization, pollution loadings and regional circulation patterns makes the quantitative assessment of this effect quite difficult. Therefore, detailed multi-scale modeling studies will continue to be the principal tools to develop attribution and detections associated with land-use and precipitation changes. It remains to be seen if satellite products can detect this change following techniques similar to those discussed in Shepherd (2005) and Cai and Kalnay (2005) using blended re-analysis data sets.

5. Tropical evergreen forest conversion to agriculture

Tropical forests, occupying approximately 800 million hectares, are being cleared at the rate of approximately 14 million hectare per year. Observational studies, spanning several decades, and numerical modeling studies both show that tropical deforestation influence cloud formation and rainfall (Sud and Smith, 1985; Meher-Homji, 1991; McGuffie et al., 1995; Costa and Foley, 2000; Cutrim et al., 1995; Pielke, 2001; Silva Dias et al., 2002; Lawton et al., 2001; Durieux et al., 2003; Sen et al., 2004; Fisch et al., 2004; Ray et al., 2006). While prior studies agree that deforestation alters cloudiness and rainfall, there is considerable disagreement on the magnitude and nature of the changes. These studies also show that the processes through which deforestation impacts rainfall, result from changes in the following characteristics: physical evaporation, transpiration, surface albedo, and aerodynamic roughness.

Observational studies (Meher-Homji, 1991; Pielke, 2001; Durieux et al., 2003; Ray et al., 2006) generally fall into two classes: 1. Comparisons of rainfall, cloudiness or satellite observed proxies to rainfall between adjacent forested and deforested sites; 2. Time series analyses of similar data over

one site that samples conditions before and after the site was deforested.

These studies report a wide range of changes in rainfall associated with deforestation (1–20% decrease), as well as the alteration of seasonality and frequency of convection. Difficulties encountered by the first type of studies are in accounting for topographic effects and natural spatial variability of rainfall. When using rain gauge data, both of these techniques have to account for differences related to the use of instrumentation in different settings, namely forested versus deforested, which could be significant (Meher-Homji, 1991). A potential strategy for developing reliable data sets for future studies would be to establish long-term rainfall monitoring sites that take into account the above discussed issues, especially in forest areas that face clearing in the near future.

The majority of the General Circulation Model (GCM) experiments that examine the effect of tropical deforestation on climate assume conversion of forests into pastures with higher albedos and lower surface roughnesses. Most of these studies find that deforestation leads to significant reduction in rainfall over the Amazon, while a few report an increase in rainfall (McGuffie et al., 1995; Costa and Foley, 2000). The impact of deforestation in southeast Asia has a seasonal dependency, with rainfall increasing during the wet season but decreasing in the dry season, while over Africa deforestation has minimal impact on rainfall according to one study (McGuffie et al., 1995). GCM simulations also indicate that deforestation has the potential to significantly modify monsoon rainfall over India (Sud and Smith, 1985), while regional-scale modeling results show that the east Asian summer monsoon is sensitive to deforestation in the Indo-China region (Sen et al., 2004).

One of the drawbacks of the GCM and coarse resolution regional models is the inability to resolve mesoscale circulation features induced by landscape heterogeneity. While heterogeneity-induced mesoscale circulations have the potential to modify cloudiness (Souza et al., 2000; Pielke, 2001; Silva Dias et al., 2002; Baidya Roy and Avissar, 2002; Werth and Avissar, 2002), an assessment of its impact on rainfall is lacking and requires well-designed, longer-term regional-scale modeling efforts.

The conversion of forest to pasture that is commonly assumed in several deforestation experiments is also unrealistic. Satellite imagery shows that the secondary growth in the deforested areas often restores the surface albedo to values close to those observed prior to deforestation in a relatively short period of time. Also, deforested areas often are replaced by a composite patchwork of varied land-use categories that include urban, agricultural crops, plantations, pasture, bare soil, and secondary growth and the aggregate effect of such patch work is not very well understood. Seasonal and diurnal observations of the surface energy budget over some the dominant categories of land-use that replace tropical forests are required for this purpose. Prior studies suggest significant water extraction by deep roots in

tropical forest. Nepstad et al. (1994) found that 75% of the water extracted by eastern Amazonian forests during the dry season originates from soil layer below the depth of 2 m. However, the rooting depths specified in regional and global atmospheric models specify rooting depths of approximately 2 m for tropical forests. Proper specification of root structures are needed to realistically simulate important hydrological processes such as precipitation recycling that can account for approximately 25%–30% of the rainfall in the Amazon region (Eltahir and Bras 1996).

There is some indication that deforestation in a continent surrounded by oceans has more potential to impact tropical circulation compared to deforestation that is further removed from ocean sources of water vapour (van der Molen et al., 2006), and this needs to be validated through the use of cloud-resolving and regional modeling experiments. Mesoscale numerical modeling experiments also show that lowland deforestation and associated increases (decreases) in the Bowen ratio leads to elevation (lowering) of the orographic clouds (Fig. 3) forced by terrain downwind (Lawton et al., 2001; van der Molen, 2002; Nair et al., 2003; Bruijnzeel, 2004; Ray et al., 2006), leading to changes in the direct harvesting of cloud water by montane vegetation. Since this “horizontal” precipitation can be a significant input to the local water budget, assessments of changes in horizontal precipitation resulting from deforestation are needed.

6. Boreal forest conversion due to fire

Fire in the boreal forest is an integral part of this ecosystem (Stocks, 1991). Fig. 4 from Vidale et al. (1997) illustrates the complex patchwork of different aged forest that occur due to the fire history in the area, as well as other disturbance such as disease, insect infestations and land management (Atlas of Canada, 2006). According to this Atlas, there are about 9000 forest fires recorded annually in Canada. An average of 2.1 million hectares are burned every year; virtually all of it is boreal forest. Lightning accounts for about 85% of all hectares burned annually, and people are responsible for the rest. The fires caused by people are more numerous, but burn a smaller area than those ignited by lightning. This fire frequency should be similar in other boreal forest areas.

As shown in Vidale et al., mesoscale circulations develop in response to the spatial variations of surface sensible heat fluxes that occur due to these fires and other landscape pattern disturbances, which also includes the ubiquitous lakes that exist in the boreal forests. Knowles (1993) reported that cumulus cloud form preferentially downwind of recent burn areas within a boreal forest landscape. These burn areas have a lower albedo than the surrounding landscape. Such a preference also indicates that subsequent lightning strikes from deep cumulus cloud will more likely initiate fires immediately downstream of a recent burn scar, rather than elsewhere in the forest. After a period of time, as the forest regenerates, the burned area may be lighter than

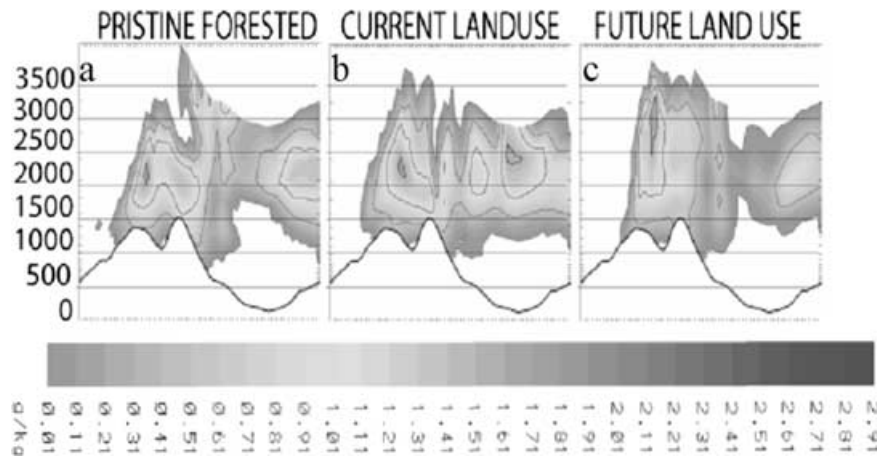


Fig. 3. Numerical simulation of orographic cloud formation for three lowland deforestation scenarios in Northern Costa Rica: a) pristine forests; b) current land-use; c) complete deforestation of the lowlands. East-West cross sections of cloud water mixing ratio, simulated by the Regional Atmospheric Modeling System, show orographic cloud bank form at lower elevation when there is forest cover in the lowland areas (Nair et al., 2006).

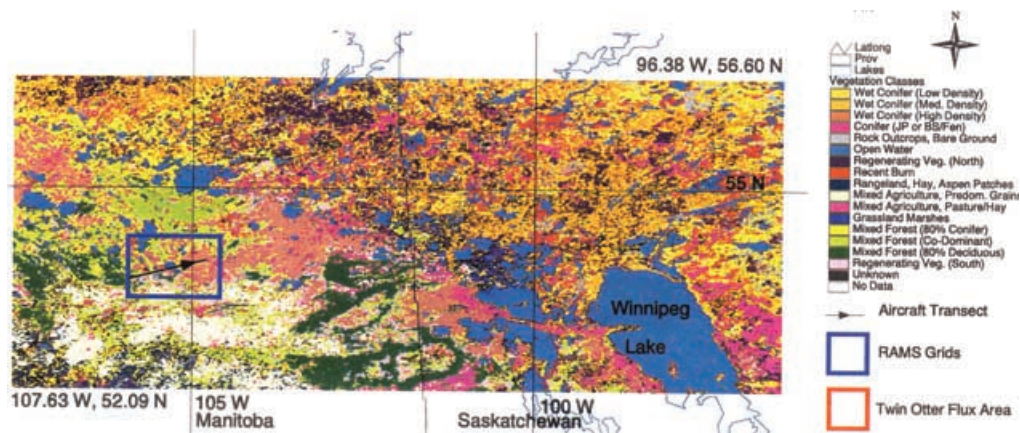


Fig. 4. AVHRR 1 km BOREAS regional land-cover classification. Dark lines and letters indicate latitudes and longitudes with an interval of 5 degrees. The map corresponds to RAMS grid 52.09° to 56.60°N and 107.63° to 96.38°W. From Vidale et al. (1997).

the surrounding unburnt landscape, as aspen and other second growth forest and shrubs grow.

The environment of the boreal forest, particularly in the spring, is very conducive to fires as the roots are embedded in cold, or even frozen soils, yet the air temperature and solar insolation at this time of the year is high. Almost all of the net radiation received at the surface is transferred back into the atmosphere as sensible, rather than latent heat flux (Sellers et al., 1995). Pielke and Vidale (1995) show that one consequence of the resultant large heating of the atmosphere by the sensible heating is a particularly deep planetary boundary layer, as well as a preference for the summer polar front to often situate along the boreal-tundra ecotone boundary. The region south of the polar front in the spring (and also in the summer) is a weather location where thunderstorms occur. The spatial variations of the landscape will provide focused regions for thunderstorm development that would otherwise not occur.

7. Urbanization

While the fraction of the Earth's surface currently classified as urban accounts for less than 2% of available land surface, over 45% of the Earth's population is concentrated there (Arnfield, 2003). Furthermore, future projections indicate this percentage will soon exceed more than 50% (Cohen, 2003). Observational studies over the past three decades have demonstrated that urban areas radically restructure the local energy budget and thus lead to different boundary layer structure (Oke, 1988; Arnfield, 2003; Shepherd, 2005). The anthropogenic influence also includes altering the aerosol environment. These changes likely lead to alterations in urban precipitation frequency, intensity, and patterns. Shepherd (2005) presents the most recent and complete review of urban precipitation issues.

Observational studies of urban precipitation stretches back over three decades (e.g. Huff and Changnon, 1973). Studies in

the U.S. have been especially concentrated in three urban areas, St. Louis, Atlanta, and Houston. These studies have generally demonstrated increases in rainfall over and downwind of urban areas (Shepherd and Burian, 2003; Dixon and Mote, 2003; Bornstein and Lin, 2000; Huff, 1986; Huff and Changnon, 1973) attributed largely to the Urban Heat Island (UHI) initiated convergence zone and to a lesser extent, to the increased surface roughness. Evidence also suggests an increase in heavy rain events (Huff, 1986). In addition, the UHI has been found to decrease the likelihood of freezing rain events in urban areas (Changnon, 2003). Bornstein and Lin (2000) also found urban influence on established thunderstorms approaching an urban area.

Laboratory and mesoscale modeling studies have demonstrated that the UHI impacts local mesoscale circulation patterns. These changes to the circulation should be expected to have an effect on precipitation as these circulation features are often accompanied by or provide the forcing mechanism for precipitation. Studies have demonstrated significant changes to the convective boundary layer structure (e.g. Hildebrand and Ackerman, 1984; Baik et al., 2001) and even influence the behavior of cold fronts (Gaffen and Bornstein, 1988). But the most completely studied influence is that of a UHI with a coastal sea breeze (Yoshikado, 1992; Kusaka, 2002; Ohashi and Kida, 2002, 2004; Cenedese and Monti, 2003; Martilli, 2003) where nearly all found a significant influence of the urban area on the development (timing and intensity) of the sea breeze front.

Most recently, urban parameterizations schemes (e.g. the Town Energy Balance (TEB) model; Masson, 2000; Lemonsu and Masson, 2002) have allowed mesoscale models to simulate the impacts of urban areas on mesoscale flow. This has permitted detailed sensitivity experiments (van den Heever 2005; Molders and Olson, 2004; Nobis, 2006) plus more sophisticated studies of actual events (Craig, 2002; Rozoff et al., 2003; Gero et al. 2006; Nobis 2005; Niyogi et al., 2006). Results from these studies have further confirmed the role of urban areas in precipitation modification and could be utilized in land-use planning and operational forecasting.

The biggest single unknown to the urban induced precipitation problem is the role of urban aerosols. Studies like van den Heever and Cotton (2005) have begun to examine the relative sensitivities involved, but as Shepherd (2005) points out, while evidence points strongly to a role for aerosols in urban precipitation modification, the details of that role remain highly uncertain.

8. Tropical forest fires and resultant biomass burning effect on rainfall

The role of tropical evergreen forest conversion to agriculture was discussed in Section 5. In this section, the specific focus is on the role of fires in the tropics, often ignited by direct human intervention, on rainfall. Deforestation in the tropics has been ac-

celerating over the past several decades (Skole and Tucker, 1993) and is often accomplished by biomass burning (Setzer et al., 1994) as a means for land clearing (Crutzen and Andreae, 1990). Biomass burning produces smoke plumes with large quantities of aerosols (tiny particles) which can potentially affect the regional, even global hydrological cycle (Ramanathan et al., 2001).

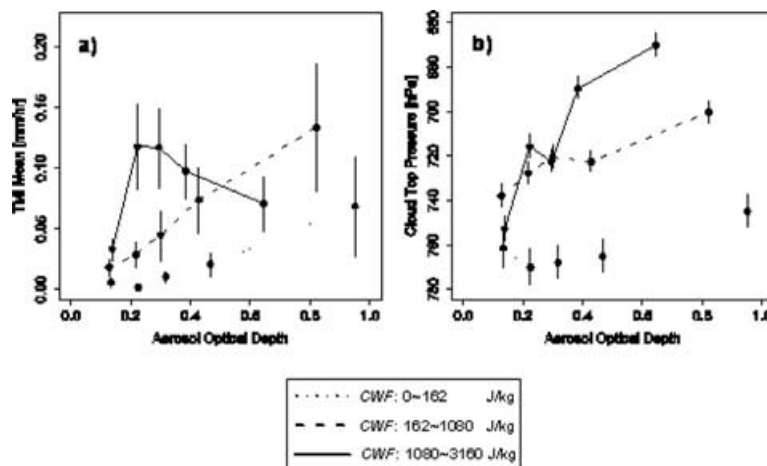
Aerosols serve as cloud condensation nuclei which affect the formation of cloud droplets (Cotton and Anthes, 1992); thus the extensive input of aerosols from fires could significantly affect cloud properties and rainfall. The first study to suggest this connection was based on significantly elevated concentrations of cloud droplets observed downwind of sugarcane fires (Warner and Twomey, 1967). Later, aerosols were hypothesized to reduce precipitation by increasing the number of small cloud droplets but suppressing their coalescence into larger, precipitation-sized drops (Albrecht, 1989).

In addition to affecting cloud microphysics, aerosols absorb and scatter radiation. Increased aerosols in the atmosphere from forest fires thus reduce the radiative energy reaching the Earth's surface (Ramanathan et al., 2001). Because surface radiative input drives evaporation, which is balanced with precipitation on a global scale, aerosol-induced reductions in surface radiation are likely to decrease global rainfall (Lohmann and Feichter, 2005).

The arrival of satellite-based observations of aerosols and rainfall in the 1990s first enabled monitoring over large spatial scales. Kaufman and Fraser (1997)—based upon satellite observations from the AVHRR sensor—have shown a reduction in the cloud droplet radius with aerosol loading from biomass burning in the Amazon, but did not quantify the effect on precipitation. Rosenfeld (1999), using satellite-based precipitation observations from the Tropical Rainfall Measuring Mission (TRMM), illustrated for a single day that aerosols from a biomass burning event in Indonesia shut down warm-rain processes.

While research concerning the impact of biomass burning on rainfall—such as Rosenfeld (1999)—have generally been limited to short time periods and have been focused on warm-rain processes, recent studies have revealed the crucial role of ice processes and dynamical feedbacks that go beyond the previous simple microphysical and radiative considerations. For instance, Andreae et al. (2004) analyzed Amazonian aircraft observations, and found delays in the onset of precipitation within smoky clouds. However, the lack of early warm-rain processes enabled updrafts to accelerate and reach higher altitudes and produce stronger storms. Thus the authors concluded that the net effect of aerosols on total precipitation “remains unknown.” A recent modeling study incorporating state-of-the-art spectral microphysics (Khain et al., 2005) has shown that while elevated aerosol concentrations initially lead to lowered precipitation efficiency due to warm-rain suppression, the delay in raindrop formation decreases the drag on updrafts by falling raindrops and increases the latent heat release by the additional water that reaches higher altitudes, where freezing takes place (Khain et al., 2005).

Fig. 5. (a) Relationships of rainfall measured by the TRMM-TMI sensor with MODIS aerosol optical depth (τ_a), for the year 2003. The data are binned by τ_a , with each bin spanning 20%tile of the τ_a values and further stratified into different *CWF* regimes. The dotted, dashed, and solid lines range from the lowest third to the top third values of *CWF*, respectively. The error bars denote the standard errors (σ/\sqrt{N}) of the bin-average. (b) The relationship between MODIS cloud top pressure and τ_a . Note the y-axis has been inverted to indicate the fact that higher cloud is associated with lower cloud top pressure.



Lin et al. (2006), based on satellite observations during the entire Amazonian biomass burning season, empirically confirmed the aforementioned results of Andreae et al. (2004) and Khain et al. (2005). Increased aerosols from fires were correlated with increased observed total (warm+ice-phase) rainfall from the TRMM-TMI sensor (Fig. 5a), even after accounting for the atmospheric stability environment through the cloud work function (*CWF*). Changes in cloud properties were also correlated with aerosol loading; i.e. higher cloud tops (Fig. 5b), increased presence of ice, and enhanced cloud cover.

Thus aerosols from biomass burning have been demonstrated to have complicated effects on rainfall. This topic will likely continue to attract significant scientific interest in the coming years (National Research Council, 2005), especially as tropical precipitation provides three-fourths of the energy that drives the atmospheric wind circulation through latent heat release (Kummerow et al., 2000); thus perturbations to tropical rainfall from biomass burning can have potentially critical climate consequences. Such alterations to the hydrological cycle can also have major effects for the biogeochemical cycles (Niyogi et al., 2004), and the fresh water supply (Ramanathan et al., 2001; Lohmann and Feichter, 2005).

9. Afforestation and reforestation

Afforestation and reforestation (A&R) are proposed as possible tools to mitigate desertification (FAO, 1989) and to reduce atmospheric concentrations of CO_2 by sequestering carbon in forest biomass (UNDP, 2003). Afforestation refers to locations that did not have natural forest cover, while reforestation replaces a forest that was removed. In southern South America, A&R plans started in Chile, Uruguay and Argentina in the last two decades, supported through government economic incentives or subsidies (World Bank, 2000). In southwestern Australia, A&R are also seen as ways to ameliorate salinity (Walker et al., 2002).

Overall, conversion from grasslands or croplands to forest leads to a decrease in albedo and increases of LAI, roughness

length and rooting depth (Sellers, 1992; Jackson et al., 1996; Pitman, 2003). Changes in these parameters can modify the near-surface energy fluxes, which can influence temperature and humidity (Pielke, 2001). In general, observations and modeling studies agree that A&R would decrease near-surface temperature and increase latent heat (e.g. Fahey and Jackson, 1997; Xue and Shukla, 1996; Nosetto et al., 2005). Simulated impacts on precipitation are not so clear, and they depend on geographical location, regional atmospheric characteristics, extent of the afforested-reforested area (Xue and Shukla, 1996; Pitman and Narisma, 2005) and biophysical parameters involved in the land-use/land-cover change (Xue et al., 1996).

Regional and global modeling studies have usually addressed deforestation effects on precipitation (e.g. Xue and Shukla, 1993; Chase et al., 1996; Kanae et al., 2001; Baidya Roy and Avisar, 2002; Narisma and Pitman, 2003; Oyama and Nobre, 2004; Avissar and Werth, 2005, and other references in this paper). Therefore, the effects of A&R could be inferred from those studies. For instance, land-cover changes, from trees to grass or crops over southwestern Australia, from the mid 1700's to present, could partially explain the observed decreases in winter precipitation (Pitman et al., 2004). Thus, large-scale reforestation could increase rainfall in the long-term. However, Xue and Shukla (1996) found that the effects of afforestation-deforestation over the Sahel on precipitation were not linear, but depended on the location of the perturbed area with respect to the position of large-scale circulation features (i.e., subsidence branch of the Hadley circulation).

In general, modeling exercises show rainfall increases in an A&R scenario with respect to a current land cover. Those changes can be attributed to changes in moisture convergence (due to changes in roughness length and displacement height) and latent heat (Xue and Shukla, 1996; Pitman et al., 2004; Beltrán, 2005). Using a global circulation model (COLA-GCM), Xue and Shukla (1996) found increases of 0.8 mm day^{-1} (or 27%) on the Sahel precipitation over the afforested area, and decreases south of it. Using a regional coupled

atmospheric-biospheric model, GEMRAMS, over the central Pampas in southern South America, Beltrán (2005) found that afforestation led to increases of 1 mm day^{-1} on average in simulated summer precipitation (i.e. December–January). In both studies, the impact of afforestation on precipitation was relatively high for a “dry year”. In a recent study, Jackson et al., (2005), using a U.S. projected afforestation scenario based on the response to payments for carbon sequestration and a regional climate model (RAMS), found that changes in summer precipitation were not noticeable, and depended on site location. A general decrease in rainfall was found in afforested areas located in the northern states. Precipitation increased in a few areas such as in Florida and southern Georgia and in other grid cells not directly affected by the land-cover change. In this case, the shift of available energy from sensible to latent heat in the afforestation experiment reduced the convective precipitation in the temperate regions.

Observational (Jackson et al., 2005) and several modeling studies (i.e. Xue et al., 1996) have shown that tree plantation establishment may affect the hydrological cycle. Precipitation processes depend on local, regional and large-scale atmospheric characteristics, and therefore regional atmospheric modelling represent an important tool to study the impacts of realistic patterns of A&R on precipitation.

10. Conclusions

This paper documents the diverse role of land-use/land-cover change on precipitation. Since land conversion continues at a rapid pace (e.g. see Table 1 in Pielke et al., 2006), this type of human disturbance of the climate system will continue and become even more significant in the coming decades. The regional alteration of landscape also has global climate effects through teleconnections as concluded in National Research Council (2005); a conclusion which is bolstered by studies such as that of Chase et al. (2000) and Fedemma et al. (2005).

The National Research Council (NRC) had the following conclusion and recommendations in recognition of the important role of land-use/land-cover change on the global climate, including the alteration of precipitation processes:

‘Regional variations in radiative forcing may have important regional and global climatic implications that are not resolved by the concept of global mean radiative forcing. Tropospheric aerosols and landscape changes have particularly heterogeneous forcings. To date, there have been only limited studies of regional radiative forcing and response. Indeed, it is not clear how best to diagnose a regional forcing and response in the observational record; regional forcings can lead to global climate responses, while global forcings can be associated with regional climate responses. Regional diabatic heating can also cause atmospheric teleconnections that influence regional climate thousands of kilometers away from the point of forcing. Improving societally relevant projections of regional climate impacts will require a better understand-

ing of the magnitudes of regional forcings and the associated climate responses.

The NRC Report recommended that we should

Use climate records to investigate relationships between regional radiative forcing (e.g. land-use or aerosol changes) and climate response in the same region, other regions and globally.

Quantify and compare climate responses from regional radiative forcings in different climate models and on different timescales (e.g. seasonal, interannual), and report results in climate change assessments.’

The report also concluded

“Several types of forcings—most notably aerosols, land-use and land-cover change, and modifications to biogeochemistry—impact the climate system in nonradiative ways, in particular by modifying the hydrological cycle and vegetation dynamics. Aerosols exert a forcing on the hydrological cycle by modifying cloud condensation nuclei, ice nuclei, precipitation efficiency, and the ratio between solar direct and diffuse radiation received. Other nonradiative forcings modify the biological components of the climate system by changing the fluxes of trace gases and heat between vegetation, soils, and the atmosphere and by modifying the amount and types of vegetation. No metrics for quantifying such nonradiative forcings have been accepted. Nonradiative forcings have eventual radiative impacts, so one option would be to quantify these radiative impacts. However, this approach may not convey appropriately the impacts of nonradiative forcings on societally relevant climate variables such as precipitation or ecosystem function. Any new metrics must also be able to characterize the regional structure in nonradiative forcing and climate response.

Improve understanding and parameterizations of aerosol-cloud thermodynamic interactions and land-atmosphere interactions in climate models in order to quantify the impacts of these nonradiative forcings on both regional and global scales.

Develop improved land-use and land-cover classifications at high resolution for the past and present, as well as scenarios for the future.

Our summary of research on land-cover/land-use change on rainfall also results in the following recommendations:

- Future work for each biome must address scaling issues (from local to regional to global), future land use and land cover scenarios, and the resultant altered energy and water dynamics as they effect ecosystem function;

- Tropical and higher latitude deforestation appears to result in different responses to precipitation. This could be due to the role of cold and warm frontal (i.e. baroclinic) dynamics in the higher latitudes in precipitation processes even in the summer; . . .

- More research is needed to better understand the role of landscape patterning on precipitation. Such studies need to discriminate topographic effects and natural spatial variability of rainfall, from changes due to human land-use/land-cover change.

- Long-term rainfall monitoring sites in forest and grassland areas that face future land-cover/land-use change (e.g. such as deforestation or expansion of irrigation in grasslands) should be a priority. This will provide an observational basis to document the actual role of the landscape conversion on precipitation.

- A major under-monitored climate variable is rooting depth of the vegetation. We need accurate rooting depth specification in the models.

– All of the vegetation and soil information needs to be monitored and made available using the same parameters as required in the models. For example, rather than specifying that a soil is sandy loam, we need the soil density, conductivity, heat capacity, etc, expressed in quantitative units.

– Since model domain size, grid spacing and parameterizations (e.g. convective schemes) significantly influence simulated precipitation (e.g. Castro et al., 2005), more comprehensive regional modeling sensitivity studies are needed to assess those impacts.

Since it is now recognized that land-use change influences precipitation on the regional and the global scale, this important climate forcing should be elevated as a research priority within the climate community.

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