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GREAT LAKES LARGE BASIN RUNOFF MODEL

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INTRODUCTION

Agencies concerned with managing water resources of large watersheds, particularly over large time intervals, must be able to assess expected hydrology of an area. Large-scale watershed models are required to estimate basin runoff to the Great Lakes for use in long-term routing determinations, water resource operation decisions, operational hydrology studies, and long-term forecasting. These models must be designed as continuous-time flow representations for assessing water resource questions over the long term (as opposed to flood prediction over the short term). The models must satisfy limited-data requirements, mandated by data availability for large areas such as the Great Lake basins. Allowable data inputs are limited to daily precipitation and air temperature. Also allowed are any data that can be abstracted easily from available maps or climatic summaries. Model concepts must be physically based, so that understanding of watershed response to natural forces is facilitated, and so the models are economical to use.

The Great Lakes Environmental Research Laboratory (GLERL) built its Large Basin Runoff Model (LBRM) for modeling river systems within the Laurentian Great Lakes Basin. This interdependent tank-cascade model is a *lumped-parameter* model of runoff at the mouth of a watershed and has been tested on the 121 watersheds of the Great Lakes. It was developed from large-scale (climatological) concepts and designed for weekly or monthly volumes of runoff. The model consists of water and heat balances, as do other water-budgeting models, but with alternative physical interpretations given to its components. The model is physically based and uses climatological considerations not possible with small watersheds. In particular, evapotranspiration losses for large areas may now be considered as a function of readily available data. Analytical solutions are used instead of numerical solutions to bypass associated numerical error. The model is described and applied in an example watershed.

LARGE BASIN RUNOFF MODEL

The LBRM is an interdependent tank-cascade model that uses physical concepts outlined in Figure 1 (Croley, 2002). The main feature of this arrangement is that it is described by strictly continuous equations; none of the complexities associated with inter-tank flow rate dependence on partial filling are introduced. For a sufficiently large watershed, these nuances are not observed due to the spatial integration of rainfall, snowmelt, and evapotranspiration processes. Since the solution is analytically tractable, large time steps may be employed without numerical error or excessive computational requirements. The integration of data inputs over large time steps may introduce errors that can only be assessed by example applications in the selection of the appropriate time step. However, for large watershed areas, there is some temporal integration of inputs that may make the approximation of uniform inputs over each time interval inconsequential.

The upper soil zone is the void space in the surface soil layer to a depth that can be considered to control infiltration, usually a few centimeters. The lower soil zone is located beneath the upper soil zone and above the water table. The groundwater zone is located beneath the water table.

These definitions are inexact since the water table fluctuates in time, implying that these zones are not static. Likewise, all moisture in these zones may not be involved in basin outflow. For example, moisture beneath the water table is part of the groundwater zone only if it is part of the flow toward the stream channel network on the watershed surface. Moisture that flows from the watershed as groundwater movement is not considered part of this groundwater zone. (No provision is made for water flowing in or out of the watershed as groundwater.) While the location and extent of these zones may be poorly defined, conceptually they are zones that give rise to flow rates as pictured in Figure 1.

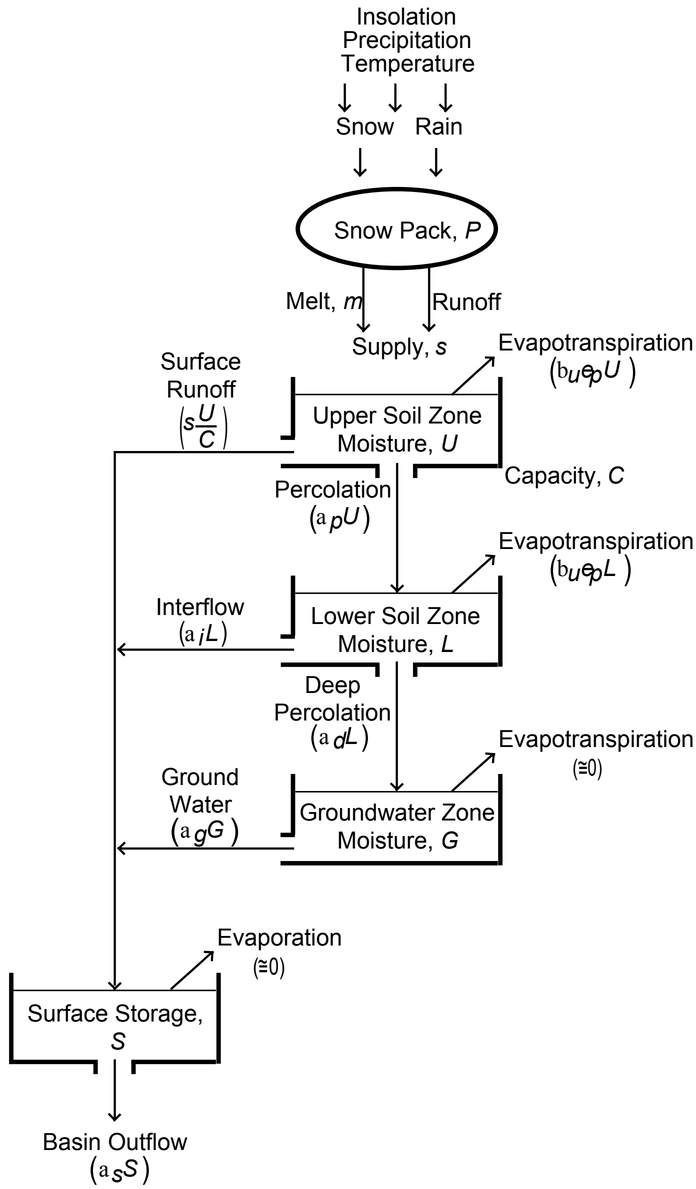

Figure 1. Watershed component tank cascade mass balance.

Net Supply: Precipitation falling onto the watershed surface and snowmelt constitute the net supply to the watershed. Interception can be considered as part of evapotranspiration, and surface depression storage is too transient for consideration since peak flow rates are not of interest. Both are well within the error of measurement for average area1 precipitation and are neglected. Snow accumulation is governed by the concept that precipitation under warm air temperatures occurs as rainfall and under cold temperatures as snow or ice, which accumulates in the snow pack. Snow accumulation is thus governed by the following concept:

$$\begin{aligned} \frac{d}{dt} P &= -m, \qquad & T > 0 \\ &= p, \qquad & T \leq 0 \end{aligned} \tag{1}$$

where $t =$ time (d), P = equivalent water volume present in the snow pack (m^3), m = snowmelt rate (m^{3}d^{-1}), p = precipitation rate (m^{3}d^{-1}), and T = air temperature (ºC). Daily air temperature is estimated typically as the average of daily maximum and minimum temperatures. The simplification of allowing melt only during above-zero air temperatures appeared realistic in example comparisons for volumetric determinations over the week or month (Croley, 1982a). Ignoring evaporation from, and condensation to, the snow pack is justified by the limited data requirements for which the model is designed. The net supply rate is then given as

$$\begin{aligned} s &= p + m, \qquad & T > 0 \\ &= 0, \qquad & T \le 0 \end{aligned} \tag{2}$$

where s = net supply rate (m^3d^{-1}). Snowmelt is determined from the simple concept that there are no heat additions from which melt could later occur during periods of sub-zero air temperatures. For periods of above-zero air temperatures, snowmelt results from absorbed insolation and precipitation. However, it is constrained by the available snow pack,

$$\begin{aligned} m &= m_p, \qquad & m_p \mathrm{d} \le P_0, \\ &= P_0/\mathrm{d}, \qquad & m_p \mathrm{d} > P_0 \end{aligned} \tag{3}$$

where m_p = daily potential snowmelt rate (m^3d^{-1})and the zero subscript on snow pack refers to its initial value at the beginning of the day (at time zero). It is given as:

$$\begin{aligned} m_p &= 0, \qquad & T \le 0, \\ &= a\,h, \qquad & T > 0 \end{aligned} \tag{4}$$

where a = proportionality constant for snowmelt per degree-day (m^3 °C^{-1} d^{-1}]) and h = degree-days per day (°C d d^{-1}), computed as the integral of air temperature with time over those portions of the day when it is above freezing. Since the fluctuation of air temperature during the diurnal cycle is unknown, a triangular distribution is assumed (to approximate an expected sinusoidal variation) for ease of computation. The resulting expression for degree-days is:

$$\begin{aligned} h &= 0, \qquad & T_{\max} \le 0, \\ &= T_{\max}^2 / \left(T_{\max} - T_{\min}\right)/2, \qquad & T_{\min} < 0 < T_{\max}, \\ &= T, \qquad & 0 \le T_{\min} \end{aligned} \tag{5}$$

where $T_{\max}$ = maximum daily air temperature (°C) and $T_{\min}$ = minimum daily air temperature (°C). [Note that snowpack heat storage (warming and cooling) are neglected in (3)—(5).]

Infiltration: At any instant, the net supply rate is divided between surface runoff and infiltration. Surface runoff is proportional to the relative size of the contributing "wetted" area of the watershed (partial-area concept), as well as to the net supply rate

$$r = s\frac{A_w}{A} \tag{6}$$

$$f = s - r \tag{7}$$

where r = surface runoff rate (m^3d^{-1}), A_w = area of wetted contributing watershed portion (m^2), A = area of the watershed (m^2), and f = infiltration rate (m^3d^{-1}). By further approximating the relative size of the contributing area as the relative content of the upper soil zone (a good assumption for a very thin zone), area1 infiltration becomes

$$f = s\left(1 - \frac{U}{C}\right) \tag{8}$$

where U = upper soil zone water volume (m^3) and C = upper soil zone capacity (m^3). Equation (8) indicates that infiltration is proportional to the volume remaining in the upper soil zone. This is the basis for Horton's infiltration-capacity relationship at a point (Croley, 1977, pp. 168-170), although Horton's model uses volume remaining beneath the point (small area), not over a large area. Equation (8) also indicates that infiltration is proportional to the net supply rate. This is an area1 concept for infiltration that has been empirically verified (Kumar, 1980); it does not work for infiltration at a point, which is better described by infiltration-capacity concepts.

Tank Outflows: Since hydrograph recessions are described successfully by exponential decay relationships (Linsley, Kohler, and Paulhus, 1975, pp. 225-229), the linear reservoir concept is deemed appropriate for describing outflow rates from the various storages within the watershed. The concept describes an outflow rate as proportional to the storage remaining. It is expanded here to describe basin outflow, percolation, and deep percolation, as well as the traditional descriptions of interflow and groundwater flow. The form of the equation is

$$z = \alpha Z \tag{9}$$

where z = outflow rate from a storage (m^3d^{-1}), α = linear reservoir constant (d^{-1}), and Z = volume of water in storage (m^3). In (9), Z is U and α is α_p for z equal to percolation; Z is L (lower soil zone water volume, m^3) and α is α_i or α_d for z equal to interflow or deep percolation, respectively; Z is G (groundwater zone water volume, m^3) and α is α_g for z equal to groundwater flow; and Z is S (surface zone water volume, m^3) and α is α_s for z equal to basin outflow. Small parameter values for a tank outflow imply small releases and large storage volumes; large values imply small storages and outflows nearly equal to inflows. The linear reservoir concept is modified when considering evaporation or evapotranspiration (evaporation plus transpiration) from any zone of the watershed.

$$e = \beta Z e_p \tag{10}$$

where e = evaporation or evapotranspiration rate (m^3d^{-1}), β = partial linear reservoir constant (m^{-3}), and e_p = rate of evaporation or evapotranspiration, respectively, still possible (m^3d^{-1}). In (10), evaporation or evapotranspiration is taken as proportional both to the potential rate, determined from heat balance considerations over the watershed, and to the available water volume (reflecting both area1 coverage and extent of supply). This is in agreement with existing climatological and hydrological concepts for evapotranspiration opportunity. In (10), Z is U and β is β_u for e equal to upper zone evapotranspiration, Z is L and β is β_ℓ for e equal to lower zone evapotranspiration, Z is G and β is β_g for e equal to groundwater zone evapotranspiration, and Z is S and β is β_s for e equal to surface zone evaporation.

Mass Balance: By combining (8), (9), and (10) with the definitions given above, a one-dimensional mass continuity equation may be written for each zone; see Figure 1.

$$\frac{d}{dt}U = s\left(1-\frac{U}{C}\right) - \alpha_p U - \beta_u e_p U \tag{11}$$

$$\frac{d}{dt}L = \alpha_p U - \alpha_i L - \alpha_d L - \beta_\ell e_p L \tag{12}$$

$$\frac{d}{dt}G = \alpha_d L - \alpha_g G - \beta_g e_p G \tag{13}$$

$$\frac{d}{dt}S = s\frac{U}{C} + \alpha_i L + \alpha_g G - \alpha_s S - \beta_s e_p S \tag{14}$$

The analytical solutions of (11)—(14) are "continuous"; that is they are amenable to ordinary solution techniques. Furthermore, solutions may proceed for either flow rates or storage volumes directly without the complication of constraint consideration. All derivatives of the solutions with respect to individual parameters exist and are continuous; therefore, analytical gradient-search procedures are possible in parameter determination. The solutions are physically satisfying; non-negative flow rates and storage volumes are guaranteed with any physically plausible set of inputs. The solution equations are unchanged for other time increments; the daily time interval, d, would be simply replaced in the equations. The net supply and potential evapotranspiration are considered to be uniform over the time interval and the choice of time interval must assess the validity of this treatment.

Evapotranspiration: All incoming heat is considered here to be released by the watershed surface by ignoring heat storage and the energy advected by evaporation. The release consists of atmospheric heating (composed of short-wave reflection, net long wave exchange, sensible heat exchange, net atmospheric advection, and net hydrospheric advection), snowmelt and evaporation-evapotranspiration (referred to herein jointly as evapotranspiration). At any instant, the evapotranspiration rate is proportional to the amount of water available as in (10) (reflecting both areal coverage and extent of supply), and to a "potential" rate, e_p, associated with the non-latent heat released to the atmosphere (atmospheric heating), dH/dt (Croley, 1982b):

$$e_p = \frac{dH}{dt} \Big/ \left(\rho_w \gamma_v\right) \tag{15}$$

where γ_v = latent heat of vaporization (596 - 0.52 T cal g^{-1}) and ρ_w = density of water (10^6 g m^{-3}). Potential evaporation is the evaporation that would occur if adequate moisture were available. It is often taken as the amount expected from an open water surface and is used as an estimate of potential evapotranspiration over land and vegetative surfaces (Gray, 1973, pp. 339-353). Very often, engineering calculations of potential evapotranspiration use climatic indicators of temperatures, wind speeds, humidities, etc., by assuming that these quantities are independent of the actual evapotranspiration that does occur. This is adequate for estimates over small areas where evapotranspiration has only a small effect on these quantities. However, over a large area, climatological observations suggest that actual evapotranspiration affects these quantities and hence affects potential evapotranspiration (evapotranspiration opportunity or capacity); the heat used for evapotranspiration reduces the opportunity for additional evapotranspiration (complementary evapotranspiration and evapotranspiration opportunity concept). Morton (1965) made use of this concept to compute regional evapotranspiration from climatological observations. Witherspoon (1970) used an approximation of Morton's work to compute basin evapotranspiration in a flow model for Lake Ontario. Bouchet (1963) postulated that the potential evapotranspiration energy is the absorbed insolation less the energy used for regional evapotranspiration.

This concept is modified here for a smaller-than-regional scale by considering that a portion of the net heat balance after absorbed insolation is available for either potential or actual evapotranspiration. That is, part of it is used in evapotranspiration and the rest of it determines the potential evapotranspiration. Thus, the total heat available for evapotranspiration over a day is composed of the heat actually used for evapotranspiration and that used for atmospheric heating.

$$\Psi = H + \rho_w \gamma_v \left(E_u + E_\ell + E_g + E_s \right) \tag{16}$$

where Ψ = total heat available for evapotranspiration during the day (cal) and H = non-latent heat released to the atmosphere during the day (cal). The value of e_p is determined by simultaneous solution of (11)—(14) and the following complementary relationship between actual evapotranspiration and potential evapotransipiration, derived from (15) and (16):

$$\int_0^{\mathrm{d}} \left[e_p + \left(\beta_u U + \beta_\ell L + \beta_g G + \beta_s S \right) e_p \right] dt = \Psi / \left(\rho_w \gamma_v \right) \tag{17}$$

The evaporation from stream channels and other water surfaces (surface zone) in a large basin is very small compared to the basin evapotranspiration; groundwater evapotranspiration is also taken here as being relatively small. By taking e_p as uniform over the day and ignoring evapotranspiration from the surface and groundwater zones, (17) yields:

$$e_p \cong \frac{1}{\mathrm{d}\, \rho_w \gamma_v} \frac{\Psi}{1 + \beta_u \bar{U} + \beta_\ell \bar{L}} \tag{18}$$

where $\bar{U}$ = average water volume in the upper soil zone (m^3) over the day and $\bar{L}$ = average water volume in the lower soil zone over the day (m^3). As expected, both potential and actual evapotranspiration depend upon the available water supply. If the water supply is large, actual evapotranspiration approaches the limit of the water supply or $\Psi / \rho_w \gamma_v$ and potential evapotranspiration approaches zero. If the water supply is small, actual evapotranspiration approaches zero and potential evapotranspiration approaches $\Psi / \rho_w \gamma_v$ (Croley, 1982b).

The determination of Ψ from observable meteorological variables is difficult (recall the limitation to daily precipitation and air temperature). Daily air temperature is taken here as an integrated reflection of the portion Ψ of the remaining heat balance after absorbed insolation. This concept is satisfying in that air temperature is considered an indicator of the heat balance, rather than an independent variable in the determination of potential evaporation as is done classically. At low temperatures, it is expected that Ψ is small since potential and actual evapotranspiration are low at low temperatures. Over the daily cycle, this energy is rarely negative (net condensation) and is considered here as strictly positive. The heat available for evapotranspiration is estimated empirically from the average air temperature as follows:

$$\Psi = k \exp\left(T/T_b\right) \tag{19}$$

where k = proportionality constant (cal), and T_b = a base scaling temperature (°C). The constant, k, is determinable from the following boundary constraint on the long-term heat balance:

$$\sum \Psi_i = \sum \left(\sigma_i - m_i \rho_w \gamma_f\right) \mathrm{d} \tag{20}$$

where σ = daily solar insolation at the watershed surface (cal d^{-1}), γ_f = latent heat of fusion (79.7 cal g^{-1}), and the subscript, i, refers to daily values. Equation (20) conserves energy in that all insolation not used for snowmelt appears sooner or later as other components of the heat balance that determine Ψ. Daily insolation at the surface of the watershed may be estimated from extraterrestrial radiation and cloud cover:

$$\sigma = 10000 A \tau \left(b_1 + b_2 x\right) \tag{21}$$

where τ = cloudless daily insolation (langleys d^{-1}) available in standard climatological summaries as a function of latitude and time of the year, b_1 and b_2 = empirical constants, and x = daily ratio of hours of bright sunshine to maximum possible hours of bright sunshine, estimated from daily air temperatures (Gray, 1973). In the absence of cloud cover data, x may be estimated (Crawford and Linsley, 1966, p. 50) from

$$x = \mathrm{MIN}\left[\left(T_{\max} - T_{\min}\right)/15,\ 1.0\right] \tag{22}$$

There were several alternatives to the "heat balance," used here to compute snowmelt and evapotranspiration, considered early in the model development, but they were impeded by the limited-data design objectives. Comprehensive heat balances that considered all advection terms through control volumes defined over the upper soil zone or upper and lower soil zones were written in the early modeling. Net long-wave radiation transfer and sensible heat transfer were estimated directly by using empirical relations. These relations required unavailable data, which were estimated based on engineering judgment. Freezing of the upper soil zone, snowpack and ice formation and decay, and Penmann's potential evapotranspiration were all computed as part of these comprehensive heat balances. The net supply and evapotranspiration models presented here resulted in a two-fold improvement in modeling over these earlier efforts (Croley, 1982a), as measured by the root mean square error of model output (basin outflow). Presumably, these models are superior because of their limited data requirements. Also, the use of air temperatures as an indicator of what has occurred in the watershed is superior to its use as an independent variable in computing potential evapotranspiration and net supply. This change in perspective is fundamental to modeling large-scale watershed hydrology from a climatological viewpoint.

APPLICATION

GLERL developed, calibrated, and verified conceptual model-based techniques for simulating hydrological processes in the Laurentian Great Lakes (including Georgian Bay and Lake St. Clair, both as separate entities). They integrated the models into a system to estimate lake levels, whole-lake heat storage, and water and energy balances for forecasts and for assessment of impacts associated with climate change (Croley, 1990, 1993a,b; Croley and Hartmann, 1987; Croley and Lee, 1993; Croley et al., 1998; Hartmann, 1990). During the application process, experience was gained that may benefit others who would apply the LBRM to large basins.

Data Preparation: For application of the LBRM to a very large drainage basin (such as that associated with a Great Lake), the basin is first divided into watersheds with areas of between 120—20000 km^2 (there are 121 watersheds in the entire Great Lakes basin); most are between 1000—5000 km^2. The following input data are required to apply the model: daily precipitation, daily minimum and maximum air temperatures, a standard climatological summary of daily extraterrestrial solar radiation and empirical constants (b_1 and b_2), and for comparison purposes, daily basin outflows. Conversion of units for precipitation from inches per day or centimeters per day to cubic meters per day and for insolation [see (20)] from langleys per day to calories per day involves the area of the watershed. The meteorological data from stations about and in a watershed are combined through Thiessen weighting to produce areally-averaged daily time series of precipitation and minimum and maximum air temperatures for each watershed. In past determinations of water supply effects from climate change scenarios (Croley, 1990, 1992, 1993a; Hartmann, 1990), GLERL used about 1,800 meteorological stations for overland precipitation and air temperature (about 15 per watershed or approximately 1 per 70 km^2). Recent experience (Croley and Hartmann, 1987; Croley et al., 1998) also suggests that 5-30 stations per watershed for overland meteorology is sufficient for operation of the LBRM at daily time intervals. Thiessen weights are determined for each day of record, if necessary, since the data collection network changes frequently as stations are added, dropped, and moved or fail to report from time to time. This is feasible through the use of an algorithm for determining a Thiessen area-of-influence about a station by its edge (Croley and Hartmann, 1985). Flow records of all "most-downstream" flow stations are combined by aggregating and extrapolating for ungaged areas to estimate the daily runoff from each watershed. Daily basin outflow is reported in either cubic feet per second or cubic meters per second and is converted to cubic meters per day. Then, the LBRM may be applied in a "distributed-parameter" application by combining model outflows from each of the watersheds to produce the entire basin runoff.

There are five variables to be initialized prior to modeling: P, U, L, G, and S as P_0, U_0, L_0, G_0, and S_0, respectively. While the initial snow pack, P_0, is easy to determine as zero during major portions of the year, these variables are generally difficult to estimate. If the model is to be used in forecasting or for short simulations, then it is important to determine these variables accurately prior to use of the model. They may be taken as the values at the end of a previous model run, preceding the time period of interest, for forecasting uses of the model. If the model is to be used for calibration or for long simulations, then the initial values are unimportant. The effect of the initial values diminishes with the length of the simulation and after 1 or more years of simulated results, the effects are absent from a practical point of view. Calibrations should be repeated with initial conditions equal to observed long-term averages until there is no change in the averages to avoid arbitrary initial conditions when their effects do not diminish rapidly.

Model Use: Since a change in C can be exactly compensated (in terms of intrabasin flows and evapotranspiration) by other parameter changes, C is set arbitrarily to 2 cm over the watershed surface. However, C affects all tank storage volumes and should be determined if boundary conditions on soil moisture (or other storage volumes) are available. Note also that β_g and β_s are taken as zeroes since evaporation from the surface and evapotranspiration from the groundwater zone are small relative to evapotranspiration for the upper and lower soil zones; see Figure 1. Finally, empirical coefficients b_1 and b_2 are taken from available climatological summaries.

GLERL calibrated the LBRM for each Great Lakes watershed with 30 years of daily weighted climatologic data. The nine parameters are determined (Croley, 2002) by searching the parameter space systematically, minimizing the root mean square error between model and actual outflows for each parameter, selected in rotation, until all parameters converge within two or three significant digits. Comparisons with other runoff models (Croley, 1983a) and climatology (Croley, 2002) show the LBRM is superior for large basins.

The LBRM captures "realism" in its structure with several advantages over other models. Basin storages, modeled as "tanks," are automatically removed as respective parameters approach their limits. Thus, the structure of the model changes within a calibration. This is achieved without the use of "threshold" parameters in the model since physical concepts are used which avoid discontinuities in the goodness-of-fit as a function of the parameters; these concepts appear especially relevant for large-basin modeling. Because the "tanks" relate directly to actual basin storages, initialization of the model corresponds to identifying storages from field conditions which may be measured; interpretations of a basin's hydrology then can aid in setting both initial and boundary conditions. The tanks in Figure 1 may be initialized to correspond to measurements of snow and soil moisture water equivalents available from aerial or satellite monitoring.

EXAMPLE APPLICATION

The Lake Superior Basin, above the locks at Sault Ste. Marie, drains about 130,000 km^2 of Ontario, Minnesota, Wisconsin, and Michigan. It is divided into 22 watersheds for use with the LBRM (see Figure 2). Watershed boundaries are based on state hydrologic unit maps from the U.S. Geological Survey (USGS) for Michigan, Wisconsin, and Minnesota, and on drainage basin map overlays from the Water Resources Branch of the Inland Waters Directorate of Environment Canada for Ontario. Watersheds not draining directly into Lake Superior were combined with those into which they drained so that all resulting watersheds have a direct outlet to the lake.

Meteorological Data: Meteorological data for stations in the United States are from the National Climatic Data Center, National Environmental Satellite Data and Information Service, NOAA. In Canada, data are from the Canadian Climate Centre, Atmospheric Environment Service, Environment Canada. For each day of available data, all meteorological stations with no missing data for that day were used to compute all Thiessen weights, which in turn were used to weight available meteorological data to determine daily watershed spatial averages. In the event that no stations have data on a given date, then the average value of the weighted data on the corresponding day-of-the-year for all available years of record was used to fill in. Thus there are no missing data values (and there must *not* be) in the meteorological data files prepared for use with the LBRM and its calibration. A daily meteorological data set for the entire Lake Superior basin (land area) was constructed by multiplying each areal-average daily data value from each watershed by the corresponding watershed area, summing all weighted values for the entire basin (from all watersheds being used), and dividing by the sum of the drainage areas actually used. The areal-averaged daily air temperatures and precipitation for each watershed are more than 99-percent complete. Also, values of average mid-monthly daily short-wave radiation received on a horizontal position of the earth's surface under cloudless skies and the coefficients in (21)were taken from Gray (1973; pp. 3.11—3.16).

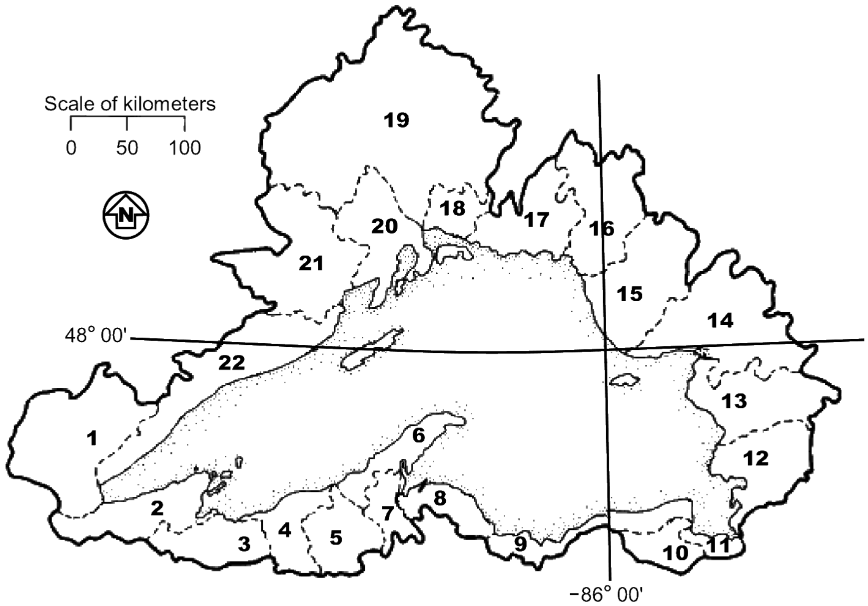

Figure 2. Lake Superior watershed map.

Hydrological Flow Data: All "most-downstream" stream flow gages are used with their drainage areas as given by the USGS or Inland Waters Directorate, while the total area in each watershed is based on the state hydrologic unit maps and drainage basin map overlays, discussed previously. Relative drainage areas for all flow gages were determined by dividing each gaged area by the total area of the watershed. All hydrological stations within a given watershed (non-overlapping drainage areas) whose records contain no missing data, for each day in question, were used to determine the watershed outflows into Lake Superior for that day. This aggregation for each day was accomplished by adding data values from each gage within the watershed and dividing by the sum of the relative drainage areas for the gages actually used, to extrapolate for the entire watershed area. Thus, missing data at a given gage were effectively "filled-in" by using data at nearby gages within the same watershed.

Model Application: The Large Basin Runoff Model is programmed in FORTRAN 95; source code and example data sets are available over the World Wide Web at http://www.glerl.noaa.gov/wr/lbrmexamples.html. The modeled watershed outflow and the measured flow are presented for comparison in Figure 3 for all Lake Superior watersheds, aggregated together. The surface zone storage half-life is 8.8 d and is larger than the Lake Ontario basin half-life (Croley, 1983b) and may reflect the boggy, swampy nature of much of the Lake Superior basin. The groundwater zone storage half-life for Lake Superior is 34.6 w and is almost an order of magnitude less than the groundwater zone storage half-life for Lake Ontario, and may reflect the presence of the Precambrian shield under much of the Lake Superior basin. The upper soil zone storage half-life of 7.6 h is smaller than that for the Lake Ontario basin, while the lower soil zone half-life is about the same at 8.2 w. This may imply that, for the Lake Superior basin, a single soil zone may be adequate to model the basin response. This is also consistent with the general structure of the Lake Superior basin—a thin layer of soil over bedrock.

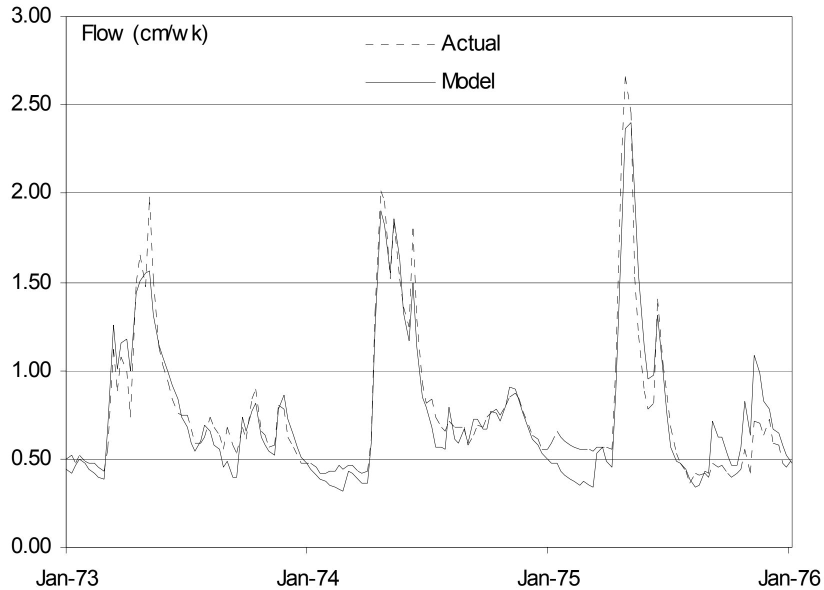

Figure 3. Monthly total LBRM Lake Superior basin runoff.

SUMMARY

The Large Basin Runoff Model developed at GLERL is an accurate, fast model of weekly or monthly (derived from daily) runoff volumes from Great Lakes watersheds; it has relatively simple calibration and data requirements. Parameters have physical significance and calibrated values appear reasonable. The net supply and evapotranspiration sub models offer limited data requirements. The Lake Superior applications illustrate spatial integration effects on model resolution and filtering of both information and data errors consequent with these applications. The distributed-parameter is marginally better than the lumped-parameter application. The lumped application yielded a correlation with observed daily flows of 0.84 and the distributed application yielded a correlation of 0.88. Applications of the model to watersheds about Lake Superior show good-to-exceptional agreement with available flow data where flows are natural and unregulated; applications to watersheds with regulated flows varied from poor to good.

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