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Hydrologic modeling of an arctic tundra watershed: Toward Pan-Arctic predictions

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Abstract. A simple land surface model is used to explore the dynamics of the hydrologic cycle operating in arctic tundra regions. The model accounts for the topographic control of surface hydrology, ground thermal processes, and snow physics. The approach described relies only on the statistics of the topography rather than the details of the topography and is therefore computationally inexpensive and compatible with the large spatial scales of today's climate models. As such, the model can easily be applied on an arctic-wide basis to explore issues ranging from the delivery of seasonal melt water to the Arctic Ocean to impacts of climate change on the hydrologic cycle.

1. Introduction

Global climate models predict that warming associated with rising levels of "greenhouse" gases such as carbon dioxide in the atmosphere will have significant effects in the Arctic during the coming century [*Maxwell*, 1992]. General circulation model (GCM) experiments simulate a maximum annual warming at high northern latitudes, associated with increased precipitation and earlier snowmelt [*Intergovernmental Panel on Climate Change*, 1995]. With this in mind, researchers are attempting to answer a number of key questions: If arctic rivers deliver less freshwater to the Arctic Ocean owing to enhanced evapotranspiration, what will be the impact on river ecology, ocean shelf dynamics, surface ocean stability, and sea ice formation? Will changes in snow cover extent and amount affect regional and global climate via changes in the surface energy balance? Will climate change augment plant growth and thus increase the uptake of CO_2 from the atmosphere? If soils become warmer, will the increased microbial activity release carbon stored in the soil? Will warmer temperatures increase the production of methane, another greenhouse gas, in regions where wetlands expand?

The uncertainty in predicting how the Arctic may be impacted by climate-change revolves around several issues related to terrestrial temperatures and surface hydrology. For example, to understand how climate change will affect the carbon balance of arctic regions requires an understanding of the relative effects on plant production (carbon storage) and microbial decomposition in soils (carbon loss). The net balance between these two processes will determine whether the Arctic is a source or sink of carbon dioxide to the atmosphere. This balance is influenced by a number of factors [*Shaver et al.*, 1992] but is strongly related to soil moisture [*Billings et al.*, 1982; *Johnson et al.*, 1996; *Oechel et al.*, 1998] and soil temperatures [*Flanagan and Veum*, 1974; *Oechel et al.*, 1998; *Parton et al.*, 1987].

The tundra landscape is unique. Infiltration and soil water movement is confined to shallow zones by permafrost, and the seasonal regimes of rivers, soil moisture, and surface-groundwater interactions are dominated by freeze-thaw processes [*Dingman*, 1970; *Hinzman et al.*, 1991; *McNamara et al.*, 1997; *Roulet and Woo*, 1988]. In contrast to the temperate regions, these shallow zones limit the baseflow typically supplied by deeper groundwater, and thus most of the water flows through a narrow zone in contact with plant roots and soil organic matter rather than through deep zones of mineral soil. This typically results in a flashier hydrograph than seen in temperate areas [*Haugen et al.*, 1982]. Finally, arctic regions are snow covered for 8–10 months of the year, making an accurate representation of snow processes critical to a successful hydrologic model.

For all the recent improvements in GCM land surface models, most are still inadequate for exploring the above mentioned questions. Originally designed for midlatitudes, most do not adequately represent either snow physics or permafrost dynamics. Furthermore, no GCM land surface model to date adequately represents the topographic control over surface hydrology. Current models still use the one-dimensional soil column as the fundamental hydrologic unit. This view, though perhaps effective in simulating such processes as the evolution of ground temperatures and the growth and ablation of snowpack at the soil plot scale, unfortunately ignores the role topography plays in the development of soil moisture heterogeneity and the critical (perhaps overwhelming) impacts of this heterogeneity on surface water and energy fluxes. This "flat earth" approach to modeling land surface physics has persisted largely because of computational constraints; the distributed hydrological models that have commonly been used to treat

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topographical variations within basins are much too expensive for any GCM effort. Some researchers [*Famiglietti and Wood*, 1994a, 1994b; *Stieglitz et al*., 1997] have recently suggested that this deficiency can be overcome through the use of a recent class of model, "Topmodel" [*Beven et al*., 1994; *Beven*, 1986a, 1986b; *Beven and Kirkby*, 1979] that defines the watershed to be the fundamental hydrologic unit and treats topography-induced soil moisture heterogeneity within the watershed statistically at little computational cost. *Stieglitz et al*. [1997] used the analytic form of the Topmodel equations to produce consistent predictions of baseflow from a watershed and the saturated fraction within it (partial contributing area), the latter having a direct impact on calculated evapotranspiration and surface runoff. By treating the effects of soil moisture variations on the surface fluxes explicitly, this Topmodel approach provides a significant conceptual improvement over current, GCM soil column models.

2. Background

2.1. Snow

While sophisticated multilayer snow models have been developed and successfully applied [*Davis et al*., 1995; *Hardy et al*., 1998; *Jordan*, 1995], the treatment of snow processes, especially those used within GCMs, have been relatively simple. Some models consider the winter snow pack only as a store of soil moisture [*Abramopoulos et al*., 1988; *Bonan*, 1996; *Koster and Suarez*, 1996], while others blur the distinction between the snow and the ground surface altogether by envisioning a composite soil and snow layer [*Dickinson et al*., 1993; *Pitman et al*., 1991]. Still others [*Slater et al*., 1998; *Verseghy*, 1991] do distinguish between separate snow and ground layers, yet represent the entire pack with a single snow layer regardless of the actual pack depth. A good review of the numerous snow models used in GCMs is found in the work of *Foster et al*. [1996]. However, most of these simple schemes have considerable flaws [*Lynch-Stieglitz*, 1994]. A. K. Betts et al. (Evaluation of the land-surface interaction in the ECMWF and NCEP/NCAR reanalysis models over grassland (FIFE) and the boreal forest (BOREAS), submitted to *Journal of Geophysical Research*, 1999) has recently shown that an insufficient representation of snow and soil processes at high latitudes leads directly to a poor evolution of the atmospheric boundary layer in weather forecasting models. Ultimately, the ability to simulate the growth and ablation of the winter pack and capture its insulating capabilities is a prerequisite to obtain a proper simulation of the timing of ground freeze-thaw processes and the evolution of the summer active soil layer. This insulation capability is most clearly seen in arctic watersheds where the difference between ground and air temperatures can be of the order of 30°–40°C. Recently, sophisticated snow treatments have been included in land surface models and demonstrate a clear improvement in the overall simulation of the hydrologic cycle, including ground freeze and thaw processes. These multilayer snow schemes [*Loth et al*., 1993; *Lynch-Stieglitz*, 1994; *Yang et al*., 1997] are now either being included or have been included in GCMs [*Loth and Graf*, 1998a; *Loth and Graf*, 1998b; *Yang et al*., 1997].

2.2. Thermal processes

To represent ground thermal processes numerous models [*Koster and Suarez*, 1996; *Pitman and Desborough*, 1996; *Yang and Dickinson*, 1996] have employed force-restore methods [*Deardorff*, 1978]. However, as noted by *Dickinson*, [1988], the force-restore formalism needs considerable modification to account for heterogeneity in soils and in snow cover. As such, many groups are moving away from force-restore formalism and toward multilayer ground schemes in which heat transport is physically modeled via diffusion along the thermal gradient [*Abramopoulos et al*., 1988; *Bonan*, 1996; *Stieglitz et al*., 1997].

2.3. Soil Moisture

While the spatial distribution of soil moisture within a watershed is far from uniform, it is somewhat predictable. Lowlands tend to be zones of convergent flow and therefore zones of high soil moisture content. Upland soils tend to be progressively drier. Because of this topographic control of surface hydrology [*Beven and Kirkby*, 1979; *Burt and Butcher*, 1985], the expansion and contraction of lowland saturated zones is predominantly determined by down slope redistribution of recharged subsurface soil water (i.e., infiltration minus evapotranspiration). When upland recharge is large, baseflow increases and lowland saturated zones expand. When recharge is small, baseflow decreases and lowland saturated zones contract. These saturated regions, in valleys and along hillslopes, represent the surface expression of the water table, and in sum, constitute the bulk of the saturated fraction of the watershed. In this way the topographic control of surface hydrology can impact such processes as sensible and latent fluxes to the atmosphere. For example, while evapotranspiration may be near the potential rates in saturated or near-saturated regions (valleys), it may be negligible in regions with large moisture deficit (uplands).

One approach to modeling the spatial distribution of soil moisture is to work with the details of the watershed topography by explicitly modeling the movement of water from the hillslopes to the valleys [*Hinzman and Kane*, 1992; *Zhang et al*., 1997]. However, while this may be satisfactory at the scales of very small watersheds, it is computationally incompatible with the scales required by climate models [*Wood et al*., 1990].

Another approach is to use empirically based models that require extensive calibration with historical discharge data [*Bowling and Lettenmaier*, 1998; *Liang et al*., 1994]. While these models have demonstrated the ability to simulate monthly discharge, they may not be adequate for simulating the spatial and temporal patterns of soil moisture and the subsequent surface water and energy fluxes resulting from this soil moisture heterogeneity on timescales relevant to today's GCMs. The reason for this is that the depth of the hydrologically active soil column is anything but static during the short summer months, and therefore a unique calibration may be impractical. Calibration of empirical models is further complicated by the fact that in a typical climate warming experiment we might expect the hydrologically active layer to deepen substantially.

We combine two methods of modeling the flow of water within a watershed. The first makes use of a soil column model framework traditionally employed in most GCMs in which the vertical movement of water and heat within the soil and between the soil surface plus vegetation to the atmosphere is simulated. This generalized land surface model is capable of operating in a wide variety of climatic regimes. The second method makes use of the statistics of the topography and allows us to track the horizontal movement of shallow groundwater from the uplands to the lowlands (a TOPMODEL or topographic approach). By combining these two approaches, we can produce a three-dimensional picture of soil moisture

distribution within a watershed without the need to explicitly model the landscape. Since the approach relies only on the statistics of the topography rather than the details of the topography, it is computationally inexpensive and compatible with the large spatial scales of today's climate models. While a brief description of this model follows, a complete description and validation at a New England watershed can be found in the work of *Stieglitz et al*. [1997].

3. Model Description

3.1. Application of Topmodel Equations

TOPMODEL formulations permit dynamically consistent calculations of both the partial contributing area and the baseflow that supports this area from knowledge of the mean depth of the water table and a probability density function (pdf) of the soil moisture deficit derived from topography statistics. At the heart of Topmodel are three basic assumptions: (1) the water table is nearly parallel to the soil surface so that the local hydraulic gradient is close to tan β, where β is the local slope angle; (2) the saturated hydraulic conductivity falls off exponentially with depth; and (3) the water table is recharged at a spatially uniform, steady rate that is slow enough, relative to the response timescale of the watershed, to allow the assumption of a water table distribution that is always at equilibrium. Given these three assumptions, an analytic relation can be derived between the mean water table depth ($\bar{z}$) within the watershed and local water table depth at any location $x(z_x)$ [*Sivapalan et al*., 1987; *Wood et al*., 1990]

$$z_x = \bar{z} - 1/f[\ln(a/\tan\beta)_x - \lambda]. \tag{1}$$

The term $\ln(a/\tan\beta)_x$ is defined to be the topographic index, χ, the ratio of the upslope dainage area a, to the local slope at that point, tan β. The mean watershed value of $\ln(a/\tan\beta)$ is λ, and the rate of decline of the saturated hydraulic conductivity is described by f. An immediate consequence of (1) can be seen by setting z_x equal to zero, i.e., locating the local water table depth at the surface. All locations associated with values of the topographic index χ greater than $\lambda + f\bar{z}$ are situated within saturated regions. Finally, following *Sivapalan et al*. [1987], baseflow (Q_b) is

$$Q_b = \frac{AK_s(z = 0)}{f} e^{-\lambda} e^{-f\bar{z}} \tag{2}$$

Equations (1) and (2) comprise the basic formulations that govern saturated flow and the distribution of the water table with respect to the topography. From knowledge of $\bar{z}$ and the cumulative pdf of the topographic index χ (obtained from Digital Elevation Models (DEMs)), the partial contributing area, and the baseflow consistent with this area can be calculated. For a complete description of Topmodel, see *Beven and Kirkby* [1979] and *Beven* [1986a, 1986b].

We incorporate the analytic form of the Topmodel equations with a single column land surface model that tracks the mean state of the watershed. The ground scheme consists of ten soil layers. Diffusion and a modified tipping bucket model govern heat and water flow, respectively. The prognostic variables, heat and water content, are updated each time step. In turn the fraction of ice and temperature of a layer may be determined from these variables. A three-layer snow model [*Lynch-Stieglitz*, 1994] has been incorporated into a modified BEST vegetation scheme [*Pitman et al*., 1991]. Radiation and atmospheric conditions determine the surface energy balance. Topmodel equations and DEM data are used to generate baseflow, which supports lowland saturated zones. Soil moisture heterogeneity represented by saturated lowlands (predicted by Topmodel equations) subsequently impacts watershed evapotranspiration, the partitioning of surface fluxes, and the development of the storm hydrograph. The model is fully described by *Stieglitz et al*. [1997]. Figure 1 depicts the current application of Topmodel equations. This approach to modeling the land surface has now been validated at several watersheds; ranging in scale from the Red Arkansas Basin (570,000 km^2 [*Ducharne et al*., 1998]), to the Sleepers River watershed (Vermont, 8.4 km^2 [*Stieglitz et al*., 1997]), to the Imnavait Creek watershed presented here (2.2 km^2).

The following changes have been implemented to deal with arctic tundra regions. In vegetated regions, model layer 1 now represents a litter layer. Further, in regions with tundra vegetation, model layer 1 represents a combined litter and moss layer, which allows for enhanced evapotranspiration. In either case, roots no longer extract water from this layer. Finally, the model now allows the inclusion of organic peat layers. Otherwise, all thermal and hydraulic processes operate exactly as before.

Recently, Topmodel [*Beven et al*., 1994] has been used at Imnavait Creek. *Ostendorf* [1996] coupled Gas-Flux [*Tenhumem et al*., 1994], a model used to simulate net ecosystem productivity, with Topmodel, in order to explore the net summer CO_2 exchange, runoff, and evapotranspiration, at Imnavait Creek. However, by neglecting any representation of thermal processes, specifically, the prognostic calculation of soil temperatures that are used to drive microbial respiration, the applicability of this approach is limited to experimental watersheds where measured soil and moss temperatures are continuously available. In addition, because snow cover is not simulated, simulations are valid only during the short growing season. However, as recently demonstrated, winter microbial respiration may account for a significant fraction of the annual soil respiration [*Oechel et al*., 1997; *Zimov et al*., 1996]. Finally, if the purpose is to incorporate such biophysical models within the confines of regional and global climate models, the dicretization of the probability density function commonly employed for most Topmodel applications is still relatively expensive from a computational point of view (see section 3.b of *Stieglitz et al*. [1997]).

4. Model Validation

4.1. The Imnavait Creek Watershed

This section summarizes extensive discussions on the Imnavait Creek watershed given by *Hinzman et al*. [1996, 1991], *Kane et al*. [1991], and *McNamara et al*. [1997].

The Kuparuk River (the basin is 8140 km^2) has its headwaters in the Brooks Range and drains through the coastal plains of Northern Alaska to the Arctic Ocean. Nested within the headwaters of the Kuparuk Basin at 68° N is the Imnavait Creek sub-watershed (2.2 km^2). Imnavait Creek is located in rolling piedmont hills where the predominant soils are 15–20 cm of porous organic peat underlain by silt and glacial till [*Hinzman et al*., 1991]. The topographic sequence of land cover ranges from wet sedge in the riparian zones to tussock tundra along the midslopes and dry heath near the ridge tops. Water tracks, regions of enhanced soil moisture that run perpendicular to the hillslope at a spacing of ~10 m, channel flow down

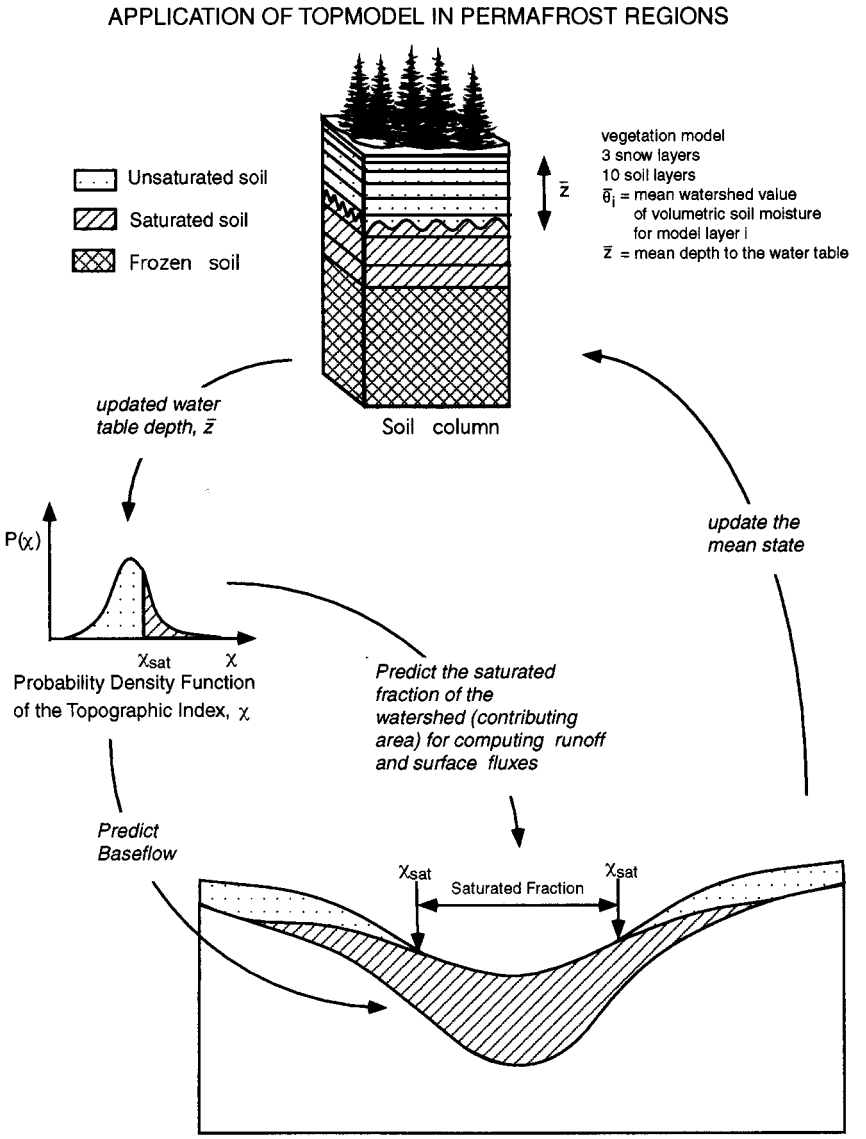

Figure 1. Coupling the analytic form of Topmodel equations with the single column model. From an update of the mean water table depth ($\bar{z}$), Topmodel equations and Digital Elevations Model data are used to generate baseflow (equation (2)) and the partial contributing area.

the slope. From 1985 through 1993 the mean annual precipitation, maximum snow water equivalent content, and air temperature averaged 34 cm, 11.53 cm, and −7.4°C, respectively, and 66% of the annual precipitation fell during the short summer season. Snowmelt accounted for 47% of annual discharge while runoff and evapotranspiration were 46% and 54% of the water budget. In this region of continuous and permanent permafrost, snowfall is possible on any given day; however, the snow season begins in earnest in September when ground freezing begins. Spring melt is in late May or early June. The maximum thaw depth in summer ranges from 25 to 100 cm depending on vegetation, aspect, slope, and soils. Hydrologic activity is constrained to the near surface because of the shallow maximum thaw depth and the fact that saturated hydrologic conductivities fall off extremely rapidly beyond the porous organic layer.

The porous organic peat soils characteristic of arctic watersheds have a unique capacity to hold water. While typical soils have porosities ranging from 40% to 50% and field capacities range from 10% to 40% by volume, peat soils may have porosities above 90% and field capacities between 60% and 70%. Therefore peat soils can not only hold large quantities of water

Table 1. Peat and Topmodel Parameters

Parameter	Description	Value
Ks	saturated hydraulic conductivity at surface,* cm/s	1.9×10^{-4}
ϕ	peat porosity,* cm^3/cm^3	0.85
θ	peat field capacity,* cm^3/cm^3	0.65
θ_r	peat residual moisture content, cm^3/cm^3	0.10
λ_{th}	dry thermal conductivity of organic peat, 10^{-3} cal/cm s C	0.15
λ	mean watershed value of $\ln(a/\tan\beta)$	8.01
f	saturated hydraulic conductivity decay factor	10.0

Hinzman et al. [1991].

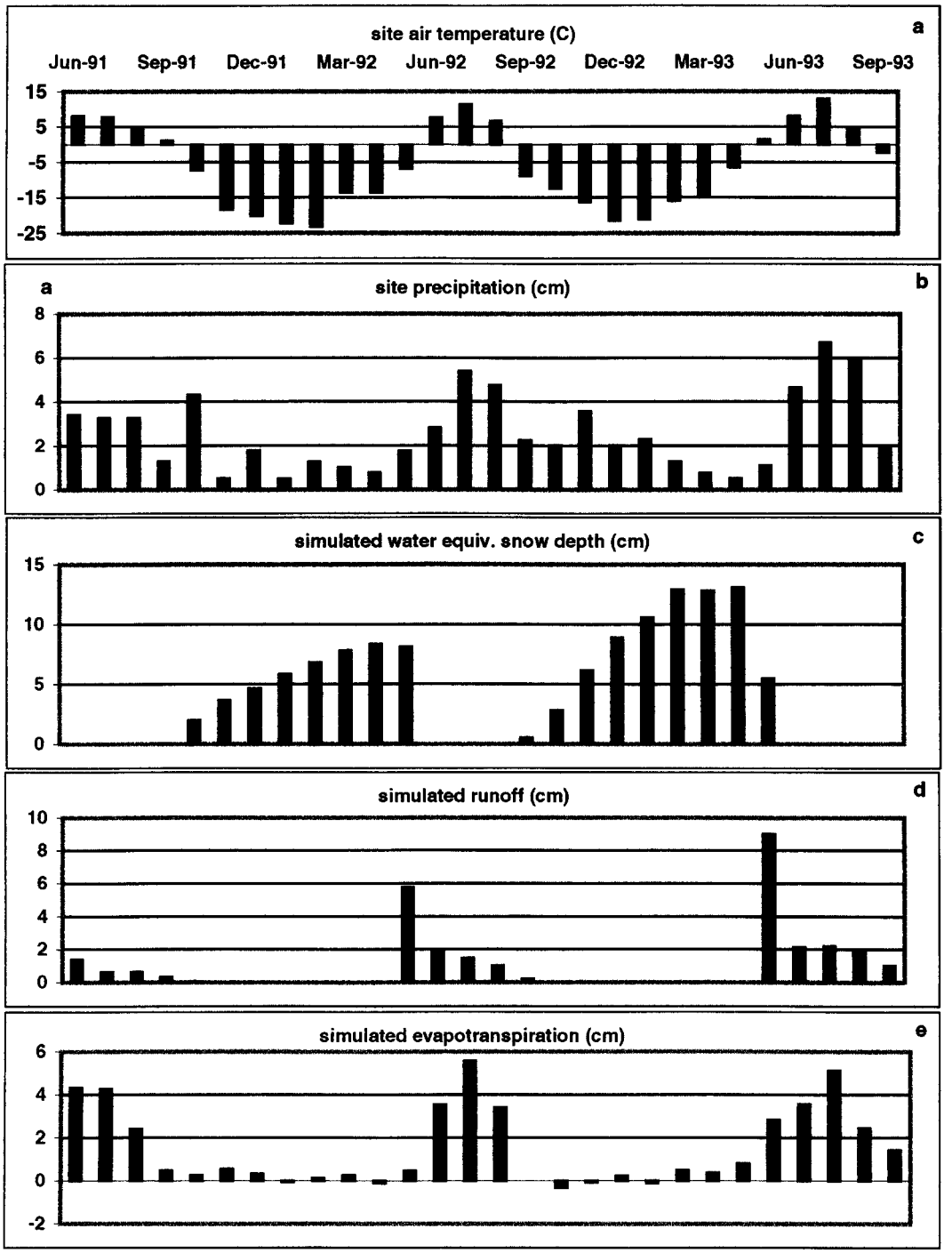

Figure 2. Monthly averages of various watershed water balance components for the period June 1991 through September 1993.

(due to high porosities) but will also retain large quantities of water (due to high field capacities). Significant baseflow occurs only when the soil moisture is relatively high of the order of the field capacity. Thus it is understandable why surface runoff is such a significant component of the annual discharge [*McNamara et al.*, 1998].

To test the model, we used the two and a half years (May 15, 1991, through October 20, 1993) of hourly meteorological data, daily runoff, and daily soil temperature data, collected by Larry Hinzman and Doug Kane of the University of Alaska. Two data sets were used. The first is the forcing variables recorded on the ridge above Imnavait Creek; hourly air and dew point temperature, precipitation, incident solar and thermal radiation, and wind speed. In this high wind regime, only 70–85% of winter precipitation may be captured using a Wyoming snow gage [*Benson*, 1982]. No correction to winter precipitation measurements is applied. The second data set is used to validate model results. Validation measurements include ground temperatures and hourly watershed discharge. To allow the model to adjust to imperfect initial conditions, we only use

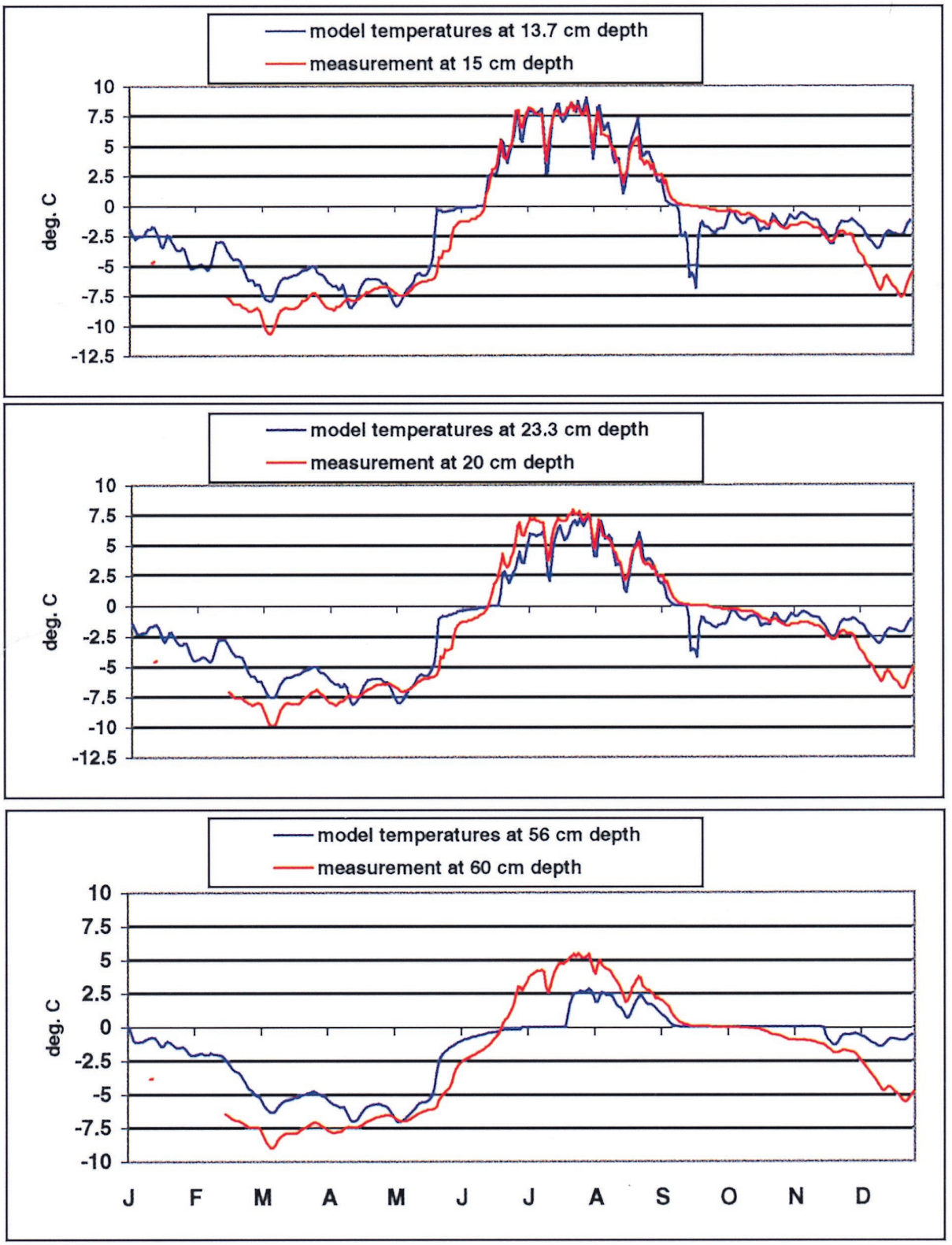

Plate 1. Model-simulated daily averaged ground temperatures and ground temperature data observed at Imnavait Creek during 1992.

1992–1993 data for validation. Soil temperature data taken at given depths are not consistently available over the entire validation period. DEM data for Imnavait are available at a resolution of 20 m × 20 m.

4.2. Model Soil Profile and Initial Conditions

The soil column, modeled by 10 layers, extends to a depth of 3 m. Below 3 m is assumed to be impermeable bedrock. Considering that the maximum thaw does not extend much below

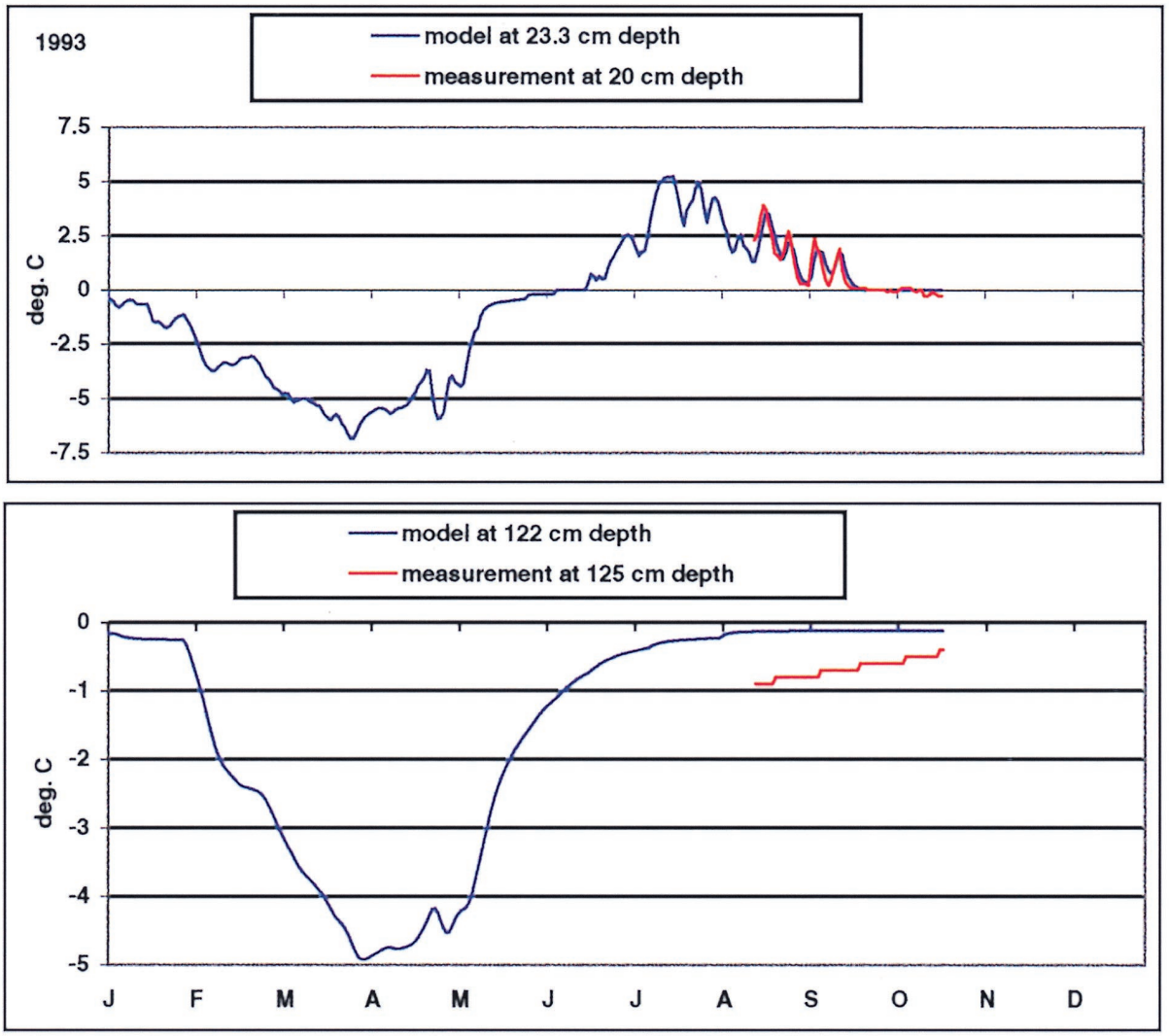

Plate 2. Model-simulated daily averaged ground temperatures and ground temperature data observed at Imnavait Creek during 1993.

1 m, this is a fairly good assumption. The first three model layers extend to a depth of 17.5 cm and are modeled as organic peat soils [*Hinzman et al.*, 1991]. The remaining layers are modeled as silty loam soils. Table 1 of *Stieglitz et al.* [1997] gives the soil water retention properties of the various soil textures. To capture the diurnal nature of the surface energy balance, we take the thickness of the first model layer to be 4 cm. The May 15, 1991, initial volumetric moisture was set to saturation and initial ground temperatures were either obtained from measurements, or for deep temperatures, from the spinup described below.

Using the first two full years of hydrometeorological data to drive the model, we generated the May 15, 1991, initial conditions for the deep temperatures following the procedure of *Lynch-Stieglitz* [1994]. The annual deep temperature determined in this manner is ~4.5°C greater than the long-term average air temperature because the ground is insulated from the atmosphere by the winter snowpack.

The watershed is modeled as 100% tundra vegetation, whose physical characteristics are given in Table 2 of *Stieglitz et al.* [1997]. The only correction to the values given in this table is that the canopy water storage coefficient for tundra vegetation, $a(I)$, is 0.1, and the minimum and maximum stomatal resistance has been reduced by an order of magnitude. This change in the stomatal resistance reflects the fact that for sphagnum moss, which comprises ~55% of the vegetation biomass at Imnavait [*Chapin and Shaver*, 1996; *Chapin et al.*, 1995], transpiration is not governed by stomatal control but instead determined by ground water levels [*Spieksma et al.*, 1997]. The statistics of the topographic index were calculated from DEM data. Table 1 gives the values of the remaining parameters necessary to run the model.

Although the meteorological data used to drive the model are input hourly, the energy fluxes at the ground and snow surface mainly determine the models internal time step. During the period May 1991 through October 1993, the modeled determined time step averaged 10 min, and the computation time was ~7.5 seconds per simulated year on a 400-MHz Pentium PC.

4.3. Results

Figures 2a and 2b (precipitation and air temperatures) present model-input variables while the growth and ablation of

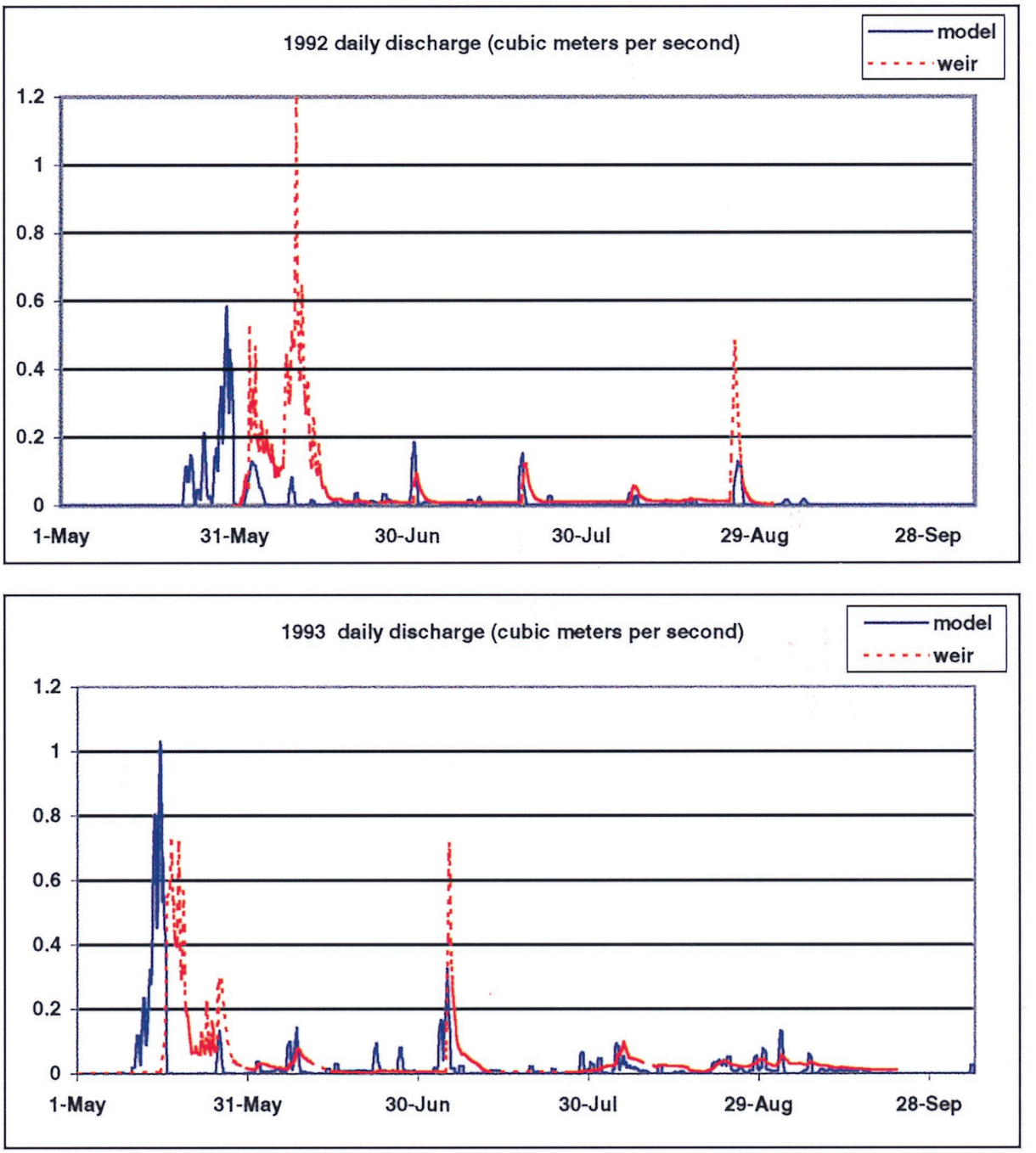

Plate 3. Model-simulated daily runoff and observed runoff measured at the Imnavait Creek weir for the period of April through October during 1992 and 1993.

the snowpack (Figure 2c), total runoff (Figure 2d), and evapotranspiration (Figure 2e) are model simulated. Prior to onset of the snow season the mean water table falls to a minimum depth of ~10 cm (Figure 3c). With freezing of the soil column beginning in early fall, soil moisture does not change significantly until the onset of spring melt. As the pack ablates in late May and early June, melt waters infiltrate the still frozen ground. The soil is recharged, and the mean water table depth rises from the previous summer value nearly to the surface. The associated partial contributing area (Figure 3d) increases from 20% to almost 40% (in good agreement with *McNamara et al*. [1997]. Surface runoff generated over the rapidly expanding saturated regions quickly enters the stream system. In fact, the surface runoff spike (Figure 3a) from the two snowmelt seasons account for a full 59% of the annual discharge. As the soil active layer deepens in the summer, evapotranspiration (and the latent heat flux) begins to increase, peaks in July and August, and falls rapidly as the snow season approaches. This peak in July and August evapotranspiration is associated with a maximum thaw depth of ~80 cm and a seasonal increase in the precipitation. The surface of the ground begins to freeze in early September at the onset of the snow season and the entire soil column is frozen by mid January. Finally, annual precipitation is partitioned 47% into runoff and 53% into evapotrans-

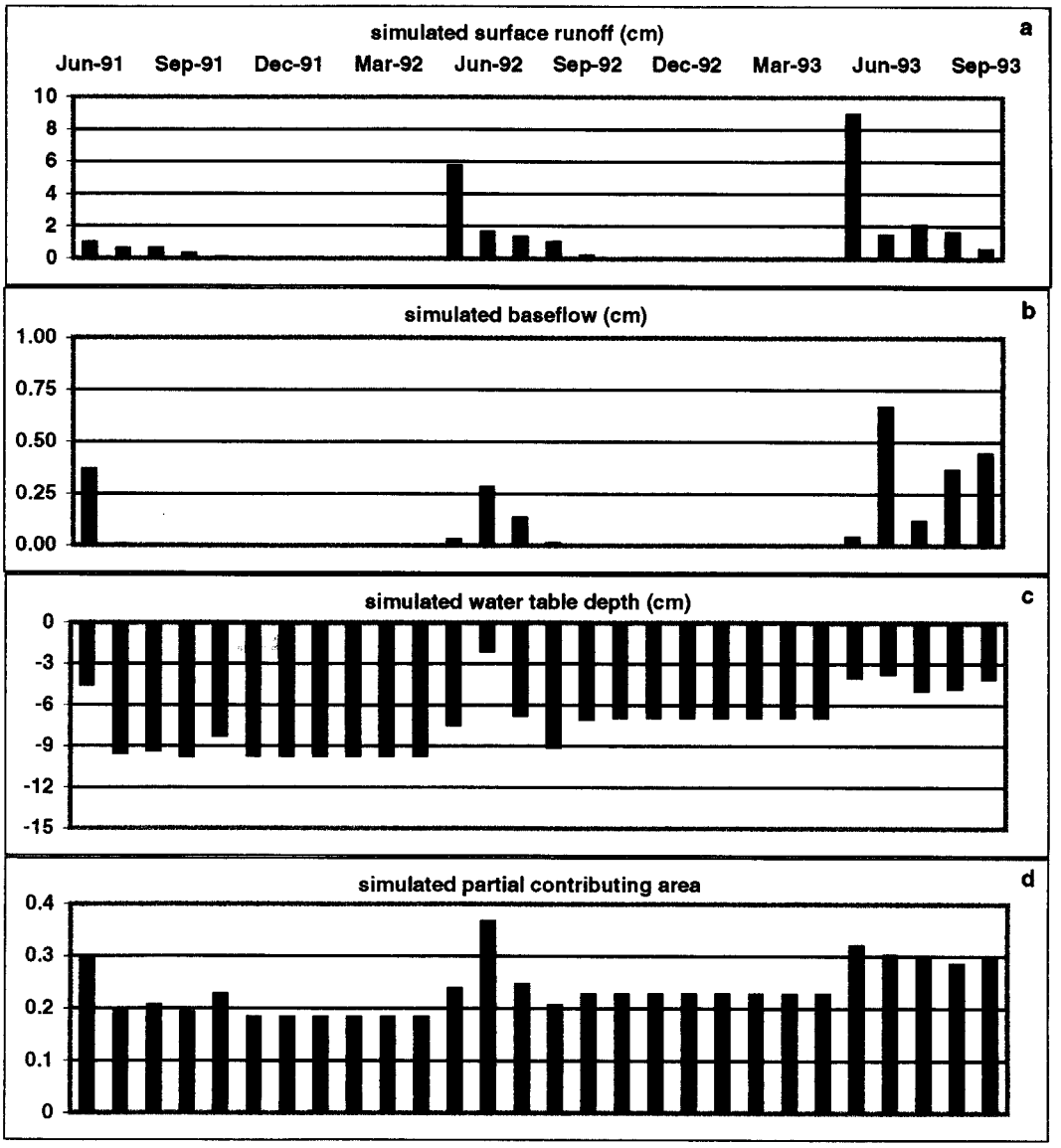

Figure 3. Monthly averages of total model runoff separated into its surface runoff and baseflow components, the mean water table depth, and the expansion and contraction of the partial contributing area, for the period June 1991 through September 1993.

piration, close to the partitioning measured in the long-term field record discussed above.

Ground temperature comparisons for 1992 and 1993 are shown in Plates 1 and 2. For the most part, modeled ground temperatures show good agreement with site data. These include the postsnowpack springtime increase of temperatures in the upper layers. At depth, however, the 1992 comparisons show the model to be too warm in the winter and slightly too cold in the summer. At least one possible explanation can be surmised from the graph. As the deep ground thaws, site data show a linear temperature warming through the 0°C point. This is not the case with modeled temperatures. For at least 4–6 weeks (Plate 1c) temperatures are at 0°C. The implication is that the model is holding excess water and needs that much extra energy to get past the heat of fusion barrier. Considering the fact that this modified Topmodel representation only keeps track of the mean state of the water table when calculating ground temperatures, this is to be expected. On the other hand, the 1992 site data shown here were collected at the ridge top, where the local water depth would be expected to be below the mean value. The net effect of holding excess water is twofold; the unrealistically high heat capacity associated with the excess water dampens the seasonal cycle, and the heat of fusion barrier results in a phase shift such that melt and freeze up occur later than reality.

Overall, both the spring hydrographs resulting from the collapse of the winter snowpack and the summer storm hydrographs are reasonably simulated (Plate 3). However, several model deficiencies are readily apparent.

First, because the spatial distribution of snow cover is not represented in the model framework, modeled snowmelt consistently leads site data by 5–10 days. With high winds and low vegetation height, snow in this region of the Arctic tends to blow into valleys and build up [*Kane et al.*, 1991; *Liston*, 1986; *Liston and Sturm*, 1998]. As such, it takes considerably longer to melt a snowpack whose depth is substantially increased over

a reduced area compared to a pack that is uniformly distributed over the landscape. Further, as pointed out by *Hinzman et al*. [1996], where the snowpack is thick and dense on the valley floor, it functions as a dam and holds back the water until the bonding strength of the snow is overcome. Employing a spatially distributed hydrologic model that also does not account for snow heterogeneity, *Zhang et al*. [1997] finds similar results at the Imnavait Creek watershed.

Second, in 1992, the model underestimates the quantity of meltwater discharge. Again, we believe snow heterogeneity is the culprit. As discussed, the partial contributing area expands rapidly during the melt period. Given that this expanded partial contributing area is almost by definition in the valleys, melt water generated in these regions will rapidly enter the stream system as surface runoff. However, model-generated meltwater is uniformly applied over the entire watershed, and therefore an unrealistically high fraction of the melt water recharges the unsaturated uplands. With the high field capacities indicative of the surface peat, much of this water is retained in the soil matrix and subsequently frozen. It could be argued that the uniform application of melt water over the entire watershed would artificially enhance the expansion of partial contributing area and thereby enhance surface runoff. However, the enhancement of runoff due to a somewhat artificially high partial contributing area is small compared to the dominant effect that most of the melt is taking place in saturated valley regions. Snow distributions for Imnavait creek are shown in *Liston and Sturm* [1998] and *Hinzman et al*. [1996]. *Luce et al*. [1998] discuss in detail the problems related to simulating the timing and quantity of snowmelt discharge in regions where snow heterogeneity is significant. In addition, the inability of the Wyoming snow gage to fully capture winter precipitation may also play a role in the underestimate of meltwater discharge.

Third, modeled summer storm discharge consistently leads site data. Here the lead-time is relatively small and the cause of the problem is the "beaded" stream system. Beads, or small but deep ponds, that act as small reservoirs, are abundant in the meandering Imnavait Creek [*Kane et al*., 1991; *McNamara et al*., 1998]. These ponds receive stream water, retain it, and release it only when full. Depending on the earlier soil moisture condition, this will result in the delayed hydrograph signal shown in Plate 3.

The unique nature of arctic hydrologic and thermal processes is best brought out by comparison. The Sleepers River watershed, a small experimental watershed located in the eastern highland of Vermont, is hydrologically representative of most upland regions in the northeast United States. While snow cover is significant, ground freezing is rare beyond 10 cm. The vegetative cover is primarily a mixed coniferous and deciduous forest. Figure 7 of *Stieglitz et al*. [1997] shows model-predicted and measured discharge at Sleepers River. It is clear that the model simulation at Sleepers River is superior to that for our arctic watershed. To begin with, the assumption of spatial uniformity with respect to snow cover is more valid in a New England watershed. Because the tall and rough vegetation acts to reduce wind speed near the ground there is little wind-blown snow. As a result, both the timing and quantity of the melt water discharge are simulated well. Further, because the stream at Sleepers River has no storage capability, the timing of summer storm discharge is also well simulated. A final difference between these two watersheds is that while surface runoff is the dominant discharge mechanism at Imnavait Creek, baseflow dominates total discharge at Sleepers River. Owing to the fact that the soils at Sleepers River have much lower field capacities, are relatively deep, and are hydrologically active year-round, the majority of discharge occurs via the slow relaxation of the hydrologic gradient under the influence of gravity and topography.

While the overall simulations of discharge is adequate, even at these small spatial scales snow heterogeneity significantly impacts the timing and quantity of snowmelt related discharge and poses a real obstacle toward application on an arctic wide basis. One possible solution begins with recent work by *Liston and Sturm* [1998], who have developed a spatially explicit model that accounts for the effects that wind, vegetation, and topography on the distribution of snow cover in a tundra landscape. While not explicitly compatible with the statistical treatment of topography presented here, the empirical equations governing wind blown snow can be used to treat snow distribution in much the same way we currently treat soil moisture heterogeneity through a statistical representation in which valleys are regions of snow accumulation and uplands are regions of snow ablation.

5. Summary

There has been a marked increase in interest in arctic hydrology in the past 2 decades. Within the global warming community, the overriding question is how global warming will impact the arctic hydrologic cycle and how will this altered hydrologic cycle impact local and regional atmospheric dynamics, vegetation dynamics, net primary productivity, carbon and methane fluxes, and the flux of river waters to the Arctic Ocean. However, the majority of land surface models used to study such questions have been primarily designed for lower latitudes, and as such, are not capable of realistically simulating the physical processes operating in this extreme climate. In this paper we have presented a computationally efficient model that can simulate the overall water and energy balance of a small arctic watershed. The overall simulation adequately captures the evolution of ground temperatures, summer storm hydrographs, and with some reservations, the growth and ablation of the seasonal snowpack and the spring snowmelt hydrograph. Because we rely only on the statistics of the topography, this approach can be applied on an arctic wide basis. Hopefully, this work then represents another step towards ultimate implementation throughout the Arctic, as well as the eventual integration of hydrologic, thermal, and biologic processes with a unified watershed framework.

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