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Putting aquifers into atmospheric simulation models: an example from the Mill Creek Watershed, northeastern Kansas

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Abstract

Aquifer–atmosphere interactions can be important in regions where the water table is shallow (<2 m). A shallow water table provides moisture for the soil and vegetation and thus acts as a source term for evapotranspiration to the atmosphere. A coupled aquifer–land surface–atmosphere model has been developed to study aquifer–atmosphere interactions in watersheds, on decadal timescales. A single column vertically discretized atmospheric model is linked to a distributed soil-vegetation–aquifer model. This physically based model was able to reproduce monthly and yearly trends in precipitation, stream discharge, and evapotranspiration, for a catchment in northeastern Kansas. However, the calculated soil moisture tended to drop to levels lower than were observed in drier years. The evapotranspiration varies spatially and seasonally and was highest in cells situated in topographic depressions where the water table is in the root zone. Annually, simulation results indicate that from 5–20% of groundwater supported evapotranspiration is drawn from the aquifer. The groundwater supported fraction of evapotranspiration is higher in drier years, when evapotranspiration exceeds precipitation. A long-term (40 year) simulation of extended drought conditions indicated that water table position is a function of groundwater hydrodynamics and cannot be predicted solely on the basis of topography. The response time of the aquifer to drought conditions was on the order of 200 years indicating that feedbacks between these two water reservoirs act on disparate time scales. With recent advances in the computational power of massively parallel supercomputers, it may soon become possible to incorporate physically based representations of aquifer hydrodynamics into general circulation models (GCM) land surface parameterization schemes.

1. Introduction

Climatic change may affect regional hydrologic balance by altering runoff, soil moisture storage, lake levels, aquifer levels, streamflow, and water quality (e.g. [18,29]). For example, under a CO_2 doubling scenario, the Max-Planck-Institute for Meteorology model [14,36] predicts a winter soil moisture increase over mid-latitude continents and a summer soil moisture decrease over most of the same area. However, confidence in regional changes simulated by general circulation models (GCMs), such as soil moisture, humidity and precipitation increases or decreases, remains low [36]. One reason for the difficulty with regional climate change prediction is that even at their best resolution (about 100 km), GCMs cannot resolve local surface contrasts in vegetation, soil and topography [35]. We propose here that aquifer systems represent another source of land surface heterogeneity [32] to be considered in studying the regional effects of climate change. Indeed, Bonan [7] in his work on GCM subgrid parameterization, noted that "in addition to the atmospheric sources of water (precipitation, dew) currently used in land surface models geologic water sources (e.g. aquifers) are needed if one is to simulate both runoff and surface air temperature accurately". Aquifer systems can potentially interact with the global climate system.

By aquifer systems, we mean a one or more saturated geologic units of varying composition (e.g. sand, clay) and hydraulic conductivity [27]. Aquifers are separated by confining units that have low hydraulic conductivity relative to aquifers. Aquifers are distinct from the unsaturated zone, lying above the aquifer, where the soil and other deposits are normally not saturated. While the

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hydraulic conductivity of the aquifer varies according to the type of deposit, the hydraulic conductivity of the unsaturated zone depends both on the type of deposit and the time-varying water content profile.

There are several reasons why it is important to study the interaction between aquifers and the atmosphere on a watershed scale, using a regional climate model (RCM). The first reason for including aquifers in a regional climate simulation is to determine how water resources may be affected by global climate change. A change in precipitation and temperature will alter recharge. Thus, water table levels, if the phreatic aquifer is in good hydrologic connection with surface water bodies, will be affected (e.g. [10,71,72]). Lake levels and stream discharge will be modified as groundwater levels fluctuate (e.g. [56,57,69]) (Fig. 1). Changes in surface-ground water flow systems may affect municipal and agricultural water availability, quality, and pumping costs, as well as flood frequency (e.g. [8,18,29]). A second reason for including aquifers in a regional climate simulation is to determine how groundwater affects latent and sensible heat fluxes to the atmosphere. The position of the water table affects soil moisture and the availability of water to plants and therefore is directly related to latent and sensible heat fluxes to the atmosphere (e.g. [15,17]) (Fig. 1). Under climate change, variation in evapotranspiration may alter convective precipitation and thus influence the local and regional atmospheric water balance [19]. Finally, there is growing interest in studying the linkages between ecosystem dynamics and the hydrologic cycle. Accurate tracking of nutrient fluxes through the hydrologic cycle will require the incorporation of physically based representations of aquifers in land surface parameterization schemes of atmospheric models.

Aquifers within watersheds react to changes in hydrologic stress on a timescale of years to hundreds of years. In the Murray Basin, Australia clearing of deeply rooted native vegetation and subsequent replacement with shallow rooted crops, has led to an increase in the recharge rate by approximately two orders of magnitude. Consequently, the water table has slowly been rising since land clearance approximately 50 years ago [2]. The saline water table has risen into the root zone in some low lying areas, resulting in land salinization [1]. In the High Plains Aquifer, Kansas, groundwater pumping for irrigation has resulted in water table drops of 3–60 m below pre-irrigation levels [9]. This has resulted in a significant reduction in the perennial stream-network density (Fig. 2(a)). Whether the cause of water table declines is climate change or land use, a variation in input to the groundwater system affects water levels and influences streamflow, evapotranspiration, sensible heat flux and potentially convective precipitation.

A conceptual model is presented in Fig. 1 (after [25,70]) which illustrates how aquifers interact with the atmosphere under climate change conditions. In general, the relatively long residence time of groundwater in aquifers introduces a delay in the adjustment of flow systems to atmospheric forcing [32,72]. The response time of an aquifer to an impulse is usually much longer (years to decades) than for the atmosphere (days to

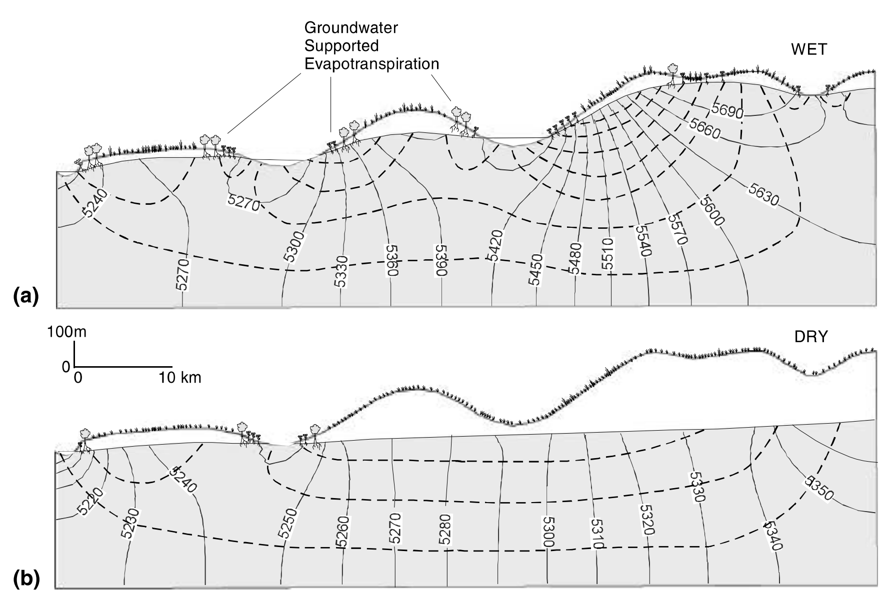

Fig. 1. Schematic diagram illustrating groundwater supported evapotranspiration (areas where plant root system can access the water table directly) during wet and dry climatic periods. During wet conditions, the water table encroaches the land surface in topographic depressions within the uplands. The flow system has components of both local and regional flow. During dry climatic periods, the watertable falls below local topographic depressions in the uplands significantly reducing the amount of groundwater supported evapotranspiration. Regional groundwater flow dominates and lateral groundwater flow supports enhanced evaptranspiration in regional lowlands.

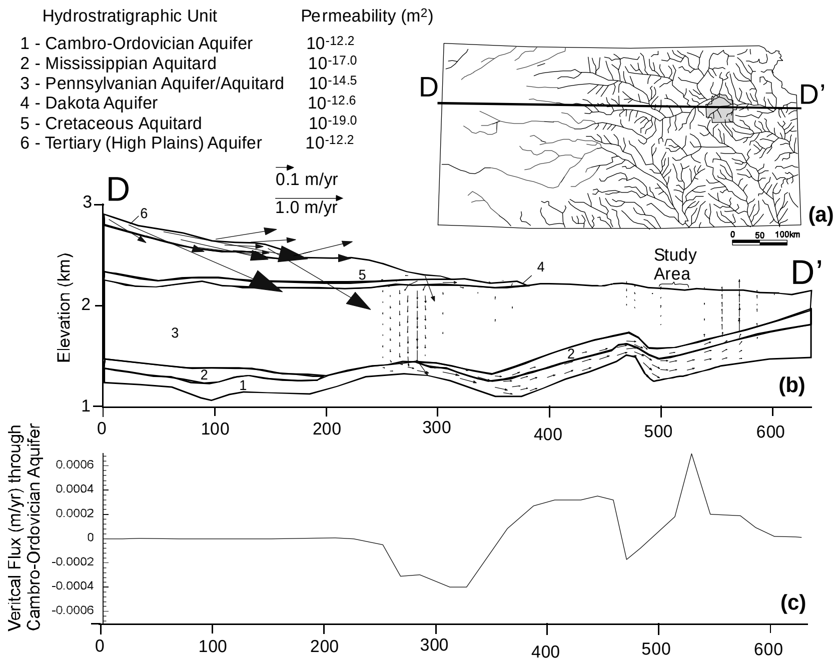

Fig. 2. (a) Comparison of perennial surface water drainage network between 1964 (light plus bold lines) and 1991 (bold lines only). Change in stream network is largely due to groundwater withdrawls and as sociated 4–60 m watertable declines within High Plains aquifer in western Kansas (stream network data from Kansas Geological Survey Website). Also shown in figure is the location of a east–west geologic cross-section across Kansas and the location of Wabunse County where our study area is located. (b) Calculated groundwater velocities using cross-sectional groundwater flow model RIFT2D. The magnitude of the vectors ranged between 1.4 and 0.0001 m/yr (stratigraphy after 51); (c) vertical groundwater flux into and out of Cambro-Ordovician aquifer.

years). Assuming a homogeneous aquifer with a parabolic water table profile, the response time of the aquifer, defined as the change in volume (ΔV) over the discharge (ΔQ) may be written e.g. [45]

$$t = \frac{\Delta V}{\Delta Q} = \frac{S_y L^2}{3T}, \tag{1}$$

where t is the time for the aquifer to drain after a period of recharge (s), S_y is the specific yield (dimensionless) of the aquifer, L is the length of the flow system (distance from the stream to the groundwater divide, m), and T is the transmissivity of the aquifer (the product of hydraulic conductivity and saturated aquifer thickness, m^2/s). Even for a relatively small watershed, where, for example, $S_y = 0.20$, $T = 1 \times 10^{-3}$ m^2/s and $L = 2$ km, the characteristic time to respond to a recharge event would be about eight years. For a larger catchment, with, for example, $L = 10$ km, this time goes up to about 200 years.

While it may be intuitive that aquifer recharge, water levels, and evapotranspiration interact, quantification of this feedback relies on detailed study of the properties of individual aquifer systems [46]. In studying the reaction of watersheds to climate change hydrologists need to consider the long residence time of groundwater, as well as the fact that variations in geology, topography, and anthropogenic water use result in considerable spatial and temporal heterogeneity in the response of a particular watershed. For example, a shallow water table, through capillary rise and plant uptake, can moderate the soil moisture and evapotranspiration deficit during dry periods [17,37]. As the water table drops, the phreatic aquifer has less influence on soil moisture and evapotranspiration. When the water table adjusts to new climatic conditions, aquifer thickness (T) and interconnectivity (L) to surface water bodies are modified. In the glacial topography of Fig. 1(a), water levels drop under drier conditions, evapotranspiration and evaporation decrease, and local flow systems are eliminated. Without the connection to the groundwater, the center lake dries out and the stream becomes ephemeral. This type of terrain is found throughout glaciated watersheds of the upper midwest USA, where changes in water table elevation up to 2 m, over a timescale of less than a decade, have been documented [71]. In the semi-arid region of Fig. 1(b), a drop in the water table results in increased irrigation costs. With the decrease in aquifer thickness, depletion of the groundwater reservoir is more likely (e.g. [26]). In both types of watersheds (Figs. 1(a) and (b)), once the water table drops, the interconnectivity to the surface decreases (larger L), and the remaining regional flow system will have a longer response time to climatic impulses (Eq. (1)) than the local flow systems did.

The representation of land surface processes in GCMs and in RCMs, is evolving toward a physically based watershed approach (e.g. [15,23,24,44,47,58,64]). The previous generation of GCM land surface models (LSMs) represented runoff as rainfall minus evapotranspiration. Excess soil water was removed from the domain without utilizing a surface water drainage network or representing groundwater–surface water interactions (e.g. [47]). But in reality, this excess water becomes river discharge and aquifer recharge and may affect fluxes to the atmosphere [42]. The watershed approach to LSMs includes lateral transfers of water through rivers and aquifers. Streamflow and overall hydrologic budget improve when a river routing scheme is incorporated into a regional model (e.g. [3,34,53]). These models also allow for the development of AGCMs that could incorporate biogeochemical cycling and ecosystem dynamics.

Water is transported both laterally and vertically through aquifers, and aquifers interact with the atmosphere through their influence on surface flow systems and soil zone fluxes. Aquifers have traditionally been ignored in LSMs. Only recently [13,32,48,62] has the influence of groundwater been investigated in LSMs. Studies to date have not included a physically based groundwater model, because it is not computationally practical to incorporate one into a GCM. Simplified representations of groundwater and surface flow, such as the statistical–dynamical method of TOPMODEL (e.g. [6,68]) have been successfully used in surface-vegetation–atmosphere models (e.g. [23,24,64]). While TOPMODEL is more computationally efficient than physically based distributed-parameter hydrologic models, its empirical nature prevents it from being able to represent the long-term hydrodynamics of aquifers. Other types of conceptual rainfall–runoff models (e.g. [5,12]) and physically based groundwater–river models (e.g. [66]), incorporate varying degrees of aquifer properties, but interaction with the atmosphere is usually not considered. Groundwater levels and streamflow are sensitive to groundwater pumping, recharge rates, and aquifer properties [8,26,61], and thus a 3-D groundwater model, like MODFLOW [50] is needed. Three-dimensional groundwater models such as MODFLOW have been used in an off-line manner e.g. [38,61] to investigate the potential effects of climate and land use change on watersheds. However, this approach does not allow changes in groundwater-influenced latent and sensible heat fluxes to feedback to the atmosphere.

We have developed a coupled land atmosphere simulation program (CLASP II) model (Fig. 3) to investigate decadal timescale impacts of global climate change on watersheds. The CLASP II combines aspects of both traditional LSMs and groundwater models within a watershed scale system. This model is unique in its coupling of a physically based three-dimensional

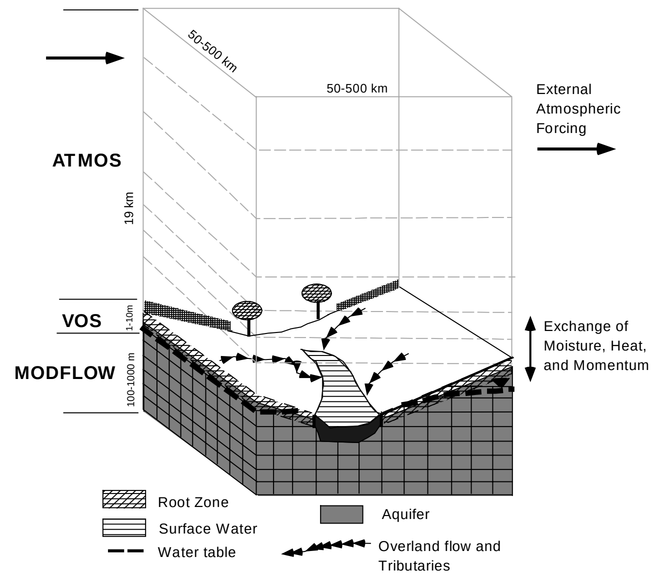

Fig. 3. Schematic of the coupled land atmosphere simulation program model, version 2 (CLASP II).

groundwater model to a LSM–atmosphere system. With this atmosphere–surface/groundwater model, we can represent long-term (decades to centuries) atmosphere–watershed (up to 500×500 km in size) feedbacks. The model is designed to utilize GCM output for watershed scale simulations. The watershed is nested within the larger scale atmospheric boundary conditions. The first version of the model, CLASP I [32], utilized an idealized, quasi two-dimensional representation of aquifers, with restricted linkages between groundwater and surface water. CLASP II contains a full three-dimensional groundwater model and has the ability to represent aquifer–stream interaction in a physically based manner. Because the groundwater model is physically based (MODFLOW; [50]), changes in groundwater flow systems, aquifer and lake levels, and stream discharge can be calculated.

We first describe the model and present a field application of CLASP II to a watershed in northeastern Kansas. An eight-year simulation (1985–1993) is presented to illustrate the model's ability to simulate the entire hydrologic cycle on decadal-time scales. We compare this base run, which incorporates a single-layer aquifer, to a model with the aquifer "removed". Although we do not present a climate change simulation here, the historical record chosen includes both very dry (1988) and very wet (1993) years. This provides insight into the type and distribution of aquifer–atmosphere interactions that occur under varying atmospheric conditions. Results from a long-term (40 year) drought simulation is presented to illustrate the long-time scales required for aquifers to equilibrate to climate change conditions.

In this initial application of the CLASP II model, our primary objectives are: (1) to demonstrate that the model results are consistent with field measurements and (2) to assess whether or not groundwater supported evapotranspiration is significant in a small Kansas catchment and (3) to demonstrate that a physically based groundwater model can provide insights into groundwater–atmosphere interactions on decadal timescales, especially under conditions of climate change.

2. Modeling approach and model description

The CLASP II model was developed as an aquifer–soil-vegetation–atmosphere model for use with GCM or reanalysis boundary input forcing [31,32]. Fig. 3 is a schematic of the atmosphere–surface–aquifer system. It is distinguished from traditional hydrologic surface–subsurface models in its coupling with the atmosphere. It is distinguished from traditional GCM land surface schemes by its physically based groundwater model and focus on watershed-scale processes. The atmosphere is represented by a single column, with vertical discretization. We chose to use an atmospheric model rather than historical data because otherwise it would be difficult to obtain the driving data for the surface model. The type of detailed observations needed for the Penman–Monteith equation (temperature, radiation, humidity and wind speed) are generally not available hourly, over a number of years. Also, observed (wind, temperature, humidity profiles) external boundary conditions are used for the atmospheric model. This can be contrasted with running a land-surface model alone, where we would have to input radiation, for example, by inferring it from surface observations or by using someone else's radiation computation (there are few direct measurements). We could, for example, use National Center for Environmental Prediction (NCEP) reanalysis radiation fields, but then they would be partly affected by the land-surface characteristics of the NCEP model. In addition, we wanted the capability to study feedback between groundwater and the atmosphere. We cannot change the large-scale circulation input, but the properties of the land surface can alter the net water input to the region and the character of the precipitation. The single column atmospheric model, meant to emulate a GCM gridbox, prevents us from capturing horizontal atmospheric heterogeneities in the domain, although precipitation disaggregation can be parameterized and explored in CLASP II. However, uniform precipitation has been used in this study. This is a shortcoming of CLASP II as it is presently applied, because it has been demonstrated that surface hydrology is particularly sensitive to the subgrid areal distribution of precipitation in GCMs [32,54]. Fine lateral resolution can be achieved with mesoscale atmospheric models and such models have been used to simulate climate change time slices [20,33]. However, use of a mesoscale model continuously on a decadal timescale would be computationally prohibitive at present. Without decadal scale simulations, it is not possible to capture long-term aquifer forcing.

The soil and the single-layer aquifer model in CLASP II are horizontally discretized. We neglect the effects of interactions with deeper aquifers in this model (see discussion in Section 3). This spatially distributed approach for the land surface and aquifer allows for explicit specification of soil, vegetation, and aquifer properties for each cell. We do not incorporate the detailed mechanics of soil moisture redistribution or surface water hydrodynamics and this precludes us from predicting the transient response of these systems on a timescale of days. The river routing scheme uses Manning's equation e.g. [16] to calculate river stage and accumulates water from every cell and routes it to the outlet in one timestep. Thus, CLASP II cannot resolve, for example, the propagation of a flood wave downstream. We have chosen to use a simple reservoir approximation of the soil zone rather than a more complicated multi-layer, explicit water-energy balance formulation e.g. [24]. The CLASP II takes a different approach than existing LSMs for representing fields such as evapotranspiration and soil moisture because of our focus on groundwater–atmosphere interactions. This model has been designed for the purposes of exploring the water resources implications of global climate change and the interaction between the water table and latent heat fluxes. Thus, the CLASP II formulation, with its simplified atmosphere and soil-vegetation zone, is appropriate because of our focus on yearly decadal timescale watershed feedbacks. Decadal timescale simulations can be carried out with CLASP II on a conventional workstation because of the relatively simple representations of the atmosphere and soil-vegetation zone.

Atmospheric boundary conditions were obtained from operational analyses produced by the nested grid model (NGM) of the US NCEP. Pressure, humidity, temperature and wind gradients at the atmospheric model horizontal boundary [31] are supplied. The resolution of the analysis data [30] is approximately 150×150 km. Atmospheric radiation, wind speed, pressure, humidity, temperature and precipitation are simulated internally. Land surface sensible and latent heat flux, streamflow and soil moisture as well as aquifer levels are also simulated internally. We use a 15 min timestep in CLASP II. This is a normal timestep for atmospheric models but it is short for a groundwater model, where timesteps of days or weeks are typical. The short timestep is needed in order to adequately characterize latent and sensible heat fluxes as well as the dynamics of atmospheric processes, such as cumulus convection.

2.1. ATMOS

The atmospheric model is a single column with 25 layers in the vertical extending up to 19 km. Vertical discretization varies between 10–300 m, with more resolution in the boundary layer. This structure mimics a single GCM gridbox. The model is governed by the primitive equations for the conservation of momentum, thermodynamic energy, mass and water vapour (Table 1). Precipitation is distributed uniformly over the surface and takes the form of either supersaturation precipitation or cumulus convection. The Emanuel scheme [21] is used for cumulus convection. Surface and atmospheric radiation are computed using broad-band radiative transfer, whereas cloud cover and unresolved turbulent transport in the boundary layer are parameterized [31]. The sensible heat flux and momentum exchange between the surface and atmosphere are computed using drag laws. Lateral heat, water, and mass fluxes are imposed as boundary conditions.

2.2. VOS-MOD

The vegetation–overland flow–soil model/modular groundwater flow model (VOS-MOD) consists of soil-vegetation zone routines integrated into the USGS groundwater flow model, MODFLOW [50]. The soil-vegetation zone is discretized laterally in the same finite-difference grid as the underlying aquifer. The soil-vegetation zone interacts with the atmosphere through sensible and latent heat as well as momentum fluxes. The soil-vegetation zone interacts with the aquifer through recharge to the aquifer and evapotranspiration directly from the water table [46].

In the single layer soil-vegetation zone, type of soil and vegetation are specified [67]. The thickness of the soil-vegetation zone is defined by the rooting depth. The Penman–Monteith equation is used to calculate evapotranspiration as a function of radiation, air temperature, humidity and canopy conductance e.g. [16]. The maximum possible canopy conductance is modified by air temperature, vapor-pressure deficit, solar radiation and soil moisture deficit resistance functions [32]. The soil moisture is a function of precipitation, evapotranspiration, and the position of the water table. The soil moisture cannot go above field capacity or below the wilting point. When the water table is in the root zone the soil moisture is at field capacity and the evapotranspiration is at a maximum. If the water table is in the root zone, the evapotranspiration is assumed to come from the water table and therefore any precipitation recharges the aquifer. If the water table is not in the root zone, the change in soil moisture is calculated as the difference between precipitation and evapotranspiration. Any soil water above field capacity becomes recharge to the aquifer.

Both hortonian and saturation overland flow are incorporated into the model [41]. When the precipitation rate exceeds the infiltration capacity of the soil, the excess water is routed as hortonian overland flow to the nearest stream cell. When the water table rises above the land surface, the water is routed as saturation overland flow to the nearest stream cell.

The 3-D, finite-difference, modular groundwater flow model, MODFLOW, forms the groundwater flow portion of the code [50]. The equation for groundwater flow in a heterogeneous and anisotropic medium where the principal axes of hydraulic conductivity are aligned with the coordinate directions is solved (Table 2). The model stream network is specified [55]. Streamflow at the outlet is equal to the sum of the upstream contributions from stream cells plus/minus the sum of the leakage from/to the aquifer. Baseflow into or out of the stream is proportional to the difference in head between the stream and the aquifer and the conductance of the streambed. Streamflow is assumed to be instantly available to downstream reaches within a timestep. This approxi-

Table 1
Governing atmospheric hydrodynamic and thermodynamic equations

$$\frac{\partial}{\partial t}(\rho u) = -\nabla \cdot \rho u \mathbf{V} - \frac{\partial}{\partial z}(\rho u w) + \rho f v - \frac{\partial}{\partial x} p + F_u$$

$$\frac{\partial}{\partial t}(\rho v) = -\nabla \cdot \rho v \mathbf{V} - \frac{\partial}{\partial z}(\rho v w) + \rho f u - \frac{\partial}{\partial y} p + F_v$$

$$\frac{\partial}{\partial t}(\rho \theta) = -\nabla \cdot \rho \theta \mathbf{V} - \frac{\partial}{\partial z}(\rho \theta w) + F_\theta$$

$$\frac{\partial \rho}{\partial t} = -\nabla \cdot \rho \mathbf{V} - \frac{\partial}{\partial z}(\rho w)$$

$$\frac{\partial p}{\partial z} = -\rho g$$

$$\frac{\partial}{\partial t}(\rho q) = -\nabla \cdot \rho q \mathbf{V} - \frac{\partial}{\partial z}(\rho q w) + F_q$$

t is time (s), x is the west–east coordinate (m), y is the south–north coordinate (m), z is the vertical coordinate (m), ρ is air density (kg/m^3), $\mathbf{V}$ is the horizontal wind vector (m/s), u is the west–east component of $\mathbf{V}$ (m/s), v is the south–north component of $\mathbf{V}$ (m/s), w is the vertical wind (m/s), f is the Coriolis parameter (s^{-1}), p is air pressure (N/m^2), the F terms represent parameterizations of various unresolved physical processes, θ is the atmospheric potential temperature (°C), g is the acceleration due to gravity (N/kg), and q is the atmospheric water-vapor specific humidity (dimensionless).

Table 2
Governing groundwater equation

$$\frac{\partial}{\partial x}\left(k_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(k_{yy}\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(k_{zz}\frac{\partial h}{\partial z}\right) - W = S_s\frac{\partial h}{\partial t}$$

k_{xx}, k_{yy} and k_{zz} are the values of the hydraulic conductivity along the x, y and z axes, respectively (m/s); h is the potentiometric head (m); W is sources and sinks (volumetric flux per unit volume, s^{-1}); S_s is the specific storage (m^{-1}); and t is time (s).

mation is acceptable for looking at integrated runoff on monthly and yearly timescales.

3. Catchment description

A small northeastern Kansas catchment was chosen for the first application of the CLASP II model. Data from the catchment was used to verify that the model was simulating the physical system within reasonable bounds. The domain is the Mill Creek Watershed in Wabaunsee County, northeastern Kansas (Fig. 4). The normal annual precipitation in Wabaunsee County is about 90 cm/year, the mean annual Mill Creek discharge is about 2×10^8 m^3/year, and the mean annual snowfall is about 50 cm [39]. The snow does not accumulate into a snowpack but rather melts within a week or two of deposition [G. L. Macpherson, University of Kansas, personal communication, 1997]. The approximately 40×40 km^2 domain is a closed catchment with origins in the highlands and termination at the junction of Mill Creek and the Kansas River. The landscape consists of low hills, where elevations range from 270–450 m. The groundcover is primarily grassland, with some trees in the stream floodplain. The domain is close to the Konza

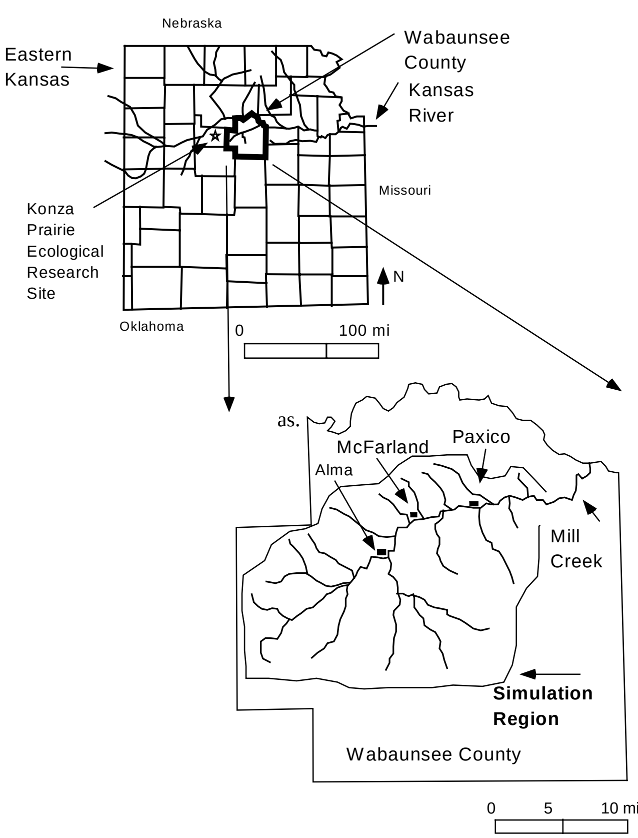

Fig. 4. The Mill Creek Watershed, Wabaunsee County, Kansas. The precipitation station is at McFarland. The USGS stream gauging station is at Paxico.

Prairie Natural Research Area (KPRNA), a 35 km^2, tallgrass prairie ecological research site (Fig. 4). The KPRNA lies within the 15×15 km^2 site of the First International Satellite Land Surface Climatology Project Field Experiment (FIFE), which took place in 1987–1989 [59]. The objective of the FIFE project was to study the exchange of radiation, water, and carbon dioxide between the land surface and the atmosphere using satellite, airborne and surface observations. The availability of soil moisture and evaporation [4] data from the FIFE experiment was the major reason for choosing northeastern Kansas as the study domain. The FIFE soil moisture and evapotranspiration measurements, which were scattered over the area in upland and lowland locations, were averaged over the site. There is a limited amount of groundwater data available adjacent to and in the study area [40,49]. Daily historical Mill Creek at Paxico streamflow measurements for the years 1953–1993 [65], precipitation data (1984–1994) at McFarland [43], temperature data (1984–1994) from a station next to the domain [43], and limited aquifer water level measurements [40] are available for model calibration and validation.

The hydrogeology of Wabaunsee County is characterized by Permian and Pennsylvanian Systems as well as some Pleistocene glacial drift deposits, underlain by a deep Paleozoic aquifer system. The upper aquifers and confining units consist of Permian age alternating layers of limestone (1–2 m) and shale (2–10 m) [60], with some Pennsylvanian age sandstone–limestone units. The deep Paleozoic aquifer system is recharged in Colorado near the Rocky Mountain Thrust Belt. We have developed a simple, west–east, cross-sectional model of this deeper flow system (Fig. 2) using the groundwater flow model RIFT2D [53]. This cross-sectional model was constructed using stratigraphic information presented by [52] and permeability data from [28]. The cross-section neglects the effects of erosion on underpressure development within the confining units as well as recharge in western Colorado. Throughout most of Kansas, there is a modest component of vertical flow of groundwater between the Cambro–Ordovician aquifer system and the overlying Pennsylvanian deposits. Within the Mill Creek study area, these flow rates are several orders of magnitude smaller than flow rates in the shallow aquifer system. However, Fig. 2(b) indicates that in some regions of the state, this deep groundwater flow component may be a significant (5% <) fraction of the overall subsurface hydrologic budget. Within the study domain, groundwater occurs largely in consolidated limestone aquifers in the Permian deposits and in stream channel alluvium. [49] suggests that some of the Permian limestone in the KPRNA may be solution enlarged. Slug tests have indicated that the hydraulic conductivity in the KPRNA limestone and shale layers ranges from 10^{-8} to 10^{-3} m/s [49]. Macpherson's [49] study of

KPRNA hydrogeology shows annual variability of water levels in alluvium and limestone wells of 1–2 m. A rough estimate of water withdrawals has been made in the Mill Creek domain using USGS data [J. Kenny, Kansas USGS unpublished data, 1997]. Yearly ground and surface water withdrawals for rural domestic, livestock, municipal, irrigation, and industrial uses add up to at most 1–3% of the yearly streamflow.

In Wabaunsee County, the Kansas River and alluvial aquifer systems are the principal sources of water in the region [9,63]. The Kansas river is not within the model domain, but we assume the alluvial properties are representative of Mill Creek alluvium. Fader [22] gave the transmissivity of the Kansas river valley alluvial–aquifer system as 0.006–0.05 m^2/s and the long-term storage coefficient as an average of 0.15. Jian et al. [38] the average hydraulic conductivity of the Kansas river alluvium as 2.4×10^{-3} m/s and the vertical hydraulic conductivity of the Kansas river streambed as 3.5×10^{-6} m/s.

4. Model implementation

The Mill Creek land surface and aquifer are divided into cells 2 km on a side. The soil-vegetation zone cells overlie the aquifer cells. The thickness of the soil-vegetation zone is set equal to the rooting depth. The thickness of the aquifer depends on the elevation of the base of the aquifer and the surface topography. The typical thickness of the aquifer is 50–100 m while the overlying unsaturated zone varies in depth from zero to 100 m. The watershed is contained within the rectangular domain. Cells outside of the watershed boundaries are inactive. A single-layer aquifer is used here, but multiple layers can be specified. Representing the limestone–shale–alluvium system of the Mill Creek aquifer as a single layer is a great simplification. However, it was decided that attempting to resolve the multiple 1–10 m thick layers of limestone and shale in the aquifer would be an unwarranted level of complexity, for this study. We lack the groundwater level data to test a multi-layer aquifer model. We seek to capture the bulk behaviour of the system while realizing that many details, such as rapid macropore response to precipitation events or perched water tables, will be beyond our reach.

In addition to the simulations using the full CLASP II model, a simulation was also constructed in which the aquifer was removed. This was accomplished by (1) setting the initial water table low so that it would not interact with the land surface, (2) routing directly to the stream water that would have gone as recharge to the aquifer, (3) not allowing baseflow from the stream to the aquifer. This results in a surface water routing scheme that, like most GCMs, does not resolve groundwater storage and flow. In the no-aquifer run, the land surface retained the same properties and lateral discretization it had before the aquifer was removed. We did not separately calibrate the no-aquifer run. Our intent was to isolate the aquifer contribution to the system by comparing the aquifer versus no-aquifer runs. We could have calibrated the no-aquifer run to produce similar results to the aquifer run, but our desire was to study model hydrodynamics, not to calibrate to observations. Comparison of model results with and without an aquifer provided insight into the role of groundwater supported evapotranspiration. Actually, use of the word 'calibration' in the studies presented here is misleading and we henceforth we will avoid this term. We used observations as a check that the model was behaving realistically and consistently. Reproducing observations was not our primary objective here but rather we sought to examine potential aquifer–atmosphere interactions in general.

4.1. Initial and boundary conditions

The initial state of the atmosphere is determined from the large scale boundary conditions at the start of the run. The single column atmospheric model cannot compute horizontal gradient terms so four NGM analysis points are used to specify these boundary conditions. The NGM analysis data is read every 6 h with input at intermediate times provided through linear interpolation in time [31]. The one-way nesting of the CLASP II model inside the NGM analysis forcing means that CLASP II physics cannot affect the global atmosphere, and thus some inconsistency between the larger scale forcing and the nested model is inevitable

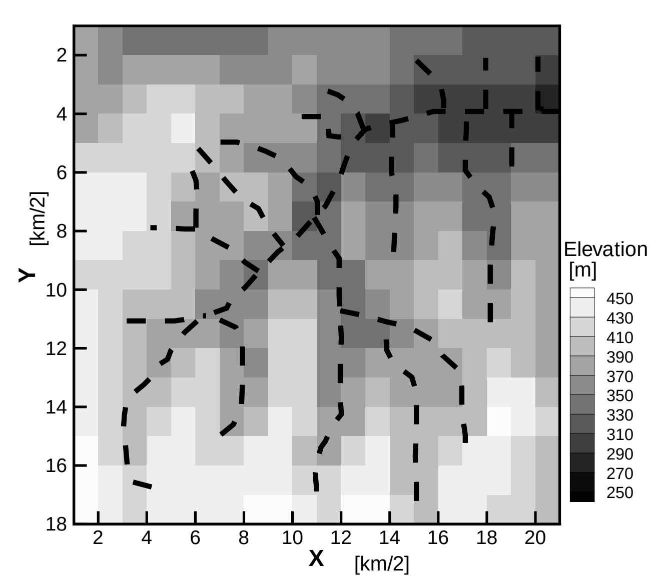

Fig. 5. Topography and stream network in the Mill Creek Watershed.

[20]. Also, the accuracy of CLASP II is limited by the accuracy of the larger scale forcing [30].

For the land surface, soil and vegetation type are specified for each cell. The rooting depth, field capacity and wilting point are functions of soil and vegetation type. Soil density and heat capacity are fixed. The Mill Creek stream network overlain on catchment topography is shown in Fig. 5. For each stream cell, the elevation, conductance, slope, and roughness of the streambed are fixed. The stream has no water in it at the start of the simulation. Streamflow is allowed to build up through baseflow and overland flow. For each aquifer cell, the surface elevation, base elevation, hydraulic conductivity and storage coefficient are fixed. The initial groundwater head is specified at the start of the simulation. A no-flow condition for the groundwater is imposed at the topographic groundwater divide boundaries of the catchment. A constant groundwater head condition is imposed at the Mill Creek outflow, representing the intersection of Mill Creek with the Kansas river.

4.2. Model parameters

Vegetation, soil, and aquifer parameters are specified in Tables 3–5. Properties representative of generalized prairie grass and mixed woodland/crops were assigned to non-stream cells and stream cells, respectively [39,49,67]. It was assumed that each grid cell was completely covered by the vegetation. The average soil texture in the domain was represented as silty–clay–loam soil [11,32,67]. Aquifer properties were chosen based on Kansas river sand–gravel alluvium [22,38] and FIFE Permian limestone–shale [49,60] studies.

Table 3
Vegetation parameters

Vegetation parameter	Vegetation type	
	Grass	Forest/crop
Rooting depth	1.0 m	2.0 m
Roughness length	0.04 m	2.0 m
Coverage	100%	100%
Albedo	0.2	0.1
Leaf area index	1.5 m^2/m^2	9.0 m^2/m^2

Table 4
Soil parameters

Soil parameter	Soil type
	Silty–clay–loam
Field capacity	0.38
Wilting point	0.11
Soil density	1800 kg/m^3
Soil heat capacity	1260 J/kg K

Table 5
Aquifer parameters

Aquifer parameter	Aquifer type
	Single layer
Hydraulic conductivity of aquifer cells	2×10^{-5} m/s
Hydraulic conductivity of stream cells	2×10^{-3} m/s
Specific yield	0.20
Conductivity of streambed	0.020 (tributaries) 0.080 (main stream) m^2/s
Vertical saturated hydraulic conductivity of soil	3.0×10^{-7} m/s
Elevation of base of aquifer	100.0 m

5. Model parameters: Mill Creek catchment with a single layer aquifer

Model parameters were adjusted, within the range of physical bounds, to try to reasonably match available observed data for the years 1987–1989. For the atmosphere, a range of convective and supersaturation precipitation parameters were tested [32] against precipitation observations. The evapotranspiration canopy conductance parameter was adjusted to match available daily FIFE [4] evapotranspiration measurements. For the land surface and aquifer, infiltration capacity of the soil, aquifer conductivity and streambed conductance were adjusted. In modifying these parameters, we were trying to match the observed magnitude and variability of streamflow and annual groundwater fluctuations. The initial water table configuration was arrived at by repeatedly running the model with 1986–1987 (representing average atmospheric conditions) NGM forcing. These spin-up runs provided a base water table map. In the spin-up runs, the initial guess for the water table did not affect the final result, because if the initial water table configuration were far out of equilibrium with the forcing, it would just take longer to achieve a 'steady-state' water table. Imposed on this base water table are seasonal and annual variations, within a few meters.

We did not attempt to reproduce exactly fields such as streamflow and evapotranspiration. Our intent was to study atmosphere–aquifer interactions in a heuristic sense, not reproduce observations. On the other hand, we wanted to utilize climatic forcing and watershed properties/conditions that are physically realistic. In addition, given uncertainty in model boundary forcing [30], observed fields, distribution of precipitation, soil properties and model parameterizations, it would be misleading to attempt detailed calibration. Model testing focused on comparison of observed and simulated monthly–yearly timescale trends of the Kansas catchment. A nine year simulation, from 1985 to 1993 is presented. This period was chosen because of available atmospheric and watershed forcing and validation data. The first few months of 1985 represent a spin-up period for the coupling between model components, and thus some error is introduced into the results for 1985.

6. Results

6.1. Base run

6.1.1. Comparison of simulated results with observations

The annual simulated and observed precipitation trends are similar (Fig. 6). The simulated precipitation tends to be less than observed. This is consistent with the fact that we are comparing a model driven by atmospheric boundary forcing on a scale of about 150 km with a point precipitation measurement (McFarland). Comparison of simulated precipitation with an average of precipitation taken over stations within about 200 km of the model domain, shows slightly better agreement. However, we choose to show the McFarland precipitation because it is the only station actually in the domain and the precipitation at McFarland is closely related to the observed discharge. The observed surface water discharge tracks the simulated trend fairly closely (Fig. 6). The model does not capture extreme high flows in 1986, 1992, and 1993. By examining monthly precipitation and discharge (Fig. 7), shown for the years 1989–1993, we can see that simulated discharge peaks tend to be too low when simulated precipitation is too low. We cannot expect to reproduce the details of precipitation in such a relatively small domain (40 km on a side) with a single column atmospheric model. Our model did not allow for spatial disaggregate of precipitation [32]. Also, in this complex carbonate–shale aquifer, it is difficult to

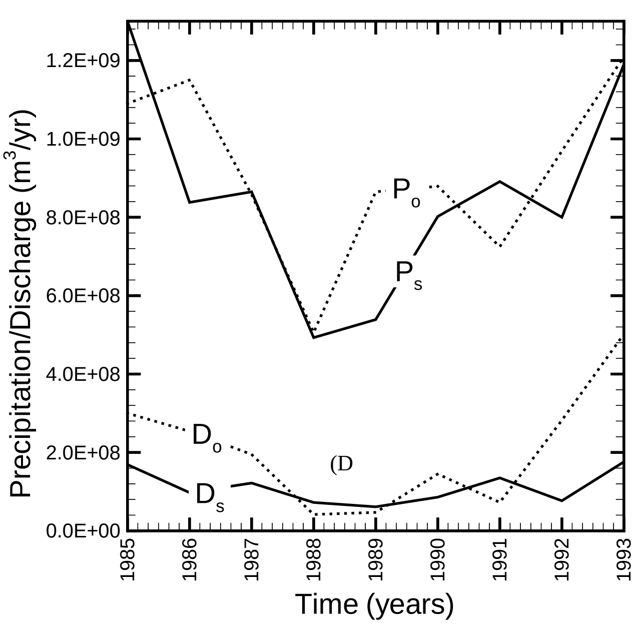

Fig. 6. Simulated and observed annual precipitation (P_s, P_o) and stream discharge (D_s, D_o).

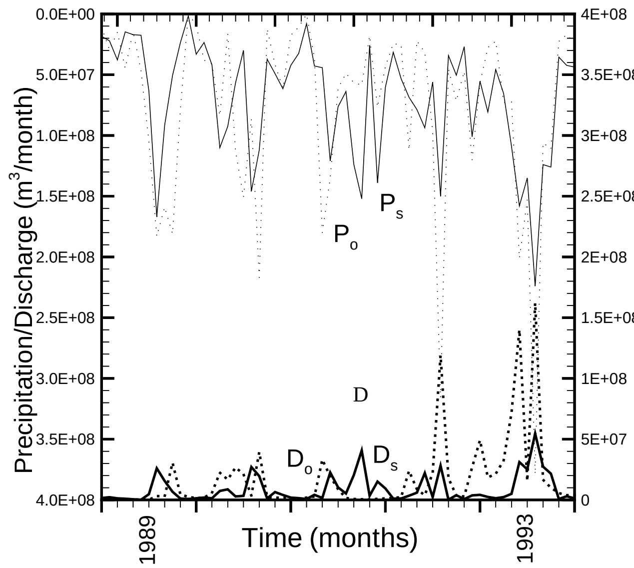

Fig. 7. Simulated and observed monthly precipitation and (P_s, P_o) and stream discharge (D_s, D_o).

match the observed discharge exactly because of potential macropore flow. The model does capture overall yearly trends in precipitation and stream discharge, even in a small domain with complex geology.

Comparison of FIFE average soil moisture over a 1 m column [4] with the simulated soil moisture from sample stream and upland cells (Fig. 8(a)) shows agreement in simulated and observed variability in 1987. The FIFE soil moisture is only available for certain portions of 1987–1989. In 1987, the FIFE soil moisture is bracketed by the wetter stream cell and the drier upland cell (Fig. 8(a)). The 1988 FIFE soil moisture seems high given that 1988 was a dry year (Fig. 6) and this may represent measurement error or localized soil moisture anomalies. The tendency toward lower soil moisture than FIFE is present in soil moisture simulated in all upland cells of our model. This discrepancy may be a result of the simple soil model allowing the soil moisture to drop too low in dry periods, as is common in some LSM [11]. The model captures the annual variability in the soil moisture reservoir in response to atmospheric forcing (Fig. 8(b)) which are similar to the FIFE measurements. The observed FIFE evapotranspiration [4] is slightly higher than the simulated domain average evapotranspiration (Fig. 9). This is consistent with the drier simulated soil moisture. Some of the difference between simulated and FIFE soil moisture and evapotranspiration may be caused by the fact that the model soil moisture and evapotranspiration are averaged over a larger area than the FIFE observations. The simulated and observed evapotranspiration show the same type of daily variability in magnitude. Overall, the model soil moisture reservoir and evapotranspiration flux are physically reasonable but a more detailed soil moisture

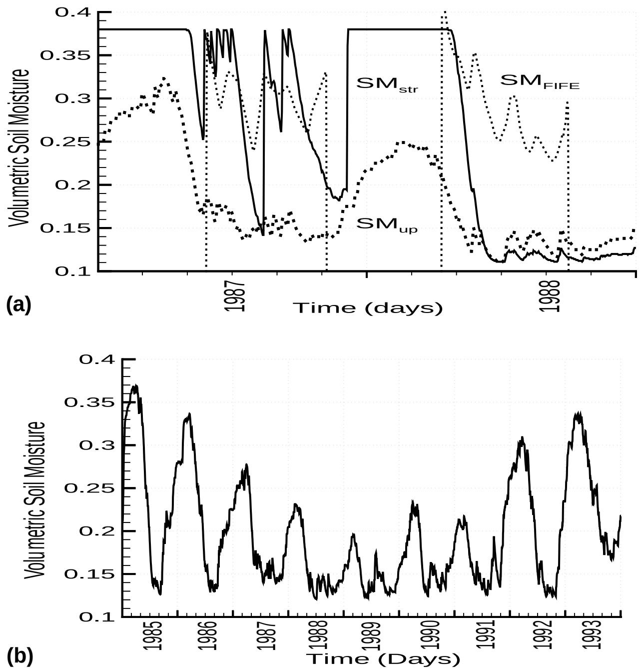

Fig. 8. Volumetric daily soil moisture: (a) simulated for an upland cell (SM_{up}) and a stream cell at Alma (SM_{str}) and FIFE measured (SMFIFE); (b) simulated domain average. FIFE measurements are only available over a portion of the simulation period.

model might be better able to reproduce observations, particularly during dry periods.

The daily groundwater head (Fig. 10) at a stream cell near Alma (Fig. 4) and an upland cell (6 km south of Alma) show, superimposed on the annual cycle, a decline in head up to 1989–1990 and then a rise. This trend tracks the annual precipitation. This shallow inter-annual variation in groundwater levels does not reach steady state but rather evolves according to initial conditions and annual changes in groundwater storage. Annually, at the stream cell, the head oscillates between about 0.5–2.0 m. This is consistent with Macpherson's [49] observations of annual variability of 1–2 m in the limestone–shale aquifers of the KPRNA (Fig. 4). On an annual cycle, the groundwater head in the stream cell increases in late winter and early spring and declines in summer. This cycle is controlled by the precipitation and evapotranspiration. The root zone at Alma is between 308 and 310 m above sea level and the water table tends to be within the root zone in spring and winter and then falls below the root zone during summer and fall. The evapotranspiration is higher when the water table is in the root zone. In the upland cell, the water table stays below the root zone during the entire simulation. The upland cell acts as a source term for lowland water during the dry period.

We have very limited groundwater head observations for Paxico and Alma [40]. The observations are one-time measurements made upon completion of wells in 1995–1996. These measurements fall outside the model simulation period. Thus, we cannot directly compare these point measurements to the simulated head in the model 2×2 km cells for the period 1985–1993. However, the observed groundwater heads serve as a check that our heads are within reasonable bounds. Table 6 is a com-

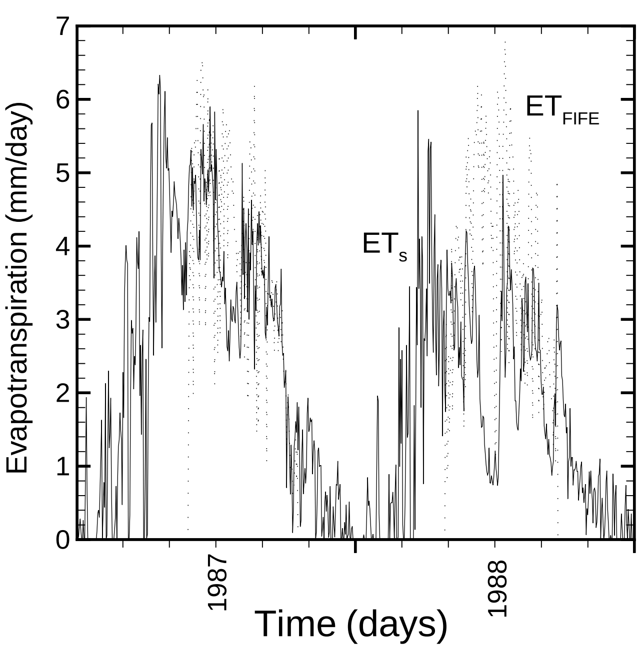

Fig. 9. Average simulated daily evapotranspiration (ET_s) and FIFE evapotranspiration (ET_{FIFE}). FIFE measurements are only available over a portion of the simulation period.

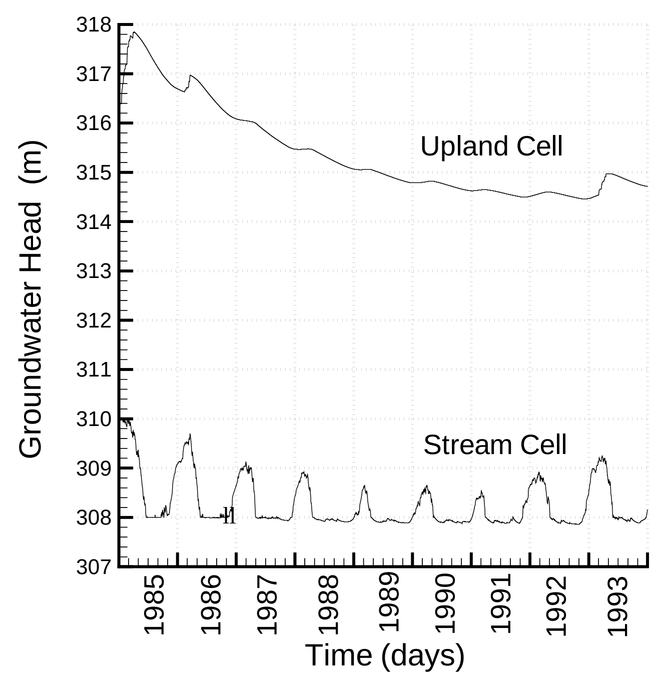

Fig. 10. Simulated daily groundwater head (meters above sea level) at a stream cell (Alma) and an upland cell (about 6 km south of Alma).

Table 6
Groundwater levels (depth below surface)

	Depth to water (m)	
	Simulated (1993)	Measured
Paxico, well 1	6.9	4.9 (1995)
Alma, well 1	17.8	12.2 (1996)
Alma, well 2	8.9	6.1 (1996)

parison of simulated and observed groundwater heads at Paxico and Alma. The simulated heads were obtained by selecting the MODFLOW cell containing the measurement location for December, 1993. The observed and simulated depths to water agree within 45%, which is acceptable given the complex limestone–shale aquifer structure and the thickness of the aquifer (150–350 m) relative to water level changes. Calculated water levels were consistently lower than those observed (between 2 and 5.6 m). If these measurements are representative of hydrologic conditions within the watershed during the period of model calibration, then the hydraulic conductivity of the limestone aquifer would need to be lowered to produce better agreement with this data.

6.1.2. Model budget and seasonal cycle

In the dry years (1988–1989) of the nine year simulation period, evapotranspiration exceeds rainfall (Fig. 11). The evapotranspiration (ET_s) and stream discharge (D_s) track the precipitation (P_s) trend. The changes in soil moisture (ΔS_{sm}) increase to a peak in 1991, when the reservoir refills after the dry years of 1988–1989. The maximum amount of change in aquifer storage (ΔS_{gw})

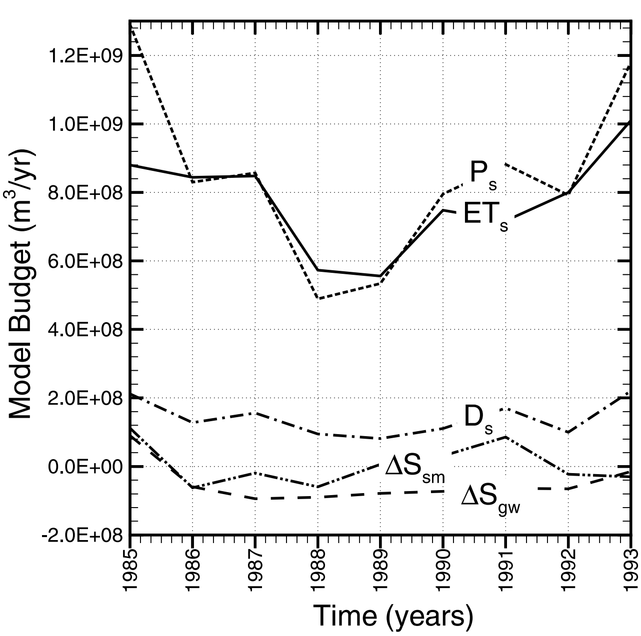

Fig. 11. Simulated annual model budget showing precipitation (P_s), evapotranspiration (ET_s), stream discharge (D_s), change in soil moisture (ΔS_{sm}), and change in aquifer storage ΔS_{gw}.

occurred during 1987–1989 (change in storage is mostly negative). With successively wetter years, the change in aquifer storage becomes almost zero in 1993.

The plan view seasonal cycle of simulated evapotranspiration for the sample year 1990 (Fig. 12) illustrates the spatial heterogeneity of model fluxes. The evapotranspiration is high over the domain in spring. In summer, the evapotranspiration decreases in the uplands but is maintained at a higher rate in the stream network. The evapotranspiration decreases in fall but enhanced evapotranspiration in the stream network is still evident. The evapotranspiration is maintained at a higher rate in the stream cells because the deeper rooted trees access more moisture than the upland grassland cells. More importantly, the highest evapotranspiration occurs in the stream cells because the shallow water table encroaches the root zone. An examination of water table depth for spring and summer 1990 indicates that the water table is in the root zone in only about 5% of the domain. The number of cells where the water table is in the root zone drops from 13 in spring to 11 in summer (Fig. 13). The seasonal variability of the position of the water table relative to the root zone and the fact that the water table is not within the root zone in all stream cells, demonstrates the need for a physically based groundwater model. Comparison of evapotranspiration between upland and lowland cells (Fig. 14) suggests how important groundwater supported evapotranspiration can be. Evapotranspiration in cells which the water table encroach the root zone is up to three times higher than upland cells with deep water tables during summer months.

6.2. Comparison of simulated results with and without an aquifer

Fig. 15 compares precipitation, evapotranspiration and stream discharge for model runs with and without an aquifer. Fig. 16 shows the differences in precipitation, evapotranspiration, stream discharge, and change in aquifer storage between the "No Aquifer" and "Aquifer" simulations (Aquifer minus No Aquifer). The annual precipitation in the 'No Aquifer' run is slightly less than the 'Aquifer' run in the wetter years (1986, 1991–1993). The evapotranspiration is lower for the 'No Aquifer' run, indicating that groundwater evapotranspiration makes a significant contribution. The percent change in evapotranspiration for the 'Aquifer' verses 'No Aquifer' runs ranges from about 20% in 1988 (dry year) to about 5% in 1993 (wet year). Thus, the groundwater supported fraction of evapotranspiration is larger during drier years. This is supported by the fact that the greatest amount of water is drawn from the aquifer (change in storage most negative) in the dry years of 1987–1989. For the 'No Aquifer' run, the change in aquifer storage is zero and thus this reservoir

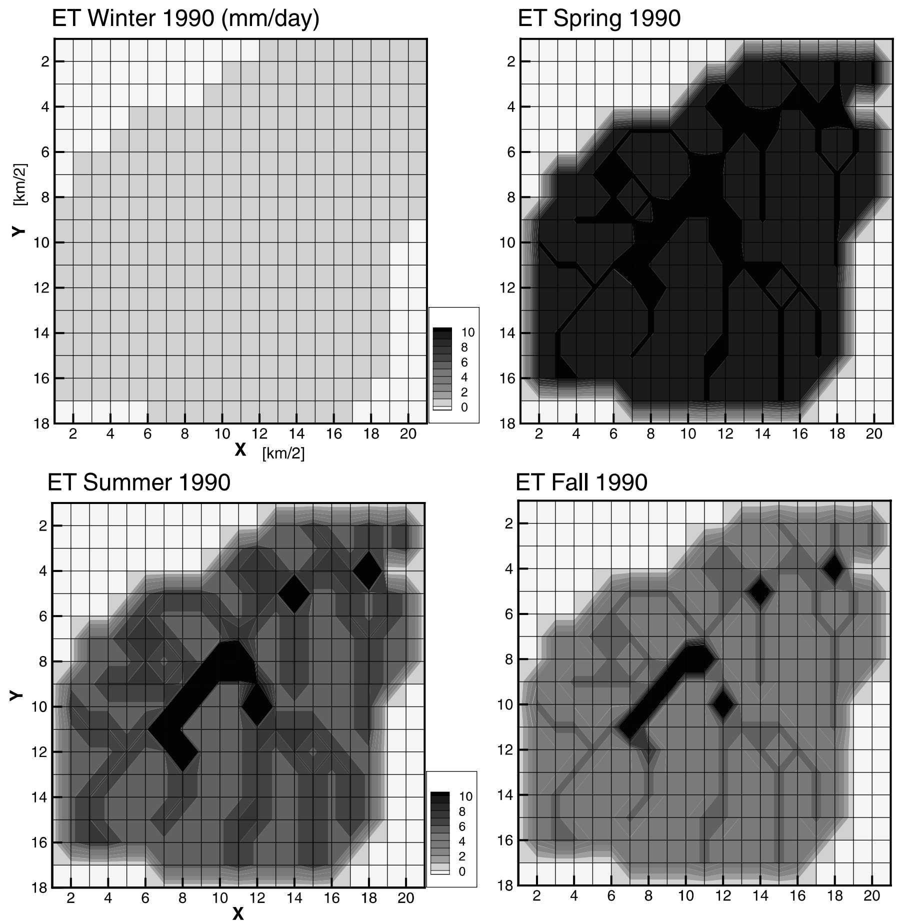

Fig. 12. Plan view simulated seasonal evapotranspiration for 1990.

of water is unavailable. The streamflow in the 'No Aquifer' run (Fig. 15(b)) is slightly less than in the 'Aquifer Run' because of the differences in precipitation and the fact that saturation overland flow is not available in the 'No Aquifer' run.

6.3. Results of 40 year drought sensitivity study

We constructed a longer term (40-year) simulation by recycling 1989 atmospheric forcing. This scenario is meant to represent drought conditions similar to the 1930 dust bowl conditions in this area. A west–east water table cross-section through row 10 of the domain at the end of forty years of drought (Fig. 17) shows the drop in the water table. The initial water table represents dynamic equilibrium conditions for the result from 1986–1993 climatological forcing. This water table clearly does not obey the common assumption that the hydraulic gradient can be approximated by the slope of the topography e.g. [73]. At the end of the forty year drought, the decline in the water table ranges from about 15 m in the uplands to no decline in one lowland area. The water table flattens as water flows from the uplands to the lowlands during the drought. In one portion of the cross-section, the water table moves out of the root zone. By the end of the 40-year simulation, the water table has not yet fully equilibrated to the new climatic forcing suggesting that an even longer simulation period is needed for water table levels to stabilize.

7. Discussion

In this work we have started to explore interactions between aquifer systems and the atmosphere. We have demonstrated that "groundwater supported" evapotranspiration can contribute up to 20% of the annual amounts during periods of drought. We argue here that aquifers have their own characteristic response time to changes in climatic forcing that can be on the order of hundreds of years. Calibrated LSMs with more sophisticated soil-vegetation zones could also capture many of the observed fields, such as evapotranspiration and soil moisture, perhaps with greater accuracy than CLASP II. There are also more sophisticated groundwater–surface water models available, that would represent fields such

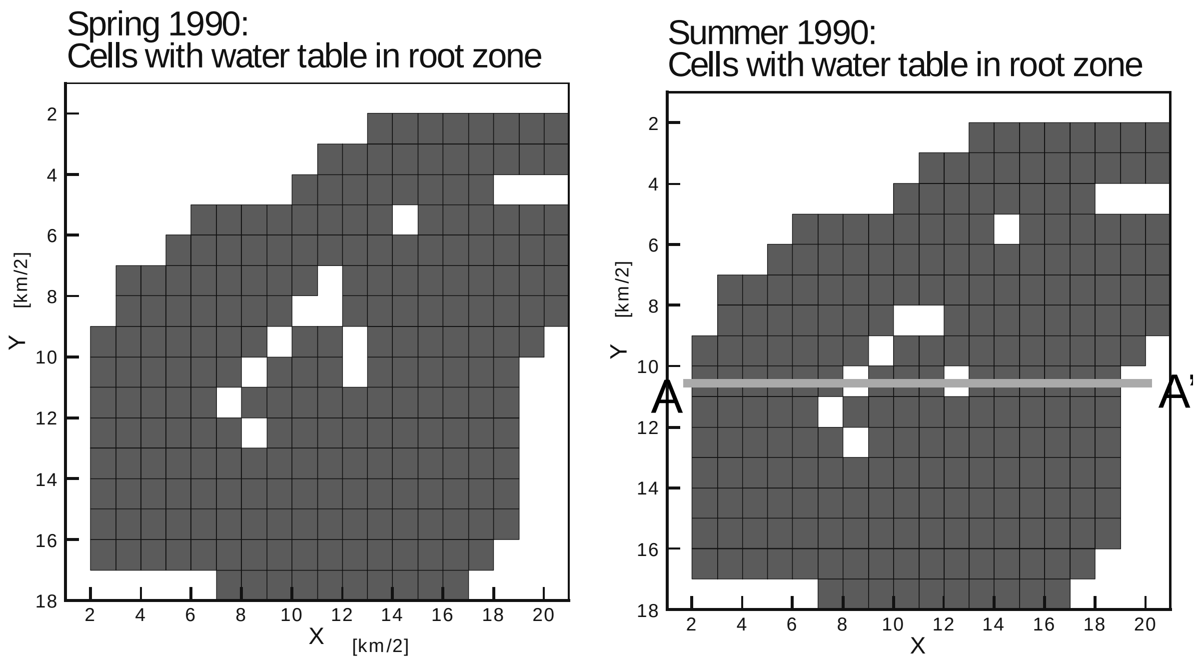

Fig. 13. Plan view seasonal watertable position for 1990. Within the watershed, the white cells indicate areas where the water table is within the root zone. The location of the cross-section A–A' presented in Fig. 17 is also listed.

as river discharge better. The CLASP II model takes a different approach, designed to address water table–atmosphere interactions. The CLASP II model is a compromise between complexity of model components and computational practicality. We chose to simplify the model in some respects in order to be able to carry out decadal-timescale simulations on a watershed spatial scale.

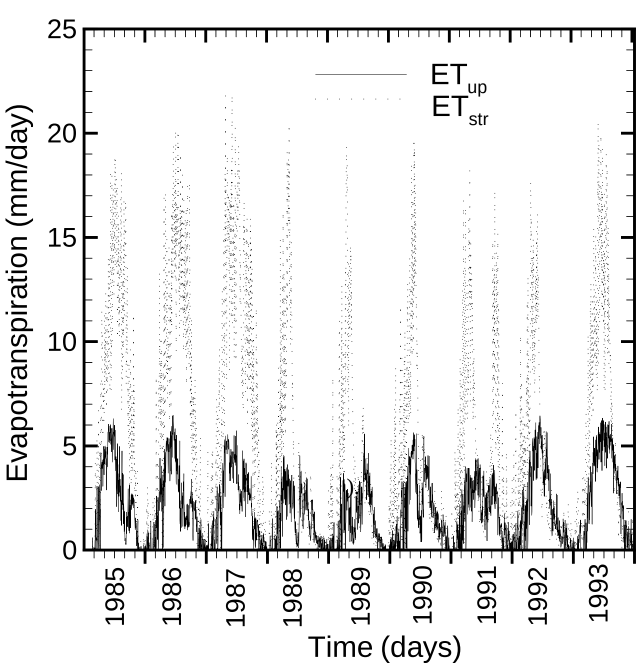

Fig. 14. Simulated daily evapotranspiration for an upland cell (ET_{up}) and the stream cell at Alma (ET_{str}).

We believe that it may soon be computationally feasible to incorporate a physically based groundwater model into GCMs land surface parameterization schemes using supercomputers. This is because hydrologic processes in individual watersheds can be run on individual CPUs within a massively parallel supercomputer. For the Mill Creek Watershed, the portion of CPU time on a conventional UNIX workstation (Silicon Graphics™ Origin 200) taken up by the groundwater model (not including the soil-vegetation zone) is about 20%. In experiments with a larger watershed, having approximately ten times as many cells as the Mill Creek domain, this figure rises to about 40% of CPU time taken up by the groundwater model. A more computationally practical watershed based approach to land surface modeling in GCMs is the topographic index formulation [24,64]. However, our results argue for additional field and quantitative studies to better constrain feedbacks between aquifer and atmospheric water reservoirs. Our work could be useful in improving simplified representation of aquifers in GCMs. Furthermore, even if a GCM does not have a physically based groundwater model, we can gain valuable insight by nesting models like CLASP II within GCM output. This downscaling of GCM output to the watershed scale, utilizing a 3-D groundwater model, allows us to examine water resources implications of global climate change.

As noted by [72], there is a paucity of information on the potential effects of greenhouse warming on regional

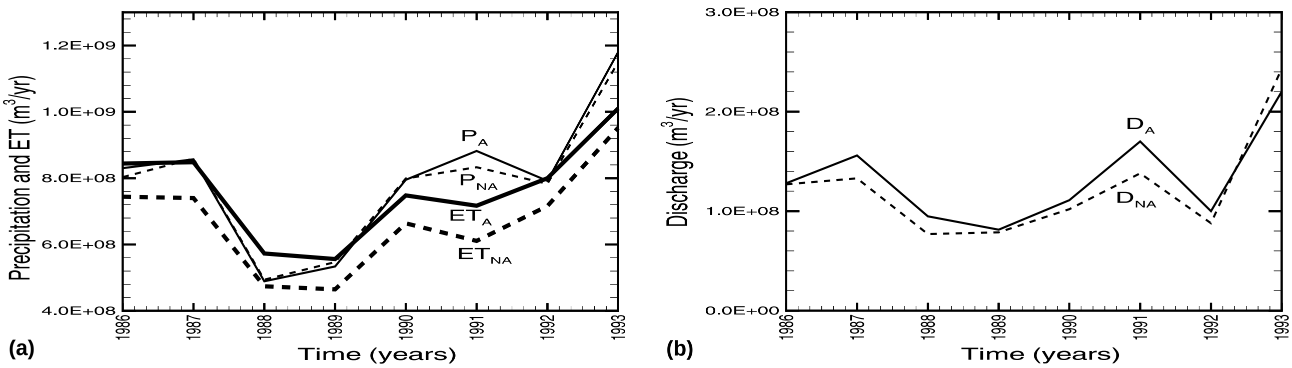

Fig. 15. A comparison of simulated annual: (a) precipitation (P_A, P_{NA}), evapotranspiration (ET_A, ET_{NA}); (b) stream discharge (D_A, D_{NA}) for 'Aquifer' and 'No Aquifer' runs.

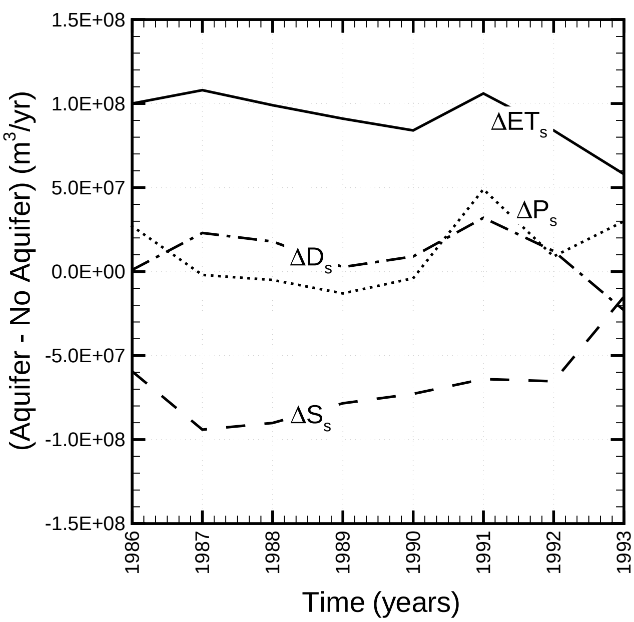

Fig. 16. The difference between simulated annual precipitation (P_s), evapotranspiration (ET_s), stream discharge (D_s), and aquifer storage (S_s) for 'Aquifer' and 'No Aquifer' runs.

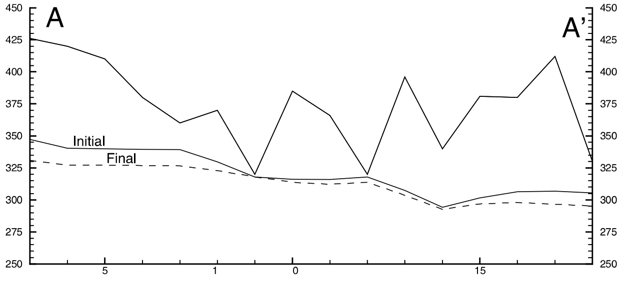

Fig. 17. Comparison of water table at start and end of 40-year "drought simulation across an east–west cross-section through the Mill Creek watershed, Kansas. The location of the cross-section line is shown in Fig. 13.

groundwater flow. Future work will consider in what type of geologic settings groundwater interacts with the atmosphere and how this interaction affects water resources. The Mill Creek, Kansas watershed has been an observation-rich test case for developing CLASP II. However, it is not ideally suited to the study of aquifer–atmosphere interactions because of the complex geology of the aquifer, the relatively deep water table over much of the catchment and the lack of lakes. We expect that in glaciated watersheds, where the water table is shallow and the topography is hummocky, we will see even stronger interaction between the atmosphere and the aquifer.

8. Conclusions

A coupled atmosphere–surface water–groundwater hydrologic model (CLASP II) has been developed and applied to the Mill Creek catchment in northeastern Kansas, using a nine year historical record. The model is designed to operate on a decadal timescale and watershed spatial scale (10^2–10^5 km^2). The physically based groundwater model included in CLASP II allows us to study the temporal and spatial response of groundwater levels to climatic forcing. Through comparison with available observations, we have demonstrated that the CLASP II model captures monthly and yearly hydrologic trends in streamflow, aquifer levels, evapotranspiration and soil moisture. Aquifer levels were found to affect evapotranspiration. The differences between evapotranspiration in model runs with and without an aquifer represented were substantial indicating the potential importance of groundwater supported evapotranspiration. Annually, simulation results indicated that from 5% (wet year) to 20% (dry year) of evapotranspiration was drawn from the aquifer. Significant simulated changes in aquifer storage, soil moisture, and evapotranspiration between wet and dry years point to the linked nature of the atmosphere–surface water-groundwater system. Our results suggest that seasonal and inter-annual feedbacks between water levels and atmospheric forcing exist. Under extended drought conditions, only a physically based groundwater model can capture the behavior of the water table and resultant influence on land-atmosphere moisture feedback. A long-term (40-year) "dust-bowl" simulation was run in which drought conditions were represented by recycling 1988 climatic forcing. During this simulation, the water table declined by over 15 m and was observed to move out of some lowland area root zones. However, even a longer simulation period is necessary (perhaps as much as 200 years) for the water table to equilibrate to this extreme climatic forcing. This illustrates the disparate response times that aquifers and atmospheres have.

Results from this study suggest that it may soon be feasible to incorporate physically based representations of aquifers into land surface parameterization schemes of GCMs. For the Mill Creek catchment, the portion of CPU time taken up by the groundwater model was about 20%, on a conventional Unix workstation. This percentage rises with increasing watershed size or finer lateral discretization. While it is not presently computationally practical to include a physically based groundwater model in a GCM, it is certainly reasonable to incorporate a 3-D groundwater model in regional studies.

Acknowledgements

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