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A HYDROLOGY TEMPERATURE MODEL FOR A SMALL MOUNTAIN WATERSHED

by

Charles Wilson Pettee

A thesis submitted in partial fulfillment
of the requirements for the degree

of

MASTER OF SCIENCE

in

Watershed Science

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Charles W. Pettee

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ABSTRACT

A Hydrology Temperature Model for a

Small Mountain Watershed

by

Charles Wilson Pettee, Master of Science

Utah State University, 1976

Major Professor: Dr. Richard H. Hawkins
Department: Forestry and Outdoor Recreation

A small mountain watershed located in the Wasatch Mountains of North Central Utah is calibrated to a lumped, deterministic simulation model which is capable of predicting daily streamflow and stream temperature. The input information required is daily precipitation and maximum and minimum air temperatures.

In this study, the area of watershed modeling is reviewed in general and as it specifically applies to the study watershed.

The degree of correlation between observed data and predicted output is only mediocre. The model remains unverified for streamflow prediction and is poorly verified for stream temperature.

(70 pages)

INTRODUCTION

The general idea behind a modeling attempt is to apply a systems approach to that part of the hydrologic cycle acting within a watershed. In general, a system can be defined, as Dooge (1973, p. 4) has, as "any structure, device, scheme or procedure, real or abstract, that interrelates in a given time reference, an input, cause, or stimulus, of matter, energy, or information, and an output effect or response, of information, energy, or matter." Specifically, watershed system modeling can be interpreted as a procedure consisting of mathematical relations which interrelate in a given time reference, an input of information, and an output of information.

There are three essential parts which comprise a model: coefficients, structure, and initial or boundary conditions. The structure of a model is the representation within the model of the pertinent processes or conditions which relate the relevant inputs to the desired outputs. Associated with these processes and conditions are one or more real or empirical coefficients. The magnitude of these coefficients determine the rate and relative importance of each component of the structure. The initial and boundary conditions provide a starting point for those structure components which are continuous functions and a mass or energy exchange between the modeled system and the outside world respectively.

Models vary considerably in the way their structure represents the system, the number of coefficients used, the inputs used, and the outputs desired. Models with simple structure requiring few coefficients are termed low resolution, while those with complex structure and many coefficients are termed high resolution models. Low resolution models characteristically require little input and less effort to calibrate, but will generally not yield as abundant or as accurate an output as will a high resolution model.

Models can be further categorized as deterministic or stochastic and either lumped or distributive. A deterministic model is one which uses functions which result in quantitative relationships among the processes represented in the structure of the model which can be related to physical characteristics of the watershed. Stochastic models are those which relate the input and the output through the use of random or statistical relationships.

When modeling small areas, it is easiest to apply the modeled processes to a single unit of area in the spatial dimension. This type of model represents the entire area as one point and is called a lumped model. Lumped models will introduce an averaging effect, that is, the coefficients will reflect the average characteristics of the area. Under some conditions there may be a significant lack of homogeneity in the watershed characteristics within the area under study. It may be more appropriate, then, to divide the area into two or more parts and consider the appropriate processes independently in each section, then

integrate their separate contributions to arrive at the overall result. Each of these sections is treated as an independent lumped subsystem, and the sum of these lumped subsystems is called a distributive system model.

In theory, models attempt to simulate the real world as closely as possible. There are two gaps, however, in the transmission of real world processes into a working model. These gaps are areas of information losses. The first area of information loss is a result of the development of a conceptual representation of the naturally complex physical laws governing watershed behavior. Processes in general are well understood, but detailed understanding of some processes is lacking. The second gap is the transition between this conceptual representation of the hydrologic processes and assembling them into a working model with accurately measured inputs. Even though a process is understood in concept, it is not always possible to describe it as accurately using mathematical functions.

The degree to which the structure and coefficients used in the model represent the real world has direct consequences on how much insight into the watershed system can be realized. If the model structure is presented simply as a set of empirical relationships, then only the inputs and outputs found are of any significance. The question of why an output found was produced from a particular input is not answered. In this case there can be no parallels drawn between the model structure and its coefficients and the hydrologic processes

occurring in the real world. Crawford (1971) calls this "black" box technology. At the other extreme is "white" box technology, or pure deductive science. This is demonstrated by a model which has a structure directly analogous to the hydrologic process acting within the watershed. Coefficients in this structure are directly measureable parameters which characterize the watershed. Most models lie somewhere in between these two extremes as "grey" boxes.

STUDY PROBLEM AND OBJECTIVES

With the current emphasis on environmental impact, the ability to forecast the hydrologic effects of any particular land use or manipulation is becoming an essential part of a hydrologist's responsibilities. Another area of concern involves gaining an understanding of the processes which effect runoff from a watershed. A need exists, therefore, for some type of predictive procedure with the capability of incorporating land condition dependent variables. Inherent with the development of such a procedure is a study of the hydrologic processes acting within a watershed system, their relative importance and magnitudes. A deterministic watershed model can provide the hydrologist with such a predictive procedure.

Most models currently being investigated have relatively high resolutions. Most wildland situations do not have the data collection apparatus to handle the input requirements of these high resolution models. For this reason only daily precipitation and maximum and minimum air temperature data will be required for the model in this study.

The objectives of this study will be to:

1. Review previous types of modeling efforts.

2. Review the modeling procedure in general and specifically for the study area.

3. Create and make operational a deterministic watershed model which uses only daily inputs of precipitation and maximum and minimum air temperatures to predict daily streamflow and stream temperature.

4. Examine the resulting model structure, coefficients and initial or boundary conditions and gain some insights into the functional aspects of the study watershed system.

REVIEW OF LITERATURE

Hydrology

Because of the natural complexity of the processes involving water movement through a watershed, water resource models generally have very detailed and lengthy mathematical process representations. In addition to this, the calibration step is very repetitious. For these reasons, the area of system modeling is a recently initiated one, and is tied very closely with the development of computational systems.

The general processes acting within a watershed are widely agreed upon. A result of this is that most hydrology models consist of a similar set of streamflow producing processes. The combination of processes vary depending upon the purpose for which the model is being developed. An example of a specific purpose model is the flood frequency model by Hauth (1974) which routes the precipitation in excess of infiltration to predict flood peaks. In this case only three processes are considered, antecedent soil moisture, infiltration, and surface runoff routing. The model of the following study is a general purpose model. That is, the entire streamflow regime is of importance and so every hydrologic process is considered.

A major distinction among hydrology models can be made with respect to the size of the area being modeled. The kinds of modeling

problems associated with watershed size are generally similar to those of other types of analytical hydrology.

Large scale watershed systems involve many miles of stream channels carrying large quantities of water. The storage and travel times characteristic of these systems necessitate the use of channel hydraulics as the major structural factor used in arriving at the streamflow. When the size of the watershed is large in comparison with the variation of other watershed characteristics such as precipitation patterns, geologic type, vegetation type, etc., then it is not a valid assumption to consider the area as being homogeneous in these respects. Models of large scale systems, therefore, are nearly always the distributive type. One of the first models was developed by D. M. Rockwood (1958) for the U.S. Corps of Engineers. This model was developed to predict river flow from large watersheds for purposes of reservoir and channel routing, storage, and design. The purpose of this model reflects the fact that it models large watersheds and thus channel effects on the streamflow are dominant.

On small scale watersheds, channel storage effects are less important, although they are sometimes still present. These channel effects in small area models are a function of the time increment being used. As the time increment becomes smaller, the channel travel time effect becomes more important. Since channel flow is less dominant, more emphasis is placed on the activity of the water before it reaches the stream channel. While small scale watersheds

tend to be more homogeneous in their characteristics than large watersheds, they are not always modeled as lumped systems. This decision is based on the particular watershed as variations in factors such as elevation, precipitation, and temperature can be large over small areas which have steep slope gradients.

Crawford and Linsley began doing modeling research at Stanford University in 1959. A result of this effort is a series of models, one improving on the previous, which are summarized in their publication on the Stanford Watershed Model IV (1966). This general model used a lumped representation and employed channel routing while using a fifteen minute time increment. On the other hand, Bowles, Riley, and Shih (1975) used no channel routing in their application of the Utah State University Watershed Simulation Model with a time increment of one day. This demonstrates the fact that the inclusion or deletion of channel routing is also based on the length of time increment used.

Small watershed areas can be further grouped into urban or wildland conditions.

Urbanization, hydrologically speaking, is basically a procedure where large areas are covered with impervious surfaces. This results in excessive overland flow and the concern here is to get this overland flow to the nearest drain or permeable surface in such a manner as to prevent large peak flows and water ponding. The main effort of urban hydrology models is routing this overland flow. In this type of small watershed, channel routing and hydraulics become

the dominant factors effecting storm runoff. Narayana, Riley, and Israelson (1969) and Lumb, Wallace, and James (1974) have routed precipitation in excess of infiltration and storage in their models.

While not all wildlands have totally permeable surfaces, overland flow is not as dominant on wildland areas. Precipitation excess routing is still a valid method for use in wildland situations (Dawdy, Lichty, Bergman, 1972 and Hauth, 1974). Another method used commonly on wildlands and the one used in this study is the "compartment" type structure, in which the watershed is separated into its different storage areas (interception, soil moisture, groundwater, etc.). These storage areas are mathematically represented as compartments, and functions describing the movement of water among these compartments are used in the model.

Watershed models have been used in a variety of problem solving situations. Leaf and Brink (1975) have developed their model to consider both short and long term hydrologic impacts of timber harvesting and weather modification, or a combination of the two. This can aid in developing management studies for varying planning intervals. Hauth (1974) used his model to route overland flow and predict flood peaks for small drainage areas throughout Missouri. Using his model, which was calibrated on a relatively short period of observed records, he was able to reconstruct the long term flood records for the streams. Riley and Hawkins (1975) developed their general purpose model for a forested watershed and then simulated rangeland conditions by

varying the coefficients.

Although this may give the impression that models hold the answer to all of our problems, caution should be used when dealing with them. Watershed models have only recently been experimented with as management and research tools and are not adequate yet to stand as the sole source of information upon which a decision can be based. Furthermore, when a model is used, the user should be completely familiar with it in order to be aware of its abilities and limits. Care must be taken to avoid blindly using a model's output.

Stream Temperature

Stream water temperature is a fairly well understood phenomenon. In his studies of the water temperature of small streams in Oregon, Brown (1969) reports that solar radiation accounted for over 95 percent of the heat input during the midday period during summer. This realization does not leave much choice on modeling procedures for stream temperature. Either a heat budget approach is used, or some type of empirical index for solar radiation. The important factors explaining the temperature of a stream can be grouped again according to the stream size. In large rivers, evaporation and heat conduction from the streambed are significant temperature influences. As has been discussed earlier, the major portion of water entering any particular reach is introduced through the channel rather than seepage through the streambed. This suggests that lateral seepage temperatures are

less significant. In small streams the temperature is a result of the amount of heat input through net radiation and that heat already present in the streamflow as it enters the stream channel. Due to the fact that most small watershed streams originate as cold ground water, net radiation can be considered as the main phenomenon influencing stream temperature in small watershed streams.

The most accurate method is to use a heat budget on the stream section being modeled. Brown (1969) uses solar radiation in the development of his small stream temperature prediction model. These, however, require the measurement of either solar radiation or cloud covers. Most water temperature regimes consist of some type of sinusoidal response on a daily, and a yearly basis. This has prompted the use of sine functions in empirical equations (Ward 1963). In the study which follows, daily maximum air temperature, in combination with an empirical coefficient which varies sinusoidally with the time of year, will be used to index solar radiation. While these empirical functions avoid the requirement of measured radiation they require, themselves, an extensive temperature data base in order to determine the coefficients used.

DESCRIPTION OF THE STUDY AREA

The watershed being modeled in this study is the West Branch of Chicken Creek. This watershed is located on the Davis County Experimental Watershed in the Wasatch Mountain Range of North Central Utah. The Davis County Experimental Watershed was established in 1930 as a United States Department of Agriculture Forest Service administered research area and is under the direction of the Intermountain Forest and Range Experiment Station.

The watershed is 217 acres in area, lies with a northwest aspect with an elevation range of between 7,550 and 8,396 feet. The slope of the watershed is relatively low with an average of about 19.5 percent.

The average yearly precipitation here is about 45 inches per year. The major portion of this precipitation falls during the winter months in the form of snow. During the summer months potential evapotranspiration generally exceeds precipitation.

The drainage system consists of one well defined perennial stream channel with several poorly defined intermittent and ephemeral contributaries. Peak flows occur in the spring as a result of snowmelt.

The soils found on the watershed vary considerably. They range from deep loamy alluvial soil in the valley bottom to deep clayey colluvium soils on the side slopes and shallow gravelly loamy soils on the ridges. In general, the soils are deep and have good moisture

holding capabilities.

The primary vegetation type on the study watershed is Aspen (Populus tremuloides Michx). This type occupies about 60 percent of the watershed area. The rest of the watershed is in grasses, forbs, and brush. Conifers comprise only about 3 percent of the entire watershed area.

The watershed is located within a research area and is therefore relatively well instrumented. Measurement devices on the watershed include an "H" type flume stream gauge, a network of weighing type and storage type precipitation gauges, and more recently a climatic station with maximum minimum thermometers and an air temperature recorder in addition to a water temperature recorder.

A more complete description and hydrologic analysis of this watershed has been made by Johnston and Doty (1972).

MODELING PROCEDURE

The modeling procedure can be described in general by the following series of steps:

1) Problem Identification
2) System Identification
3) Data Accumulation and Reduction
4) Model Formulation
5) Model Calibration and Verification
6) Interpretation of Results

While a specific order of approach is suggested by these steps, there are feedbacks between all of these steps which are an important part of the modeling procedure. For example, if during the Calibration and Verification step the model cannot reproduce the desired outputs, then one of the first four steps has not been defined properly. This may be a result of the modelers concept or mathematical representation of the problem or the system, or data inaccuracy. It may also be that there is simply too much loss of information between the real watershed and a working model in order for it to be possible to represent the system. In any case, the modeler must reconsider the previous steps and modify the model.

Problem Identification

One of the primary areas of concern of any type of watershed management, manipulation, analysis, or long range planning, is the protection of the integrity of the streamflow's quality, quantity, and regime.

The movement of water through a watershed is dependent upon many different physical phenomenon. Some of these are well understood and some are not. In order for a watershed manager to predict the effect of any of these aforementioned practices, it is necessary to have a thorough understanding of these physical phenomena acting within a watershed and their relative importance. In order to grasp the relationships between different land uses and how streamflow will respond to them, it is helpful to think of the watershed as a continuous and dynamic system. The best way to investigate this cause and effect relationship is to carefully examine the streamflow characteristics with respect to the watershed characteristics producing it.

This streamflow-watershed condition relationship concern overlaps the major areas of hydrologic endeavor. Researchers are primarily interested in the cause and effect relationships which exist on a watershed. Practicing land managers want to incorporate the land use which yields the most benefits for several years to come. These benefits and land uses are directly effected by streamflow and temperature.

One of the best uses for watershed models may be as a teaching device. A hydrology student is interested in understanding these watershed processes and relationships as well as cause and effect relationships and land use consequences on streamflow and temperature.

If there is a process of particular importance in a study, that process can be refined and expanded, if necessary, even to the point of adding more kinds of input.

A hydrology temperature model of the type which follows provides considerable insight into the individual physical characteristics, water storage and movement processes, and their interrelationships, which summed together into a continuous and dynamic system, we call a watershed.

System Identification

The system this problem deals with is that part of the hydrologic cycle which acts within a watershed. This includes the various inputs and outputs of mass and energy as well as the hydrologic phenomena influencing water movement and temperature throughout the watershed.

Data Acquisition and Reduction

The data used in this study has been provided by the United States Forest Service Intermountain Forest and Range Experiment Station, Logan, Utah. The study area is generally well instrumented. The location of instruments on the watershed are shown in Figure 1. There

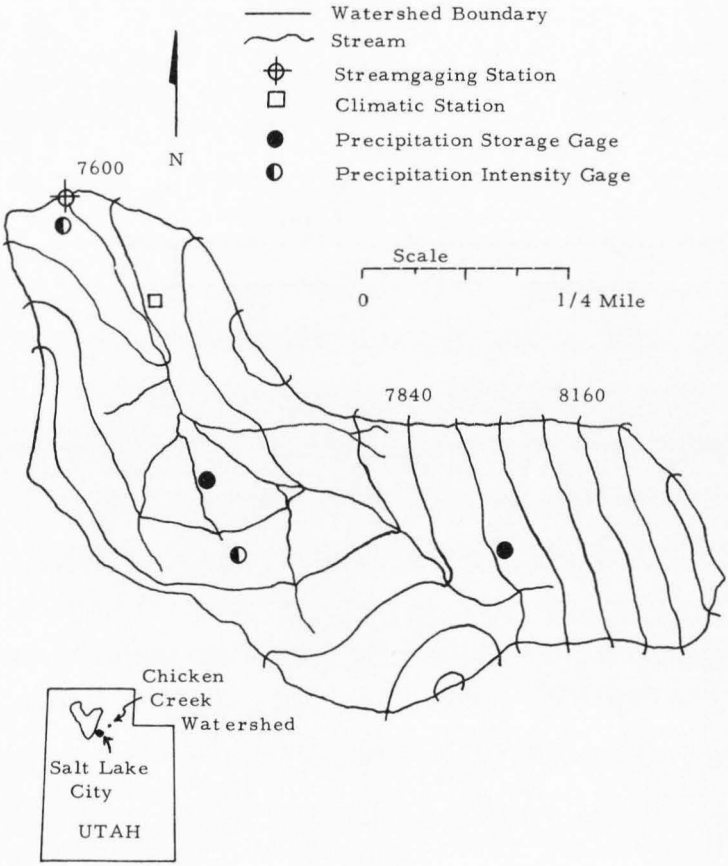

Figure 1. Topographic map of the West Branch Chicken Creek Watershed showing instrument locations.

are four precipitation gauges on the West Branch Chicken Creek watershed. There is only one temperature recorder in the area on the lower part of the watershed.

Streamflow measurements are complete starting in 1965 and continuing until present. Precipitation records are sprinkled with missing data, but storage gauges help keep the yearly totals fairly accurate. With the precipitation gauge density in the area these records should be fairly reliable; however, there has been some concern of snow being blown over the watershed divide from the west. This effect creates considerable variation in snowpack water contents between the lower and upper portions of the watershed. With only one temperature recorder in the area and an elevation range of 3,000 feet, the temperature data situation is not excellent, but not bad either. The proximity of the watershed to the Rice climatic station nearby at 6,900 feet elevation in the same drainage line to the west makes the situation good. The records at the Rice station are nearly complete for as far back as 30 years. Stream temperature data has been collected since 1971 for only the period of May through November.

Using the watershed instrument records together with the Rice station records, data for daily precipitation, minimum and maximum air temperature, streamflow and stream temperature has been assembled. Missing temperature data for the site were regressed from Rice station records using a linear regression equation with different coefficients for each month.

Data acquisition is an important step in the modeling procedure. With sufficient foresight in watershed instrumentation, considerable effort and error can be eliminated in a modeling attempt. Modeling is a relatively recent addition to a hydrologist's repertoire of tools and hence very little instrumentation has taken place with a watershed model in mind. In this model, snowmelt is the major streamflow contributor. Neither solar radiation nor cloud cover data is available for this watershed, therefore, a less accurate degree-day method is used. A result of this shortcoming is that the processes included in model structures and their accuracy are dependent upon the availability of data.

Model Formulation

With the previously mentioned purpose in mind, and in keeping within the constraints of the system and available data, a model can now be constructed. The model presented in this study is basically the storm model developed by Riley and Hawkins (1975) which was adapted to consider snow using a different time increment. In addition, the model was expanded to predict daily water temperature.

The watershed area is small, vegetation type is consistent, and geological origin of the soils homogeneous. For these reasons, the model represents a lumped watershed system in both the spacial and temporal dimensions on a daily basis.

Watershed characteristics, inputs, and responses vary considerably from place to place throughout the world. This variability should also be reflected in the structure of watershed models attempting to simulate them. For example, when modeling an area of relatively low infiltration rates one would expect a rather more sophisticated infiltration function than in an rea of high infiltration capacities. As has been noted earlier, the area modeled in this study receives most of its precipitation in the form of snow and most of the runoff is a result of snowmelt. This means that the modeling problem in this case will be one of developing a snowmelt function.

The movement of water through a watershed can be described as a series of storages with transfer functions describing the movement of the water among these storages. The model is assembled by routing the precipitation input through each hydrologic process in the same order the precipitation would be affected in the real watershed system. A diagrammed water routing flow chart is shown in Figure 3. Figure 2 shows how these processes in the flow chart relate to an actual watershed cross section.

Channel interception

A lumped model assumes that the inputs are uniformly distributed over the watershed. Under these conditions, precipitation intercepted by the stream channel can be considered as the same fraction of the total precipitation as the surface area of the stream is to the total area of the watershed.

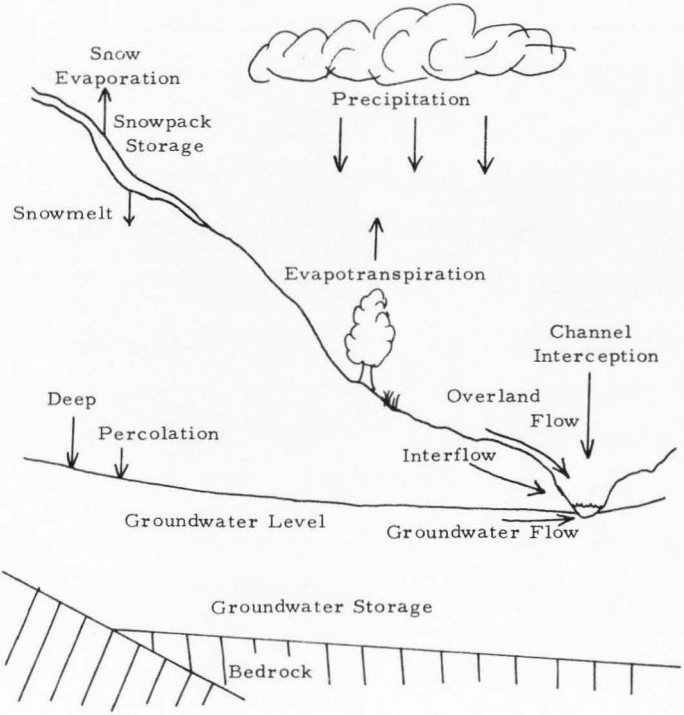

Figure 2. Watershed cross section showing model structure components as they relate to the real watershed system.

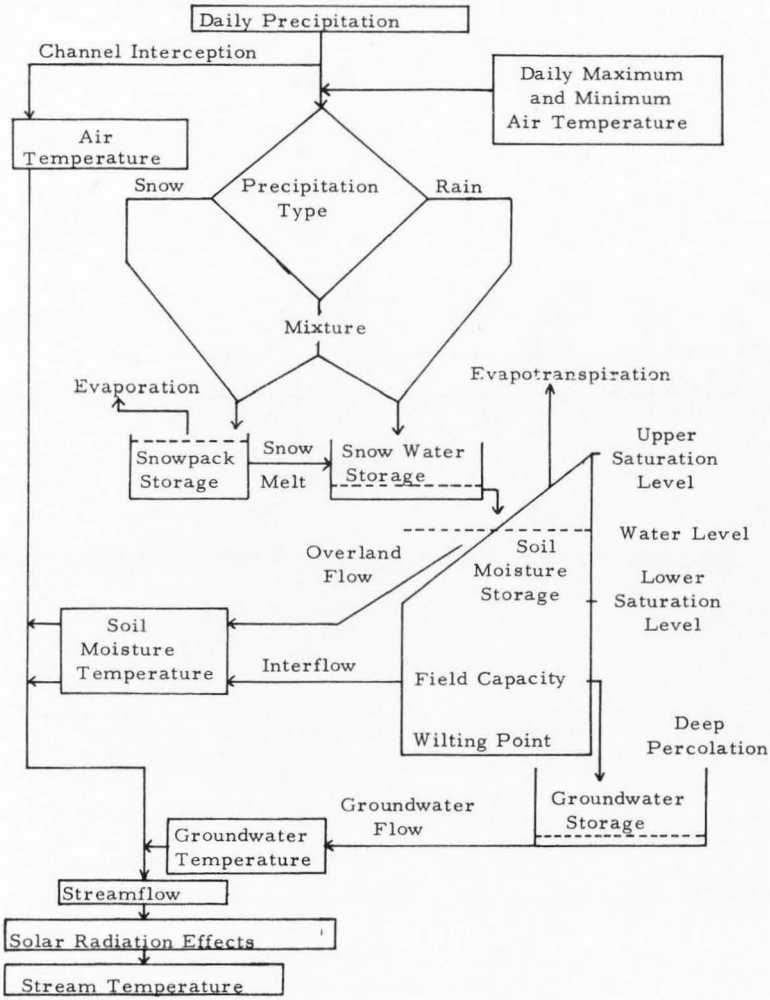

Figure 3. Water routing diagram showing model structure components for daily streamflow and stream temperature determinations.

$$\text{Channel Interception} = \text{ACHP} * \text{Precipitation} \qquad (1)$$

where: ACHP = Channel Interception Rate

This percentage is treated as a coefficient which will be evaluated in the calibration procedure later. The remainder of the precipitation will react with the watershed surface.

Precipitation type

To determine the form of the precipitation, a routine is used which was developed by the Army Corps of Engineers (1956).

$$\text{When Air Temp } 30^{\circ}\text{ F } \%\text{ Rain} = 0 \qquad (2)$$

$$\text{Air Temp } 30^{\circ}\text{F}$$

$$\text{and } 38^{\circ}\text{ F } \%\text{ Rain} = (\text{Air Temp} - 30)/_{8}$$

$$\text{Air Temp } 38^{\circ}\text{ F } \%\text{ Rain} = 100$$

This function is shown in graphical form in Figure 4. The precipitation is assumed to fall at the daily average temperature. This routine uses a straight line to approximate the percentage of precipitation which is rain or snow between two critical temperatures. At 30° F. precipitation is entirely snow and at 38° F. entirely rain. These two temperatures are assumed to be correct and thus are not calibrated and are not optimized coefficients.

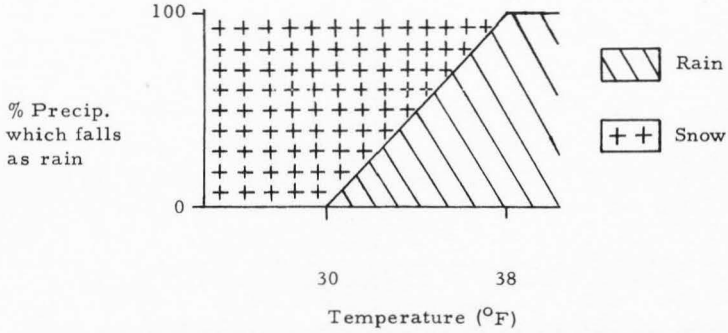

Figure 4. Precipitation type graph showing how differential is made between snow and rain events. Source, U. S. Army Corps of Engineers (1956)

Snowmelt

This precipitation is then collected in two storages at the soil surface. In one the water equivalent of the snowpack is stored as snow, the other stores the free water content of the snowpack. If snow is present, then the snowmelt is estimated. For this, the degree day method is used. For every degree the daily maximum temperature is above a base temperature, the snowmelt will increase by a fixed quantity called the degree day melt coefficient.

$$\text{Snowmelt} = \text{SMELT} * (\text{Max. Air Temp} - \text{T BASE}) \qquad (3)$$

where: SMELT = Degree Day Melt Coefficient

T BASE = Base Temperature

The United States Army Corps of Engineers has done research on this type of snowmelt index (1956). Their results show that this degree day melt coefficient varies through the accumulation and snowmelt

season. This variation is due mostly to the changes in the albedo of the snow surface. As the snow undergoes metamorphism, the amount of radiation absorbed is increased. In accordance with this phenomenon, the degree day melt coefficient is used as a function of the free water content of the snowpack. As more free water is held in the snowpack, the albedo is reduced and the degree day melt coefficient used in the model is increased. The Corps of Engineers study shows that on the average the albedo will vary from 80 percent on fresh snow to 40 percent on very old, ripe snow. According to these figures, the melt coefficient varies by a factor of two throughout the winter season. The coefficient value stated in Table 1 is the maximum value for a ripe snowpack. The minimum value is one half of this and occurs when the amount of free water held in the snowpack is very small relative to the total snowpack water equivalent.

Snowpack cooling

When the daily maximum temperature is below the base temperature, then the snowpack is considered to be cooling off at a rate of the cold content coefficient.

$$\text{Snowpack Cooling} = CC * (T\ BASE - \text{Max. Air Temp.}) \quad (4)$$

where CC = Cold Content Rate

This cold content of the snowpack is accumulated from day to day. The cold content of the snow pack must be brought to zero before the snow can start to melt. Once snowmelt is initiated the amount of snowmelt is transferred from snow storage to water storage. This

melting and cooling process adds three more coefficients, the melt, cold content, and base temperature coefficients. In addition to this melt, groundmelt is taken from snow storage and added directly to the soil moisture storage. The amount of groundmelt is constant for every day snow is on the watershed, and is optimized as a coefficient.

When snow is present, the surface water storage represents the free water content of the snowpack. The snowpack, however, will only retain a small percentage of its weight as free water, and the remaining water will drain from the snowpack. This maximum water content is a coefficient and the amount of surface water storage which exceeds this maximum water content is routed into the soil moisture storage.

Soil moisture storage

The soil moisture storage is considered to be a lumped linear reservoir. Water is drained from this storage either as interflow into the stream channel, or as deep percolation to the groundwater storage, or as overland flow when the soil becomes completely saturated. To determine when these different drainages occur, the amount of water in storage is considered relative to the amount held in storage by the physical forces present.

There are three general types of water retention in soil (Hewlett 1969): 1) gravitational water held between saturation and field capacity pressures which will drain under the force of gravity,

2) available water held between field capacity and wilting point pressures, this water will not drain under the influence of gravity, but is available for plant use and evaporation, and 3) hygroscopic water held by molecular attraction to the soil particle even against the highest root potentials. When the amount of water supplied to the soil moisture storage exceeds the amount the soil is capable of storing, saturation is exceeded and the extra water runs off immediately as overland flow.

The major runoff period results as snow melts on saturated soil. The importance of this overland flow in producing peak runoff has prompted the refinement of the saturation process from that used in the Riley and Hawkins model (1975). The level of saturation is considered to vary linearally over the watershed. Once the excess water has run off, water exceeding field capacity on higher parts of the watershed can drain down producing more excess flow the following day without further input to the soil moisture.

The water held between saturation level and field capacity level is treated as a lumped linear reservoir with two outlets, one to the stream channel and one to groundwater. Water held below field capacity cannot drain with the force of gravity. This moisture is, however, available for evapotranspiration processes.

Evapotranspiration

Evapotranspiration is indexed by the product of an optimized

coefficient and the daily maximum air temperature. This quantity is removed daily from the soil moisture storage unless a snowpack is present. This is assumed to occur at its potential rate as long as the soil moisture is above field capacity. Below this it becomes increasingly harder for plant roots to extract water from the soil until the rate becomes zero at the wilting point. In the model, a linear function is used with the evapotranspiration rate at the potential rate at field capacity, then linearally dropping to zero at the wilting point (see Figure 5).

$$\text{Potential Evapotranspiration} = \text{EVAPO} * (\text{Max. Air Temp} - 32) \quad (5)$$

where EVAPO = Evapotranspiration Rate

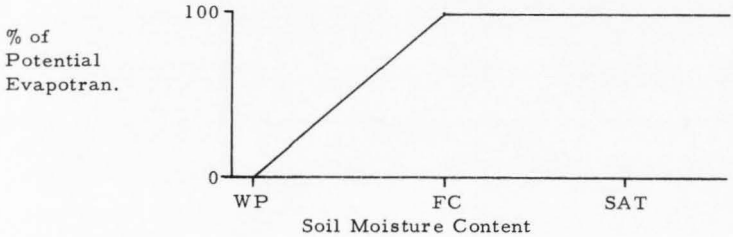

Figure 5. Evapotranspiration vs. soil moisture content.

Snow evaporation

When a snowpack is present no evapotranspiration is taken from the soil moisture but evaporation occurs from the snowpack. This amount of evaporation is indexed using the product of a snow evaporation

coefficient and the daily maximum air temperature above 32° F.

$$\text{Snow Evaporation} = \text{SNEVAP} * (\text{Max. Air Temp.} - 32) \quad (6)$$

where: SNEVAP = Snow Evaporation Rate

Flow through a porous substance has been described as being proportional to the cross sectional area of flow, the head per unit length, and a flow resistance coefficient which is a function of the porous substance (Dooge, 1973). Considering the entire watershed as a lumped linear reservoir, the area becomes unity, the head per unit length becomes total head, and the coefficient remains a coefficient which is a function of the soil type and other watershed characteristics.

Interflow and deep percolation

The two flows from soil moisture storage are then found by the product of a coefficient and the head of water above field capacity.

$$\text{Interflow} = \text{FQF} \times (\text{SM} - \text{FC}) \quad (7)$$

$$\text{Deep Percolation} = \text{FK} \times (\text{SM} - \text{FC}) \quad (8)$$

where: FQF = Inteflow Coefficient

FK = Deep Percolation Coefficient

SM = Soil Moisture Level

FC = Field Capacity Level

Groundwater flow

Deep percolation water is then added to groundwater storage and the same flow logic is applied to the groundwater reservoir to determine the groundwater flow.

$$\text{Groundwater Flow} = AGW \times GWL \qquad (9)$$

where: AGW = Groundwater Flow Coefficient

GWL = Groundwater Level

The daily channel streamflow consists of the sum of the channel interception, saturation excess flow, interflow from soil moisture, and groundwater flow. In the model presented by Riley and Hawkins, channel detention for the West Branch Chicken Creek watershed was found to be about one and one half hours using a one half hour time increment. No detention time or other channel routing is used in this twenty-four hour time increment model.

Soil moisture temperature

Stream temperature in this model is a function of two phenomena, net solar radiation and heat contained in the runoff as it enters the stream channel. In order to accomplish this the temperature of each contributor to streamflow must be found. As in the precipitation type function, the precipitation, and therefore the channel interception, is assumed to have the same temperature as the mean air temperature for that day. The two contributors to streamflow from the soil moisture reservoir are assumed to be at the soil temperature. The problem of finding this temperature is one of determining just where this flow occurs. Temperature fluctuations in the soil only occur in the top several feet below this the temperature remains steady (see Figure 6). If we assume this runoff temperature has an effect on influencing the

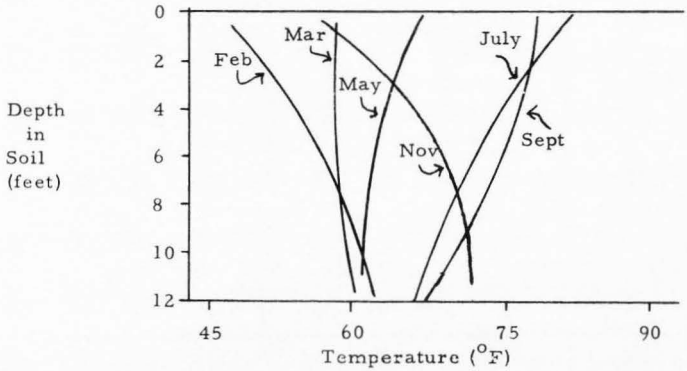

Figure 6. Temperature vs. soil depth. Shows variation in soil temperature with depth on a yearly basis. Source, Hausenbuiller. (1972) p. 146.

stream temperature, it must occur in this top several feet. We can then simply model the variation in soil temperature. From Figure 6 it can be seen that soil temperature fluctuations are similar to air temperature fluctuations on a yearly basis. Soil temperature extremes are not as great as the air temperature extremes and are lagged slightly. In the model, the soil moisture temperature is warmed or cooled according to the product of a coefficient and the difference between the average daily air temperature and the soil temperature.

$$SMT = SMT + FSMT * (\text{Ave. Air Temp.} - SMT) \qquad (10)$$

where: SMT = Soil Moisture Temperature

FSMT = Soil Temperature Variation

If snow is present, then 32° F. is used in lieu of the average daily air temperature.

Groundwater temperature

Groundwater, which is stored deep in the soil is assumed to be at a constant temperature. This temperature is optimized in the calibration procedure as a coefficient.

The temperature of the streamflow, before radiation is considered, is the average of the temperatures of the processes which are producing flow on that day, weighted by how large a portion each flow contributes relative to the total streamflow.

Solar radiation

Net radiation is indexed by the daily maximum air temperature. The coefficient used here is not constant throughout the year like the other coefficients. It is a function of the time of year and has a maximum value during the summer and minimum value during the winter. Another coefficient is calibrated to determine what day of the year the radiation coefficient is at its maximum. Figure 7 shows how the radiation coefficient varies throughout the year.

$$FMRC = (Max.\ FMRC) * (1/2) * (1 - SIN(2\ ((D-FMRCS)/365))) \quad (11)$$

$$\text{Solar Radiation} = FMRC * (Max\ Air\ Temp - 32) \quad (12)$$

where FMRC = Solar Radiation Rate

D = Number of Days Since Oct 1

FMRCS = Solar Radiation Shift

There are conditions and processes which have been left out of this model structure . In their description of the soils on the watershed,

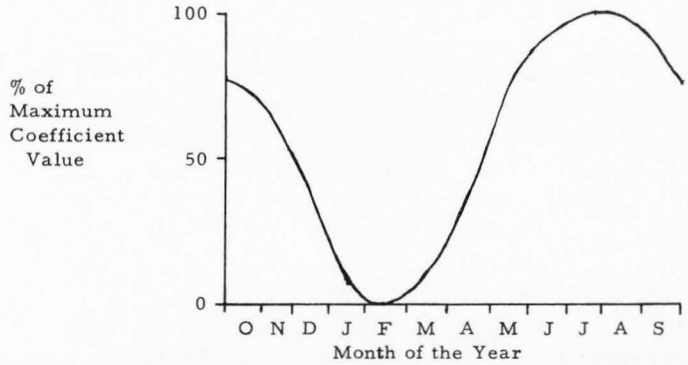

Figure 7. Variation in solar radiation index coefficient with time of year. Used in determining solar radiation effects on stream temperature.

Johnston and Doty (1972) judged the soils to be well drained. In the application of the Riley and Hawkins model in a classroom situation no periods of precipitation in excess of infiltration were found. These applications were for summer rainstorms which produce the most intense rainfall periods during the year. Even on a very warm spring day, snowmelt rarely exceeds one inch per day in the model. For these reasons, infiltration limits were not included in the model. Interception was also not considered. The reason for this was that it was deemed insignificant as far as the final result was concerned. The snowpack evaporation constant will account for these losses. One condition which was not included because of the inability to document its presence was frozen soil. Under snowmelt conditions, frozen soil can severely limit infiltration and, as a result, produce huge

fluctuations in streamflow. The inclusion of this condition was not within the resolution of this model. The results of the model indicate that this problem was not of significant consequence.

These processes and conditions which have been left out of the structure are still occurring on the watershed and will influence the resulting values of the coefficients. For example, the snow is assumed to melt evenly over the entire watershed. This will not occur in the real system however and, as a result, the degree day melt factor and the slope of the saturation line will be influenced by this effect. It can be seen from this that if all of the major water routing processes acting within the watershed are not included, the meaning of coefficients which intend to represent certain physical watershed parameters will tend to become obscure, or the model becomes more of a black box.

Model Calibration and Verification

It is in this step that the modeler gains an intimate understanding of the workings of the model and the system it attempts to simulate. The coefficients used in the structure of the model must now be evaluated.

There is no predetermined procedure for doing this. One technique used is a patterned trial and error process in which the coefficients are simply varied until the best fit is found between the calculated output and the measured observed data.

An objective function is used to determine just what the best fit means. This function is a statistical comparison of the calculated and observed outputs in order to get a number, representing the degree of best fit, which can be compared from trial run to trial run. The objective function used is similar to a standard error calculation. It uses the same equation but does not have a statistical basis of being identical to a statistical standard error. The object, then, in the calibration procedure is to minimize this standard error. Other forms of this objective are mentioned in Table 2. These functions are in the form of average standard errors. They are used to compare the best fit of calibrated models used in modeling watersheds with different flow regimes.

This seemingly random trial and error approach can be patterned. The coefficients are tested one at a time and only changed to values which result in a lower objective function. The pattern is then repeated until a point of diminished return is reached with respect to lowering the objective function.

As can be imagined, this procedure can become very time consuming as the number of coefficients becomes large. With 14 coefficients in this model, the objective function then is a 14 dimensional function. If a graph of this objective function could be drawn, it would show a lot of local maximum and minimum points. One of the most criticized areas of this approach is the possibility of getting stuck in one of these local minimums and not finding the overall minimum. However,

if the full range of reasonable coefficient values is considered, the likelihood of this occurring can be reduced considerably.

In a model of this type, the coefficients are not quantities which can be directly measured on the watershed but rather a function of several of these measureable characteristics. On the other hand, there are some practical limits which can be placed on them. For example, none of the coefficients used in this model can have a negative value. This would mean negative flows would occur, which is not practical. The sum of the two coefficients determining flow from the soil moisture reservoir and also the groundwater coefficient by itself cannot be greater than unity. If this occurs, then more water will drain than is available to drain and the model will be creating water.

Apart from these obvious limits, some limits can be made with a little hydrology logic. Maximum values for snowmelt and evapotranspiration days are generally known and the coefficients should not result in too much snowmelt or evapotranspiration. A rough estimate of the channel interception coefficient can be made by finding the surface area of stream channels on the watershed. Groundwater inflow temperature has a generally known range. Degree day coefficients for snowmelt have been studied extensively by the Corps of Engineers (1956).

One of the advantages of a model which has a structure that is analogous to the hydrologic system is that these different processes can be analyzed independently. For example, the level of groundwater

storage should remain fairly constant from year to year. It should not increase by an order of magnitude or run dry on this watershed. These factors also were considered in the calibration process.

In addition to the objective function, the coefficient of determination (R^2) and mass balance between the calculated outputs and observed outputs were calculated. These values were used as secondary criteria in determining the best fit.

This calibration is repeated for each year of data set aside for calibrating the model. Now the modeler will have a minimum objective function with a set of coefficients for each year calibrated. The first indication that the model is working will be that these sets of coefficients will be fairly consistent. They will not be the same, even for an excellent model. They will reflect peculiarities in the inputs and watershed characteristics which vary from year to year or are not accounted for in the model. For example, the direction of storm approach or air temperature lapse rate may vary considerably creating slightly different streamflow reactions on a small watershed, while recorded data at the instruments may be the same. If this variation in coefficients is minimized then the model will yield better results because all of the major processes affecting streamflow have been included.

If a model is going to be used in any management or research endeavor, it must accurately predict outputs of streamflow and stream temperature for any set of inputs. This means the model must be general enough to handle the full range of variation of inputs and outputs

for any year. In order to do this one set of coefficients must be found from the group of calibrated sets. The procedure for choosing this common set of coefficients in this study was simply to use the average of the coefficients found in each calibrated set of coefficients.

The verification of the model tests its ability to be applied to one of these situations. The data set aside for verification is run using the common set of coefficients. The model should be able to simulate this independent data satisfactorily.

Interpretation of Results

The coefficients found in the calibration procedure and the statistical results they represent are listed in Table 1 and Table 2. These results indicated that only a mediocre degree of fit has been achieved.

The main reason for this lack of correlation seems to be the result of inaccurate air temperature data recreation. For example, during January of the first streamflow calibration year, there is a sudden rise in streamflow to a level nearly one half of that of the peak flow for that year in only two days. During this time, the mean air temperature never rises above freezing temperature. The air temperature data during this period is regressed from the Rice climatic station. This inconsistency suggests that something is wrong either with the streamflow records or the regression curves. In addition to this obvious inconsistency, during the calibration to model structure feedback, different snowmelt indexes were tried with varying degrees of

Table 1. Table of coefficient symbols, descriptions and values found during calibration.

Coefficient Symbol	Coefficient Description	Coefficient Value
	STREAMFLOW	
TBASE	Base Temperature	36° F
UPSAT	Upper Saturation Level	31.4 in.
LOSAT	Lower Saturation Level	16.0 in.
FC	Field Capacity	10.0 in.
WP	Wilting Point	0.0 in.
FQF	Interflow Rate	0.028 in./in.
FK	Deep Percolation Rate	0.0025 in./in.
AGW	Groundwater Rate	0.002 in./in.
SMELT	Maximum Snowmelt Rate	0.032 in./F° Day
SNEVAP	Snow Evaporation Rate	0.001 in./F° Day
EVAPO	Evapotranspiration Rate	0.0021 in./F° Day
ACHP	Channel Interception Rate	0.0012
GMELT	Groundmelt Rate	0.0085 in./Day
CC	Cold Content Rate	0.0015 /F° Day
FMWC	Maximum Snowpack Water	0.065
	STREAM TEMPERATURE	
FMRC	Maximum Solar Radiation Rate	0.51 °F/°F
FMRCS	Solar Radiation Shift	20 Days
FSMT	Soil Temperature Variation	0.007 °F/°F
FGT	Groundwater Temperature	40.0 °F

Table 2. Table of statistical results found while using the calibrated coefficients.

STREAMFLOW

STATISTIC	WATER YEAR 1970 - 1971	WATER YEAR 1971 - 1972
Initial Soil Moisture Level	8.5 in.	6.0 in.
Standard Error	0.4941 cfs	0.4913 cfs
Coefficient of Determination	0.870	0.882
Average Standard Error	69.46 %	74.56 %
Mass Balance	-8.36 %	-3.79 %
Total Precipitation	38.7 in.	41.6 in.
Total Streamflow	28.5 in.	26.4 in.

STREAM TEMPERATURE

STATISTIC	WATER YEAR 1971 - 72	WATER YEAR 1972 - 73	WATER YEAR 1973 - 74
Initial Soil Temperature	32° F	32° F	32°F
Standard Error	4.1°F	3.1°F	5.2°F
Coefficient of Determination	0.863	0.885	0.490
Average Standard Error	25.45 %	16.60 %	31.82 %
Total Precipitation	41.6 in.	29.0 in.	37.8 in.
Total Streamflow	26.4 in.	16.9 in. *	21.7 in. *

*Measured data not available, values predicted by the model

expected accuracy. The mean daily temperature was tried as well as fitting a sine function to the daily maximum and minimum temperatures and figuring degree days both above and below a base temperature. Despite these different methods tried, the resulting objective functions did not vary significantly.

Another problem was encountered when precipitation was assumed to fall at the average daily air temperature. In some cases, rain fell when there was no reaction from the streamflow hydrograph or snow fell when the hydrograph responded. This problem seems to exist mainly in late fall and early spring when the daily temperature fluctuates widely about a mean in the middle 30° F. range. Another minor problem arose when precipitation in the form of rain fell late at night just before midnight. In the model, the first runoff from this came on that same day while, in reality, the runoff occurred the following day. This is a problem in not allowing for the short lag of maybe an hour or two in the streamflow response to rain.

Streamflow temperature predictions also were mediocre. Again the assumption that channel interception fell at the mean daily air temperature caused some aberrations in streamflow temperatures. The observed streamflow temperatures were less variable than the model represents. Using the temperature of the inflow to the stream channel to determine stream temperature before the effects of net radiation, one might think that the stream temperature would always correlate weaker than streamflows. It was my experience, however,

that, on this watershed, inflow temperatures of interflow did not have a very sizable influence on the stream temperature correlation. Only the summer stream temperatures were modeled in this study and nearly all of the runoff occurs in the early spring. As a result, most of the correlation is done for the low flow summer season which is the time when temperature is critical. This again indicates either the failure of the degree day to index streamflow temperature or a lack of accurate air temperature or stream temperature data.

The model consistantly predicts higher stream temperatures than are observed during the short time the runoff and observed temperatures occur simultaneously. This is a result of considering the stream discharge and surface area as being constant when the radiation effects are calculated. In reality the stream temperature will be directly proportional to the surface area of the stream and indirectly proportional to the discharge. Discharge was the only one of these two variables for which data was available. Better results were found by assuming the ratio of the surface area to discharge for the stream was constant rather than only varying the discharge.

Six years of streamflow, precipitation, and air temperature data were available for calibration and verification. The first four years of this data relied on regressed air temperature data for the entire year. Originally, the first four years were to be calibrated and the last two years used for verification of the model. When attempting to derive one common set of coefficients for the first few years, a

total lack of consistency was found from year to year. The problem is the weak correlation of the regressed air temperature data from the Rice climatic station. For this reason, the last two years of data with mostly on site air temperature data were considered more indicative of the true situation, therefore these were used to calibrate the model. As a result of this decision, there was no data to verify the model.

At this point, the model is not calibrated or verified to the desired accuracy. In keeping with the general modeling procedural steps, this calls for a reexamination of the previous steps. These steps have been reevaluated in the preceeding discussion. Application of these reevaluations will require further data measurement and reduction, and time, which is the reason this study ends here.

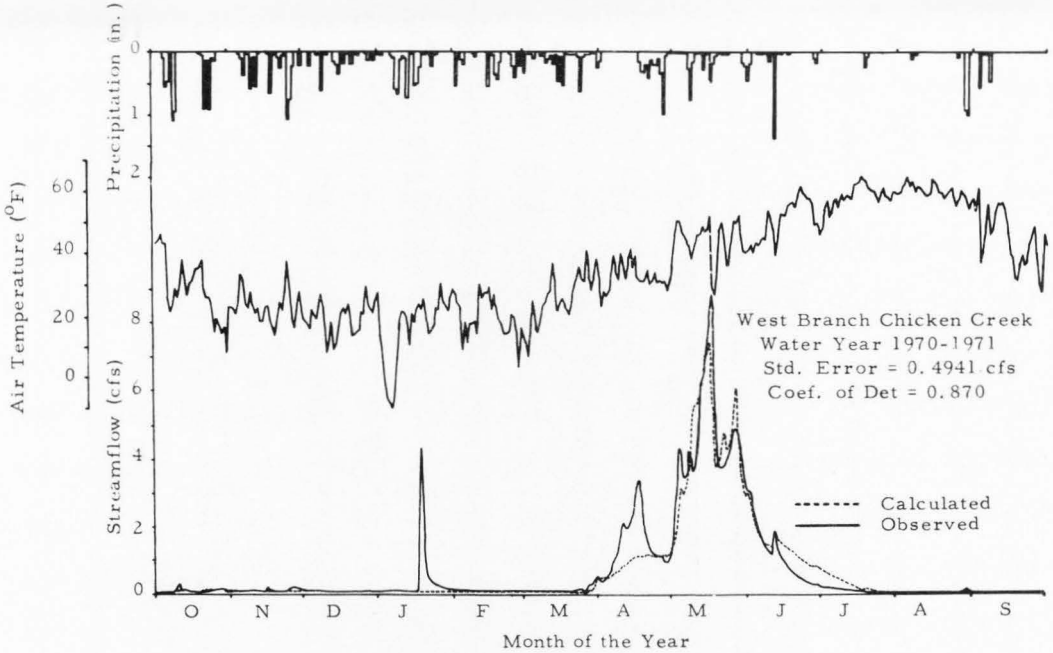

Figure 8. Observed and calculated streamflow, mean daily air temperature, and daily precipitation vs. time of year for the water year 1970-1971 calibration using a common set of coefficients.

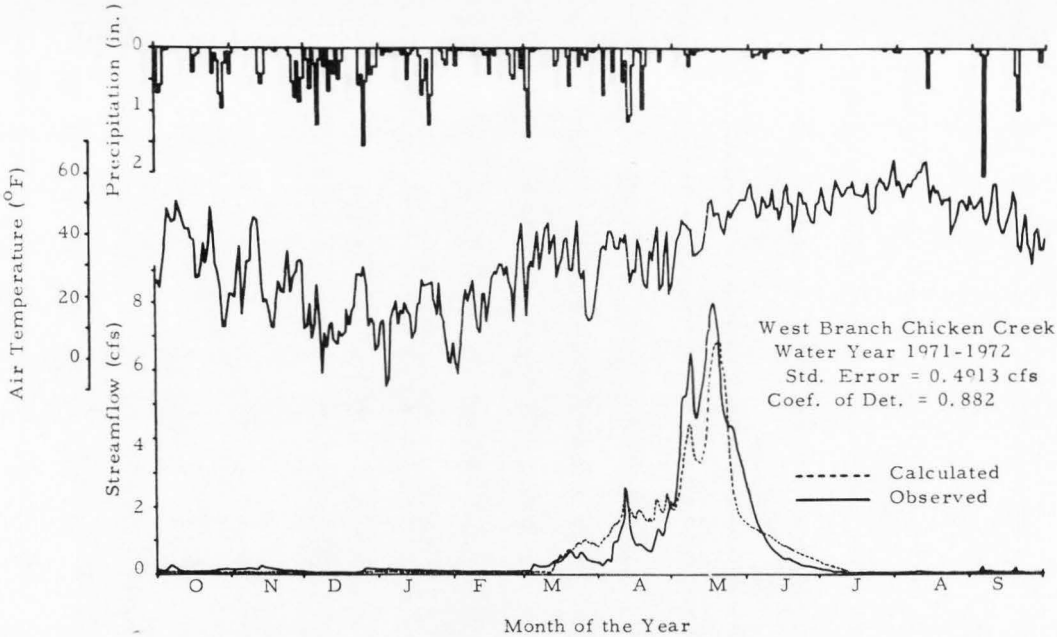

Figure 9. Observed and calculated streamflow, mean daily air temperature, and daily precipitation vs. time of year for the water year 1971-1972 calibration using a common set of coefficients.

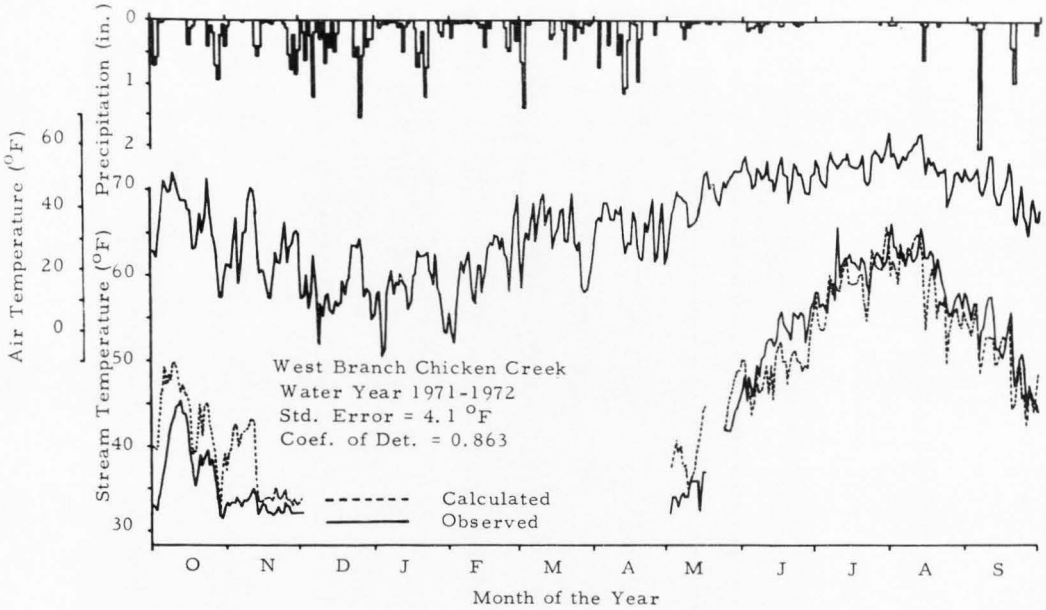

Figure 10. Observed and calculated stream temperature, mean daily air temperature and daily precipitation vs. time of year for the water year 1971-1972 calibration using a common set of coefficients.

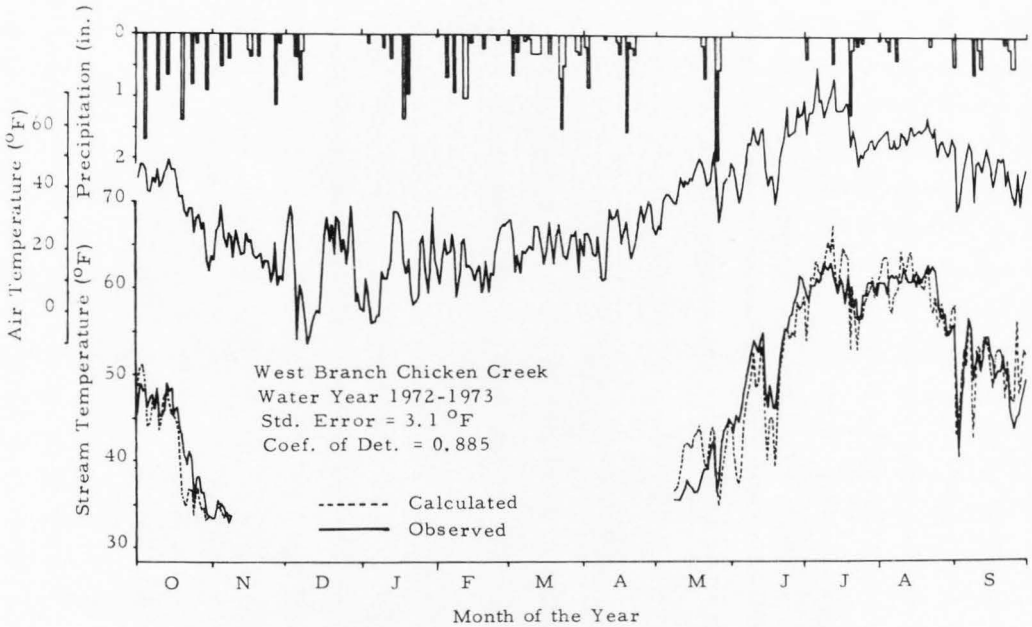

Figure 11. Observed and calculated stream temperature mean daily air temperature, and daily precipitation vs. time of year for the water year 1972-1973 calibration using a common set of coefficients.

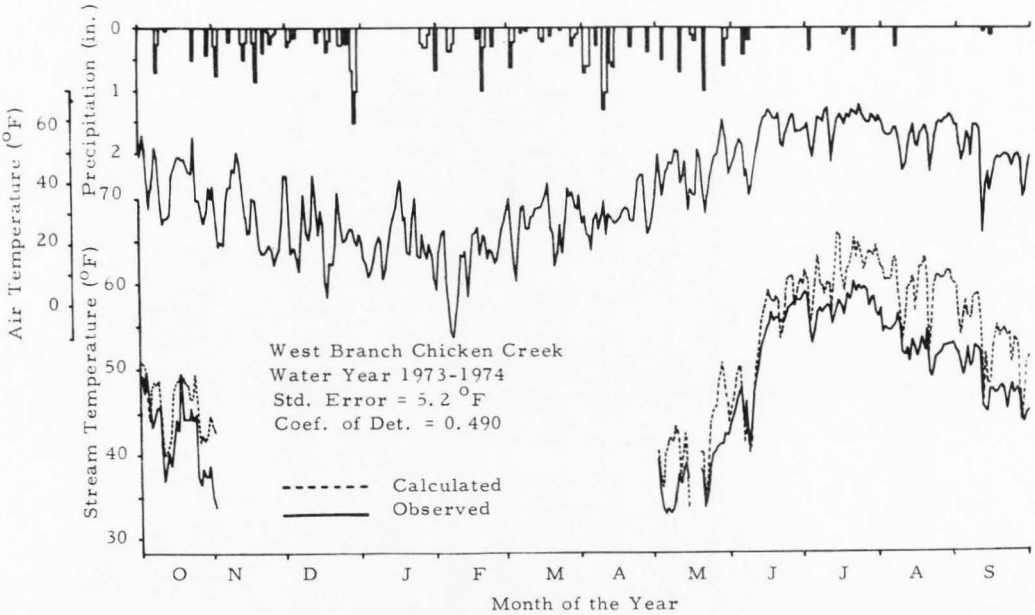

Figure 12. Observed and calculated stream temperature, mean daily air temperature, and daily precipitation vs. time of year for the water year 1973-1974 verification using a common set of coefficients.

INSIGHTS INTO THE SYSTEM

One of the objectives of this study was to provide some insight into the hydrologic processes as they occur on the watershed. It is appropriate now to make some comments on this subject. The soil moisture storage, groundwater storage, and channel interception structure in this model and the one presented by Riley and Hawkins (1975) are identical. The basic difference between the two models is the length of the time increment used. Having had experience with both of these models, I will also make comparisons between the two.

Considerable information can be found about the soil moisture storage and drainage properties in general on the watershed by examining the individual and relative magnitudes of the coefficients involved in this process.

The relative capacities for the various moisture holding storages in the soil can be determined. In this calibration, it was found that considerably more moisture was held in available status compared to that which would drain. Also, the moisture level was seldom above field capacity during the summer months and only during the spring snowmelt season was it anywhere close to the saturation point. This suggests that the watershed soil has considerable storage capacity when exposed to intense summer rainstorms, and the only precipitation producing runoff during summer months is that falling directly on,

or adjacent, to the stream channel. This also was found to be the case in the application of the Riley and Hawkins model to this watershed.

By summing the two soil drainage coefficients, the fraction of the amount of moisture available to drain that will drain during one day can be found. Further relations can be derived, as follows, which yield the fraction of the total soil moisture above field capacity in excess of evapotranspiration which drains directly into the stream channel as interflow and to groundwater storage as deep percolation respectively.

$$\text{Interflow} = (FQF) / (FQF + FK) \tag{13}$$

$$\text{Deep Percolation} = (FK) / (FQF + FK) \tag{14}$$

In this twenty four hour time increment model, these fractions are .918 and .082 respectively, as compared to .009 and .991 in the one half hour time increment model.

Figure 2 indicates definite positions and boundaries for the storage components contributing to streamflow within the watershed. In the model, these storages are actually not defined as specific places on the watershed, but are defined in terms of how long it takes water to travel from the storage area to the stream channel with respect to the length of the time increment used. Channel interception is that precipitation which reaches the channel within the length of the time increment. Interflow is delayed slightly and groundwater delayed considerably with respect to the length of the time increment. This means that these storage area boundaries will change with changes in

the time increment. In the model by Riley and Hawkins, most of the runoff from summer rainfall was found to be a result of channel interception, and most of the precipitation entering the soil percolated to long term storage as groundwater. This suggests that the routing of precipitation through the watershed takes longer than the one half hour time increment. In this model, when the time increment is increased to twenty four hours, the channel interception percentage of precipitation is increased because more water can reach the channel in twenty four hours than one half hour. It is interesting to note that this increase is not in the same proportion to the increase in the time increment. This is because in the model the precipitation contributing to the channel interception is not all falling directly on the channel. Some is contributed from the soil in the immediate vicinity of the channel. This means that there are two rates of delay included in this one process. The delay due to the resistance to flow of the soil in the immediate vicinity of the channel and the instantaneous response of the precipitation falling directly on the stream. The proportion of water in soil storage contributing to interflow is nearly one hundred percent. From this we can deduce that the delay to water entering the soil moisture storage and not evaporating is somewhere in the order of the length of the time increment, which is twenty four hours. In the half hour incremented model, runoff which occurred was nearly all channel interception. This concurs with the fact that most of the precipitation which fell went into long term (with respect to one

half hour) storage as groundwater. In the twenty four hour incremented model the storms for which the Riley and Hawkins model predicts channel interception flow and interflow produce only very small channel interception flows.

CONCLUSION

One of the shortcomings of a piece by piece study of a system is that in reality, it is a conglomerate of many interdependent factors with infinite variety. It is almost impossible to separate these factors so that each can be studied as an independent process. A well constructed watershed model is a step closer to being able to study a watershed system as it is continuously functioning.

Models have just recently been developed commensurate with the introduction of computational systems. With continued research in the modeling field, models will grow more dependable and accurate as both the computational systems and modeling theory and techniques become more refined.

In this study, a model has been created and calibrated to a small mountain watershed. The results indicate that there is not a linear, not even a direct proportionality between the length of time increment used and the coefficients determined with a given structure. They vary with the length of calculation increment, but the way they vary depends on the characteristics of the watershed.

Further study into the effect of using different time increments is needed. A model using twelve or six hour time increments would be interesting. This also would correct some problems encountered in

this study, concerning the temperature of the precipitation as it falls. The smaller the time increment, the more accurate is the assumption that precipitation falls at the mean air temperature during that time increment.

The model should also be calibrated to differing watersheds to understand more clearly the resulting coefficients for different conditions. This is needed especially before the model can be reliably used to predict long term planning activities.

On this watershed, using the temperature of the influent streamflow in calculating the stream temperature may be no more accurate than simply using a solar radiation index on a constant temperature stream. Nearly all of the runoff occurs in the early spring as snowmelt which is not when the critical stream temperature period occurs. If the influent flow's temperature is deemed of importance on a watershed, a study of the temperature of water as it enters the stream would be a valuable bit of information.

The results of this calibration are not accurate enough for the model to be used as a management or research tool. The reason for this failure is thought primarily to be a result of the inaccurate air temperatures recreated from regression equations. With more data becoming available, another try should be made with a few changes, as suggested in this study, in order to come up with a usable small watershed model.

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APPENDIX

PROGRAM LISTING IN WANG BASIC

```
10 COM A(36,32)
20 RLM  ********      COEFFICIENTS      ********
30T1=36:T2=30:X1=25:Y1=36:V=9:T5=36:L1=10:L2=16.0:L9=.065:L3=6.0:F3=.0
28:F4=.0025:F6=.002:M3=.032:F9=.0015:K=.001:K1=.0021:F7=.065:A2=.0012:
M4=.0085:C5=0.51:C8=.007:C9=40.0:T8=32
40 R9,C1,L7,L8,P6,N=0:Z6=1/L1:Z7=1/L9:Z2=.5*Z7:Z9=L2+Z7:G1=.0031/F6:FO
R I=X1TO Y1:V=V+1:IF V[13THEN 60:V=V-12
50 REM ********      INPUT ROUTINE      ********
60 FOR J=1TO 32:IF A(I,J)[0THEN 540:Q=(INT(A(I,J)*.000000001))*.1:P=(I
NT((A(I,J)*.000000001-Q*10)*1000))*.01:T=INT(((A(I,J)*.000001)-INT(A(I
,J)*.000001))*1000)*.1:R9=R9+1
70 B8=((A(I,J)*.001)-INT(A(I,J)*.001))*100:IF T[90THEN 80:T=-1*(T-90)
80 IF Q[90THEN 90:Q=(Q-90)*(-1)
90 T3=T-(B8*.5): T4=T+(B8*.5)
100 REM ********      CHANNEL INTERCEPTION      ********
110 Q8=A2*P:  P3=P-Q8
120 REM ********     PRECIPITATION TYPE      ********
130 IF T[=T2THEN 140:IF T[T1THEN 150:P1=P3:P2=0:GOTO 170
140 P2=P3:P1=0:GOTO 170
150 P1=((T-T2)*.125)*P3:P2=P3-P1
160 REM *********      SURFACE STORAGE      ********
170 L8=L8+P2:  L7=L7+P1
180 REM ******** GROUNDMELT AND SURFACE TEMP      ********
190 Q9=0:T9=T:IF L8[=0THEN 350:T9=32:Q9=M4:IF (L8-Q9)]=0THEN 210:Q9=L8
:L8=0
200 REM ********      RADIATION SNOWMELT      ********
210 L8=L8-Q9:L3=L3+Q9:M1=M3:IF L7]=(L8*F7)THEN 230:M1=M3*.5*(1+(L7/(L8
*F7)))
220 REM ********      SNOW EVAPORATION      ********
230 E=K*(T4-32):IF E]=0THEN 240:E=0
240 IF E[=L8THEN 250:E=E-L8:L8=0:L7=L7-E:IF L7]0THEN 270:L7=0:GOTO 270

250 L8=L8-E
260 REM ********      DEGREE DAY DETERMINATION      ********
270 D1=T4-T5:IF D1]=0THEN 280:C1=C1+F9*D1:D1=0:GOTO 290
280 M2=M1*D1:IF C1]=0THEN 300:C1=C1+M2:IF C1[=0THEN 290:M2=C1:C1=0:GOT
O 300
290 M2=0
300 IF M2[=0THEN 320:IF M2[L8THEN 310:L7=L7+L8:L8=0:GOTO 320
310 L7=L7+M2:L8=L8-M2
320 IF L7[=(F7*L8)THEN 330:R=L7-F7*L8:L7=F7*L8:GOTO 370
330 R=0:GOTO 370
340 REM ********      EVAPOTRANSPIRATION      ********
350 R=L7:L7=0:E=K1*(T4-32):IF E]=0THEN 360:E=0:GOTO 370
360 IF L3]=L1THEN 370:E=E*(L3*Z6)
365 REM ********      OVERLAND FLOW      ********
370 L3=L3+R-E:IF L3[L2THEN 390:IF L3[Z9THEN 380:Q3=(L3-(Z9)+.5*Z7):L3=
L3-Q3:GOTO 400
```

```
380 Q3=(L3-L2)*(L3-L2)*(L9)*(.5):L3=L3-Q3:GOTO 4 00
390 Q3=0
400 IF L3[L1THEN 410:H=L3-L1:GOTO 430
410 H=0
420 REM ********      INTERFLOW      ********
430 Q4=F3*H
440 REM ********      DEEP PERCOLATION      ********
450 Q5=F4*H
460 REM ********      GROUNDWATER FLOW      ********
470 L3=L3-Q4-Q5:G1=G1+Q5:Q7=G1*F6:G1=G1-Q7
475 REM ********      SUM THE FLOWS      ********
476 Q2=Q3+Q4+Q7+Q8
480 REM ********      SOLAR RADIATION EFFECTS      ********
490 T8=T8+C8*(T9-T8):T6=((Q3+Q4)*T8+T*Q8+C9*Q7)/(Q4+Q3+Q8+Q7):C6=T4-32
:IF (T4-32)]=0THEN 500:C6=0
500 C7=C5*.5*(1-SIN(((R9-20)/365)*2*#PI)):T6=T6+C7*C6:PRINTUSING 510,P
,Q2,T3,T,T4,(T6-32)*(5/9),Q,V,J
510%P=#.## Q=#.#### T=-##.#-##.#-##.#  C=-##.# -##.# =0  DATE ## ##
520 REM ********      STATISTICS      ********
530 IF Q]80THEN 540:Q3=(T6-32)*(5/9):C=C+Q:B=B+Q3:D=D+(Q*Q):S=S+((Q-Q3
)*(Q-Q3)):P6=P6+P:N=N+1
540 NEXT J:NEXT I:S1=(S/N)!.5:S2=((D-((C!2)/N))/N)!.5:R2=1-(S1/S2)!2:P
RINTUSING 550,S1,R2,S1/(C/N),G1-(.0031/F6),P6,N:END
550%STND DEV= #.#####      R!2= -##.###      AVE SE = ##.#####          DE
L SS = -##.####     TOT PRECIP = ###.##        N = ####
```

VITA

Charles Wilson Pettee

Candidate for the Degree of

Master of Science

Thesis: A Hydrology Temperature Model for a Small Mountain Watershed.

Major Field: Watershed Science

Biographical Information:

Personal Data: Born January 7, 1951, in Indianapolis, Indiana, son of Horace and Winifred Ermine Pettee; married Penny Sue Demaree Pettee May 19, 1973.

Education: Attended elementary school in Indianapolis, Indiana; graduated from Thomas Carr Howe High School in 1969; received Bachelor of Science Degree in Aeronautical and Astronautical Engineering from Purdue University in 1973; completed requirements for Master of Science degree in watershed science at Utah State University in 1976.

Professional Experience: 1974 to present, Graduate Research Assistant, Utah State University, Logan, Utah; 1973 to 1974, Mass Properties Engineer, Hughes Aircraft Company, Space and Communications Group, Culver City, California.