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Evaluating the seasonal and interannual variations in water balance in northern Wisconsin using a land surface model

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[1] We evaluated the performance of the Integrated BIosphere Simulator (IBIS) land surface model in the temperate forests of northern Wisconsin (46°N, 89°W) to determine whether model formulations, driven with daily historical precipitation, temperature, relative humidity, solar radiation, and wind speed data, were capable of simulating water flow and storage within a seasonally cold climate regime. We focused concurrently on understanding seasonal and interannual variations of both the water fluxes to the atmosphere and water partitioned into surface runoff and groundwater infiltration, with special attention to the transitions from cold-dominated (snow, ice) to warm-dominated (rain, liquid soil moisture) hydrology. Results showed when compared with a suite of field observations IBIS simulated water and energy cycling at daily to interannual timescales with reasonable accuracy. Because of errors associated with field observations, the accuracy with which we simulated each component of the water balance is not easily quantified. By investigating the complete land surface water balance, however, we increased the likelihood that all components were being captured. The modeled monthly energy balance, annual water balance, and drainage rates were generally within 5–15% of the observed values. Modeled and observed soil temperatures generally differed by less than 3°C and had r^2 values that were greater than 0.9. Soil moisture values were within 5–20%, and freeze and thaw timing was within a few days of observations. Modeled snow dynamics captured the observed snow arrival and departure (accumulation on the surface) within a few days of observations, but overestimated the average maximum depth by 86%. Because model formulations were subjected to varying soil conditions and water phases, this evaluation exercise enhanced our understanding of northern Wisconsin's water balance and increased model credibility for applications in seasonally cold climates.

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1. Introduction

[2] Ecosystem processes depend on water. Consequently, our understanding of biogeochemical and ecological processes is limited by our ability to track the flow and storage of water within ecosystems. To describe hydrological processes occurring within terrestrial ecosystems, at the scale of an entire landscape, continuously through time, often requires knowledge beyond what can be characterized by research at a single site [*Turner et al.*, 2001; *Greenland et al.*, 2003; *Rastetter et al.*, 2003; *Wisconsin Academy of Science, Arts and Letters*, 2003]. Coordinated observations of water stocks and flows are needed across space and time.

[3] Process-based models provide a means to examine hydrological and ecological processes across broad spatial and temporal scales [*Coe*, 1998; *Turner et al.*, 2001; *Hobbie*, 2003; *Rastetter et al.*, 2003]. These models must, however, be thoroughly tested and evaluated against a suite of field observations that adequately represents the region and timeframe of interest [*Levis et al.*, 1996; *Gardner and Urban*, 2003; *Kucharik et al.*, 2000; *DeAngelis and Mooij*, 2003; *Kucharik et al.*, 2006].

[4] In the last 2 decades, land surface models have been used to examine the hydrological processes occurring within ecosystems, ranging from tropical to boreal forests, and from desert to grassland [*Foley et al.*, 1996; *Delire and Foley*, 1999; *Pitman et al.*, 1999; *Bowling et al.*, 2003]. Generally speaking, these models have succeeded in describing the gross features of the terrestrial water balance, including patterns of evapotranspiration, runoff, and soil moisture changes. However, recent studies indicate that these models may face particular challenges in cold climates.

[5] The ability of land surface models to simulate the behavior of regional hydrological systems in cold climates has become an international research priority. Recent studies include those coordinated under the Project for Intercomparison of Land Surface Parameterization (PILPS)

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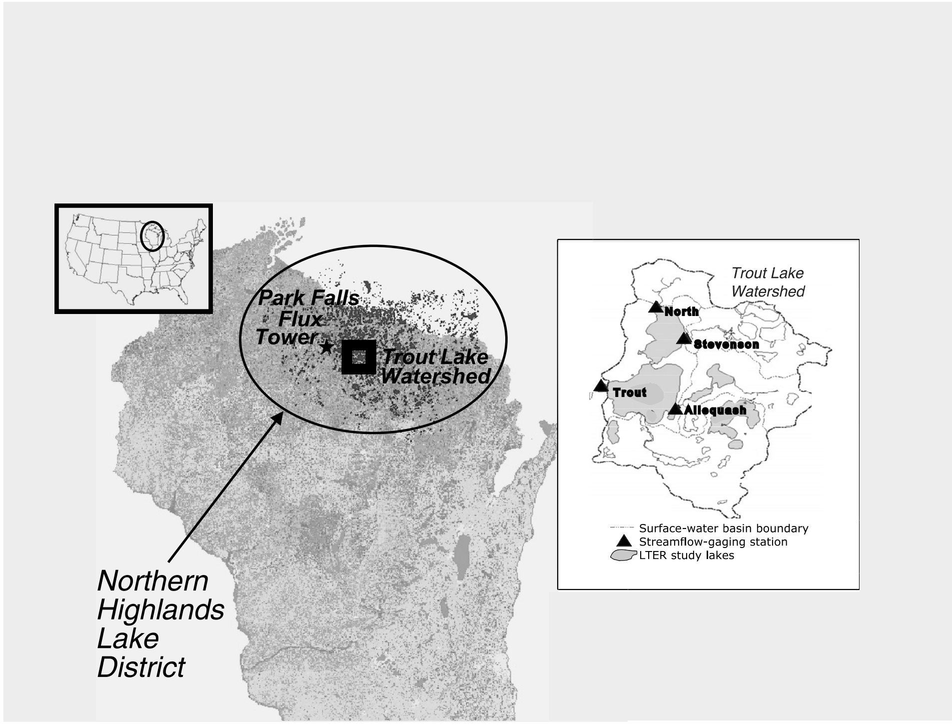

Figure 1. Northern Highlands Lake District. The Lake District is approximately 3500 km^2. It contains the flux tower site at Parkfalls and the Trout Lake watershed (12,000 ha). North Creek basin is the smallest (half the size of Stevensons, 4 times smaller than Allequash, and 13 times smaller than the entire Trout Lake watershed). The watershed map is modified from http://infotrek.er.usgs.gov/doc/webb/icons/Trout.Lake.map.gif.

Phases 2(d) [*Luo et al.*, 2003], PILPS 2(e) [*Nijssen et al.*, 2003], the Rhône-Aggregation Land Surface Scheme Intercomparison Project [*Boone et al.*, 2004], and Boreal Ecosystem-Atmosphere Study (BOREAS) [*Sellers et al.*, 1997]. These projects varied in location, spatial extent, and vegetation cover, but all found that snow properties, soil moisture, and soil temperature were important to understanding seasonal hydrology, and stressed the complexity and importance of the spring snowmelt period [*Betts et al.*, 2001; *Boone et al.*, 2004; *Luo et al.*, 2003; *Nijssen et al.*, 2003]. Earlier versions of IBIS were used in the PILPS 2(e) [*Nijssen et al.*, 2003] and BOREAS [*El Maayar et al.*, 2001]. These studies showed that IBIS captured broad ecosystem dynamics but indicated the need for further investigation of seasonal cycles in cold, snowy climates.

[6] Here we examine the behavior of a land surface model across a drainage basin in the Upper Midwest region of the United States. Our study focuses on the temperate forests of northern Wisconsin (46°N, 89°W). The study region, commonly known as the Northern Highlands Lake District, covers approximately 3500 km^2 of low relief, glacial outwash terrain [*Riera et al.*, 2000] approximately 50 km south of Lake Superior (Figure 1). The region is roughly covered by ~65% upland forest, ~10% lakes, and ~25% wetlands [*WISCLAND*, 1992]. The region's temperature ranges from summer maxima of 32°C to winter minima of −34°C with winter air temperatures typically being below freezing for at least 3 months of the year. Annual snowfall accumulation can be up to ~240 cm/yr, with snow covering the ground for ~140 days or more (WSCO, Wisconsin State Climate Office, Climate of Wisconsin, 2005, adapted from Climatography of the United States Number 60, NOAA, available online at http://www.aos.wisc.edu/~sco/stateclimate.html). The region's temperate climate, with high summer rainfall and high winter snowfall, provides ideal conditions to study the effects of snow, ice, and changing soil conditions on seasonal hydrology.

[7] Northern Wisconsin has been the location of numerous long-term research projects on both limnology and forestry (Chequamegon Ecosystem Atmosphere Study, 2005, available online at http://cheas.psu.edu; NTL-LTER, North Temperate Lakes Long-Term Ecological Research, 2005, University of Wisconsin, Madison, available online at: http://lter.limnology.wisc.edu; and NTL-WEBB, North Temperate Lakes Water, Energy, and Biogeochemical Budgets, 2005, United States Geological Survey, Madison, Wisconsin, available online at: http://infotrek.er.usgs.gov/doc/webb) (hereinafter referred to as ChEAS, 2005, NTL-LTER, 2005, and NTL-WEBB, 2005, respectively). These projects have investigated different components of the hydrologic cycle (and associated flows of energy and material) that, in combination, capture nearly all components of the surface water balance.

[8] Previous investigations with land surface models in northern Wisconsin have largely focused on turbulent energy fluxes from the land surface to the atmosphere, namely evapotranspiration and sensible heat, to evaluate

simulations of the surface energy and water balance. A well-known eddy-covariance flux tower near Park Falls, Wisconsin, has been the main source of model validation data [*MacKay et al.*, 2002; *Davis et al.*, 2003]. Studies include: *Denning et al.* [2003], who coupled a biosphere model (SiB-2) to a regional atmospheric model to investigate diurnal cycles of energy and atmospheric CO_2 fluxes; *Baker et al.* [2003], who also used SiB to simulate latent and sensible heat fluxes and net ecosystem exchange seasonality between 1997 and 1999. These studies, however, only account for short-term variations in water, energy, and CO_2 exchange between the forest and atmosphere, which are studied primarily in the summer when variations in energy fluxes are the greatest. Long-term variations in hydrological stocks and flows have not been considered by the previous studies and they have not examined water that runs off the surface or infiltrates into groundwater.

[9] Other hydrological research in northern Wisconsin has focused on water flow through lakes and groundwater. For example, *Lenters et al.* [2005] used an energy balance approach to estimate evaporation of a representative inland lake. *Dripps* [2003] examined land-surface and groundwater interactions, accounting for groundwater recharge rates using both field measurement and modeling techniques. Also, numerous studies have focused on the interactions between groundwater and lakes [*Hurley et al.*, 1985; *Krabbenhoft et al.*, 1990; *Anderson and Cheng*, 1993; *Hunt et al.*, 2003; *Pint*, 2002]. These studies carefully investigated groundwater recharge and flow, but did not focus on evapotranspiration, surface runoff, and other land surface energy and water fluxes.

[10] To evaluate seasonal and interannual changes in hydrology across northern Wisconsin, with the degree of detail appropriate for ecological and natural resource applications, we must address two challenges.

[11] First, we must concurrently simulate the components of the surface water balance. For a more complete understanding of the water balance we must simultaneously address land-atmosphere (evapotranspiration), land-stream (surface runoff) and land-groundwater (drainage) water movements. The amount and timing of these atmospheric fluxes and land surface water flows are all important aspects of the region's hydrologic cycle and are absolutely crucial to surrounding ecosystem processes such as lake chemistry and biology [*Hagerthey and Kerfoot*, 1998; *Webster et al.*, 1990; *Hurley et al.*, 1985], flooding associated with spring snowmelt [*Doesken and Judson*, 1997], and vegetation-atmosphere carbon exchange [*Davis et al.*, 2003; *Werner et al.*, 2003].

[12] Second, we must represent the entire seasonal cycle. Understanding the seasonal transitions from cold-dominated hydrology (snow, ice) to warm-dominated hydrology (rain, liquid soil moisture) requires special attention to the complexity of the seasonal soil freeze and thaw and the snow mass and energy balance. Because of soil freezing and snow accumulation, the hydrological cycle undergoes a massive transformation (phase change) that creates a delay in the release of wintertime precipitation, thereby creating a sizeable springtime pulse of water from snowmelt. Additionally, the fate of snowmelt, whether it runs off the surface or infiltrates into soil and groundwater, depends on the soil conditions. Understanding the magnitude and timing of the springtime pulse, which relies on a complex array of biophysical factors, requires careful investigation into multiple hydrological and biophysical processes. This is particularly important because the seasonality of the water cycle is also likely to change in response to both land-use change [*Twine et al.*, 2004] and future climate change [*Kling et al.*, 2003].

[13] To address these challenges and increase our understanding of the full complexity of land surface hydrological processes, we evaluate our land surface modeling results against a diverse range of available field observations related to water balance. These include observations of evapotranspiration (FLUXNET, Gap-Filled Flux Products Compilation, 2005, available online at http://www.fluxnet.ornl.gov/fluxnet/gapzips.cfm#Anchor4) (hereinafter referred to as FLUXNET, 2005), streamflow (NTL-WEBB, 2005), groundwater recharge [*Dripps*, 2003], soil temperature (NTL-LTER, 2005; ChEAS, 2005), soil moisture (ChEAS, 2005), and snow properties (NCDC COOP, National Climate Data Center Cooperative Observer Program, National Water Service, 2005, available online at http://www.weather.gov/om/coop) (hereinafter referred to as NCDC COOP, 2005) at sites throughout the temperate mixed forest of northern Wisconsin.

2. Model Overview: IBIS Land Surface Model

[14] In this study we use the Integrated BIosphere Simulator (IBIS) land surface model, which has been used to investigate biophysical, ecological, and hydrological processes on local, regional, and global scales [*Foley et al.*, 1996; *Kucharik et al.*, 2000]. In IBIS, physical and ecological processes are represented in a hierarchy of hourly, monthly, and yearly time steps: a framework that allows for an integrated, physically consistent investigation into both atmospheric fluxes and water partitioning into surface runoff and groundwater infiltration [*Foley et al.*, 1996].

[15] In previous studies relevant to this region, IBIS has been shown to accurately simulate ecosystem processes across the whole Mississippi River basin [*Lenters et al.*, 2000; *Donner et al.*, 2004; *Twine et al.*, 2004], across a small watershed in northern Wisconsin [*Dripps*, 2003], and across agricultural fields in southern Wisconsin [*Kucharik et al.*, 2001; *Kucharik and Brye*, 2003]. On a regional scale, IBIS was shown to be a useful tool to investigate hydrologic processes important to resource management issues. For example, applying IBIS to the Mississippi River basin demonstrated the importance of land cover to seasonal changes in water balance [*Twine et al.*, 2004] and how climatic variability and agricultural practices impact the basin's hydrology and nutrient export [*Donner et al.*, 2004].

[16] For this study, we use an updated version of IBIS that contains improved formulations for soil infiltration [*Li et al.*, 2005] and an updated algorithm for simulating natural vegetation phenology, which accounts for growing degree-days, precipitation accumulation, moisture stress within the vegetation, and root-zone temperature [*Twine et al.*, 2004].

[17] Within IBIS, the vegetation cover is represented by an upper plant canopy (trees) and a lower plant canopy (shrubs and grasses), which are prescribed according to individual plant functional type characteristics such as biomass, leaf area index, and growing degree-day require-

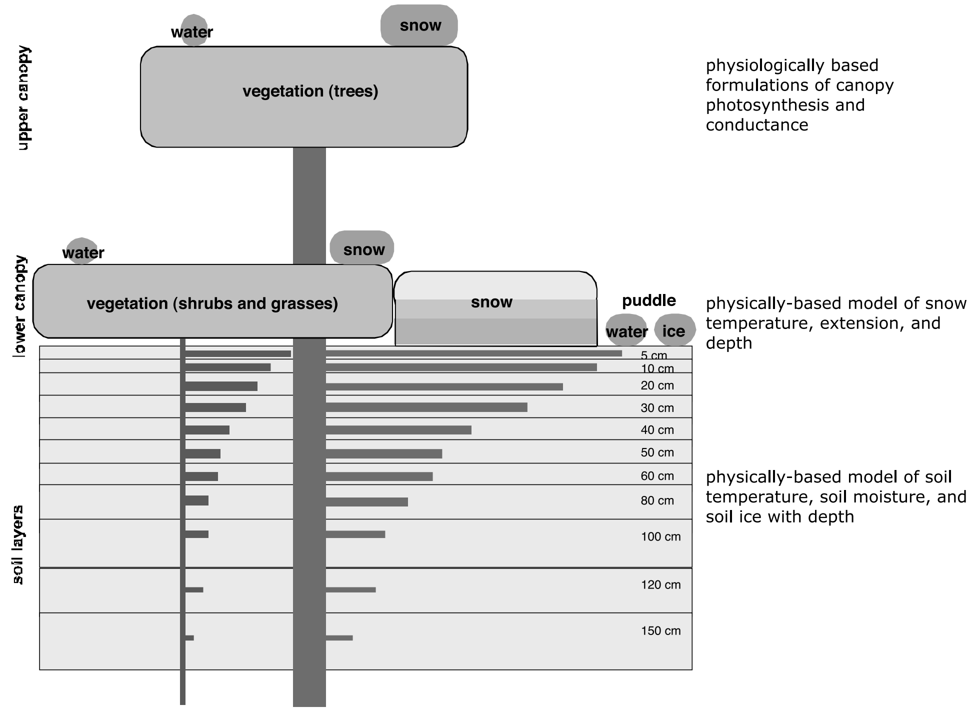

Figure 2. Integrated BIosphere Simulator (IBIS), figure adapted from *Kucharik et al.* [2000].

ments [*Foley et al.*, 1996; *Kucharik et al.*, 2000]. The region's current vegetation is over 65% temperate forest, comprising both deciduous and coniferous plant types (ChEAS, 2005) [*WISCLAND*, 1992]. Therefore, for this study, we designate the vegetation to contain plant functional type characteristics of a temperate mixed forest.

[18] Soil and snow are represented with multiple layers. The soil layers are characterized by diverse textural classes that vary as a function of depth. Within the model, we assigned eleven soil layers to a total of 1.50 m of soil depth (Figure 2). Each layer is represented with unique temperature, volumetric water content, and ice content. For this study, the first 0.05 m of soil was designated as organic, with properties outlined by *El Maayar et al.* [2001]. The soil type for depths between 0.05 to 1.5 m was designated as sandy loam, containing the physical and hydrological characteristics including sand, silt, and clay fractions, wilting point, and field capacity outlined by *Campbell and Norman* [1998]. This soil profile is representative of the soils within the study region (ChEAS, 2005) (STATSGO, State Soil Geographic Database, http://www.ncgc.nrcs.usda.gov/products/datasets/statsgo/, 1994, U.S. Department of Agriculture, Natural Resources Conservation Service, Fort Worth, Texas). The model simulates snow dynamics using a three-layer model that explicitly represents mass balance and snow thermodynamics (Figure 2) [*Thompson and Pollard*, 1995; *Foley et al.*, 1996]. In this study, snow density was designated as 0.075 m^3 water/m^3 air based on regional snowfall density (NCDC COOP, 2005).

[19] We run IBIS using daily input derived from a combination of sources (details in section 4.2). The model is run from 1948 to 2000. The simulated results are evaluated for the same time period as observations (1989–2000).

3. Four Model Evaluation Criteria

[20] To evaluate the ability of the IBIS land surface model to simulate the hydrological balance of a region with prolonged periods of below freezing temperatures, we set forth and examine four criteria.

[21] 1. Does the model simulate the energy balance of the landscape? For example, accurately simulating radiative, latent and sensible heat fluxes is critical to knowing whether the amount of water released to the atmosphere is reasonable. To determine this, we compare IBIS latent and sensible heat fluxes to those observed at an eddy-covariance flux tower.

[22] 2. Is the annual average water balance and interannual variability reasonably simulated? Water that is not evapotranspired eventually leaves the region via stream or groundwater flow. Therefore, to ensure a closed water budget, the total water leaving the terrestrial landscape in the model should be comparable with observed water flows. To evaluate this, we calculate surface runoff and groundwater infiltration values to compare with stream gauge observations.

[23] 3. Is the partitioning between surface runoff and soil infiltration to groundwater realistic? Whether water infiltrates into the soil or runs directly off on the surface strongly influences the character of aquatic ecosystems, lakes, and rivers. Furthermore, knowing the route the water travels influences water chemistry and quality and is important to many resource management applications. We compare the partitioning between surface runoff and soil infiltration simulated by IBIS with groundwater recharge observations.

[24] 4. Does the model capture the seasonal timing of water flows? The seasonality of the climate influences water availability, particularly in the springtime when snowmelt creates a significant pulse of water to groundwater and streamflow. Understanding the timing and magnitude of this pulse is key to understanding the region's overall hydrologic functioning. To evaluate the seasonality of changing water flows, we compare multiple data sources including soil temperature, soil moisture, and snow depth.

[25] Addressing these criteria requires understanding the timing and magnitude of water flows like evapotranspiration, surface runoff, soil infiltration, and groundwater recharge throughout the entire year. In the four "Model Evaluation Criterion" sections to follow, we address each of these evaluation questions by comparing IBIS simulations with field observations.

4. Data Sets

[26] Northern Wisconsin's landscapes and waterscapes have been studied extensively, and the region has a variety of ongoing projects. The diversity of scientific questions motivating these studies has created a wealth of information that can be used to provide input and independently validate land surface models.

4.1. Field Data for Model Evaluation

[27] Field studies provide the opportunity to evaluate model processes and parameterizations at varied times and locations. From four research projects, we have field observations of evapotranspiration, evaporation, groundwater recharge, soil and snow water storage, and streamflow with which to address whether IBIS can satisfactorily meet the aforementioned model evaluation criteria for cold climates.

4.1.1. ChEAS Project

[28] The Chequamegon Ecosystem Atmosphere Study (ChEAS) (http://cheas.psu.edu/) is a multi-organizational research effort started in 1997 to study biosphere-atmosphere interactions in the Chequamegon National Forest, near Park Falls, Wisconsin. The site, also involved in AmeriFlux and FLUXNET, uses eddy-covariance fluxes collected on a 447-m tower (45.95°N, 90.27°W) to measure CO_2, water vapor, trace gases, and energy exchange [*Werner et al.*, 2003]. Micrometeorological sensors mounted on a local TV tower (WLEF) gather information regarding the area's energy and carbon fluxes. Because of the extreme height, this tower has an unusually large footprint of up to several square kilometers, which is approximately two orders of magnitude larger than most other AmeriFlux sites [*Baker et al.*, 2003; *Denning et al.*, 2003]. Micrometeorological observations including net radiation, soil temperature, and soil moisture measurements are also collected as outlined by *Cook et al.* [2004].

4.1.2. NSF NTL-LTER Project

[29] The National Science Foundation research site known as the North Temperate Lakes Long-Term Ecological Research (NTL-LTER) site was established in 1981 to study the ecology of lakes over long timescales and broad spatial extents in the Northern Highlands Lake District [*Magnuson et al.*, 2005]. The project collects a variety of physical, chemical, and biological data related to the lakes and their surrounding landscape and climate (http://lter.limnology.wisc.edu). The project uses these long-term records to study the seasonality and interannual variability of a diverse range of hydrological processes including lake evaporation [*Lenters et al.*, 2005] and flow between lakes [*Cardille et al.*, 2004].

4.1.3. NWS NCDC Cooperative Observer Program

[30] The National Climate Data Center's Cooperative Observer Program (NCDC COOP) is a nationwide weather and climate-monitoring network started in 1890 and run by the National Weather Service (http://www.weather.gov/om/coop). It relies on volunteers who take daily temperature, precipitation, snowfall, and snow depth measurements [*Doesken and Judson*, 1997]. Within Vilas County, Wisconsin, there are six stations that have relatively complete records from 1948 to 2000.

4.1.4. USGS NTL-WEBB Project

[31] The U.S. Geological Survey project known as the North Temperate Lakes Water, Energy, and Biogeochemical Budgets (NTL-WEBB) project began in 1992 with research priorities of understanding the region's hydrological and biogeochemical cycling [*Elder et al.*, 1992; *Walker and Bullen*, 2000]. The project monitors groundwater wells and stream gauges throughout the Trout Lake basin (http://infotrek.er.usgs.gov/doc/webb). As part of the NTL-WEBB and NTL-LTER projects, *Dripps* [2003] used field measurements and groundwater modeling to estimate groundwater recharge from 1996 to 2000.

4.2. Model Inputs

[32] The IBIS model is forced with daily historical meteorological data derived from multiple sources. Monthly values of temperature, relative humidity, solar radiation, and wind speed are derived from the Climate Research Unit (CRU05) at the University of East Anglia [*New et al.*, 2000] and daily values from the National Center for Environmental Prediction (NCEP) climate reanalysis [*Kalnay et al.*, 1996; *Kistler et al.*, 2001]. As in the study by *Lenters et al.* [2000], we use a combination of these two $0.5° \times 0.5°$ resolution data sets to create daily climate inputs for the model. The monthly CRU05 data are used because of their higher quality at the monthly timescale. The NCEP climate reanalysis daily trends are added to the CRU05 monthly means to interpolate daily mean values. In this study, we use daily values from 1948 to 2000 at a single grid cell (46.25°N, 89.75°W).

[33] For IBIS' daily precipitation inputs, we averaged values from six NCDC COOP stations (details in section 4.1.3) in Vilas County, Wisconsin, that had relatively continuous records from 1951–2000 (NCDC COOP, 2005). These precipitation values were used instead of the NCEP and CRU05 values because they better represented the local rainfall in the region. The CRU05 data set tended to consistently underestimate the region's precipitation during the 1990s.

5. Model Evaluation Criterion One: Energy Balance

[34] We first determined whether IBIS could adequately simulate the energy balance of the heavily forested Northern Highlands region. To do this, we compared our simulated short-timescale (30 min) sensible and latent heat fluxes with

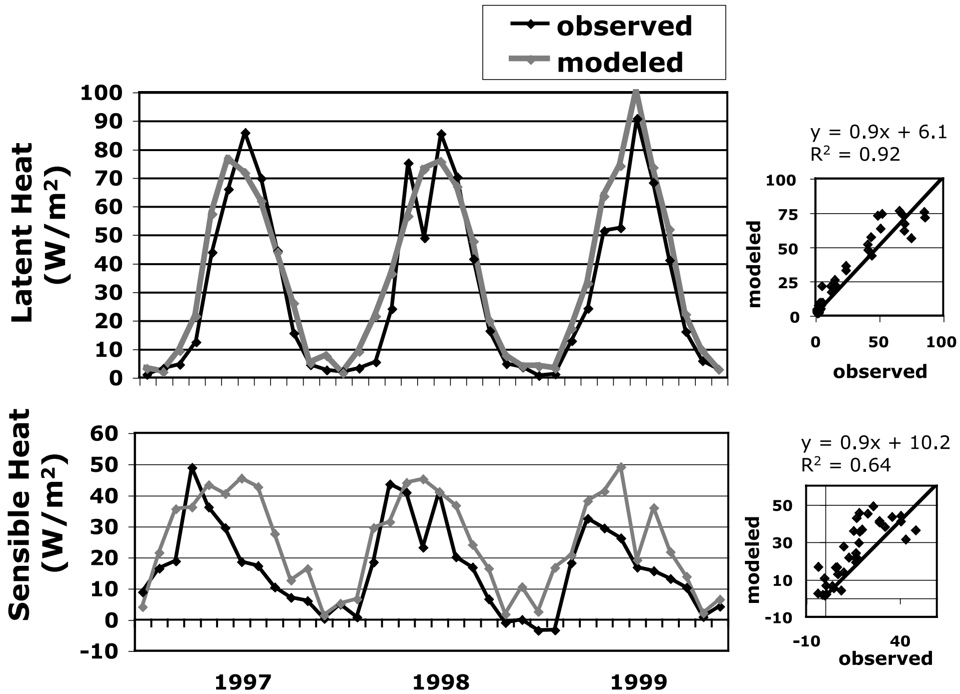

Figure 3. Energy balance comparison. Latent and sensible heat fluxes simulated in IBIS were compared with observations measured at the WLEF Flux Tower at Parkfalls, Wisconsin. The bivariate plots on the right show the correspondence between modeled results and observations over the entire time series depicted on the left. The solid line on the correlation graphs is 1:1. The model captures the seasonality of latent heat (r^2 = 0.92, n = 36) and sensible heat (r^2 = 0.64, n = 36). Observed latent and sensible heat fluxes tend to underestimate fluxes by 10 to 30% [*Twine et al.*, 2000; *Davis et al.*, 2003].

eddy-covariance observations collected for the ChEAS project from 1997 to 1999. The gap-filled data were processed as part of the FLUXNET's Marconi data sets [*Falge et al.*, 2001] (see also FLUXNET, 2005). Net radiation measurements were recorded at Willow Creek, an upland deciduous forest meteorological site 22 km southeast of the WLEF tower, beginning in 1999 [*Cook et al.*, 2004]. Monthly values in 1999 from IBIS simulations and the Willow Creek site compared well (r^2 = 0.98, n = 12). The differences in monthly averages ranged from IBIS simulations being 23 W/m^2 higher in July and 7 W/m^2 lower in December than the observations.

[35] The monthly average latent heat simulated by IBIS approximated the seasonal cycle observed at the WLEF flux tower (r^2 = 0.92, n = 36) (Figure 3). Overall, the observed monthly average latent heat flux during 1997–1999 was 30.5 W/m^2, while the simulated value was 34.9 W/m^2, approximately 15% higher for the annual average. IBIS simulated a peak latent heat flux during the month of July at 82.4 W/m^2, and observations also peaked during July at 87.2 W/m^2, a difference of 5.4% (Figure 3). The modeled values were slightly higher than observations in March, April, and October. Overall, the model simulates latent heat fluxes well, and was within the assumed measurement error (20%) of the flux tower observations.

[36] While flux towers are assumed to have an observational error of roughly ±20% [*Twine et al.*, 2000], other possible sources of differences between IBIS simulations and the WLEF flux tower observations could be attributed to the model's simplified representations of Leaf Area Index (LAI) heterogeneity and vegetation phenology in the region [*Kucharik et al.*, 2006]. Our prescribed summertime maximum LAI values were 4.0 m^2/m^2 for the upper canopy and 0.5 m^2/m^2 for the lower canopy. Comparatively, in 1999 observed LAI within the study region was 3.51 ± 0.89 m^2/m^2 for the 3 × 2 km area centered on the WLEF flux tower [*Burrows et al.*, 2002]. Stand level observations of LAI in 2000 of the four forest types that cover over 80% of the flux tower footprint were: conifer (3.6 ± 0.5 m^2/m^2), northern hardwoods (3.8 ± 0.7 m^2/m^2), aspen (3.5 ± 0.8 m^2/m^2), and forested wetlands (4.1 ± 0.5 m^2/m^2) [*Ewers et al.*, 2002].

[37] We also compared simulated and observed vegetation phenology at a hardwood site near the WLEF flux tower in 1999 and 2000 (observations were not available before 1999). Although IBIS simulated the timing of leaf-on and leaf-off within the range of actual values, the simulated duration of spring leaf out (April) and fall defoliation (October) always lasts for 14 days in IBIS, while the observed duration from 1998 to 2003 measured at the ChEAS site varies greatly and can last over a month. These small errors in duration could potentially account for the latent heat discrepancy between modeled and observed values around April and October.

[38] Sensible heat fluxes simulated by IBIS did not compare as well as latent heat fluxes, but are the same order of magnitude and had seasonal changes similar to the observed fluxes (r^2 = 0.64, n = 36) (Figure 3). The observed monthly average sensible heat from 1997 to 1999 was 16.4 W/m^2, whereas IBIS simulated 24.5 W/m^2 (49% overestimate in the 3-year average). Observed sensible heat

peaked in April at 41.5 W/m^2, while simulated sensible heat peaked in June at 44.8 W/m^2 (8% overestimate in peak values). Also, generally, the model had greater than observed sensible heat throughout the year, except in April.

[39] Several studies have found systematic energy budget closure problems with observed latent and sensible heat fluxes measured with eddy-covariance systems; these systems tend to underestimate fluxes by 10 to 30% [*Twine et al.*, 2000; *Davis et al.*, 2003]. Therefore, given the small magnitude of the sensible heat flux at the WLEF flux tower site (maximum values of ~40 W/m^2 in summer), a 20% error in energy closure would yield about a 4–8 W/m^2 error depending on the season, which is about what IBIS simulates as an absolute error (i.e., the average monthly error in sensible heat was 8 W/m^2). *Baker et al.* [2003] also found that SiB-2.5 overestimated the peak sensible heat flux values. They attributed the error to either net radiation overestimation in the model or underestimation of soil-surface heat, or both, and they hypothesized that differences between observed and simulated fluxes may be linked to wetlands.

[40] Overall, these energy flux comparisons suggest IBIS reasonably simulates atmosphere-land interactions in this landscape. Other studies have used similar comparisons of model simulated flux values with eddy covariance values for conclusive model validation [*Baker et al.*, 2003; *Denning et al.*, 2003]. We, however, take IBIS validation several steps further to understand in greater detail what happens to the water that is not returned to the atmosphere through evapotranspiration.

6. Model Evaluation Criterion Two: Annual Water Balance

[41] We assessed how the simulated 7-year average, annual mean, and interannual variability of the water balance compares with the observed stream gauge values. Over 35% of our study area is covered by lakes and wetlands [*WISCLAND*, 1992] that are well-connected by groundwater flow [*Attig*, 1985]. Therefore tracking water flow throughout the ecosystem is complicated by lags created by lakes and underground water reservoirs. On this water-rich landscape, comparisons with stream gauges are not straightforward because groundwater flow and open water evaporation are difficult to quantify. To improve the accuracy of our stream gauge comparisons, we modeled stream gauge long-term average and annual values with the water balance equation below, which includes variables from both IBIS model output and results from previous studies.

$$\text{Outflow} = (\text{SW} * \text{mSR}) + (\text{GW} * \text{mD}) + (\text{W} * (\text{P} - \text{E})) + \text{Inflow}, \quad (1)$$

where

Outflow modeled stream outflow volume (m^3/time), assumes all water from a basin leaves via the stream (no groundwater outflow from lakes);
SW surface watershed land area, m^2;
mSR modeled surface runoff, m/time;
GW groundwater watershed land area, m^2;
mD modeled drainage, m/time;
W water surface area, m^2;
P precipitation, m/time;
E open water evaporation, m/time;
Inflow stream inflow volume (m^3/time) Trout River (used in outflow only).

Throughout this study, references made to annual measurements correspond to the water year calendar (October through September).

[42] We modeled stream outflow volume (equation (1)) using a combination of IBIS output values of surface runoff (mSR) and drainage (mD) and values from the literature of open water evaporation (E) and land and water areas (SW, GW, W). *Lenters et al.* [2005] derived lake evaporation rates, which were used to account for open water evaporation (E) from 1989 to 1998, using the energy budget method. The NTL-WEBB project delineated both surface and groundwater watersheds and summed water surface areas (SW, GW, W in equation (1)) for the watersheds that fed North Creek, Allequash Creek, Stevensons Creek, and Trout River [*Elder et al.*, 2000] (Figure 1). We limited comparisons to 1992–1998 because observed stream gauge data began in 1992 and observed evaporation data ended in 1998.

[43] In general, the modeled stream outflow value (equation (1)) compared well with observed streamflow data from four stream gauges in the Trout Lake basin (Figures 4 and 5), particularly given uncertainties in measurements of precipitation, evaporation, stream discharge, groundwater, and overland flow [*Winter*, 1981], and the fact that IBIS was not calibrated to stream outflow values as is commonly done in other land surface models [*Nijssen et al.*, 2003]. The modeled 7-year averages for Allequash Creek (1.22×10^7 m^3/yr), North Creek (3.28×10^6 m^3/yr), Trout River (4.28×10^7 m^3/yr), and Stevensons Creek (6.36×10^6 m^3/yr) are 13% more, 6% less, 15% more, and 121% more than observed values, respectively. It is important to note that Stevensons Creek, the creek in which IBIS overestimates the 7-year average by 121%, is the second smallest basin (~7 times smaller than the Trout River basin); whereas Allequash Creek and Trout River have greater

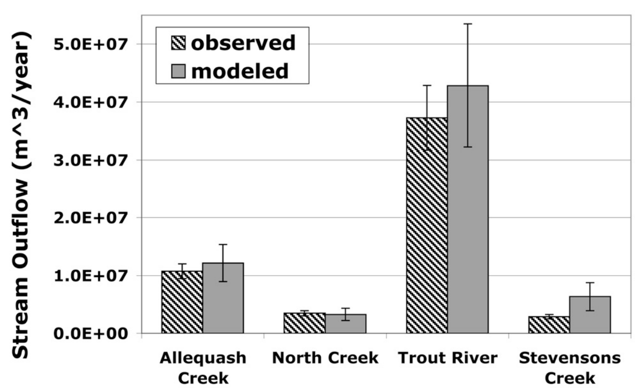

Figure 4. Stream outflow comparison (1992–1998). Modeled versus observed stream outflow for four stream gauges in the Trout Lake watershed (locations indicated on Figure 1). The observed stream gauge data were collected by the USGS-WEBB project. Data are averaged over 7 years, and bars indicate 1 standard deviation (n = 7).

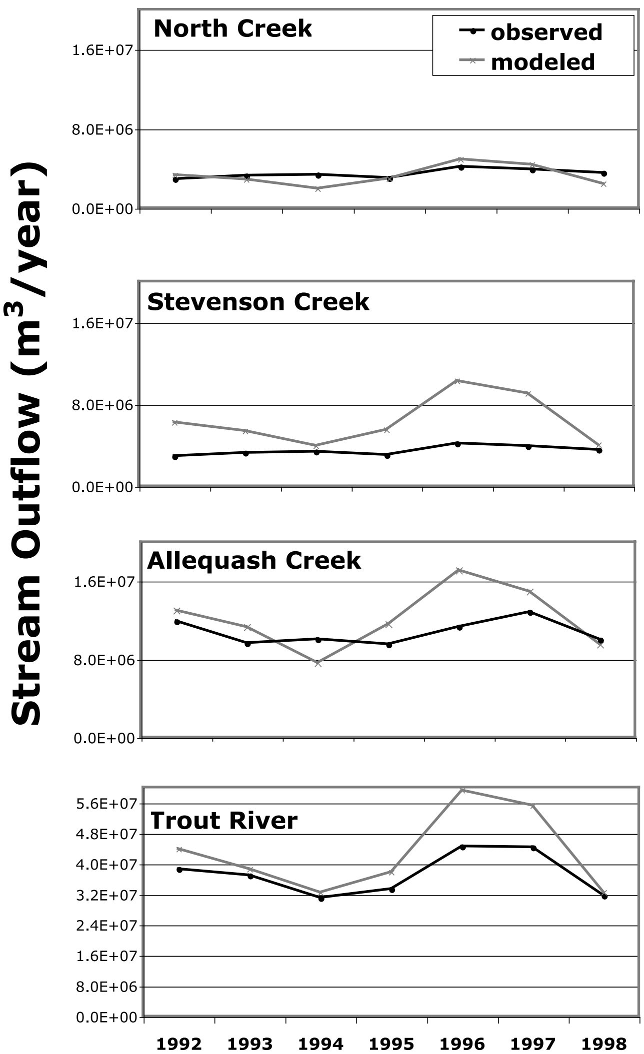

Figure 5. Interannual stream outflow comparison for Trout River (r^2 = 0.95, n = 7), Allequash Creek (r^2 = 0.45, n = 7), North Creek (r^2 = 0.45, n = 7), and Stevenson Creek (r^2 = 0.41, n = 7). Note the scale for Trout River is 3 times greater than the other creeks. Modeled stream outflow values for North Creek and Trout River have the highest level of agreement with observed stream outflow.

basin areas and thus less error relative to their 7-year averages (13% and 15%). Additionally, North Creek has a small (3 ha) lake surface area (the other watersheds have at least 80 times more open water) [*Elder et al.*, 2000].

[44] The tendency of the modeled stream outflow to be greater than the observed outflow could be attributed to our assumption that all water from a basin leaves via the stream. Groundwater flow paths are difficult to define because they are more diffuse and dynamic with time than surface flow paths [*Winter et al.*, 2003]. Consequently, there could be groundwater flow leaving the basin that is not captured in the stream (water infiltrates below the local flow system to deeper regional flow system).

[45] The modeled values had greater interannual variability than observations (Figures 4 and 5). This may result from longer-term groundwater buffering that IBIS does not simulate. Annual modeled stream outflow correlates well with the variation in year-to-year flow of the Trout River (r^2 = 0.95, n = 7). Other modeled stream outflows also show correlation: Allequash Creek (r^2 = 0.45, n = 7), North Creek (r^2 = 0.45, n = 7), and Stevensons Creek (r^2 = 0.41, n = 7). Simulated streamflow was largest for 1996 and 1997, years that are consistent with the largest values in the observed precipitation record.

[46] The modeled stream outflow values for North Creek and Trout River had the highest level of agreement with the observed stream gauge. Because North Creek basin is the smallest watershed it is less vulnerable to errors in watershed and water surface area approximations and less dependent on our assumption that one lake's evaporation rates are representative of the entire region. Alternatively, Trout River, because it is the largest and its values are averages across the entire basin, is also likely to be less sensitive to area approximations or omitted model processes. If we assume water pathways are determined by the surface watershed only (not distinguishing between watersheds being either groundwater or surface) both North Creek's and Trout River's 7-year averages only increase by 5%, whereas Allequash Creek's 7-year average decreases by 32% and Stevensons Creek's increases by 26%. In other words, both Stevensons and Allequash Creek are more sensitive to uncertainties associated with their watershed delineations, making modeled streamflow from these watersheds more uncertain.

7. Model Evaluation Criterion Three: Surface Runoff Versus Soil Infiltration to Groundwater

[47] The third model evaluation criterion assesses how surface water and groundwater are partitioned. Understanding whether water is routed across the surface (i.e., carrying sediments and phosphorus to lakes and streams) or travels through the groundwater system (i.e., carrying silicate and nitrate) is important to the timing and magnitude of flows and, ultimately, has implications for water chemistry, water quality, aquatic ecosystem dynamics, and resource management [*Jordan et al.*, 1997].

[48] We investigated the partitioning of soil infiltration versus surface runoff by comparing IBIS outputs with observations of groundwater recharge in the Trout Lake basin. *Dripps* [2003] studied recharge rates, one of the most complex and uncertain hydrologic parameters [*Jyrkama et al.*, 2002; *Dripps*, 2003]. *Dripps* [2003] calculated annual recharge (defined as water that crosses the water table) from 1996 to 2000 using a combination of five field measurement techniques: time domain reflectometry, temperature profile analysis, water table fluctuation method, recession curve displacement method, and baseflow estimation method.

[49] We compared modeled groundwater drainage from IBIS with these observed recharge values (Figure 6). Drainage from IBIS is defined as the volume of water infiltrating past the 1.5 m soil depth, an important definition because recharge depth influences the timing of drainage seasonality. The 5-year average simulated value was 31.6 cm/yr compared to an observed recharge of 28.5 cm/yr, an 11% overestimate. Annual values for 1996, 1997, and 1998 (underestimated by 8%, 15%, and 5%, respectively) were well within the range of field measure-

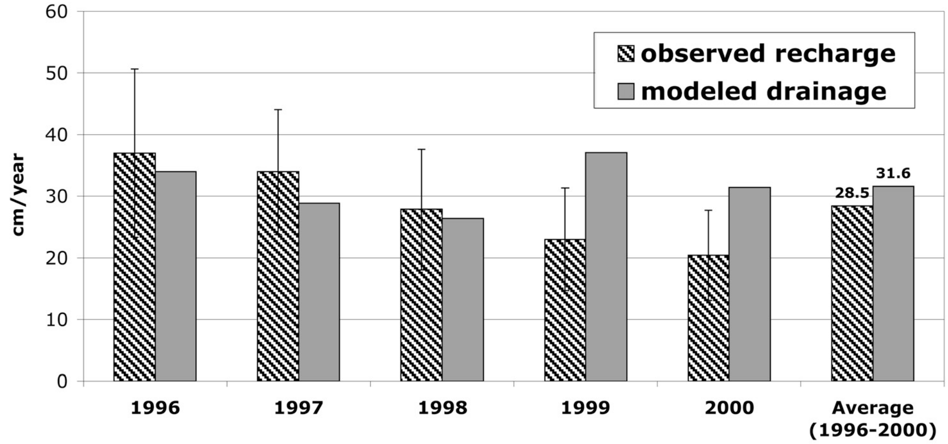

Figure 6. Annual recharge comparison. Modeled drainage is defined as the volume of water infiltrating past the 1.5 m soil depth, whereas observed recharge is water that crosses the water table. These values are comparable, especially on an annual basis. Observed recharge data are from *Dripps* [2003]. The bars represent the range of recharge values calculated using multiple field measurement techniques. Over the 5 years, the model overestimates observed drainage by 11%.

ments, which range from 29–37% [*Dripps*, 2003]. In 1999 and 2000, simulated values were not within the range of field measurement errors. These discrepancies may be the result of (1) atypical recharge field measurements in 2000, from an increased frequency of summer thunderstorms and higher intensity rainfall [*Dripps*, 2003], and (2) model sensitivity to the precipitation data used to drive IBIS.

[50] We note that when IBIS simulations were driven with the same local hourly meteorological station data compiled by *Dripps* [2003], modeled values were almost all within the range of field measurement error (the exception being 2000 when it exceeded the range by only 1.3 cm). The differences in recharge rates due to different climate data sets emphasizes the sensitivity of simulated soil infiltration to climatic drivers, particularly the intensity, duration, and temperature of precipitation events (i.e., heavy rain versus constant drizzle versus snowfall). It is important to recognize that the gridded 0.5° latitude × 0.5° longitude climate data set and mean precipitation of six NCDC COOP stations represent a spatial average for the region. Therefore, because precipitation has a greater likelihood of varying over short distances (compared to the other meteorological quantities), capturing the detailed temporal variability of sub-grid-scale meteorology is problematic.

8. Model Evaluation Criterion Four: Seasonal Timing of Water Flows

[51] The final criterion we used to evaluate IBIS is whether it has the ability to capture the seasonality of changing water flows. These flows are by far the most challenging quantities to simulate because they are the end result of numerous other processes and formulations, most notably snow thermodynamics and soil physics. Therefore, to properly simulate seasonal hydrology, the model must accurately represent snow-cover changes, snowmelt, soil temperature, soil moisture and ice, and the seasonal soil freeze and thaw. In evaluating the seasonal timing within IBIS, we used corroborating data from all projects in section 4.1. We compared the seasonality of soil temperature, soil moisture, and snow depth.

8.1. Soil Temperature

[52] We compared daily soil temperatures simulated by IBIS with field observations from the NTL-LTER from 1989 to 2000 and ChEAS site from 1997 to 2000. Hourly averaged observations of soil temperature at NTL-LTER performed at 5, 10, and 50 cm were used to calculate daily average values from 1989 to 2000. Our comparisons excluded observed and modeled values from June 1995 to August 1996 as a result of a broken soil probe and probe repositioning approximately 100 m southeast because of an airport runway expansion.

[53] We compared the NTL-LTER soil temperature observations with daily (Figure 7) and seasonal (Table 1) averages simulated by IBIS at similar depths. The largest differences in temperature occur in the spring (March to May) and summer (June to August) when the model simulates cooler soils than what are observed. Specifically, IBIS underestimated the 12-year spring and summer average temperature by 1°C at 5 cm to nearly 4°C at 50 cm (Table 1). In the fall (September to November), the modeled 12-year seasonal average soil temperatures are closest to observations with IBIS slightly overestimating temperatures by 0.3°C, 0.5°C, and 0.3°C, for 5 cm, 10 cm, and 50 cm depths, respectively. IBIS simulated the seasonal variability at 5 cm (r^2 = 0.94, n = 4186), 10 cm (r^2 = 0.91, n = 4208), and 50 cm (r^2 = 0.74, n = 3911) with a high level of success. With the 1996 adjustment, the model values more closely simulated the observations. IBIS showed an increase in overall fit of about 0.5°C with the observed values in 1996–2000, with similar trends in underestimating in the winter, spring, and summer and slightly overestimating in the fall. In May–July at 10 cm the observed soil temperatures lag the modeled soil temperatures by approximately two weeks. At 50 cm the lag is even greater (a month) and it occurs for a longer duration (April to August). Because these temperature lags only occur once the soil has warmed

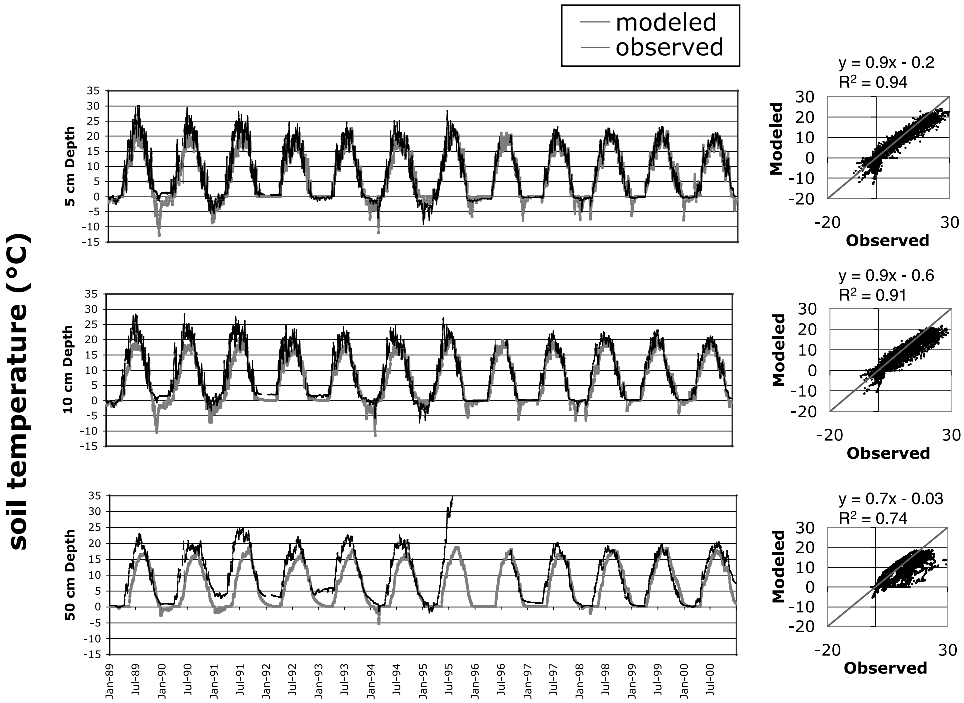

Figure 7. NTL-LTER soil temperature comparison: modeled and observed daily soil temperatures at three depths. The bivariate plots on the right show the correspondence between modeled results and observations over the entire time series depicted on the left. The solid line on the correlation graphs is 1:1. The correlation between observations and model results is greatest at 5 cm (r^2 = 0.94, n = 4186), but decreases slightly with depth (r^2 = 0.91, n = 4208 at 10 cm, r^2 = 0.74, n = 3911 at 50 cm). The peak in observed soil temperatures and the subsequent data gap at 50 cm in 1995 are the result of a soil probe malfunction.

above freezing, differences between modeled and observed soil temperatures do not affect the timing of the spring pulse.

[54] We also compared modeled soil temperatures with data from the ChEAS site. Thirty-minute-averaged soil thermocouple observations from the Mixed Upland site were used to calculate daily averages from December 1998 to November 2000 at the soil surface, 5, 20, and 100 cm (Figure 8). We compared all four depths with daily average soil temperature IBIS output. Like the NTL-LTER comparison, overall seasonal averages matched well. The 2-year averages show IBIS simulated soil temperatures to be warmer in the summer and colder in the winter, similar to what was found at three other AmeriFlux site [*Kucharik et al.*, 2006]. Unlike the NTL-LTER comparison, the 2 years (n = 788) showed no notable time lag in seasonality: surface (r^2 = 0.93), 5 cm (r^2 = 0.98), 20 cm (r^2 = 0.99), and 100 cm (r^2 = 0.99).

[55] At both sites the model simulated temperatures that were colder in the winter than observed soil temperatures. This bias increased with depth as noted by the increasing slope of the correlation line and decreasing r^2 values. These errors could be due to our assumption of a very simplified soil medium that is purely sandy loam with no imperfections. Because soil layers are defined to be thicker with depth and small simulation errors in each layer eventually accumulate over the profile, the top layers are less susceptible to error from soil type heterogeneity. Furthermore, the soil textural profile is approximated in the model by a fixed-depth structure that does not account for where soil horizons

Table 1. NTL-LTER Soil Temperature Comparisons[a]

	0.05 cm Depth			0.10 cm Depth			0.50 cm Depth		
	Modeled	Observed	Difference	Modeled	Observed	Difference	Modeled	Observed	Difference
Dec–Feb	−1.4	−0.7	−0.7	−1.0	0.1	−1.1	0.5	1.9	−1.4
March–May	4.2	5.2	−1.0	3.1	5.8	−2.7	1.4	5.3	−3.9
June–Aug	14.4	16.1	−1.6	13.5	16.2	−2.7	11.5	15.0	−3.5
Sep–Nov	7.4	7.1	0.3	8.3	7.8	0.5	10.2	9.9	0.3

[a]Soil temperature (°C) averaged from January 1989 to December 2000, excluding June 1995 to August 1996 because of soil probe malfunction. Modeled values are typically colder than observed values, and differences increase with depth.

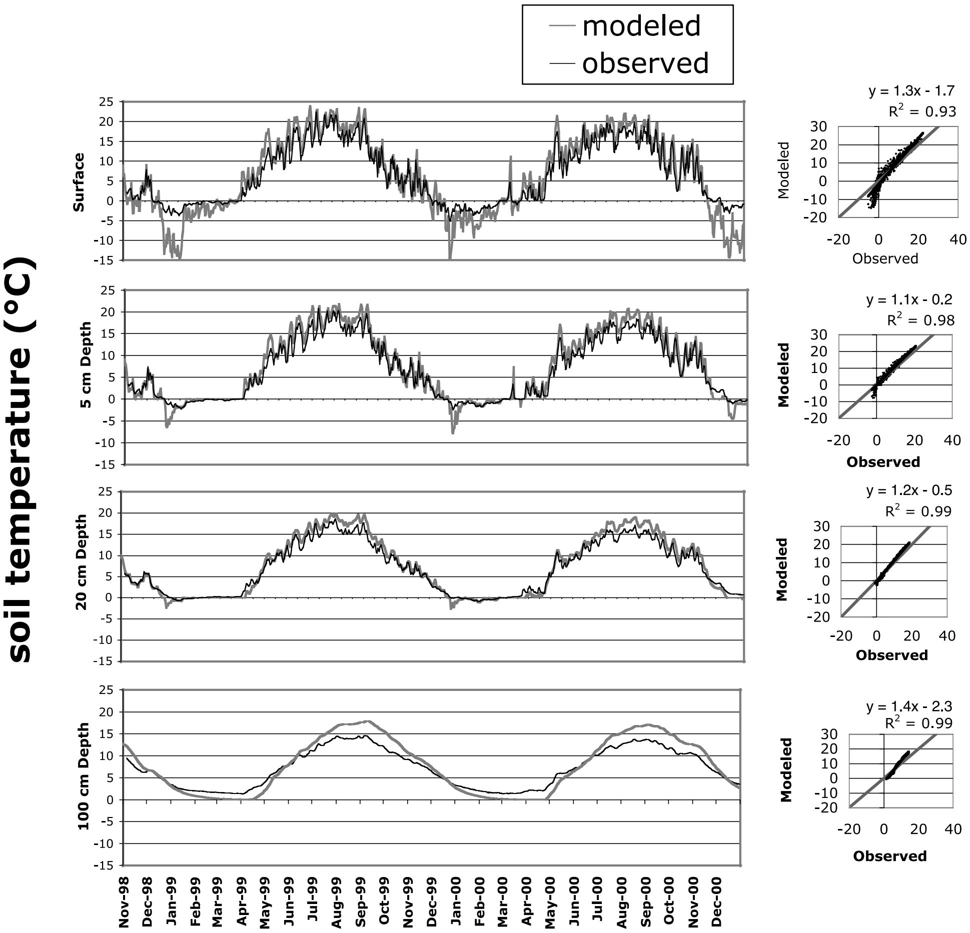

Figure 8. ChEAS soil temperature comparison: modeled and observed daily soil temperatures at four depths. The bivariate plots on the right show the correspondence between observations and modeled results over the entire time series depicted on the left (n = 788): soil surface (r^2 = 0.93), 5 cm (r^2 = 0.98), 20 cm (r^2 = 0.99), and 100 cm (r^2 = 0.95). The solid line on the correlation graphs is 1:1. The seasonal amplitude of soil temperature is smaller at lower depths in both the model and field observation record, as would be expected. At 100 cm, the amplitude of the modeled soil seasonal cycle is somewhat larger than the observations, possibly owing to small simulation errors accumulating with depth.

change. The larger differences between observed and modeled soil temperatures at increased depths is most pronounced at the ChEAS site when comparing 20 cm with 100 cm depths (Figure 8). At 100 cm, comparing the 1:1 line with the correlation indicates that IBIS is simulating warmer summertime and cooler wintertime temperatures than actual observations; these trends are not as apparent at the 20 cm depth.

[56] Capturing winter soil temperatures was improved by having a 5-cm organic layer in IBIS. The effect of the organic layer is apparent in the comparison between simulated temperatures at the surface and simulated temperatures at 5 cm depth (Figure 8). The model versus observed soil temperature seasonality at 5 cm depth (r^2 = 0.98) compares better than the surface temperature seasonality (r^2 = 0.93). The simulated 2-year wintertime (December–February) average soil temperature is 2.8°C less than the observed at the surface, but only 0.3°C less at 5 cm depth. The exact placement of the surface temperature thermocouple is difficult to determine. In reality, the observed surface temperature is more likely to be under some organic materials then exposed (i.e., covered by leaves). At the soil surface boundary, a few millimeters difference in the actual level that a thermocouple is placed versus the reported level can lead to very different results because of the tight thermal gradient at the soil surface litter layer.

[57] Overall, good agreement of the simulated soil temperature and freeze and thaw dates with observations at two independent field sites indicates that IBIS simulates the complex soil energy balance well. Adequately simulating the soil temperature is important to soil biogeochemistry because warmer temperatures can promote wintertime decomposition and nutrient release [*Strum et al.*, 2001]. And, capturing the time of the freeze and melt is especially important because and impacts the physical state (liquid or solid) of the soil moisture.

8.2. Soil Moisture

[58] Soil moisture is less homogenous and therefore more difficult to measure than soil temperature. In addition to the

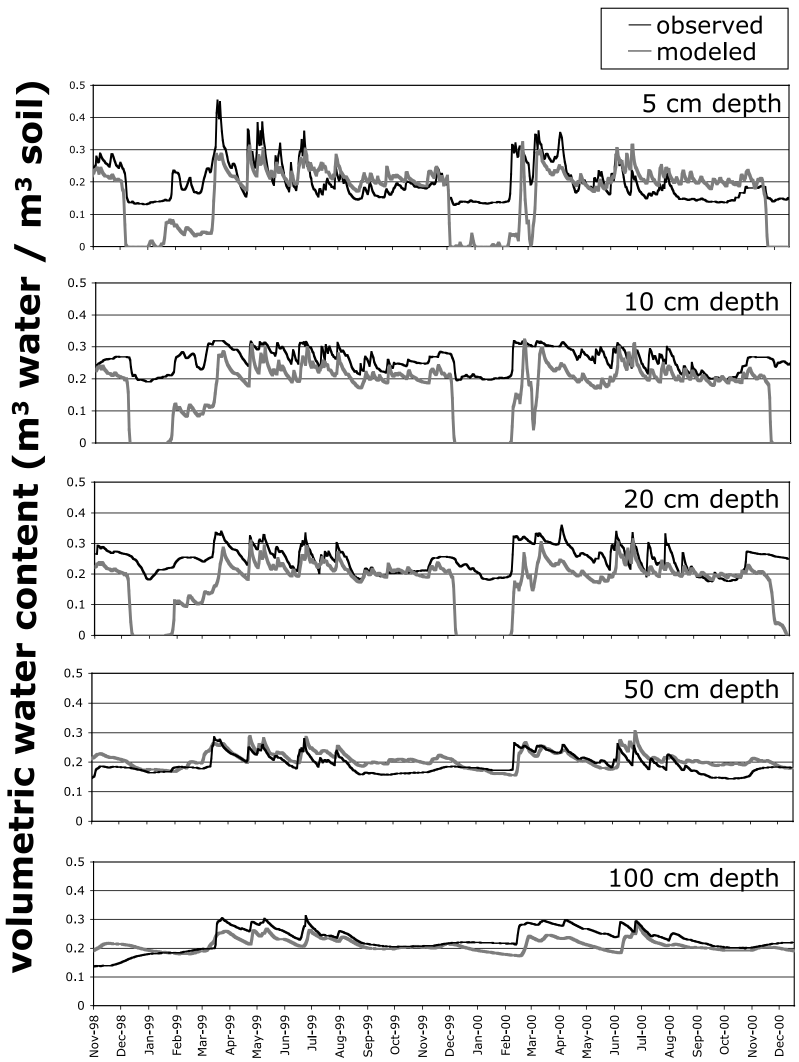

Figure 9. ChEAS soil moisture comparisons: daily soil moisture model results and observations at five depths. The correlations between model results and observations are greater for soil depths below the freeze line (n = 788): 5 cm (r^2 = 0.25), 10 cm (r^2 = 0.37), 20 cm (r^2 = 0.21), 50 cm (r^2 = 0.58), and 100 cm (r^2 = 0.44).

previously mentioned errors from the model's simplifications of the soil profile, soil moisture field observations are also vulnerable to probe orientation and pore structure surrounding the probe. We compared modeled values to soil moisture at a point measurement without modifications for the specific site characteristics other than soil texture (not accounting for macropores, root growth, soil microorganisms, etc., that can change how water moves through the profile). Nonetheless, this approximate comparison of soil moisture gives further information on how accurately IBIS generally simulates soil infiltration timing and magnitude.

[59] We compared daily soil moisture values simulated by IBIS to soil moisture monitored at the ChEAS site from October 1998 to December 2000. The ChEAS site has volumetric water content (m^3 water/m^3 soil) recorded at 10-min intervals by horizontally installed Campbell Scientific CS615 water content reflectometers for soil depths of 5, 10, 20, 50, and 100 cm [*Cook et al.*, 2004]. We use daily averages derived from the 10-min observations to compare with daily IBIS output. Qualitatively, the model adequately simulates the magnitude and timing of changes in soil moisture at all layers (Figure 9). The biggest discrepancy occurs in the winter when modeled values drop to zero in the top three layers and the observed values do not. This is likely due to how phase change is measured. The model specifies soil ice as distinct from soil moisture, therefore as soil ice increases the simulated soil moisture decreases, eventually declining to zero. The instrument, however, uses an electric current that does not differentiate ice from liquid water, and registers values even in frozen soils. Therefore the best comparison between simulated and observed soil moisture occurred when the soil was unfrozen or at soil depths below the frost line. If we include ice, instead of just water, in the calculation of IBIS' volumetric water content (m^3 water/m^3 soil), simulations are closer to observations in total magnitude in the winter. By adding ice, however, we lose the clear signal for when the soil freezes and thaws. Because of this, and because what the soil moisture probe is measuring in the winter is unclear, we focus on the timing

Table 2. ChEAS Soil Moisture Comparisons[a]

Soil depth	Modeled	Observed	Difference
5 cm	0.218	0.204	0.014
10 cm	0.216	0.263	−0.046
20 cm	0.216	0.248	−0.032
50 cm	0.216	0.194	0.022
100 cm	0.217	0.239	−0.022

[a]Volumetric water content (m^3 water/m^3 soil) averaged from November 1998 to November 2000, excluding wintertime values (December to March). Modeled soil moisture remains relatively consistent throughout the profile, whereas observed values have more variation. Variations in observed values are likely due to soil heterogeneity or soil probe orientation.

of the transition periods rather than on winter soil moisture magnitudes.

[60] IBIS captures the seasonal pulses in soil moisture. Below the frost line, IBIS results compare well with observations both in seasonal variations, at 50 cm (r^2 = 0.45, n = 788) and at 100 cm (r^2 = 0.43, n = 788), and in total magnitude of volumetric water content, at 50 cm IBIS overestimates the 25-month average of observations by 9% and at 100 cm IBIS underestimates the observations by 8%. Even in the top three layers IBIS captures some of the seasonality (n = 788): 5 cm (r^2 = 0.25), 10 cm (r^2 = 0.37), and 20 cm (r^2 = 0.21). The 2-year averages of volumetric water content when wintertime values (December–March) are removed show that the modeled values remain relatively consistent throughout the profile, whereas observed values have more variation with depth (Table 2). The differences are small; the simulations were within 5–20% of the observations, which is likely due to soil heterogeneity or how the soil probes are orientated. Because IBIS adequately simulates the timing of soil thawing in the spring and soil moisture magnitude (even during the growing season), it appears that IBIS represents seasonal soil infiltration well.

8.3. Snow Properties

[61] The physical structure of snowpack has complex behaviors that result in a wide range of density values (0.01 to 0.40) due to aging, compaction, and temperature fluctuations [*Strum et al.*, 1995; *Doesken and Judson*, 1997]. Furthermore, because snowfall can occur at a wide range of air temperatures (i.e., 2°C and below), the liquid water to snow ratio and snowfall density can vary considerably at any given location. This variability in snowfall density is evident in northern Wisconsin. Snowfall density recorded at the Minocqua Dam NCDC COOP weather station from 1989 to 2000 has an average of 0.075 m^3 water/m^3 air, a median of 0.04, and ranges from 0.01 to 0.10 88% of the time (but with extreme density values of 0.94 recorded) (NCDC COOP, 2005). Then, once this snowfall lands on the ground, its snowpack density continues to change, typically increasing from aging and melting [*Strum et al.*, 1995; *Doesken and Judson*, 1997].

[62] The physical representations of snow processes in IBIS are simplified compared to reality. IBIS does not account for complex snow density variations. Instead, it assumes a constant snow density, and snowfall density is the same as the snowpack density. Therefore, as time progresses, the snow simulated by IBIS does not compact. Comparing snow water equivalent (SWE) instead of snow depth could improve model validation (by avoiding snow density complications), but unfortunately SWE measurements are time consuming and are infrequently performed.

[63] We used snow depth measurements obtained at six NCDC COOP stations in Vilas County to better understand whether the simplified modeling approach in IBIS can adequately simulate the changing physical structure of snowpack or whether the model representation is contributing to simulated errors in the seasonality of stream discharge. We were specifically interested in comparing observed and simulated values of the onset of snow accumulation and the timing of snowmelt because those dates strongly control the soil temperature profile. The 12-year daily average modeled snow depth reached a maximum of 82 cm on 18 February, whereas the observed values reached a seasonal maximum of 44 cm on 30 January. This difference equates to a three-week lag and an 86% overestimate in magnitude by IBIS (Figure 10). Because the snow density parameterization more closely approximates snowfall density than variable snowpack density, we expected that the maximum snow depth should be greater than the observed maximum depth. Furthermore, we would expect that the simulated error (overestimate) in snow depth would increase in magnitude over the season because the simulated snowpack does not change in density with time and real snowpack typically becomes denser as the season progresses.

[64] The simulated temporal behavior of the snowpack in IBIS compares well to the observed timing of snow accumulation onset and departure. If we classify snow cover arrival and departure as three consecutive days with over 1 cm of snow depth accumulation, the 12-year average simulated snow cover arrival date was less than a day different than observed values in early November. Snow departure in mid-April occurred an average of 4 days earlier in the model; 135 days had snow depth greater than 1 cm in the model compared with an observed average of 144 days.

[65] In summary, IBIS still captures snow cover timing adequately, even though the model makes several assumptions and simplifications that result in an overestimation in snow depth. The physical properties of snowfall and snowpack vary dramatically across the landscape throughout the year. Instead of rigorously accounting for these variations, IBIS uses a single snow density that more closely approximates snowfall density than snowpack density. By making this assumption, IBIS accurately simulates snow cover at the expense of not accurately simulating snow depth, especially late in the snow season. Snow cover, however, has more important biophysical implications to the seasonal hydrologic cycle than snow depth. Because snow is a good insulator, its presence or absence in the fall determines how cold the soil becomes and thus plays a key role in wintertime soil decomposition and nutrient release and in the timing of spring snowmelt. This importance of snow cover was especially evident in model simulations where snow cover timing was not adequately simulated (when snow density was estimated on the basis of snowpack density, not snowfall density). Inadequate snow cover in the fall resulted in modeled soil temperatures that where consis-

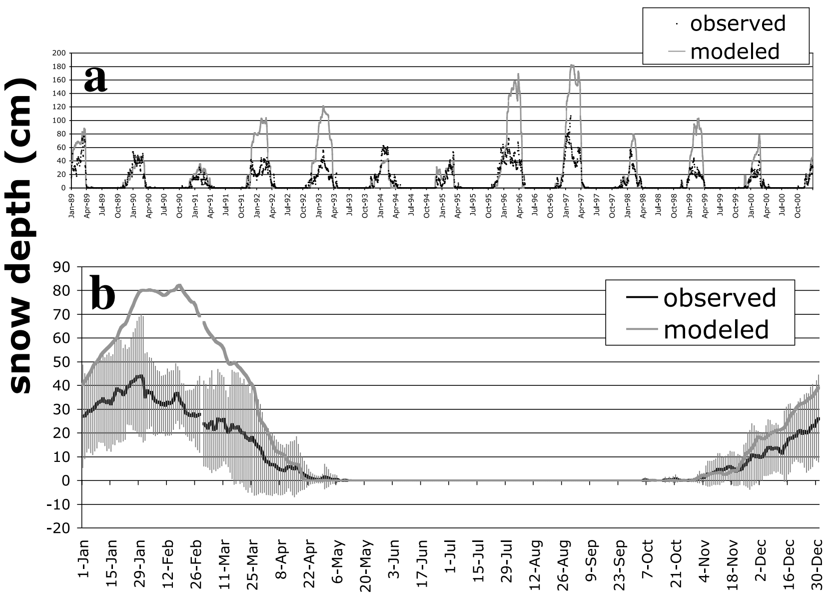

Figure 10. Snow depth comparison. (a) The continuous daily time series from snow depth model simulations and observations from 1989 to 2000 and (b) the same daily snow depths averaged across the 12 years show that IBIS realistically simulates the timing of snow cover but frequently overestimates wintertime snow depths, especially from January to March. The bars indicate 1 standard deviation (n = 12).

tently 5°–10°C colder than observations in the winter months.

9. Model Synthesis: Simulations of Northern Wisconsin's Water Balance

[66] As a summary of the region's water balance, we examined how IBIS simulates each water component's 12-year average (Figure 11), interannual variability (Figure 12), and daily averages (Figure 13) from 1989 to 2000. These comparisons highlight the variability and magnitude of evapotranspiration, surface runoff, and drainage rates and how these components relate to one another. The following is an example of how our overall understanding of interannual and seasonal variations in water balance in northern Wisconsin is enhanced by using a land surface model.

9.1. Annual Averages

[67] Throughout the 12 years, each water balance component contributes by varying degrees to the region's overall water flux. The long-term, 12-year average annual precipitation (814 ± 105 mm/yr, 1 standard deviation) was simulated to be 74% rain (602 mm/yr) and 26% snow (212 mm/yr) (Figure 11). Most of the water (58%) leaves the terrestrial environment by evapotranspiration. The non-evapotranspired residual is apportioned into surface runoff (6%) and drainage (36%) (Figure 11). Additionally, each water balance component fluctuates to varying extents from year to year (Figure 12). Evapotranspiration rates are the most constant between years, with a maximum range of 121 mm/yr (from 396 mm/yr in 1994 to 516 mm/yr in 1991) and standard deviation of 31 mm/yr. Precipitation and drainage are highly variable between years. Precipitation ranges from 610 mm/yr in the driest year to 950 mm/yr (56% more) in the wettest year, just 3 years later. Drainage also undergoes a dramatic increase from 127 mm/yr in 1990 to 411 mm/yr two years later in 1991, an increase of over three fold. The frequency of surface runoff events appears to be log-normally distributed. For 10 of the 12 years annual surface runoff values are less than 30 mm/yr. In 1995 and 1996, surface runoff increases eightfold to just over 160 mm/yr. The infrequency of years with values of surface runoff over 30 mm/yr indicates that water usually infiltrates, and events where land surface conditions are such that substantial amounts of water run off the surface are rare.

9.2. Seasonal Variability

[68] Seasonal variability differs for each water component. Precipitation generally occurs as single day events that vary from trace amounts to maxima of 44 mm/day. During much of the year precipitation exceeds evapotranspiration

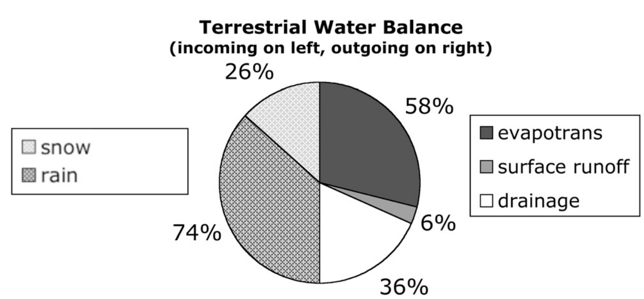

Figure 11. Long-term average terrestrial water balance (1989–2000). The 12-year averages of water apportionment into water budget components for a temperate mixed forest on sandy loam soil. Evapotranspiration is consistently the largest outgoing component, followed by drainage to groundwater.

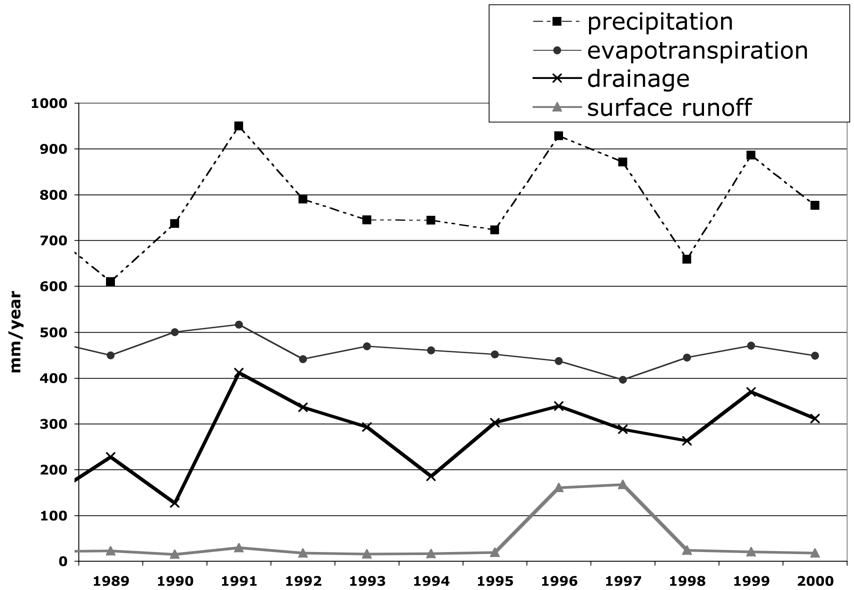

Figure 12. Interannual variability of water budget components. Twelve years of annual water apportionment of the water budget for a temperate mixed forest on sandy loam soil as simulated by IBIS. Drainage and surface runoff rates vary the most from year to year relative to their long-term averages.

and surplus water contributes to soil moisture, surface runoff, and subsurface drainage (Figure 13). Evapotranspiration rates remain below 1 mm/day from mid-October through early April. In April, evapotranspiration rates steadily increase until they level off around 3 mm/day in July, after which rates decline until mid-October. This pattern in evapotranspiration is similar to the precipitation's 15-day running average ($r^2 = 0.73$, $n = 366$), which is lowest in December and peaks in July. On average, mid-July is the only time a water deficit occurs (evapotranspiration exceeds precipitation).

[69] Surface runoff and drainage rates are influenced by the region's freeze-thaw seasonality and magnitude of precipitation events. Water stored as ice and snow decreases drainage and surface runoff rates throughout the winter and increases these fluxes in the spring. The surface runoff pulse is greatly influenced by extreme events, specifically in 1996 and 1997 when daily surface runoff exceeds 5 mm/day for

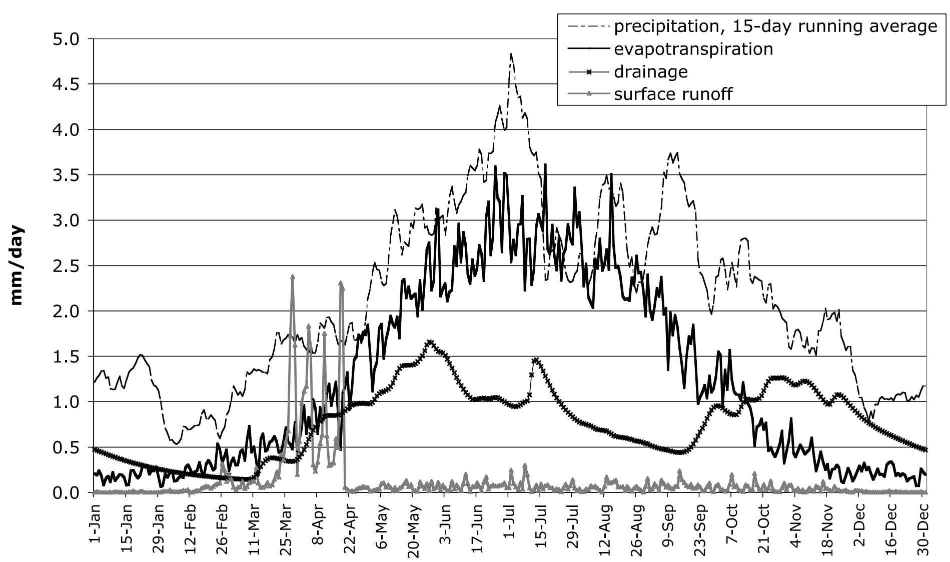

Figure 13. Daily seasonal averages (1989–2000). Twelve-year daily averages of water apportionment for a temperate mixed forest on sandy loam soil simulated by IBIS. The distribution of water to different components of the water budget changes throughout the year: Evapotranspiration rates correlate with precipitation, surface runoff rates pulse in February to April, and drainage rates increase the most in May and October.

more than a week in March and April. Drainage usually occurs in mid-March through early June and again to a lesser extent after fall defoliation in October. Drainage also occurs in the summer when there are intense rain events, such as in the summer of 2000. Extreme summer precipitation events in 2000 led to the major summer drainage event where drainage exceeded 5 mm/day for 6 days. The daily drainage values in the summer of 2000 account for the entire mid-July drainage pulse in the 12-year daily averages (Figure 13).

10. Summary and Conclusions

[70] For comprehensive evaluation of the terrestrial water balance, we outlined four key criteria as guidelines for model evaluation: (1) Does the model simulate the energy balance of the landscape? (2) Is the annual average water balance and interannual variability reasonably simulated? (3) Is the partitioning between surface runoff and soil infiltration to groundwater realistic? (4) Does the model capture the seasonal timing of water flows? These criteria are set forth as a structure for future investigations of seasonal land surface hydrology in cold climates. They are intended to increase the confidence in which land surface models can be applied to generate a more general understanding of the complex interactions between regional climate, water resources, limnology, and geology.

[71] In our study, we addressed these four criteria by comparing IBIS to a suite of field observations. The results showed that the physically consistent modeling framework of IBIS simulated, with a reasonable degree of accuracy, the water and energy fluxes to the atmosphere and the water partitioned into surface runoff and groundwater infiltration at daily to interannual time steps.

[72] Because of errors associated with field observations, the accuracy with which we simulated each component of the water balance is not easily quantified. By investigating the complete land surface water balance, we have, however, increased the likelihood that all individual components were captured. The modeled monthly energy balance had about a 5–15% error compared with observations. The model error of annual water balance, as evident by streamflow comparisons, was generally 15% or less, but appeared to be dependent on the size of the basin. Integrated observations on a stream that drains a larger land area appeared to agree better than values from small basins. Modeled drainage rates were generally within 5–15% of the observed average and were less than the reported measurement errors of 29–37%. Modeled and observed soil temperatures generally differed by less than 3°C and had r^2 values that were greater than 0.9. The capturing of the soil seasonality was improved with the addition of an organic layer, which corrected for an overall cold bias in winter. Soil moisture values were within 5–20%, and freeze and thaw timing was within a few days of observations. Seasonality of soil moisture was more difficult to capture than soil temperature; r^2 values were $\sim$0.4 to 0.5. Modeled snow dynamics captured the observed snow arrival and departure (accumulation on the surface) within a few days of observations, but overestimated the average maximum depth by 86%, and the maximum simulated depth occurred 3 weeks later than the observed.

[73] Through these comparisons, we gained insights into the sensitivities of biophysical processes and improved our understanding of the complex hydrologic and energy fluxes in seasonally cold, snowy climates. One key insight was an improved understanding of the importance of simulated soil temperature. Many biological processes like carbon and nitrogen fluxes depend on soil temperature [*Strum et al.*, 2001; *Luo et al.*, 2003]. Soil temperatures also determine the rate and extent to which water infiltrates [*Cherkauer and Lettenmaier*, 1999]. Initially, simulated winter soil temperatures in IBIS were too cold. By modifying snow cover and organic matter to increase soil insulation, we were able to more adequately simulate soil temperature, which ultimately impacted the timing of spring melt. This insight highlights the complexity of cold climate biophysics, thus illustrating the importance of expanding comparisons beyond typical model validation techniques. Without careful investigation and long-term data, the model's misrepresentation of soil temperature and its implications on the springtime pulse may have gone unnoticed.

[74] Because extensive field observations like flux towers and recharge rates are not available in many locations, we suggest focusing on two relatively simple comparisons that can increase the accuracy with which land surface models capture water cycling seasonality in cold climates. First, simulated soil temperatures should be compared to field observations to verify that the model is adequately simulating soil temperature throughout the year. Soil temperature measurements are straightforward, inexpensive, and robust across the landscape, and they give important insights into the system's sensitivities and seasonality, particularly the seasonal freeze and thaw dynamics (as previously mentioned). Second, we recommend making comparisons between modeled and observed snow cover a priority. Because snow water equivalence measurements are seldom available and snow depth measurements do not account for variations in density, we used readily available snow depth data to identify when snow cover begins and ends and compared this timing with modeled snow cover. The presence or absence of snow insulation is critical to wintertime soil temperature and consequently to the timing of spring melt. Additionally, when a land surface model assumes a single snow density, it is more important to simulate snow cover correctly than snow depth. We suggest using a lower snow density than the seasonal snowpack average to insure soil is adequately insulated. This lower density will cause an overestimation in modeled snow depth, but for seasonal water cycling, it is the water amount in the snow and the temperature of the soil that matters most, not the simulated depth.

[75] Overall, a more complete understanding of the terrestrial water balance can encourage cross-disciplinary research and help anticipate future change. Knowing where water is within the ecosystem, and how it changes through the seasons, helps improve our understanding of hydrological processes relevant to ecology, limnology, geology, and resource management. By improving the ability and confidence with which land surface models can simulate multiple water fluxes throughout the year, land surface models can generate more suitable inputs to other models that examine lake hydrology, groundwater flow, landscape biogeochemistry, and carbon cycling.

[76] By increasing our understanding of complex land surface biophysical mechanisms and existing natural fluctuations in seasonally cold climates, we increase our ability to explore the resilience of the entire ecosystem to future stressors. For example, through simulations we know that the presence or absence of snow cover in the fall affects the timing of spring thaw. We can therefore speculate about the effects of variation in precipitation and temperature seasonality that are a likely outcome of future climate change. For instance, with less snowfall in October and November, the soil would not be insulated as well as it has been in the past. This change would profoundly alter the water storage both above and belowground, the complexities of which we are just beginning to understand. Thus, through further investigations that address biophysical mechanisms and natural fluctuations in cold climates, we can better understand potential implications of change and better anticipate and manage future variability.

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