ANALYSIS OF THE RESPONSE AND THE RAINFALL DISTRIBUTION
IN A MOUNTAINOUS WATERSHED

By

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Abstract

The hydrologic response of a small, steep and forested watershed is investigated with the use of a simplified watershed model. The watershed model utilises the linear storage routing technique and treats the runoff as fast and slow. The separation of the rainfall input to each of the components of runoff is achieved with the use of a non-linear exponential equation (Eq. 3.1). Hourly rainfall data averaged from five raingauges and hourly streamflow are used as input to the model.

The model is found to perform well. Analysis of the variation of model parameters and the resulting linear form of the rainfall separation equation (Eq. 5.1), after the application of the model, indicates that the hydrologic response of the watersed is reasonably linear except for intense summer rainstorms under dry soil conditions. The linearity of the response of the watershed can be attributed to the runoff generation mechanism which is the subsurface pipeflow. Studies of the dynamics of pipeflow are reviewed and validated for the study watershed and it is shown that the parameters of the model take values that are consistent with their physical representation.

Reviewing the literature, it has been recognized that the rainfall distribution with elevation is different from linear for mountainous areas with humid climate. Analysis of the rainfall distribution in the study watershed reveals that the linear increase of precipitation with elevation is not valid for rainfall. The rainfall depth per event increases up to the mid-elevation and then decreases at the upper elevations. Furthermore, the hourly intensity decreases with height and the two dominant hillslopes do not receive the same amount of rain. This variation in rainfall is explained by meteorological precesses and the physiographic features of the watershed.
The performance of the model is highly affected by the accuracy and representativeness of the rainfall data. For example if weighted averages of several precipitation stations are used, the model performs well. However, if only a single point rainfall is used, the model performance becomes less good. As a result, unless the rainfall and runoff are known with great precision, it is difficult to make an accurate determination of the storage characteristics of the basin, and consequently it is argued that it may not be possible to distinguish between the accuracy of different routing models.
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Chapter 1

INTRODUCTION

Runoff from steep mountainous watersheds is a major contributor to total streamflow response in the Pacific Northwest. In addition, forestry activities may modify the watershed response, and fisheries and engineering projects may be affected by the volume and distribution of runoff and the resulting sediment transport.

Usually, it is suggested that the response of a small watershed to rainfall input is rather non-linear, and increases with rainfall intensity. This non-linear response is attributed to the infiltration process and the interaction of the unsaturated-saturated water flow in the soil.

Also, questions arise about the effects of precipitation distribution and magnitude on the streamflow. The precipitation can be either rain or snow and previous studies have shown that these two forms of precipitation have different distribution patterns (Fitzharris, 1975, Storr and Ferguson, 1972). Furthermore, the knowledge of rainfall distribution in space and time is very significant in the assessment of the resulting streamflow. Confusion exists among hydrologists about the rainfall distribution in mountainous regions. Many hydrologists suggest that the rainfall increases linearly with elevation up to the crest of the mountain and others claim that the rainfall increases up to a certain elevation and then levels off or even decreases at the upper elevations.

In contrast with rain, snowfall variation with elevation seems to be more consistent making the estimation of snow distribution easier. But the problem in snow hydrology
is different, because the accumulation of the snowpack during many events is more significant than its areal and temporal distribution during a particular storm. Also, the melting rates of snow can reasonably be predicted using formulas based on temperature indices.

Examination of the watershed behaviour can be achieved by the use of a hydrologic model. Some more complex physically based models require extensive information on the watershed, as well as a reliable meteorological database, much of which is difficult to obtain (Beven, 1989). This requirement prohibits their use in mountainous areas where data are usually, very limited. On the other hand, regression analysis between precipitation and streamflow usually gives no indication about the distribution of flow with time. It is important, to use a method which can be used to examine both volume and time of the response. This study sets out to examine the usefulness of a simple watershed model which has been designed to operate with a limited database and uses the linear reservoir routing technique.

The objectives of this study are:

a) to identify possible changes in the response of a small, forested and steep watershed under a variety of large rainfall events, and to examine the rainfall distribution in the watershed. An event based storage routing model will be used to simulate streamflow and to examine the watershed response. Accurate data taken at short-time intervals are required for testing the performance of the model and for evaluating the watershed response. Actually, this research examines the validity of modelling assumptions for steep mountain watersheds. The linear response of the watershed which is the basic assumption in the development of many hydrological models will be tested.

b) to examine and analyse the effects of rainfall representativeness on the streamflow simulation and to test the accuracy of the rainfall distribution using predictor equations.

Confusion has arisen because there are at least two quite different runoff mechanisms
and a clear distinction needs to be made between these two processes. One mechanism, saturated overland flow, is of primary importance for low slope watersheds with deep soils and gives rise to a non-linear runoff behaviour. The second mechanism, pipeflow, only develops on steep hillslopes and it will be argued and demonstrated that its response is much more linear. This distinction is important and forms the basis for much of the discussion in Chapter 2.

The model designed for the purpose of this thesis and the method of analysis used will be presented in Chapter 3. This model was made deliberately simple and was intended for a preliminary analysis before designing a more complex model. However, the analysis showed that the model was sufficiently representative of the watershed behaviour. The real difficulty turned out to be that of accurately characterising the rainfall input.

Chapter 4 will give the necessary information about the study watershed. The results of the application of the method of analysis to the study watershed will be presented and discussed in Chapter 5. The results of the analysis of rainfall in the study watershed and the effects of rainfall representativeness on the simulation of runoff will be presented, discussed and compared with a survey of the literature on the distribution of precipitation in mountainous areas in Chapter 6.
Chapter 2

BACKGROUND - LITERATURE REVIEW

2.1 INTRODUCTION

In this chapter the models of streamflow generation are presented. Also observations and field experiments in mountainous forested areas and the available methods for the analysis of rainfall-runoff events are reviewed.

To evaluate the hydrologic response of a basin, certain fundamental knowledge is desirable, for example the pathways that water follows in its way from the hillslopes to the stream channel must be known. Observations from field experiments help the researcher to understand the physical processes which produce runoff in a watershed, and these runoff mechanisms can then be tested by designing theoretical computer models of the runoff processes.

2.2 MODELS OF STORMFLOW GENERATION

A difficult task in hydrology is to evaluate and recognize the pathways that water follows as it flows to the outflow point of a basin. Researchers have suggested a variety of different models. This variety may be due to observations being made in different climatic and hydrologic regions. Also, as the knowledge about the physical processes has increased, the explanation of the runoff mechanisms has become more complicated. These mechanisms can be classified in the three models presented here.
2.2.1 Hortonian overland flow model

In early 1930's Horton (1933) presented his theory of infiltration. According to this theory, the infiltration capacity of soil decreases with time and, at a certain time, the rainfall intensity can be larger than the capacity of soil to absorb the rainwater. At that time the excess rainwater, that fails to infiltrate into the soil, runs off the surface. In this case, the infiltration is a limiting parameter of the rainfall-runoff process. Overland flow may occur in limited areas of the watershed or perhaps even in a whole basin.

The Hortonian overland flow model may describe the production of runoff in arid or semi-arid areas where raindrops cause crusting and swelling of the surface soil which already has small infiltration capacity (Ward, 1984). In contrast, overland flow is very limited for vegetated areas in humid climates. In these areas, the rainfall intensity is usually small and the infiltration capacity of the soil is not the limiting factor.

Hortonian overland flow has been extensively used in hydrologic modeling because of its simple concept but on many watersheds this type of runoff mechanism does not happen. Some humid zone hydrologists used to dismiss this mechanism of runoff generation as a myth, but more recently, after closer examination of the physical processes in field experiments, the usefulness of the Hortonian overland flow has been re-established (Kirkby, 1988). Compaction of the soil surface and changes in the soil properties caused by changes in land use of the watershed are the main reasons for the limited presence of Hortonian overland flow in basins in humid climate.

2.2.2 Partial areas model

In 1960's using a nonlinear mathematical model, derived from integration of the infiltration equation, Betson tried to explain the mechanisms of runoff production in a basin
(Betson, 1964). He claimed that the runoff is generated by a small but relatively consistent part of the watershed which remains constant throughout the year. Later, Dunne and Black (1970), observing flow generation in a New England watershed, suggested that the storm runoff is produced by two mechanisms,

- Direct precipitation on the channel system
- Saturation overland flow from the areas around the stream channel

They noticed that the water is concentrated in the areas near the stream (riparian or partial areas) and that during a rainfall event these areas become fully saturated. Under these conditions, the riparian areas behave like concrete slabs and all the rainfall falling on them fails to infiltrate and runs off towards the stream. Thus, runoff from these partial areas is very sensitive to rainfall intensity.

The above mechanism is not the only one producing runoff. In addition, subsurface water is exfiltrated in the partial areas, since the subsurface flow fails to deliver the water towards the stream (Dunne et al., 1975).

In contrast to the constant partial areas concept proposed by Betson, the variation of the partial areas is considered to occur both seasonally and throughout an event. Furthermore, ephemeral channels may operate during high rainfall events, concentrating more flow.

For this partial areas model, the subsurface water is a minor contributor to the stormflow hydrograph. It is claimed that the response of the subsurface flow is damped by storage and transmission through soils. The subsurface water contributes to stormflow only for those areas which are fully saturated.
2.2.3 Variable source area model

In the 1960's, based on the concept that quick subsurface flow is the primary source of stormflow from forest land (Hursh, 1944), Hewlett proposed another explanation for the runoff generation. According to his model, the runoff is generated by two mechanisms,

- Subsurface flow, and

- Direct precipitation to the stream system and riparian areas

The Tennessee Valley Authority (1965) proposed a model similar to that of the variable source area, and called it the dynamic area model.

The area contributing to stormflow and the stream channel can expand up to the point where equilibrium will be established between topography and precipitation (U.S. Forest Service, 1961), and this variation in stream channel area and the riparian areas is stated to be highly nonlinear (Hewlett and Nutter, 1970). The mechanism producing quick subsurface flow is claimed to be the “translatory flow” (Hewlett and Hibbert, 1967), in which rainwater is infiltrated into the hillslopes and moving vertically and then laterally through capillary pores. Hence, the rainwater tends to displace most of the water ahead of it stored during previous events. Because of the large differences in elevation, the pressure head is the driving force of the quick response of subsurface flow.

Later, Hewlett suggested another pathway for the subsurface flow. This explanation was developed following publication of a pioneering paper by Jones (1971) which introduced and examined the hydrological behaviour of soil pipes. The highly responsive pipeflow system could serve as the mechanism for fast subsurface water delivery (Hewlett and Troedle, 1975). The development of soil pipes is now widely reported for steep forested hillslopes, especially in areas with humid climate.

According to observations in the field, steep watersheds with shallow soils produce more efficient stormflow than the low slope basins with deep soils (Hewlett and Hibbert,
1967). This difference in runoff behaviour probably explains the difference in runoff generation between the partial and the variable area model. The partial area model has been developed after observations in a watershed with deep, poorly drained soils, whereas Hewlett and his co-workers developed the variable source area model observing runoff in a steep, upland, forested watershed with shallow soils. The difference in the runoff generation mechanisms for the two models can be explained by the differences in morphology, canopy, and hydrology of the two types of experimental watersheds (Dunne and Black, 1970).

2.3 RUNOFF GENERATION IN FORESTED MOUNTAINOUS BASINS IN HUMID REGIONS

The hydrologic response is a variable factor which is affected by many parameters, and the mechanisms of runoff production will change from region to region depending upon the climate (Dunne, 1983). Attempts to predict a lumped overall basin response for large regions are likely to fail, because such regions may include a mixture of flat and steep areas. Therefore trying to simplify the response to a mean factor is not likely to give good results (Woodruff and Hewlett, 1970).

The present work will focus on the runoff generation from forested, upland, watersheds, and this type of runoff is affected by geomorphological as well as hydrological parameters. Also, these parameters are not independent and there is a close relation between many of them.

2.3.1 Factors affecting runoff from forest lands

Hewlett and Hibbert (1967) recognised the significance of geomorphological and hydrological factors on the basin response. These factors are:
• The average soil mantle depth

• The average land slope

• The average size and number of large storms

• The land use

Furthermore, rainfall intensity, soil characteristics, and antecedent conditions are considered among these factors.

In forested hillslopes the soil mantle is shallow and with depths of the order of 1 to 1.5 m. Downslope, the soil depth may increase or decrease, affecting the soil storage capacity. Usually watersheds with small valley floor slope have deep soils whereas basins with steep hillslopes have restricted valleys and usually shallow soils. For deep soils the rainwater is stored and the process of rainfall-runoff is delayed, so that the generation of stormflow is achieved either by the saturation overland flow from the partial areas or by the subsurface stormflow or by a combination of these two. The area contributing to stormflow expands and shrinks during an event, so that an equilibrium is established between topography and precipitation (U.S. Forest Service, 1961).

In hillslopes with shallow soils, the rainwater is delivered through soil channels (Whipkey, 1965) or soil matrix flow (Hewlett and Hibbert, 1967). The partial areas are very restricted by the presence of steep hillslopes and so, their significance in runoff generation remains small (Dunne and Black, 1970).

As has already been mentioned, topography plays an important role in runoff generation, affecting the way that rainwater contributes to streamflow. Tsukamoto and Ohta (1988) examine the significance of slope types in the production of runoff, and they conclude that the convergent slope unit is more active hydrologically than divergent or straight slopes, because of the convergence of the flow lines towards the base of the slope.
From field experiments Dunne and his co-workers found that the expansion of the partial saturated areas depends on the slope type. In straight or divergent slopes the saturated areas are restricted, whereas the convergent slope is more efficient in generating saturated areas and overland flow (Dunne and Black, 1970, Dunne et al, 1975).

The number and the size of storms influence the antecedent conditions in a basin and these antecedent conditions can affect both the peak discharge and the volume of the water released from the watershed. These antecedent conditions determine whether the soil is ready or not to transmit water and whether the degree of saturation in the riparian areas is high enough to produce saturation overland flow.

The antecedent conditions, along with the rainfall intensity, control the expansion of the partial areas and stream channel. This fact is more prominent in low response watersheds than in high response ones (Hewlett et al, 1977, Hewlett et al, 1984). Low response watersheds are characterized by deep soils and for these conditions saturated overland flow is usually the dominant mechanism of runoff generation (Dunne and Black, 1970). Hence, the rainfall intensity, rather than the total stormflow volume, has a larger influence on the peak discharge also giving a non-linear response. On the other hand, in high response basins, characterised by steep slopes and shallow soils, the runoff is generated through fast subsurface routes, so that the changes in rainfall intensity produce a more linear response (Hewlett et al, 1984).

Change in land use for forested watersheds is synonymous with logging practices and road construction, and very different results have been found from field experiments. These results range from 100% increase in peak flows (Harr et al, 1975) and 30% increase in stormflow (Hornbeck, 1973) to 22% decrease in peak flows (Cheng et al, 1975) and 11% increase in stormflow (Hewlett and Halvey, 1970) between the pre-logging and post-logging period. The above differences in the results can be accounted for by the different geomorphological and hydrological conditions as well as by the inadequencies in data.
measurements and methods of analysis (Hewlett, 1982).

Soil characteristics are very important for the determination of the pathways followed by the water in order to reach the stream. Existence of lateral or vertical soil pipes in forest soils and permeable organic layers control the infiltration and runoff process. Generally, the infiltration capacity of soil is effectively unlimited in forest watersheds and therefore, all the rainwater is infiltrated in the soil and no overland flow is observed except in small areas around the stream (DeVries and Chow, 1978, Hewlett and Troedle, 1975).

In the next section the significance of soil characteristics for the hydrology of a basin and observations in field experiments will be discussed.

2.3.2 Subsurface pipe flow dynamics

The classic explanation of the delivery of water from a hillslope used the concepts of overland flow, soil matrix flow, and saturated overland flow. Overland flow is unlikely to happen in forested hillsides because of the high infiltration capacity of the soil. On the other hand, it is doubtful whether the soil matrix is universally the dominant runoff mechanism, since the flashy response of this type of basin cannot be explained by the small velocities of subsurface flow. Field experiments with dye or environmental tracers reveal the pathways followed by the rainwater in a hillslope. For example, translatory flow of stored “old” water displayed by “new” water is the mechanism suggested by Hewlett and his co-workers (Hewlett and Hibbert, 1967, Hewlett and Troedle, 1975) and this has been confirmed by others in field experiments (Pearce et al, 1986, and Sklash et al, 1986). Mosley (1982) suggests that deep soils are required for such a type of stormflow generation.

The soil in a forested hillslope is shallow and very permeable. Jones (1971) was the first to make an integrated examination and presentation of the hydrologic significance of
soil pipes. He categorized them as pipes and pseudo-pipes or ephemeral pipes. Activities of small animals, insects and decaying roots are some of the reasons for the soil pipe existence. Most of the pipes are maintained by the frequent passing of water or insects and small animals. If the pipes are not maintained, they gradually disappear and become invisible discharging water only under developed hydrostatic pressure (Tsukamoto and Ohta, 1988). Also, chemical erosion can produce pipes when the flowing water through the soil detaches and removes colloids through the soil pipes (Ziemer and Albright, 1987).

The necessary conditions for piping development has been summarized by Goldsmith and Smith (1985):

- A source of water
- A surface infiltration rate that exceeds the subsurface permeability at some depth
- A zone of soil where the soil hydraulic conductivity decreases consistently with depth
- An hydrologic gradient to cause water to flow
- An outlet for the lateral flow

All the above criteria exist in many forested hillslopes. The infiltration capacity of the soil is much larger than the rainfall intensities (Cheng, 1975) and usually a relatively impermeable soil layer exists beneath the shallow surface soil. The vertical distribution of the pipes is concentrated in the area just above the underlying layer and their number decreases higher up in the soil layer (Ziemer and Albright, 1987, Tsukamoto et al, 1982). Rain water infiltrates under unsaturated conditions in the soil, which is very permeable not only because of its “open” character (Chamberlin, 1972) but also because of the existence of vertical pipes or macropores (De Vries and Chow, 1978, Oka, 1990). These
vertical pipes deliver the water to the soil rapidly, and the lateral flow starts from the area just above the imbedding layer (Whirkey, 1965), the only requirement being that the surrounding soil is fully saturated (Whipkey, 1965, Ziemer and Albright, 1987, Mosley, 1982). Because of these vertical pipes, the necessary time for soil saturation is small. More pipes can contribute to pipe flow as the saturation level increases in the soil, and this increase depends on the rainfall intensities and the flow rate that the pipes can deliver downslope.

Since the surrounding soil is saturated, pipe flow starts to deliver water quickly downslope. The velocities of pipe flow are comparable to those of overland flow (Mosley, 1979, Tanaka et al, 1988). Also, additional water is drained into the lateral pipes from the soil matrix and increases the contribution of pipeflow in the total runoff (Oka, 1990). The water through the pipes can pass through unsaturated zones, during its movement downslope, without significant losses (Mosley, 1982). The water is delivered to the stream either directly through the banks or in the riparian areas where the pipes may terminate and the water will emerge on the surface and run off towards the stream as return flow. Pipe outlets closer to the stream show more developed pipe network, older soils and ephemeral pipe development. As a result of pipe flow the hydrologic response is flashy, the base flow stage is low and ephemeral streams are developed in the basin during a rainfall event (Jones, 1971). Depending on the antecedent soil moisture conditions and the size of the event, the whole basin can contribute to fast pipeflow (Mosley, 1982).

The above description characterizes the rapid response of forested watersheds. The base flow is sustained by the drainage procedure between unsaturated and saturated areas. This process can sustain stream flow for long periods without rain (U.S. Forest, 1961, Whipkey, 1965).

The contribution of pipeflow to the total streamflow depends on the geomorphology of the hillslopes. In straight or divergent hillslopes the expansion of the saturated areas is
restricted by the existence of steep gradients. Pipe flow may account for 97% to 100% of the total discharge from the basin (Harr, 1977, Ziemer and Albright, 1987). In hillslopes with convex areas around the stream channel the pipe flow and the saturation overland flow may represent 90-95% and 0-7% respectively of the total stream flow (Tsukamoto and Ohta, 1988, Tsukamoto et al, 1982, Tanaka et al, 1988). The saturation overland flow is more significant for the peak discharge, to which it contributes approximately 25%. The expansion of the saturated areas is restricted to about 4% as a maximum (Tanaka et al, 1988). The expansion of stream channel also depends on channel slope. It can be negligible (Harr, 1977) or significant when the channel slope is large or small respectively.

Because of the flashy response of pipe flow and its high velocities the peak flow can happen approximately at the time of peak rainfall intensity (Tanaka et al, 1988). Comparison of hyetographs and hydrographs emphasizes the significance of the pipe flow in the response of forested watersheds which lie in steep and undulating terrain.

2.4 METHODS FOR THE ANALYSIS OF RAINFALL-RUNOFF EVENTS

The methods available for the analysis of rainfall-runoff events are based on three types of hydrologic models: the storage based models, the physically based models and the regression models. Every type has its strengths and imperfections. An overview of each method and a critical comparison of them will be presented in this section.

2.4.1 Storage based models

The models often used for prediction, operation and design of engineering projects are based on the idea of the unit hydrograph which assumes that, since the physical characteristics of a basin (size, shape, slope e.t.c.) remain the same, the shape of outflow
hydrograph from rainfall events of similar characteristics should remain the same.

Commonly, hydrological models of this type separate the input to the system (basin) into three (or more) components. Each of them is routed through a separate mechanism. The input (precipitation) can be derived from just one or perhaps several measurement stations.

The components of precipitation are routed through different routing procedures or some of them are considered as losses. They are usually routed through fast, medium and slow routing. Each of these mechanisms is simulated by the use of a "fictitious" reservoir or a cascade of reservoirs. The number of the reservoirs, necessary for the simulation, depends on the specific characteristics of the basin. The outflow from each of these processes is characterized as fast, medium and slow runoff. The routing procedure in each of these components can be non-linear if the relation between storage and outflow is non-linear.

The UBC watershed model (Quick and Pipes, 1976) and the Stanford watershed model (Linsley et al, 1982) are examples of sophisticated, hydrological models based on linear storage concepts and used for stream flow prediction.

Clark (1945) introduced a method which permits the development of the unit hydrograph from the area that contributes to runoff. The contributing area depends on the shape and the conditions of the watershed. The above idea is the basis of simple watershed models, for example the models presented by Melone (1986) and Zachary (1985).

2.4.2 Physically based models

In the 1960's and 1970's hillslope hydrology was developed as a distinct area of hydrology. Detailed field studies gave the insight for the development of physically based models, for the simulation of runoff from a hillslope. The development of these models is based on the general concepts of mathematical modelling of physical processes. The four steps
of this procedure are (Freeze and Cherry, 1979):

- examination of the physical process
- replacement of the physical process by an equivalent mathematical problem (partial differential equations with boundary and initial conditions)
- Solution of the mathematical problem with accepted techniques of mathematics (analytical or numerical)
- interpretation of the mathematical results in terms of the physical process

The hillslope hydrological processes can be formulated as subsurface flow (unsaturated-saturated groundwater flow and throughflow) and overland flow across the land surface. These two processes feed the stream channel flow. In order to simulate the above processes with mathematical tools and solve the derived mathematical problem, certain assumptions need to be made.

An integrated model should consist of a set of three component models, one for overland flow, one for subsurface flow and one for channel flow. Each of these can be solved separately and the solution of one can be input to the other. Betson (1964) presented a non-linear model which treated the infiltration and the vertical movement of subsurface water. Examples of coupled models are the models presented by Wooding (1965a, 1965b, 1966) and Freeze (1972a, 1972b). The first treats the overland flow and the channel flow whereas the second treats the subsurface flow and the channel flow.

2.4.3 Regression models

Regression models are mainly used in forest hydrology for the prediction of stream flow or the examination or the effects of the land use changes on the hydrologic regime of the forested basins. They are based on the analysis of the recorded hydrographs and
hyetographs to obtain the hydrologic parameters of the basin, for example, the time lag of the basin, the peak discharge, the time to peak, the stormflow volume, the duration of stormflow, the total precipitation, the spatial and temporal distribution of precipitation. In order to find these parameters, which characterize the watershed behaviour, it is necessary to separate the observed hydrograph into its components. Several methods of hydrograph separation exist, and all of them are quite arbitrary (Viessman et al., 1977). In forested mountainous watersheds it is usual to separate the hydrograph in two components: quick flow and delayed flow with the separation line proposed by Hewlett and Hibbert (1967). The slope of this straight line is $0.55 \text{ l/sec/hr/km}^2$.

After hydrograph separation, quick flow volume, peak discharge and other parameters can be correlated with hyetograph parameters and basin characteristics. Regression models can be applied between stormflow or peak flow and total precipitation (Cheng, 1975) or stormflow and a precipitation index (Fedora and Beschta, 1989).

### 2.4.4 Critical comparison of the hydrological models

From the foregoing discussion it is clear that the approach to the problem of hydrologic modelling is different for the three types of models. Comparison of the models can reveal the usefulness of each type of modelling. The comparison will be focused in the:

a) Physical representation of the model  
b) Model performance  
c) Data requirements  
d) Model calibration  
e) Form of the results  
d) Flexibility in the model application

a) The main advantage of the physically based models is that they at least partially
represent the physical processes. On the other hand, they must consider every component of the runoff and they must describe it explicitly through mathematical equations. Storage models represent the physical processes with simplified empirical relationships, usually linear. Therefore, they do not have the physical insight of the physically based models. This disadvantage of the storage models has been criticized since their development (Freeze, 1972a, Kirkby, 1988).

Nevertheless, storage models give good results, and appear to be able to simulate the processes in the watershed. The simulation of the physical processes during the generation of runoff can be represented by the separation of runoff into its components, and this subdivision of runoff can be linear or non-linear. It is usual to use linear storage routing for each component because this type of routing maintains continuity. Any non-linear behaviour is concentrated into the subdivision process.

The regression models are based in the analysis of recorded hydrographs and hyetographs and, therefore, they do not face the above implications.

b) The storage models have been developed for large catchments, and perform well. Regression models work well for small or large basins but the relations developed can not be used universally. On the other hand, physically based models have been developed for a single hillslope and it is possible to simulate runoff processes in a basin, since a basin consists of a series of hillslopes. In practice this is very difficult. Besides, their application in such a case has given poor results (Freeze, 1978).

c) The main advantage of the linear storage and unit hydrograph models over the other methods is the limited input data required by them. Only precipitation and reference streamflow data are usually required for this type of model. In contrast, physically based models require a knowledge of spatial and temporal distribution of many physical parameters. Considering also the time required by the detailed models, it makes them inapplicable for operational purposes. Regression models require a large volume of
recorded data and detailed processing to obtain the average hydrologic parameters of the watershed.

d) One of the main steps in hydrologic modelling is the calibration procedure. Physically based models and storage models use the same types of calibration techniques (trial and error, autocalibration). Regression models do not need calibration since they are based on the analysis of recorded data. The problem of parameter identification for the physically based models is very difficult not only because of the large number of the parameters but also because of their spatial and temporal variability and their interaction. In contrast, it is certainly easier to use physical reasoning to calibrate the residence time parameter of a linear storage (storage parameter) which is usually required by the linear storage models (Beven, 1989).

e) The disadvantage of most of the regression models in contrast to the other two types of models is that they do not consider the effect of rainfall time distribution on the watershed hydrograph. Therefore, they do not give the shape of the hydrograph but only the peak discharge and the average response of the watershed during an event. However, recently a regression model which correlates the streamflow with an antecedent precipitation index and gives detailed hydrographs, has been presented (Fedora and Beschta, 1989).

f) Application of the models, in areas other than those in which they have been developed, requires flexibility, and has been demonstrated for storage based models. Such models work very successfully even in different climatic regions (Quick and Pipes, 1976, 1989a).

From the above discussion it is clear that the scope of each type of hydrological modelling is different. The detailed physically based models are used mainly for research purposes in order to understand the physical processes in a hillslope of a basin. The storage based models are mainly used for operational or engineering purposes. The regression models usually predict only the peak discharge or the stormflow from a basin
without giving detailed response of the watershed with in storm variations of precipitation. These regression models, have mainly been used for predicting the effects of land use change on the hydrology of a basin by analysing its average hydrologic response.

2.5 CONCLUSION

The hydrological processes in a forested, mountainous watershed in humid temperate climate can be described with either the partial area model or variable source area model of runoff generation. The mechanisms of runoff production depends on the geomorphological and hydrological characteristics of the watershed. Nevertheless the dominance of the subsurface flow in the runoff generation process in upland basins has been justified and proved by field experiments and observations. Especially, in the region of the Pacific Northwest and other geographical regions with similar climate the dominant mechanism of runoff production is the subsurface pipe flow (Harr, 1977, Ziemer and Albright, 1987, De Vries and Chow, 1978, Tsukamoto and Ohta, 1988, Tanaka et al, 1988). The dynamics of pipe flow can explain the flashy response of forested basins in the mountainous areas of the above regions.

The methods available for the analysis of the response of the watershed are based on the modelling procedures. Each of the three types of models has its own advantages and disadvantages, and the selection of the proper modelling procedure depends mainly on the available data and the purpose of application.
Chapter 3

METHOD OF ANALYSIS

3.1 INTRODUCTION

The objective of this thesis is to examine the hydrologic response of a steep, forested watershed in the mountainous area of the Pacific Northwest. The methods that could be used for the analysis have been presented in the previous chapter. In this chapter, the method of analysis is presented explicitly.

3.1.1 General considerations

The method used for the analysis of data is based on a simple watershed model. This watershed model is designed to represent the storage characteristics of a real watershed. It is a two component model which uses a fast and a slow storage system to represent fast runoff and slow groundwater runoff, and the control between these two components of runoff is accomplished by an infiltration type relationship. Therefore, to some extent, the model represents the natural processes of the watershed.

This simple model was originally intended to be a preliminary method of examining the data. The intension was to design a more complex model which would give a better representation of the processes. However, the application of the model showed that this simple watershed model gave reasonable answers and worked rather well.

Allowing the model parameters to vary from event to event, it is possible to identify variations in physical processes and thus, changes in the hydrologic behaviour of the
basin. Using precipitation as input, the target of the present work is to achieve a best fit of the model output to the observed hydrograph. Analysis of the model parameters that gives this best fit can reveal variations in the hydrologic response of the basin. Input data for the analysis should be as accurate as possible and they should be for a time scale less than the response time of the watershed. The shorter time scale available for Jamieson Creek, the study watershed, is hourly data for both precipitation and flow, and this time interval is about half the response time of the watershed.

3.1.2 Considerations about the model

The type of the model used for the examination of the hydrologic behaviour of the basin should be an event type. Event models are developed to simulate precipitation-rainfall events with a specified set of input data. Changes in the response from event to event can be identified through changes in model parameters.

Continuous type models maintain water balance over long time periods (months or years) between water inputs and outputs in the basin. Also, initial conditions for the continuous models are unique at the beginning of simulation and they are updated by the model through the water balance. However continuous models require continuous data which limit their application in cases where data is available only for particular events.

Another consideration about the model is to keep it as simple as possible. In a simple model, the parameters are limited in number and so, it is easier to describe or to investigate the hydrologic response of the basin. Complex models require more effort to calibrate them or to best fit their results to observed hydrographs because of the possible alternate combinations of model parameters.

Another consideration is the volume of information that is necessary for the simulation. Data from mountainous areas are usually very limited, and it is necessary to compromise and to adopt a procedure that requires limited input. Such a method is
the storage based model that uses only precipitation data. Complex physically based models using physically derived equations are inapplicable because they require information not only about precipitation but also about the characteristics of soil. Furthermore, application of sophisticated models showed that their simulation strength is very limited (Freeze, 1978). Hence, there should be a balance between the complexity of the model, the data requirement and the representation of the system (Melone, 1986).

Finally, the routing procedure should be a linear one. This is of critical importance for the examination of the watershed response. In this case, large variation of model parameters, poor agreement between observed and simulated results and non-linear subdivision of runoff can reveal non-linearity in the basin response. A systematic variation of storage parameters with event size indicates non-linear relationship between the rate of input and the rate of output, pointing out non-linear watershed response (Boyd and Bufill, 1989).

3.2 DESCRIPTION OF THE MODEL

According to the previous discussion, the model adopted is a simple event model based on linear storage routing and uses an infiltration type equation for flow separation. The model has been developed to analyse rainfall events only but it is not difficult to incorporate a routine for the snowmelt component of runoff, as well. The complexity and the diversity of rain on snow, and snowmelt events led to the decision to investigate only rainfall events.

3.2.1 Rainfall input

The model simulates both the fast and the slow components of runoff. Therefore, there should be a method for separation of rainfall input to these routing procedures. The
model uses an infiltration type equation:

\[ P_s = a \cdot t^{-n} + b \]  

(3.1)

where \( P_s \) is the rainfall that goes to slow routing and \( a, n, b \) are constants which characterize the watershed and they depend upon the initial conditions, and \( t \) is the time.

The rainfall input to the fast routing is:

\[ P_f = P_t - P_s \]  

(3.2)

where \( P_t \) is the rainfall input at time \( t \) and \( P_s \) as previously described.

This type of equation for rainfall input separation, which is used also for overland flow, appears to be reasonable for describing the physical processes in the steep forested watersheds. As has been discussed earlier, the dominant runoff generation mechanism in these basins is the subsurface stormflow, and especially the pipe (or macropore) flow. But, in order for the pipe flow to respond to water input, the surrounding soil should be saturated. The rainfall separation equation estimates the infiltration of rain to groundwater so that the non-infiltrated rainwater is available for fast runoff production (pipe flow in this case).

The rainfall separation equation is quite flexible and various types of behaviour can be simulated according to the coefficients used. For example coefficients \( a \) and \( n \) of the equation Eq. 3.1 characterize the slope of the curve in Fig.3.1a. Small values of \( n \) and large values of \( a \) result in an inclined curve (Fig.3.1b). In this case large amount of water and more time is required for soil saturation per unit depth and so, more water is delivered as slow runoff than fast runoff. The input to slow routing decreases with time up to a constant value (\( b \)). Large values of \( n \) and small of \( a \) result in a step function (Fig.3.1c) and the input to slow routing takes very quickly the value of the constant \( b \) and so, decreases considerably the volume of water and the time required for soil saturation.
Fig. 3.1. Graph of the rainfall separation equation (a) and the effects of the values of $a$ and $n$ on the equation behaviour (b and c).
Chapter 3. METHOD OF ANALYSIS

It should be clearly understood that the rainfall separation equation (Eq. 3.1) does not apply at the soil surface because the forest floor soil has large infiltration capacity and all the rainwater is infiltrated in the soil. The separation of the rainfall input to fast and slow routing occurs within and at the base of the permeable soil layer and is dependent upon the physical soil characteristics.

The infiltration type equation is therefore being used to represent the separation of flow between the fast flow through the soil pipe system and the much slower soil matrix and groundwater drainage. Hydrologically, this runoff behaviour is quite similar to the earlier ideas of surface limiting infiltration and overland flow and it is not surprising that a similar infiltration equation and storage routing can be used to represent it.

Having in mind all the previous stated aspects, the model used as a tool, herein, has been designed to be simple, but includes the most important aspects of watershed behaviour. The model uses the linear storage technique for the routing of flow and a non-linear rainfall separation equation, so that it is capable of modelling non-linear response.

3.2.2 Routing procedure

As it has been mentioned previously, the model processes the rainfall input through two routings. The rainfall input which contributes to fast runoff is routed through a cascade of two linear reservoirs with the same storage coefficient (KF). Slow runoff is simulated with a linear reservoir with large storage coefficient (KS). Fig.3.2 shows schematically the main procedures of the model.

Fast routing simulates the subsurface pipe stormflow, the riparian overland flow, and the direct precipitation to the stream system. It is assumed that the slow component of runoff represents the "groundwater" flow. Each of these runoff components has a different time response and, therefore, occurs as streamflow at considerably different time scales.
Chapter 3.  METHOD OF ANALYSIS

Fig. 3.2.  The watershed model flow chart.
In steep mountainous watersheds the occurrence of extensive saturated overland flow is strictly limited by the high hillslope gradients and the existence of the soil pipe system. Also, the expansion of the saturated areas is limited (Dunne and Black, 1970). Therefore, the fast routing in this particular case represents mainly the subsurface stormflow and the direct precipitation in the stream and any overland flow, saturated or not, is probably quite limited, perhaps occurring close to the stream channels.

The slow routing represents the drainage processes in the hillslope and the interaction of saturated and unsaturated flow in the soil (Weyman, 1970, U.S. Forest Service, 1961). There may also be some slow infiltration into the low permeability underlying soil layer and this may give rise to a slow groundwater component.

Because the reservoir routing is very important and also necessary in the model development, the method is now defined.

The main assumption that underlies the linear reservoir routing is that the relationship between discharge and stage or storage is linear. This can be written as:

\[ S = K \times Q \]  \hspace{1cm} (3.3)

where \( S \) is the storage, \( K \) is the storage coefficient and \( Q \) is the reservoir outflow. The storage coefficient is determined from the storage-discharge relationship and its value equals the value of the gradient \( \frac{dS}{dQ} \).

For linear reservoir routing, the value of the storage coefficient remains constant for the whole range of flows. Using a time increment, it is possible to discretize Eq.3.3.

\[ S(J) = K \times Q(J) \]  \hspace{1cm} (3.4)

\[ S(J - 1) = K \times Q(J - 1) \]  \hspace{1cm} (3.5)

where \( J \) and \( J + 1 \) refer to successive time increments.
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Substraction of Eq.3.4 and 3.5 gives:

\[ \Delta S(J) = K \Delta Q(J) \]  \hspace{1cm} (3.6)

where, \( \Delta \) indicates an increment.

The continuity equation can be written as:

\[ I - Q = \frac{\Delta S}{\Delta t} \]  \hspace{1cm} (3.7)

where, \( I \) is the inflow in the reservoir.

The change in inflow is written as it has been used in the UBC watershed model (Quick and Pipes, 1989b):

\[ \Delta I(J) = I(J) - Q(J - 1) \]  \hspace{1cm} (3.8)

Discretizing Eq.3.7 for time step \( J \) using Eq.3.8 and adding and substracting \( Q(J - 1) \) from the left hand side, it is obtained:

\[ \Delta I(J) - \Delta Q(J) = \frac{\Delta S(J)}{\Delta t} \]  \hspace{1cm} (3.9)

Eqs.3.3 and 3.9 can be combined and rearranged to yield:

\[ \Delta Q(J) = KT \Delta I(J) \]  \hspace{1cm} (3.10)

where, \( KT = \frac{K \Delta t}{K + \Delta t} \).

Finally the outflow from the reservoir at time \( J \) is:

\[ Q(J) = Q(J - 1) + \Delta Q(J) \]  \hspace{1cm} (3.11)

Substructing Eq.3.8 into Eq.3.11 and rearranging results in:

\[ Q(J) = KT I(J) + (1 - KT) Q(J - 1) \]  \hspace{1cm} (3.12)
Chapter 3. METHOD OF ANALYSIS

Eq. 3.12 gives the final outflow from one linear reservoir. When there are cascaded reservoirs, the output of Eq. 3.12 is treated as input to the second reservoir. Repeating calculations from Eqs. 3.10 through 3.12 using $IB(J) = QA(J)$ yields:

$$QB(J) = (KT)^2 \times IA(J) + KT \times (1 - KT) \times QA(J - 1) + (1 - KT) \times QB(J - 1) \quad (3.13)$$

where, $IA, QA$ are the inflow and outflow respectively for the first reservoir and $IB, QB$ the inflow and outflow from the second reservoir respectively at time $J$.

Eq. 3.13 is used for the calculation of fast runoff whereas Eq. 3.12 gives the outflow from the slow routing procedure.

3.2.3 Evaluation of the model performance

As has already been mentioned, the scope of the present work is to evaluate the model parameters by best fitting simulated to observed hydrographs. Hence, a criterion or a parameter, which shows the fitting of the results, is necessary in the model. Such a criterion adopted in the model is the model efficiency proposed by Nash and Sutcliffe (1970). The above criterion has been used widely by researchers in hydrologic modelling for the evaluation of the performance of models (Loague and Freeze, 1985; Assaf, 1991).

Model efficiency compares the mean value of flow and the simulated flows with the observed ones and it can be written as:

$$E_f = \frac{SQ_{OBS} - SQ}{SQ_{OBS}} \quad (3.14)$$

where:

$$SQ = \sum_{J=1}^{N}(Q_{OBS}(J) - Q(J))^2$$

$$SQ_{OBS} = \sum_{J=1}^{N}(Q_{OBS}(J) - \overline{Q}_{OBS})^2$$

$Q_{OBS}$ is the observed flow at time step $J$. 
\( \overline{Q}_{OBS} \) is the observed mean flow for the event

\( Q \) is the simulated flow at time step \( J \).

If simulated flows equals observed flows at each time step then the model efficiency takes its highest value of 1. For any realistic case \( E_f \) is less than 1 and it can take negative values. A negative value of model efficiency infers that the predicted value of the model is worse than simply using the observed mean.

3.3 CONCLUSION

The model described in this section is a simple event model and will be used to investigate the hydrologic response of forested watersheds. It uses the concept of linear storage routing and a non-linear flow separation.

The runoff is treated by the model as fast and slow runoff. The former is simulated by a cascade of two linear reservoirs whereas the latter by one linear reservoir. The rainfall input to either routing procedure is determined by the use of an infiltration type exponential equation and this represents the non-linear part of the routing.

The method of analysis consists of the best fit of simulated to observed hydrographs. For this reason, calculation procedure of model efficiency is used in the model. Examination of the variance of the model parameters that give the best fit can reveal variation in watershed response.
Chapter 4

STUDY AREA

4.1 INTRODUCTION

Hourly meteorological and hydrological data were available for the Jamieson Creek watershed. It has an area of 2.99 km² and is located in the upper Seymour River basin which lies on the south west edge of the Coast Mountains (Figure 4.1). It is approximately 30 km north of Vancouver.

The basins of Seymour river, Capilano river, and Coquitlam provide the water supply for the city of Vancouver and the other municipalities in the west end of the Lower Mainland area.

In 1969 Jamieson Creek watershed and the adjacent Elbow Creek watershed were selected as treatment and control watersheds, respectively in a paired watershed research program. The objective of the above program was to examine the effects of logging on water quality and quantity.

For the period 1970-1977 Jamieson Creek watershed remained untreated. From 1978 up to 1985 19.2 % of the watershed were logged. In this thesis rainfall data from only the first period have been analysed.

4.2 CLIMATE

The climate of southwestern British Columbia is characterized as maritime with wet and mild winters and dry and warm summers. The annual range of temperature is small with
Fig. 4.1. Map showing the location of Jamieson Creek watershed.
Chapter 4. STUDY AREA

quite high mean annual temperature compared to other Canadian regions.

The proximity to the moisture source area (Pacific Ocean) and the rapid northward increase of elevation up to 1500 m within 35 km from the coast are the reasons for this type of climate for the southwestern range of the Coast Mountains.

4.2.1 Precipitation

The mean annual precipitation for the period 1972-1977 in Jamieson Creek watershed was 3657 mm with a range of 2946 to 4293 mm. Much of the precipitation (90%) occurs in the form of rainfall while snowfall is observed in the upper part of the watershed in the winter months.

Figure 4.2 shows the average distribution of monthly precipitation for the years 1972-1977. From June to September is the driest quarter with only 13% of the mean annual precipitation; the wettest period is the four months from November to February with 56% of the mean annual precipitation.

Hall (1989) made a frequency analysis of precipitation with data from two raingauges at 640 and 853 m elevation located in the hillslope with southwestern orientation. Tables 4.1 and 4.2 show the Intensity-Duration-Frequency values for the precipitation for the above two elevations.

Comparing Hall’s results for the two elevations, a decrease in peak rainfall intensity can be observed for short storm durations and almost no difference for larger durations. This matter will be discussed in chapter 6 of the present thesis where the analysis of rainfall will be presented and discussed.
Table 4.1. Intensity-Duration-Frequency values for rainfall at 640 m on Jamieson Creek watershed (in mm/hr). (after Hall, 1989).

<table>
<thead>
<tr>
<th>Return period (years)</th>
<th>Storm duration (hours)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0.5</td>
</tr>
<tr>
<td>2</td>
<td>35.4</td>
</tr>
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<td>5</td>
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<td>10</td>
<td>71.2</td>
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<td>50</td>
<td>102.5</td>
</tr>
<tr>
<td>100</td>
<td>115.7</td>
</tr>
</tbody>
</table>

Table 4.2. Intensity-Duration-Frequency values for rainfall at 853 m on Jamieson Creek watershed (in mm/hr) (after Hall, 1989).

<table>
<thead>
<tr>
<th>Return period (years)</th>
<th>Storm duration (hours)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
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</tr>
<tr>
<td>2</td>
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<td>50</td>
<td>45.4</td>
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<tr>
<td>100</td>
<td>51.4</td>
</tr>
</tbody>
</table>
4.3 PHYSIOGRAPHIC CHARACTERISTICS

4.3.1 Topography

The topographic boundary of Jamieson Creek watershed is well defined. The watershed is underlain by granitic bedrock which is considered water tight. The elevation of the watershed ranges from 305 m at the mouth of the watershed to 1310 m at the divide. Fifty percent of the area is lying above 800 m (Figure 4.3). Watershed hillslopes are very steep. The average gradient of hillslopes is 48 % with more than 20 % of the area having values larger than 60 % (Figure 4.4). The general orientation of the hillslopes is northeast and southwest.

The stream channel is steep containing large organic debris and boulders. Occasional bedrock exposures and organic debris jams are formed. The gradient of the stream channel averages 20 %, but at the upper watershed very steep slopes prevail, reaching a
Fig. 4.3. Area-elevation curve for Jamieson Creek watershed (after Cheng, 1975).
Fig. 4.4. Land slope distribution curve for Jamieson Creek watershed (after Cheng, 1975).
value of 100% accompanied with bedrock exposures.

4.3.2 Soils

The soils of Jamieson Creek watersed are shallow and "open" textured (Chamberlin, 1972). Cheng (1975) characterized the soils as "coarse-textured sands and gravelly sandy loam". There are two major soil types, steep mountain soils and valley bottom soils. The steep mountain soils which consist of ablation till and colluvial originated materials are very shallow and permeable. The valley bottom soils are generated from lacustrine and glacier-alluvial deposits and are thicker and moderately permeable.

DeVries and Chow (1978) have shown with field experiments that the soils in Jamieson Creek contain vertical and lateral soil pipes. Figure 4.5 shows a typical cross section of the soil in the hillslope. The existence of these soil pipes and the highly permeable shallow soils in the hillslopes are the reasons for the overall high infiltration capacity of the soil. Measured and estimated values of saturated hydraulic conductivity of the forest floor and the soil with and without soil pipes are larger than the most intense rainstorm with return period of 100 years at elevations 640 and 853 m (Tables 4.1, 4.2 and 4.3). The identification of the infiltration capacity and the characteristics of the soil are critical for the understanding of the runoff mechanism in the watershed.

4.4 HYDROLOGY

4.4.1 Runoff mechanism

The runoff generation in Jamieson Creek watershed is quite distinctive. The highly permeable soil and organic forest floor are critical for the identification of the runoff mechanism. As has been previously stated, the intensity of the most heavy rainfalls is lower than the saturated hydraulic conductivity of the soil. Under unsaturated conditions
Fig. 4.5. Typical cross section of a hillslope in the Jamieson Creek watershed.
Table 4.3. Measured or estimated saturated hydraulic conductivity of soil and forest floor of Jamieson Creek watershed and vicinity (after Cheng, 1975).

<table>
<thead>
<tr>
<th>Sample Description</th>
<th>Saturated hydraulic conductivity (mm/hr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Forest floor sample (Estimated from Plamondon et al, 1972)</td>
<td>200</td>
</tr>
<tr>
<td>Soil sample without pipes (Measured by O'Loughlin, 1972)</td>
<td>120</td>
</tr>
<tr>
<td>Soil sample with pipes (Estimated by Chamberlin, 1972)</td>
<td>350</td>
</tr>
</tbody>
</table>

The hydraulic conductivity is much larger.

As a result, all the rainwater falling on the forest floor infiltrates into the soil. The existing vertical soil pipes deliver the rain water very rapidly to the area just above the underlying relatively impermeable soil or bedrock. The time needed for the vertical travel of water through soil is very small because of the preferred infiltration of water through the soil pipes, the high velocities of water in the soil pipes, and the shallowness of the hillslope soil.

When the rain water reaches the imbedding layer, it starts to be concentrated and to saturate the surrounding soil matrix. Since, the surrounding soil is saturated, the latteral soil pipe flow commences. Soil saturation is a requirement for the beginning of lateral pipe flow (Ziemer and Albright, 1987). The water in the lateral soil pipes can flow through unsaturated areas without significant losses (Mosley, 1982).

The velocities of pipeflow are comparable with those of overland flow and open channel
flow as Mosley (1979) has proven with field experiments in New Zealand. As a result a large amount of rain water is rapidly delivered towards the stream channel and contributes to the flashy response of Jamieson Creek watershed.

The runoff mechanism described above is the dominant one. Streamflow also is generated by saturated overland flow in the valley bottom area, because of the poorly drained soils, the slow soil matrix drainage and direct precipitation on the expanded stream channel. Measurements of watersheds in Japan with the same physiographic characteristics and under similar climatic environment showed that subsurface stormflow accounts for more than 90% of the total streamflow (Tanaka et al, 1988). The above feature emphasizes the importance of the subsurface pipe flow for the response of the watershed.

4.4.2 Stream flow

Mean annual streamflow in the Jamieson Creek is quite stable and ranges from 0.27 to 0.31 $m^3/sec$ which when compared with the precipitation represents an annual response of 78% to 90%, respectively. Figure 4.6 shows the mean monthly distribution of flow for the period 1972-1977. The peak monthly discharge is observed in May, June, November and December. The high flows in May and June may be due to peak rainfall events. The high flows in November and December are due to rainfall events especially in late fall period while early winter high flows are generated by rain on snow events and peak rainfall events.

Maximum instantaneous flow for the period 1972-1977 ranges from 7 to 2.61 $m^3/sec$. Peak flows are generated by either peak rainfall events or peak snowmelt and rain on snow events.

In summer months the flow is very low, taking its lowest value in August or September. The minimum instantaneous flow ranges from 0.009 to 0.016 $m^3/sec$. Low flows are sustained by longterm groundwater storage during the dry summer months.
4.5 FOREST COVER

Most of the study watershed is covered by a dense mature and overmature coniferous forest. The type of species ranges with elevation. Up to 900 m elevation the dominant species are western red cedar (*Thuja plicata* Donn), western hemlock (*Tsuga heterophylla* (Rafn.) Sarg) and Douglas-fir (*Pseudotsuga manziessii* (Mirb.) Franco). Above the 900 m level the previous species are either completely absent or very scarce. In this elevation range subalpine mountain hemlock (*Tsuga mertensiana* (Bong. Carr.), yellow cedar (*Chamaecyparis nootkatesis* (D.Don) spach) and amabilis fir (*Abies amabilis* (Dougl.) Forbes) dominate.
4.6 WATERSHED INSTRUMENTATION

4.6.1 Precipitation

The precipitation in Jamieson Creek watershed is measured at five fairly evenly distributed points (Figure 4.7). The rainfall instrumentation is a Belford weighting-type precipitation gauge. The raingauges have been installed at 425 m, two at 640 m and two at 850 m elevation (Figure 4.7).

4.6.2 Streamflow

The streamflow in Jamieson Creek is measured by a 120° V-notch weir and water stage recorder. The weir consists of a 7.3 m wide structure containing a 2.1 m high V-notch and a 1.2 m rectangular spill over. The designed maximum capacity of the weir is 11.3 $m^3/sec$ which is estimated to be generated by a 1,000 year storm event (Cheng, 1975).

The water stage is monitored by a Leopold and Stevens water level recorder which uses a nitrogen-gas purge system. Water stage measurement is checked with a point gauge (hook gauge) every one or two weeks. Water velocities are measured during high flows with an Ott current meter (Cheng, 1975).
Fig. 4.7. Map showing the instrumentation in Jamieson Creek watershed
Chapter 5

APPLICATION OF THE METHOD OF ANALYSIS

5.1 INTRODUCTION

This chapter presents the application of the method of analysis to the Jamieson Creek watershed, and discusses the results.

The application of the method of analysis consists of four steps:

a) Assessment of the best-fit values of the model parameters for each event using average rain and the exponential rainfall separation equation;

b) Evaluation of the mean values of the model parameters;

c) Simulation of all the events with the mean values of the model parameters and;

d) Analysis of the variation of simulation results when the mean values of the model parameters are used.

The rainfall data from the study watersed are divided into three categories: small, medium, and large events with respect to the total rainfall storm depth. Small events had a total rainfall depth smaller than 20 mm while the medium events had a total rainfall depth between 20 and 60 mm and the large events had rainfall depth larger than 60 mm. In this study, only medium and large events have been analyzed. Using small rainfall events for analysis is difficult because of the very small change in the streamflow hydrograph during these events.

Eighteen medium and eight large rainfall events for the period 1972-1977 have been analysed. The rainfall data for all the events are the average of the data from the five rain
gauges in the watershed. The average rainfall is obtained with the use of the Thiessen polygon method.

To simulate a best-fit outflow hydrograph to the observed hydrograph, a trial and error technique is used. The model parameters that give the best results for each event are referred to as best-fit parameters. The model efficiency \( E_f \) is used as a parameter to show the success of the above procedure.

5.2 RESULTS

5.2.1 Model parameters

Application of the method to medium and large events showed that the model parameters behave in a similar fashion. Figure 5.1 shows a typical example of the application of the model to a rainfall event.

The model performs well and its parameters showed the same behaviour for almost all the events. The storage coefficient of slow runoff remained constant with a value of 750 hr. Furthermore, the exponential part of the rainfall separation equation (Eq. 3.1) was negligible for all the events. This results in the reduction of the Eq. 3.1 to the form:

\[
P_s = b
\]

i.e. the non-linear rainfall separation equation takes a linear form.

The parameter \( b \) and the storage coefficient of fast runoff (KF) varied for different events. Except for four events the coefficient \( b \) had a range of values with a mean 1.45 mm/hr and standard deviation 0.375 mm/hr. For the previously mentioned four events the parameter \( b \) took values much larger than those for the rest of the events. It took values of 4.9, 4.4, 3.9, 3.3 mm/hr.

The storage coefficient of fast runoff (KF) had a range of values with mean 2.11 hrs and standard deviation 0.637 hrs.
Fig. 5.1. Typical example of the application of the model to a rainfall-runoff event. (Event 26-28/6/1974).
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The two varied model parameters (b, KF) showed no trend for increase or decrease either with the antecedent moisture conditions, or the maximum rainfall intensity, or with the total total rainfall depth. Figures 5.2, 5.3 and 5.4 show the variation of the storage coefficient of fast runoff (KF) with the antecedent base flow, maximum hourly intensity and total rainfall depth. The antecedent base flow is used here as a parameter indicating the soil moisture conditions in the watershed just before the rainfall events. It should be noted that precipitation data were available only for distinct events, and were intermittent. However, the streamflow data was continuous, and could therefore be used as an indicator of the soil moisture condition just before the events.

Although the parameter b generally showed no systematic trend for different rainfall and soil moisture conditions, for the previously noted four events b showed different behaviour. Figure 5.5 shows the values of b for different antecedent base flow. As can be observed from this figure, the events for which b took large values happened during low antecedent base flow or dry antecedent moisture conditions. But this is not the only condition that affects b, since other events also happened during dry conditions and showed normal behaviour.

The events for which the parameter b took large values were medium and large so that the total rainfall depth does not affect the behaviour of b (Fig. 5.6). Examination of the maximum hourly rainfall intensity during these events showed that b took large values (Fig. 5.7). Hence, it is reasonable to say that the larger values of b are observed for intense rainstorms under dry antecedent soil moisture conditions.

5.2.2 Statistical evaluation of the model parameters

As it has been stated before, the model parameters b and KF varied for different events. In order to assess if this variation significantly affects the simulation of the rainfall events, all the events have been simulated with the mean values of KF and b i.e. with values
Fig. 5.2. Variation of the storage coefficient of fast runoff (KF) with antecedent base flow.

mean KF = 2.11 hours
Fig. 5.3. Variation of the storage coefficient of fast runoff (KF) with maximum hourly rainfall intensity.
Fig. 5.4. Variation of the storage coefficient of fast runoff (KF) with total rainfall depth.
mean $b = 1.45 \text{ mm/hr}$

Antecedent base flow ($\text{m}^3/\text{sec}$)

Fig. 5.5. Variation of parameter $b$ with antecedent base flow.
mean $b = 1.45$ mm/hr

Fig. 5.6. Variation of parameter $b$ with total rainfall depth.
mean $b = 1.45 \text{ mm/hr}$

max hourly rainfall intensity (mm/hr)

Fig. 5.7. Variation of the parameter $b$ with maximum hourly rainfall intensity.
2.11 hrs and 1.45 mm/hr respectively. In Appendices A and B the hydrographs of the observed and simulated flow for both the best-fit and mean values of the parameters for all the events are presented.

To assess the statistic significance of the variation of the KF and b, three basic features of the hydrograph have been examined: peak flow, total runoff volume and time to peak. These three features have been examined for both the hydrographs simulated with best-fit and mean values of KF and b against the observed ones. The significance of the difference of the intercept and the slope of the regression equation from 0 and the 1:1 slope of the perfect agreement line has been evaluated with the t-student test. Also, the average arithmetic and absolute error have been evaluated. The average arithmetic error is assessed by the equation:

\[
\overline{ARE} = \left\{ \sum_{J=1}^{N} \frac{(P(J) - P_{OBS}(J))}{P_{OBS}} \right\} \times \frac{100}{N}
\]  

(5.2)

and the average absolute error by the equation:

\[
\overline{ABE} = \left\{ \sum_{J=1}^{N} \frac{|P(J) - P_{OBS}(J)|}{P_{OBS}} \right\} \times \frac{100}{N}
\]  

(5.3)

where:

- \(P\) is the simulated feature of the hydrograph of the event \(J\),
- \(P_{OBS}\) is the observed feature of the hydrograph of the event \(J\),
- \(N\) is the number of the events.

Simulated peak flows with best-fit parameters are highly correlated \((r^2 = 0.93)\) with observed peaks (Fig. 5.8a). The slope of the regression equation is not significantly different from 1:1 slope (at 99 % confidence level) and the intercept does not differ
significantly from 0 (at 95 % c.l.). The average arithmetic error is -0.67 % while the average absolute error is 14.87 %.

Simulated peak flows with mean values of the parameters are moderately correlated ($r^2 = 0.62$) with observed peak flows (Fig. 5.8b). The slope and the intercept of the regression equation do not significantly differ (at 95 % c.l.) from 1:1 slope and 0 respectively. The average arithmetic error is -1.87 % whereas the average absolute error is 29.69 %.

Simulated total runoff volumes with best-fit parameters are highly correlated ($r^2 = 0.996$) with the observed runoff volumes (Fig. 5.9a). The slope and the intercept of the regression equation is not significantly different from 1:1 slope (at 95 % c.l.) and 0 (at 99 % c.l.) respectively. The average arithmetic and absolute error are 1.35 % and 8.12 % respectively.

Simulated total runoff with mean values of KF and b were also highly correlated ($r^2 = 0.901$) with the observed ones (Fig. 5.9b). The slope of the regression equation does not significantly differ from 1:1 slope (at 95 % c.l.) whereas the intercept differs significantly from 0 (at 95 % c.l.). The average arithmetic error and the absolute error are 25 % and 47.98 % respectively. If the four events for which the parameter b took large values are exempted, the average arithmetic and absolute error are reduced to -0.41 % and 25.75 % respectively.

Simulated time to peak for best-fit and mean values of KF and b is highly correlated ($r^2 = 0.895$ and $r^2 = 0.84$ respectively) with observed time to peak (Fig. 5.10a,b). For both cases, the slopes and the intercepts of the regression equation do not differ significantly from 1:1 slope and 0 respectively (at 95 % c.l.). The average arithmetic and the absolute error for the simulations with best-fit and mean values of the parameters are 8.14 %, 17 % and 5.26 %, 18 %, respectively.

Ninety percent of all simulated peak flows with best-fit parameters occur, within 2
Fig. 5.8. Scattergraphs of peak flow with best-fit (a) and mean values of model parameters (b).

(——regression equation and ----95% confidence limits)
Fig. 5.9. Scattergraphs of unit runoff volume with best-fit (a) and mean values of model parameters (b).
(—regression equation and ---95 % confidence limits)
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Fig. 5.10. Scattergraphs of time to peak with best-fit (a) and mean values of model parameters (b).

---regression equation and ----95% confidence limits
hrs from the observed peaks, and they usually proceed the observed peaks by an average of 0.24 hrs. The simulated peak flows using the mean values of KF and b proceed the observed peak flows by an average of 0.65 hrs. Seventy six percent of them occur within 2 hrs from the observed peaks. Both the above features are not significantly different from 0 at 95 % confidence limits.

From the previous results, it is clear that the variation of the model parameters is not significant except for the four events for which b took large values. Hence, except for these cases, simulation of rainfall events with the mean values of KF and b does not affect the results. Therefore the watershed response is reasonably linear.

The significant variation of b for the previously noted abnormal events reveals non-linear response of the watershed under very dry antecedent soil moisture conditions and intense rain storms.

5.2.3 Discussion of the results

Application of the simple watershed model to the Jamieson Creek watershed shows that the model performs well as can be seen from the results on Appendices A and B. The model gives a simple but realistic representation of the physical processes of runoff generation in the watershed. It has been already proved that the dynamics of pipe flow are very similar to those of overland flow, so that the technique of the linear storage routing developed for the overland flow can be used for the prediction of the runoff where the pipe flow is the dominant runoff mechanism. Also the rainfall separation equation reflects the physical processes of the runoff generation due to pipe flow (saturation requirement of surrounding soil).

The good performance of the model can be attributed to the above reasons but also because accurate and representative data are available for the study watershed. The rainfall intensity is measured accurately in five, quite evenly distributed points.
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The streamflow is measured with an accurate method of flow measurement as has been discussed in chapter 4.

The above discussion shows that for the good performance of any model it is not only necessary to have a good design of model but also the data must be reliable. Further discussion on this subject will be given in the next chapter where the representativeness and reliability of the rainfall will be tested and discussed. Each aspect of the model and the controlling parameters will now be discussed.

The storage coefficient of slow runoff (KS) remained constant for all the events. As has been noted previously, the slow runoff simulates the soil matrix flow. The response of the Darcian flow to rainfall is small and quite slow, so that it sustains the long term low flows. Hence, the large and constant value of KS along with the linear routing of slow runoff is a good representation of this slow, linear drainage process.

The analysis of the data shows that the rainfall separation equation takes a linear form rather than exponential. The existence of vertical soil pipes along with the high infiltration capacity of the “open” forest soil and the small soil depth can explain the linear relationship of the rainfall separation equation, because rain water infiltrates rapidly into the soil through the soil matrix and the vertical soil pipes, and the conductivity of these vertical pipes is not a function of soil moisture change during an event. Part of this water fills the small water storage, saturating the soil, and the rest flows laterally through soil pipes. Also, a small portion of the rain water infiltrates into the underlying soil layer. The soil saturation portion of rain water and the infiltrated rain water into the soil layers of low permeability are represented by b which remains constant throughout the event because of the above mentioned reasons.

Coefficient b took values that range around a mean value of 1.45 mm/hr except for four events for which b took large values. All these four events happened in summer months and during intense rainstorms. Under dry soil conditions, the vertical movement
starts when a certain amount of water is concentrated, and surpasses the obstruction of air entrapment (Whipkey, 1965). Also, a larger amount of water is required for the soil saturation and the commencement of lateral pipe flow. The above increased water requirement is depicted in the larger values of parameter b, for these four non-linear events.

The fast routing in the model simulates mainly the pipe flow and other highly responsive runoff mechanisms. The storage coefficient of fast runoff (KF) took values which range around the mean value of 2.11 hrs. This low value may be explained from the high response of the pipeflow. Field experiments show that the velocities developed during the pipe flow are comparable to the velocities of overland flow (Mosley, 1979). Furthermore, the infiltration of the rainwater into the shallow soil in the hillslopes through vertical soil pipes is another critical factor for the decrease of the response of the watershed to rainfall. The above factor has a direct effect on the value of the storage coefficient of fast runoff.

The majority of b values and the values of storage coefficient of fast runoff (KF) were not significantly changed under different conditions. They did not demonstrate systematic variations with antecedent base flow, rainfall intensity and total rainfall depth. The variation of the model parameters (KF, b) is not significant as it can be seen from the statistical comparison of the simulation results with the best-fit and the mean values of KF and b. Except for the previously discussed non-linear events the average arithmetic and absolute errors range in acceptable levels. Inclusion of the non-linear events gave large errors as in the case of the total unit volume of runoff.

The above findings along with the linear rainfall separation equation show a reasonably linear response of the watershed for medium and large rainfall events. Noting that non-linearities in hydrologic response are mainly attributed to infiltration processes, it is reasonable to assume that the observed linearity of the Jamieson Creek watershed is due
to the existence of vertical and down-slope drainage and infiltration pathways (soil pipes) and the generation of runoff through them. The linearity of the hydrologic response allows the application of linear models in small, forested and steep watersheds but with caution for the cases when the response may be non-linear. Such non-linearities have been observed during the intense rainstorm events under very dry soil conditions and small rainfall events. The first case has been discussed previously. The non-linearities during small rainfall events happen because of the generation of the runoff through mechanisms different to that of pipe flow. Besides, field experiments (McDonnel, 1990) have shown that the pipe flow mechanism works only after a threshold value of rainfall depth. Such non-linearities also, may be attributed to the infiltration processes in the hillslopes. It should be noted that the present analysis examines only medium and large rainfall events, so that the non-linearities in the response of the watershed during small rainfall events are not considered.

Usually, it is necessary to simulate large rainfall events for flood prediction and engineering projects. In this case, simple linear watershed models can be adequate.

5.3 HYDROGRAPH SEPARATION

Observing the outflow from a watershed different time scales of the runoff can be recognized. These different time scales are due to the different components of the runoff which have different response time (Sklash et al, 1986). Separation of the outflow hydrograph in its components can be achieved either with tedious field experiments using dye or environmental tracers or with easy to apply empirical methods.

The separation of the hydrograph with field experiments is achieved with the injection of dye or tracer into the soil at various parts of the watershed and continuous measurement of the precipitation, streamflow, and dye or tracer concentration at the mouth of the
basin. The changes in the concentration of dye or tracer with time give an indication of the response of the flow component which carries the dye or tracer. This field experiments require knowledge of the physical conditions of the watershed, expensive monitoring systems, and finally, after large effort, the results may not be reliable.

On the other hand, because of the above limitations of field experiments, empirical methods for hydrograph separation have been developed from the early days of hydrology. Most of the times these methods have no physical meaning, and have been derived from observations or empiricism. The empirical methods cannot replace the value of the field experiments using dye or environmental tracers, and they are just a cheap substitute.

For steep forested watersheds in humid regions it is usual to consider only two components of the hydrograph: the quick flow and slow flow (or delayed flow). The reason for this is that the flow in such watersheds is generated mainly through subsurface flow. The quick flow incorporates the fast subsurface flow which is either "translated groundwater flow" or pipeflow and other highly responsive flow generation mechanisms, like saturated overland flow and direct precipitation on the riparian areas and stream.

The slow component of runoff is the Darcian groundwater flow through the soil matrix which is the main drainage process in the hillslope in rainless conditions.

To separate the total streamflow hydrograph into quick and slow flow, a time-based separation technique proposed by Hewlett and Hibbert (1967) is used in the case of steep, forested watersheds. With this method the hydrograph separation is achieved by drawing a line upward from the point of initial hydrograph rise at a slope:

\[ S_h = 0.55 \times 10^{-3} m^3 sec^{-1} km^{-2} hr^{-1} \]

This method cannot be applied to complex rainfall events which give outflow hydrographs with multiple peaks.

The so-called Hewlett method is very popular in forest hydrology, and is widely used
to assess the duration and the volume of the quick flow. Usually, these two features, the volume and the duration of quick flow, are correlated with rainfall data (total rainfall depth or rainfall intensity) to assess changes in watershed response after partial or total logging of the watershed or different management practices. Errors in the estimation of the quick flow volume or duration may lead to erroneous results.

To assess the validity of the Hewlett method in separating the outflow hydrograph in the Jamieson Creek watershed, its results are compared with those of the model. Seventeen rainfall-runoff events with distinct single peaks are used for the comparison. The hydrographs of the slow computed component flow from the model are compared with the results from the Hewlett method. Figures 5.11 and 5.12 are examples of the hydrograph separation with the Hewlett method and the model results.

For most of the events the Hewlett separation slope is larger than the average slope of the hydrograph of the slow routing flow. But because the slope for both methods is small, the difference in the results is not large.

The quick flow volume for both methods have been computed. Figure 5.13 shows the scattergraph of the quick flow volume as it is derived from the Hewlett method against the quick flow volume computed by the model results. These features are highly correlated ($r^2 = 0.99$) with each other.

This result suggests that the Hewlett method may be used for the separation of the hydrographs in the Jamieson Creek watershed and the derivation of quick flow volume and duration. The writer does not claim that the comparison of the Hewlett method results with the model results is an accurate and certain way to assess the validity of the empirical Hewlett method. A comparison with results of field experiments is a proper way to test and validate the Hewlett or any other empirical method, but such experiments many times are not feasible. Instead the comparison with the results of the model, which has been shown to give a good simulation the natural processes in the watershed, will be
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Fig. 5.11. Separation of the 4-7/9/1976 event hydrograph with Hewlett method and model's results.

Fig. 5.12. Separation of the 18-21/9/1977 event hydrograph with Hewlett method and model's results.
5.4 SENSITIVITY ANALYSIS OF THE MODEL

It is important to know how sensitive a given model is to changes of its parameters, especially for the present study in which the change of the model performance with the change of model parameters is one of the main interests.

In order to perform the sensitivity analysis, a particular event has been selected. This is the rainfall-runoff event of 4-6/9/1976. The examined parameters are the storage coefficient of slow runoff (KS), the storage coefficient of fast runoff (KF) and the parameter b of the rainfall separation equation. The parameters of the exponential part of the rainfall separation equation is not examined since the exponential part of the equation was negligible. The objective parameter which measures the model performance is the
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model efficiency ($E_f$) presented in chapter 3.

5.4.1 Sensitivity of the model performance to the change of storage coefficient of slow runoff (KS)

As has been previously noted, the storage coefficient of slow runoff remains constant for all the events with a value of 750 hrs. Figure 5.14 shows the sensitivity of the model performance ($E_f$) to the change of KS and for certain fixed values of KF (KF=2.2, 2.5 and 2.8 hrs). From the above figure it is obvious that for all the three values of KF the model is insensitive to the change of KS when this parameter takes values larger than 500 hrs. For values of KS smaller than 240 hrs the model performance changes very rapidly showing high sensitivity to the change of KS. For values between 240 hrs and 500 hrs there are small changes in the model performance.

The model shows the same sensitivity for all the tested values of KF. For values of KS below 240 hrs, the model performs slightly better for the smaller value of KF (KF=2.2 hrs) and the best-fit value of KF (KF=2.5 hrs).

For values of KS larger than 240 hrs the model performance is better for the best-fit value of KF. For KS larger than 240 hrs the model performs better for KS=2.2 hrs than KF=2.8 hrs.

Figure 5.15 shows the sensitivity of the model performance as a function of KS and and for three fixed values of the parameter $b$. The sensitivity of the model again shows a similar trend, so that the model is insensitive to variations of KS when KS is larger than 500 hrs.

The model performance is only slightly dependent on the variations in $b$. It is interesting that for $b=1.2$ mm/hr the model performs slightly better at low KS values, but worse for large KS values. Contrary, $b=1.8$ mm/hr gives the best results for high KS values, but slightly poorer results when KS is small.
Fig. 5.14. Sensitivity of the model performance to the change of KS and KF.
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Fig. 5.15. Sensitivity of the model performance to the change of KS and b.
5.4.2 Sensitivity of the model performance to the change of storage coefficient of fast runoff (KF)

The storage coefficient of fast runoff (KF) is one of the two varied parameters of the model which is used for the assessment of hydrologic linearity of the watershed. Hence, the sensitivity of the model to KF changes is very significant.

Figure 5.16 shows the effects on the model performance to the changes of KF for three values of b (1.2, 1.5 and 1.8 mm/hr). The sensitivity curves for the three values of b have the same shape, but each one is transposed relatively to the position of the other two. The model efficiency takes its highest value for KF=2.5 hrs and b=1.5 mm/hr.

For values of KF smaller than 1.9 hrs the model performs better for b=1.8 mm/hr and worst for b=1.2 mm/hr. For values between 1.9 and 3.2 hrs the model efficiency is larger for b=1.5 mm/hr (which is the best-fit value of b). Finally, for values of KF larger than 3.2 hrs the model performs better for b=1.2 mm/hr than for the other two values of b. In the same range of values the model efficiency is larger for b=1.5 mm/hr than for b=1.8 mm/hr.

Figure 5.16 we can observe that for variation of the KF of one standard deviation ($\sigma_{KF} = 0.637 \text{ hrs}$) around the mean value ($\bar{KF} = 2.11 \text{ hrs}$), the model efficiency ($E_f$) ranges for all the three values of b from a minimum of 0.75 to a maximum of 0.96. It should be noted here that the three values of b here represent the mean value of b ($\bar{b} = 1.45 \text{ mm/hr}$) plus or minus one standard deviation ($\sigma_b = 0.375 \text{ mm/hr}$) around the mean. For KF variation of two standard deviations the model efficiency ranges from 0.45 to 0.9.

The above feature shows that for a range of values of KF and b of one and even of two standard deviations the model performance is still quite high.
Fig. 5.16. Sensitivity of the model performance to the change of parameter KF.

- Dashed line: $b = 1.2$ mm/hr
- Solid line: $b = 1.5$ mm/hr
- Dotted line: $b = 1.8$ mm/hr
5.4.3 Sensitivity of the model performance to the change of parameter b

Figure 5.17 has a flatter maximum than Figure 5.16, indicating that, when b is close to its best value, the model is not as highly sensitive to variations in b as it is to changes in KF. The parameter KF is therefore the most sensitive.

5.5 CONCLUSIONS

Application of the method of analysis to the Jamieson Creek watershed showed that the simple watershed model simulates the physical processes in the watershed quite well. Even though it uses a simple linear storage technique, the model performs well giving a good basis for the analysis of the watershed response.

Having the watershed model as the basis of the analysis, the hydrologic response of the study watershed was assessed through the analysis of the model parameters. The model parameters had a distinct behaviour. Firstly, the exponential part of the rainfall separation equation was negligible for all the analyzed events, so that the equation took a linear form:

\[ P_s = b \]

Secondly, the storage coefficient of slow runoff (KS) was kept constant for all the events taking a large value of 750 hrs.

Thirdly, the parameter b of the rainfall separation equation and the storage coefficient of fast runoff KF varied for different events. The above behaviour of the model parameters may be explained by their physical representation and the physical processes in the watershed.

The significance of the variation of KF and b was statistically tested. Three features of the hydrograph have been examined: the peak flow, the total unit runoff and the
Fig. 5.17. Sensitivity of the model performance to the change of parameter b.
time to peak. The results showed that the variation of the model parameters was not significant except for four cases. For these four non-linear events, the parameter \( b \) took large values. These four events happened when the antecedent base flow was very low, which could indicate very dry soil moisture conditions, and the rain was quite intense.

The model parameters \( K_F \) and \( b \) showed very small variation for all events except the four events discussed above. Two important conclusions can be drawn. Firstly that the hydrological behaviour of Jamieson Creek is well described by the simple model with a fast runoff storage parameter \( K_F \) value of \( 2.11 \pm 0.637 \) hrs and a parameter of the rainfall separation equation \( b \) equal to \( 1.45 \pm 0.375 \) mm/hr. Secondly, the rainfall separation equation reduces to a linear form and the routing reservoirs are also linear, so that the watershed behaviour is linear. This is quite surprising finding and it is against the common beliefs that small watersheds respond in a non-linear fashion.

The most common and widely used hydrograph separation method in forest hydrology, the Hewlett method, has been tested against the model results. The quick flow volume obtained with the Hewlett method was highly correlated (\( r^2 = 0.99 \)) with the computed quick flow volume from the model.

Sensitivity analysis of the model showed that the model is insensitive to the variation of the storage coefficient of slow runoff (\( K_S \)) if it takes a value larger than 500 hrs.

On the other hand, the model is quite sensitive to the variation of the storage coefficient of fast runoff (\( K_F \)) and the parameter \( b \). But for \( K_F \) and \( b \) values within one standard deviation of the mean the model efficiency took high values.
Chapter 6

ANALYSIS OF PRECIPITATION

6.1 INTRODUCTION

The scope of this chapter is to understand the physical processes that affect precipitation distribution in mountainous areas, analyse the rainfall distribution in the study area, and investigate the effects of rainfall representativeness on the simulation results.

The precipitation is generated when air masses are lifted upwards. A common classification of precipitation is based on the mechanisms of air lifting that cause it (Linsley et al, 1982). The main categories of precipitation are three:

- Cyclonic
- Convective
- Orographic

Cyclonic precipitation results when moist air converges in a low pressure area or cyclone. This type of precipitation is usually induced by frontal precipitation which results from the lifting of warm air over colder and denser air. This lifting process can cause extensive precipitation over large areas. For the warm front the rate of the ascent is small since the slope of the frontal surface is relatively small, and the precipitation can extend for 300 to 500 km.

On the other hand, for the cold front the resulting precipitation is in the form of showers. The rate of ascent is larger than for warm-front precipitation, and consequently
the precipitation intensities can be higher.

Convective precipitation happens on a much smaller scale than the cyclonic precipitation, and is caused by the upward movement of warm moist air. Local differential warming of moist air produces local convection which can produce local rainfall. Thunderstorms are an extreme example of this type of precipitation.

Orographic precipitation is caused by forced lifting of air when wind blows towards a range of hills or mountains. Consequently mountain precipitation is, in the long term, always greater than in the plains. In addition, once the lifting process is started, the air mass can become unstable, reinforcing the lifting process.

In the rugged terrain of mountains the precipitation processes are often so complicated and various processes may occur simultaneously. Steep slopes, harsh weather and difficulty of access are the characteristics of mountainous regions, so that data is often very limited. Efforts to measure and analyze the mountainous precipitation and the orographic effects on it can be misleading because of the above difficulties. Previous knowledge and beliefs regarding precipitation distribution in mountainous regions need to be carefully rechecked and perhaps revised.

6.2 FACTORS AFFECTING RAINFALL DISTRIBUTION IN MOUNTAINOUS AREAS

Rainfall distribution in mountainous areas is affected by many meteorological and terrain induced factors. The mountains can induce orographic precipitation within storms and also cause convective instability from differential heating or cooling of air masses over hillslopes, increase cyclonic precipitation by retarding the passage of frontal systems and cause funneling of air streams in the valleys and consequent air lifting in the hillslopes.

Whiltmore (1972) examined the distribution of the mean annual precipitation in
South Africa and the factors affecting it. He concluded that the altitude has the most powerful influence on precipitation. He found that rainfall increased by about 30 mm per 100 m up to about 1300 m, above which altitude the rainfall increases slightly. Other studies in Japan (Tateya et al, 1989) showed that the annual precipitation increases linearly with elevation. The factor that Whiltmore recognized to have the second largest effect on precipitation distribution was continentality. As the moist air passes from sea to the land, it is susceptible to convergence due to increased frictional resistance. This coupled with enforced uplift of air against coastal cliffs or the first mountain barrier accounts for the initial zone of intense and large precipitation independent of altitude. After this zone there was a sharp decrease in precipitation which continued further inland but of smaller rate. The distance from the direction of the dominant air circulation had a much smaller effect on precipitation distribution. Finally, the effects of aspect were not clearly indicated because of subjective information and inadequate data.

In a similar study Lessman and Stanescu (1972) analyzed the distribution of mean annual precipitation at the mountainous gauge region of Colombia for the design of a network. They found that the complex variation of rainfall depended not only on the effects of altitude but also on other physiographic and meteorological factors, such as the extent, slope and orientation of the barrier, its distance from the moisture source, the prevailing wind directions and velocities, and the vertical extent and stability of moist air. Furthermore, they concluded that the tropical rainfall increased only up to a certain elevation, and consequently the rainfall decreases with additional elevation. They explained this observation because the water vapor decreases with height. The height up to which the rainfall increased depended on the site specific physiographic features but it ranged around 2000 m above mean sea level compared with a total barrier height of about 3000 m.

The effect of the wind speed has been recognized by Browning (1980). Furthermore,
he suggested that the vertical air motion can greatly enhance the precipitation. In the
deep dissected valleys the high winds can cause funneling and strong uplifting of the
air masses at the head of the valley resulting in convergence and convection. Also, he
concluded that wet-bulb potential temperature has a considerable effect on precipitation
generation and distribution, since it indicates an increase in condensation.

Temperature and humidity decrease with height, and in particular, humidity drops off
significantly above the layer of maximum cloud development (1000-2000 m) (Shaw, 1972).
The previously noted vertical air motion forced by strong winds causes increase of rainfall
amounts and intensities with height. On reaching the maximum cloud development height
both rainfall and cloud amounts become less and rainfall intensity diminishes (Shaw,
1972).

Instability of the air mass has been recognized as the main factor affecting precipi­
tation distribution (Browning, 1980, Givone and Meignien, 1990). When the airflow
approaching the mountainous areas is unsaturated and rather stable, precipitation can
only occur after the air mass has been forced upwards and cooled to saturation. In this
case, higher precipitation and intensities will be observed at the higher elevations. If the
incoming air is highly unstable and moist, even small upwind obstacles can cause precip­
titation. That is a possible reason why heavier rainfall intensities are often observed in
the foothills or lower elevations of high mountains (Givone and Meignien, 1990).

From the previous discussion it is clear that the precipitation generation and distribu­
tion is highly affected by the air temperature and the air humidity. These two factors
affect the stability of the air mass, which primarily depends on its adiabatic lapse rate
and the actual lapse rate of the surrounding air, and can cause convective instability of
the air mass on a mountain slope. In addition, the saturated adiabatic lapse rate is highly
dependent on temperature, for example, at high air temperatures it is much smaller than
the dry adiabatic lapse rate. On the other hand, for low temperatures the difference
between saturated and dry adiabatic lapse rate is very small. The above variation may explain the effects of mountain barriers on the distribution of precipitation. In winter time, because of the low temperatures, the difference between the saturated and dry adiabatic lapse rate is small. The air mass is more stable and the precipitation in the form of snowfall is highly affected by the orographic effects showing large increase with height. On the other hand, in summertime the difference between the two lapse rates is large causing large instabilities. Consequently, the warm summer rainfalls are less sensitive to orographic effects (Quick and Pipes, 1989b).

Observations that confirm the above trend have been made in the foothills of the Rocky Mountains in Canada. Storr and Ferguson (1972) analysing the precipitation distribution with height, found that the rate of increase for winter precipitation, mostly in the form of snowfall, was $636 \text{ mm/km}$ while the rate of increase for summer rainfalls was only $93.9 \text{ mm/km}$. Also, they concluded that the elevation is the dominant factor which affects precipitation whereas the aspect and the gradient of the mountain slopes affect it but in smaller rate.

Summarizing the previous discussion it can be said that the precipitation generation and distribution in mountainous areas are affected by many factors so that the analysis is difficult. However, different trends of precipitation distribution have been observed in different mountain regions, so that it may be wise to say that the precipitation distribution is highly dependent on the site specific physiographic features, the climate, and the precipitation generation mechanisms.

6.3 PREVIOUS STUDIES IN THE AREA OF INTEREST

Reviewing previous studies which have been done in the study area or in the vicinity can help in the understanding of the physical processes and the trends in the precipitation
Examination of the annual precipitation in the Northern Cascades in Washington state, U.S.A., a similar environment as that of southwestern British Columbia, showed large variation of precipitation at different sites of the same elevation (Rasmussen and Tangborn, 1976).

A drawback of much of the previous work is that the precipitation is not separated into rainfall and snowfall. Investigation of the distribution of winter precipitation in the southern edge of Mount Seymour north of Vancouver showed that rainfall and snowfall have a different distribution with elevation (Fitzharris, 1975). The rainfall increased with elevation up to the level of 800 m approximately, and then it decreased at higher elevations. The elevation of 800 m is approximately two thirds ($\frac{2}{3}$) of the height of Mount Seymour.

The distribution of snowfall for the same period was completely different. The snowfall depth increased with elevation up to the height of maximum elevation where measurements are taken (1260 m).

In a later study, Melone (1986) took some of the data published by Fitzharris (1975) and analysed the precipitation distribution with elevation for particular storm events. He concluded that the precipitation increases linearly with elevation for the three events he analysed. These events experienced large amounts of precipitation, and they consisted of rain for most of the elevation range with rain and snow at the upper elevations. Hence, these events were not representative of the distribution of rain or snow. Perhaps under these low temperature conditions, the meteorological factors affecting precipitation distribution are similar to those affecting snowfall distribution.

A meteorological study for the estimation of Probable Maximum Precipitation (PMP) was conducted for the Coquitlam Lake watershed (Schaefer, 1981). The methodology used for this study was that proposed by World Meteorological Organization (1973).
Chapter 6. ANALYSIS OF PRECIPITATION

The method makes separate estimates for the orographic component of precipitation induced by the air lifting over the mountain barrier and the convergence component. The results of the above study showed that for the whole range of elevation (156-1750 m) the PMP increased linearly with elevation up to the maximum elevation of the basin.

Schaefer study was revised by Nikleva (1990). The main differences between the 1981 and 1990 PMP study for the Coquitlam Lake watershed are the different procedure for moisture maximization which resulted in different PMP values for 1-4 days. The 1990 study estimated the storm moisture by calculating precipitable water from Quillayute/Tattosh radiosonde ascent instead of the usual calculation by the 12 hour persisting dew points which was used in the 1981 study for Vancouver Airport data.

The 1990 study also considers, the possibility of 2 PMP type storm occurring on successive days (double trough scenario). Both 24 hour and double trough scenarios were used to determine the PMP values in the Coquitlam Lake watershed.

The results showed that the convergence component of precipitation decreased with elevation but the orographic component increased with elevation at a much higher rate. The net result is an increase of precipitation depth with elevation for all storm durations.

In Jamieson Creek watershed two studies have dealt with the precipitation. Both of these studies were Bachelor of Science Theses. In the older study Jones (1974) analysed the monthly precipitation from 12 rain gauges distributed over the watershed area in order to optimize the rain gauge network. As a secondary result of his work, he found that monthly precipitation increases with height up to an elevation of 500 m, and then it decreases with additional increase of elevation. He claimed that this pattern was due to the effect on the prevailing wind of the ridge between Jamieson Creek watershed and the adjacent Orchid Creek watershed. The above speculation could not be supported since wind speed and direction measurements were not available in Jamieson Creek watershed.
Table 6.1. Intensity-Duration-Frequency values for rainfall at Seymour Falls Dam (in mm/hr).
(after Cheng, 1975).

<table>
<thead>
<tr>
<th>Return period (years)</th>
<th>Storm duration (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>5</td>
</tr>
<tr>
<td>5</td>
<td>74</td>
</tr>
<tr>
<td>10</td>
<td>91</td>
</tr>
<tr>
<td>25</td>
<td>112</td>
</tr>
</tbody>
</table>

The second study was undertaken by Hall (1989), in order to find the Intensity-Duration-Frequency (IDF) curves for the Jamieson Creek watershed. He analysed data only from the rain gauges located on the hillslope facing southwest at elevations 853 m and 640 m (28A and 21A rain gauges in Figure 4.7, respectively). He found that the small duration intensities were higher at the lower elevation station (21A) for all the return periods. The results of this work are shown in the Tables 4.1 and 4.2. He gave no explanation of these findings. The author compared the previous IDF values for the two elevations with the IDF values for the rain gauge at the Seymour Falls Dam, about 12 km south of the mouth of the Jamieson Creek watershed, located at 200 m elevation (Table 6.1).

Comparing Tables 4.1, 4.2 and 6.1, one can see that the short duration rainfall intensities increase with elevation from the Seymour Falls Dam (200 m el.) up to the Jamieson Creek rain gauge 21A (640 m el.) and then they decrease for the station 28A (853 m el.). Examination of this trend will made in the next section of this chapter, where and additional analysis of the rainfall distribution will be presented.
In conclusion, there are three points that deserve some attention. Firstly, the distribution of snowfall seems to be much different from that of rainfall. Studies analysing precipitation should clearly indicate the form of precipitation, otherwise the results may be misleading.

Secondly, the rainfall in the area of interest trends to increase with height up to $\frac{2}{3}$ of the elevation range of the mountain, and then to decrease with additional increase of height.

Thirdly, it is necessary for a much more detailed analysis of precipitation in the Jamieson Creek watershed to be undertaken in order to examine the previously discussed trend of rainfall distribution.

6.4 RAINFALL DISTRIBUTION IN THE JAMIESON CREEK WATERSHED

For a detailed analysis of precipitation frequent data is required. At the scale of the Jamieson Creek watershed, the hourly database is probably adequate and can give information on intensity as well as longer term totals. In addition to the time distribution of rain, sufficient data stations should exist, so that spatial variability can also be studied. The small time step of data gives the researcher the possibility to understand the temporal and spatial distribution of rainfall and the opportunity to examine if the precipitation intensities or/and duration change with elevation. Furthermore, the analysis should be based in event data. Data measured in longer time intervals such as daily, give no indication whether the precipitation in this time period is due to one or more events.

Another important point in the analysis of precipitation is to distinguish between the forms of precipitation and presumably, between rainfall and snowfall. Previous studies, as it has already been mentioned, have shown that the snowfall has different distribution
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from that of rainfall.

In this study, the rainfall distribution in the Jamieson Creek watershed is examined with the use of hourly data for the rainfall intensities for each event. The data are for the period of late June to mid-November for the years 1972-1975 and they have been taken from the five rain gauges in the watershed (Fig. 4.7). It is important to note that the rain gauges are installed in elevations of 425 m, 640 m and 853 m with two rain gauges located in the two main hillslopes of the watershed. Thirty four small, medium and large events have been analysed.

6.4.1 Total event rainfall distribution

Regression analysis between the event rainfall depth in the rain gauge 14A and the other four stations have been performed. The reason for this type of analysis is that it permits the analysis of rainfall distribution with elevation, and it gives the ability to develop predictor equations for rainfall distribution occurring at the study watershed. For mountainous watersheds it is critical to estimate the rainfall distribution, so that runoff may be estimated as accurate as possible, since it is difficult to install rain gauges at the upper elevations. Quite often only one rain gauge is available, and that gauge is usually at a low elevation.

The relationships in Table 6.2 have been developed using the rainfall depths from rain gauge 14A as independent variable. From these equations, it can be seen that the rainfall depth for each event follows a particular trend which is quite consistent (high correlation). This analysis shows that for the hillslope facing the southwest the rainfall depth increases only slightly from the elevation 425 m (gauge 14A) to elevation 640 m (gauge 21A). The increase is only 0.42 %. On the other hand, there is a dramatic decrease in the rainfall depth between gauge 14A and the rain gauge 28A at 853 m elevation (49% decrease). This large decrease is associated with a larger variability in the rainfall depth
Table 6.2. Regression equations for total event rainfall (mm).

<table>
<thead>
<tr>
<th>Equation</th>
<th>$r^2$</th>
<th>See (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$(21A) = 0.716 + 1.00419*(14A)$</td>
<td>0.991</td>
<td>4.458</td>
</tr>
<tr>
<td>$(28A) = 11.727 + 0.49041*(14A)$</td>
<td>0.777</td>
<td>12.413</td>
</tr>
<tr>
<td>$(21D) = -1.635 + 1.16098*(14A)$</td>
<td>0.984</td>
<td>7.523</td>
</tr>
<tr>
<td>$(28C) = 2.522 + 1.05569*(14A)$</td>
<td>0.953</td>
<td>11.359</td>
</tr>
</tbody>
</table>

$r^2$ is the coefficient of determination
See is the standard error of estimate

as can be seen from the smaller coefficient of determination ($r^2 = 0.777$).

It might be thought that this large decrease in the rainfall depth may be due to an obstacle or due to wind effects on the gauge catch. All the rain gauges were installed in a clear cut area with a width of about twice the adjacent tree height which should decreases the wind and shade effects of rain catch (Shaw, 1983). Moreover, comparison of rain gauges protected by a shield or without shield, located at 850 m elevation, showed no significant variation in the rain catch, indicating that the clearing was given the necessary sheltering (Kuochi Rae, personal communication). Hence, it would seen that the large decrease in the rainfall depth in the location of rain station 28A is not because of the presence of an obstacle or undercatch of rain gauge.

From the equations in Table 6.2, it can also be seen that the rainfall depth per event increases between the rain gauges 14A (425 m above mean sea level) and 21D (640 m a.m.s.l.) by 16 % and it increases only by 5.5 % between 14A and 28C (850 m a.b.s.l.), so that the rainfall depth increases up to the point 21D and then decreases by about 10
Chapter 6. ANALYSIS OF PRECIPITATION

Table 6.3. Regression equations for the total rainfall depth (mm) for the gauges in the same elevation.

<table>
<thead>
<tr>
<th>Equation</th>
<th>$r^2$</th>
<th>See (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td>$(21D) = -3.159 + 1.15985 \times (21A)$</td>
<td>0.997</td>
<td>3.06</td>
</tr>
<tr>
<td>$(28C) = -11.68 + 1.46182 \times (28A)$</td>
<td>0.85</td>
<td>20.09</td>
</tr>
</tbody>
</table>

$r^2$ is the coefficient of determination  
See is the standard error of estimate

% at the higher elevation of the rain gauge 28C. The high coefficients of determination show that the above trend is very persistent. Both points 21D and 28C are located in the hillslope which faces northeast.

Figures 6.1 and 6.2 shows the percentage increase or decrease in the total event rainfall depth with elevation. From these figures it is seen that the rainfall increases up to elevation 640 m and then decreases at greater heights. This fact is more prominent in the hillslope facing southwest, as has been previously discussed.

The first of the equations of Table 6.3 between the rain gauges 21A and 21D shows that point 21D on the hillslope facing northeast, receives about 16 % more rain than point 21A on the hillslope facing southwest. The second of the equations shows that the hillslope facing northeast at the point 28C receives about 46 % more rain than the point 28A at the same elevation on the opposite hillslope. The very small rainfall catch in gauge 28A may depend on the physiographic features and the meteorological factors.

The above observation is of high importance for the study of the Jamieson Creek response and the simulation of the runoff, since the hillslope facing northeast is more
Fig. 6.1. Percentage change of the event rainfall in the hillslope facing southwest.
Fig. 6.2. Percentage change of the event rainfall in the hillslope facing northeast.
than twice the size of the opposite hillslope facing southwest.

Classifying the rainfall events in small, medium and large as in the Chapter 5 and performing regression analysis it can be seen that the small rainfall events are more scattered than the medium and large events. Furthermore, the small and medium events show that they are affected to smaller extent by elevation, showing smaller increase or decrease with elevation than the larger events. On the other hand, the large events had the same distribution pattern as the general pattern and their trend were highly consistent (high correlation).

6.4.2 Hourly intensity distribution

The same type of analysis made for rainfall depth has also been adopted for the analysis of the rainfall intensity. Table 6.4 shows the equations derived from the regression analysis for the rainfall. These equations were developed using hourly intensities that happened at the same time at the five rain gauge locations in the watershed. The total number of hourly rainfall intensities used in the analysis is larger than a thousand.

The analysis showed that the hourly rainfall intensity decreases with elevation for both main hillslopes. For the hillslope facing southwest the hourly rainfall intensity decreases between rain gauges 14A and 21A by about 14 % and between 14A and 28A by about 58 % (Fig. 6.3).

The same trend has also been observed for the hillslope facing northeast. The hourly intensity decreases with elevation. The decrease between points 14A and 21D is about 10 % and between 14A and 28C about 20 % (Fig. 6.4).

The variation of the hourly intensity between the rain stations is much higher than the variation of the total event rainfall depth. This can be seen from the coefficient of determination. All the coefficients of determination for the hourly intensities are lower than the coefficients of determination for the total rainfall depth. Especially it should be
Fig. 6.3. Percentage change of the hourly intensity in the hillslope facing southwest.
Fig. 6.4. Percentage change of the hourly intensity in the hillslope facing northeast.
Table 6.4. Regression equations for the hourly intensities (mm/hr).

<table>
<thead>
<tr>
<th>Equation</th>
<th>( r^2 )</th>
<th>See (mm/hr)</th>
</tr>
</thead>
<tbody>
<tr>
<td>((21A)=0.206+0.86151*(14A))</td>
<td>0.803</td>
<td>0.778</td>
</tr>
<tr>
<td>((28A)=0.456+0.4248*(14A))</td>
<td>0.378</td>
<td>0.993</td>
</tr>
<tr>
<td>((21D)=0.272+0.90835(14A))</td>
<td>0.770</td>
<td>0.272</td>
</tr>
<tr>
<td>((28C)=0.417+0.8046*(14A))</td>
<td>0.681</td>
<td>1.01</td>
</tr>
</tbody>
</table>

\( r^2 \) is the coefficient of determination
See is the standard error of estimate

noted that the correlation of the hourly intensity between rain gauges 14A and 28C is very low (\( r^2 = 0.378 \)) indicating a large variation of the magnitude of rainfall between points 14A and 28A. The same result, but to a smaller extent, happens between points 14A and 28C. These results imply that the rainfall events may not occur simultaneously at the upper and lower watershed or alternatively that the relationships for hourly intensities at the five rain gauge locations do not remain constant for various events.

Analysis of the hourly intensities for small, medium and large events showed that the previously discussed trend of intensity decrease with elevation remains relatively constant despite the size of the event. The larger variation of hourly intensities has been observed for small and medium events while this variation decreases substantially for large events.

6.4.3 Discussion of the results

The rainfall events used in this study are drawn from the summer and early fall period because the complications of combined rain and snow events were avoided at this stage.
In addition, only a limited data period of four years was analyzed and consequently the return periods of the events are moderate, usually less than a 2 year return period, but one of about 5 year and one of about 10 year return period.

Analysis of the events clearly indicates that the summer and fall event rainfall increases up to elevation of 640 m and then it decreases at higher elevations. These results are in accordance with the observations of Jones (1974) and Fitzharris (1975). One possible explanation for this trend is the decrease of temperature and humidity with height. Especially, humidity decreases significantly above the layer of maximum cloud development. For the nearby Mount Seymour, Fitzharris (1975) estimated that the layer of maximum cloud development is at 800 m a.m.s.l. but it changes from storm to storm and it depends on the area characteristics. This height is much smaller than the usual estimate of the elevation of maximum cloud development of 1000 to 2000 m.

Furthermore, another possible explanation is the convective instability of the incoming air mass. Because of the high summer and fall air temperatures adiabatic lapse rate for saturated air is much smaller than that of the dry air. The incoming air mass in this period of the year is highly saturated because of the high evaporation. The difference between the adiabatic lapse rate of the incoming moist airflow and the surrounding relatively dry air causes large convective instabilities. As a result even the foothills or the lower mountain can cause rainfall. In such a case the rainfall intensities are higher at the lower elevations than at the upper elevations.

The above statement is in close agreement with the results of the analysis of the hourly intensity which showed that the rainfall intensity decreases with elevation. Also, comparison of the two frequency studies presented by Hall (1989) and Cheng (1975) showed that the rainfall intensities of small duration storms increase from the rain station at Seymour Falls Dam at 200 m elevation to the Jamieson Creek watershed rain station 21A at 640 m, and then they decrease from the elevation of the rain gauge 21A to the
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...elevation of 853 m of 28A. It should be noted that there is no frequency analysis for the
rain station at the lower elevation of Jamieson Creek watershed 14A (425 m a.m.s.l.).
The above comparison of rainfall intensities of small duration supports the result that
the rainfall intensities decrease at the upper Jamieson Creek watershed.

Hence, the results of the present analysis and also those of the previous analyses can
be qualitatively explained by the decrease of temperature and humidity with elevation
and the instability of the incoming air masses.

It will now be shown that the distribution of the rainfall depth per event can rep­
resent the overall spatial distribution of the rainfall, and that the distribution of the
hourly intensities, because of the type of analysis, can give an indication of the temporal
distribution of rainfall.

The large coefficients of determination for the rainfall depth prove that the spatial
variation of the rainfall distribution is quite small (except for point 28A). On the other
hand, the smaller coefficients of determination for the hourly intensities indicate that the
temporal variation of rainfall is larger than the spatial one. Also, the decrease of the
rainfall hourly intensities at the upper and mid elevation of Jamieson Creek watershed
and the increase or smaller decrease of the rainfall depth at the same elevations show
that the duration of rainstorms at these points is larger. During the processing of the
data, it was observed many times that at the upper elevations the rainfall was prolonged
by one or two hours more than the rainfall duration at the lower watershed.

Another important point about the distribution of rainfall in the Jamieson Creek wa­
tershed is the larger rainfall depth that is received by the hillslope facing northeast. The
change of the wind pattern by the rugged mountain terrain may explain this distribution.
Data for the evaluation of the wind direction and speed are not available. Hence, only
speculations can be stated here. The explanation given by Jones (1974) that the larger
rainfall on the hillslope facing northeast is due to the effect of the ridge between the
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Jamieson Creek watershed and the adjacent Orchid Creek watershed on the prevailing wind, seems to be reasonable. Moreover, the gradient of the hillslope facing northeast is larger than the gradient of the opposite hillslope, so that it can cause an additional increase on the rainfall.

Finally, the point that the large rainfall events have a similar distribution to the other smaller storm events is of high importance for the examination of the response of the watershed under large rainfall events and the prediction of the flood runoff from it.

6.5 EFFECTS OF RAINFALL REPRESENTATIVENESS ON THE SIMULATION OF RUNOFF

In the rugged mountain terrain it is very difficult to install rain gauges at upper elevations. The steep hillslopes, the harsh weather and the lack of roads and transportation make the collection of precipitation data difficult. On the other hand, the physiographic features and the complex atmospheric processes modify the distribution of precipitation significantly. Hence, in the mountainous areas where large variability in the precipitation exists, the gauge network is never adequate to define the detailed precipitation distribution.

In usual practice, rainfall data from a rain station close to the study watershed or even in a neighboring region are transposed or used as they are for the assessment of rainfall distribution and the simulation of the runoff.

In order to examine the effects of rainfall representativeness on the simulation of the runoff, data from one rain gauge and average rainfall from the five rain gauges in the Jamieson Creek watershed were used. The simulation results for the one gauge rainfall and average rainfall are compared.

The watershed model developed and presented in this thesis is used. The rain gauge
Table 6.5. Effects of the rainfall representativeness on model parameters

<table>
<thead>
<tr>
<th>Event</th>
<th>One station rainfall</th>
<th>Average rainfall</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>KF (hrs)  b (mm/hr)  R (mm)</td>
<td>KF (hrs)  b (mm/hr)  R (mm)</td>
</tr>
<tr>
<td>18-25/9/1972</td>
<td>4.2    3.3    197.79</td>
<td>2.5    1.8    175.06</td>
</tr>
<tr>
<td>23-25/6/1973</td>
<td>2.0    1.0    21.08</td>
<td>2.0    1.5    26.70</td>
</tr>
<tr>
<td>19-23/9/1973</td>
<td>3.0    1.29   68.58</td>
<td>2.4    1.3    73.07</td>
</tr>
<tr>
<td>26-28/6/1974</td>
<td>2.2    1.50   35.81</td>
<td>2.2    1.5    38.42</td>
</tr>
</tbody>
</table>

R is total rainfall depth

14A is used as the main gauge, when only one set of rain gauge data is used. This gauge has been selected because it is close to the mouth of the watershed at the lower elevations so that it represents better the real practice situation where the rain gauge is either at the lower watershed or even outside the boundary of the watershed. The average rainfall is obtained by the Thiessen polygon method with data from five rain gauges (Fig. 4.7).

Four events have been selected for the presentation of the effects of rainfall representativeness on simulation of runoff. The rainfall of the event of 18-25/9/1972 was much larger for the gauge 14A than for the rest of the basin (Fig. 6.5). An attempt to best-fit the simulated hydrograph to the observed one using data only from the gauge 14A resulted in large values of the model parameters. The storage coefficient of fast runoff (KF) took a value of 4.2 \textit{hrs} and the parameter b of the rainfall separation equation took a value of 3.3 \textit{mm/hr} (Table 6.5). These values of the model parameters are considerably different from their mean values as they have been derived in Chapter 5 (mean KF=2.11 \textit{hrs} and mean b=1.45 \textit{mm/hr}).
Fig. 6.5. Effects of rainfall representativeness on streamflow simulation
(18-25/9/1972 event)
Simulating the same event with five station average rainfall the model parameters took values $KF=2.5 \text{ hrs}$ and $b=1.8 \text{ mm/hr}$, which are very close to their mean values (Table 6.5).

To show the effect of rainfall representativeness on the rainfall simulation, the rainfall-runoff event is simulated with data from one rain gauge and five rain gauge average rainfall with the mean values of model parameters i.e. $KF=2.11 \text{ hrs}$ and $b=1.45 \text{ mm/hr}$. Figure 6.5 clearly shows how large the effect of the correct representation of rainfall is on runoff simulation.

The same procedure has been followed for the other three events. For the rainfall-runoff event of 23-25/6/1973 the peak rainfall intensity was larger at the upper watershed than in the lower watershed (Fig. 6.6). Best-fitting of the simulated hydrograph to the observed one using one rain station data resulted in values of $KF$ of $2.0 \text{ hrs}$ and $b$ of $1.0 \text{ mm/hr}$. The smaller value of $b$ than its mean value is the result of the smaller peak rainfall intensity at the lower elevation.

Simulation of the same event with average rainfall resulted in values of $2.0 \text{ hrs}$ for $KF$ and $1.5 \text{ mm/hr}$ for $b$ (Table 6.5). So, the non-representative peak rainfall intensity affected only the value of the parameter $b$, and it had no effect on the storage coefficient of fast runoff ($KF$). Figure 6.6 compares the simulation results for one rain gauge and the average rainfall data using the mean values of the parameters.

For the event of 19-22/9/1973 there was a timing difference in the occurrence of rainfall between the lower and upper watershed. Rainfall at the upper watershed proceeded that at the lower watershed (Fig. 6.7). This timing difference in rainfall distribution resulted in large value of the storage coefficient of fast runoff ($KF=3.0 \text{ hrs}$) when the event is simulated with one rain station data (Table 6.5). Simulation with average rainfall gave values of $KF=2.4 \text{ hrs}$ and $b=1.3 \text{ mm/hr}$ (Table 6.5). The comparison of the simulation results with one station rainfall and average rainfall using the mean values of $KF$ and $b$
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Fig. 6.6. Effects of rainfall representativeness on streamflow simulation (23-25/6/73 event)
with the observed hydrograph can be seen in Figure 6.7.

For the event of 26-28/6/1973 the rainfall does not differ significantly in space and time (Fig. 6.8). As a result the model parameters did not change at all and remained constant for the simulation with data from one rain station or five station average rainfall (Table 6.5), and the simulation results were essentially identical (Fig. 6.8).

These results suggest that the rainfall is the most important input factor to the model. Rainfall data from one rain gauge located at the lower watershed may not be representative of the rainfall distribution except when that the rainfall is quite evenly distributed over the watershed. But rainfall is very rarely evenly distributed in the study watershed. Variation of the rainfall distribution in space and/or time results in underestimation or overestimation of the runoff using a watershed model.

The analysis have shown that one data station is usually not a direct representation of the basin precipitation. However, in most hydrological situations it is quite usual to have only one data station and sometimes this station is not even in the watershed. The question that must be considered is whether the single data station can be used as an index of basin precipitation if proper allowance is made for elevation gradient and other factors.

A prediction of the rainfall distribution is attempted here. The equations developed in section 6.4.2 of the thesis using hourly intensity data from five rain gauges (Table 6.4) are used for the prediction of rainfall distribution. Then, the average rainfall obtained with the Thiessen polygon method from the actual data has been compared to the average rainfall assessed using the predictors equations of Table 6.4. The comparison has been made for rainfall data from the years 1976 and 1977. It should be noted that the predictor equations have been developed from rainfall data for the years 1972-1975.

Appendix C includes all the figures where the measured and the predicted average rainfall is compared. In general, the results of the rainfall prediction are comparable to
Fig. 6.7. Effects of rainfall representativeness on streamflow simulation (19-23/9/1973 event)
Fig. 6.8. Effects of rainfall representativeness on streamflow simulation (26-28/6/1974 event)
the measurements. Especially, in some cases the prediction of the rainfall is very good (Fig.B2, Fig.B4, Fig.B8, Fig.B9). The coefficients of determination between predicted and observed average rainfall ranged from event to event from 0.70 to 0.99.

The above results suggest that the rainfall data from the rain gauge 14A can be used with the predictor equations developed for the hourly intensity for the estimation of rainfall over the watershed.

6.6 CONCLUSIONS

The processes affecting the distribution of precipitation in the mountainous areas are complex and need careful examination. The literature review indicated that the major factors influencing precipitation are: elevation, slope, aspect, wind direction and speed, humidity, temperature and instability of air mass.

Examination of the rainfall distribution in the Jamieson Creek watershed reveal a distinct pattern of rainfall distribution. The rainfall for each event increases up to the mid-elevation of the study watershed, and then decreases at higher elevations. Moreover, the hourly rainfall intensities decrease with height in all the elevation bands of the watershed, so that the maximum intensity is in the lower elevations. The above distribution pattern can be explained by the decrease of temperature and humidity with elevation, and the instability of the incoming air mass.

The results of the analysis herein are in agreement with the results of previous studies in the southern boundary of the Coast Mountains. Studies (Fitzharris, 1975, Storr and Ferguson, 1972) have shown that the rainfall has a different distribution from that of snowfall. Hence, when one analyses the precipitation, must be careful about its form.

The results of this study and of others suggest that the rainfall in mountainous areas with humid climate increases up to a certain elevation, which is approximately two thirds
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the mountain height, and then it decreases at higher elevations.

Also, the findings of this analysis suggest that the duration of rainfall events may be prolonged in the upper and mid-elevation of the study watershed, since the hourly intensities are decreasing with elevation while the total rainfall depth increases up to the mid-elevation of the watershed and then decreases at the upper elevations at a smaller rate than the rainfall intensity.

Furthermore, the aspect, the slope of the hillslopes, and the position of the watershed may affect the distribution of rainfall in the study watershed. The steeper hillslope facing northwest receives about 30% more rainfall than the opposite hillslope facing southwest. The distribution of the wind in the area may explain this variation. The above result is very important for the analysis of the runoff from the watershed, since the hillslope which receives more rainfall is about twice the size of the more sheltered hillslope in the watershed.

Analysis of the rainfall distribution with respect to the size of the event indicated that large events have a more consistent distribution of rainfall. This last finding is very important for the prediction of the flood runoff from the study area.

Examination of the representativeness of the rainfall showed that the rainfall data from one rain gauge at the lower elevation of the watershed may not represent accurately the rainfall distribution over the watershed. The estimation of runoff is most sensitive to rainfall input and its representativeness. It is very important to note that, with better representation of rainfall, the model parameters took values which were essentially constant. This result emphasizes the previous result that the watershed response is linear. Inaccurate representation of rainfall can lead to large errors in the determination of the basin storage, and consequently, until the rainfall input has been correctly defined, any comparison between hydrological models will be false.

The accuracy and reliability of the rainfall distribution may increase if predictor
equations developed from the analysis of the hourly rainfall intensities, and rainfall data from one rain gauge at the lower elevation are used.
Chapter 7

CONCLUSIONS AND RECOMMENDATIONS

7.1 CONCLUSIONS

The hydrologic response of the study watershed has been investigated with the use of a simplified model. This simplified model was originally intended for a preliminary assessment of the watershed. Then a more complex and better model would have been designed. However, it became clear that this simple model gave a good representation of the watershed response, provided the rainfall input could be accurately assessed.

The runoff in the study watershed is mainly generated through the mechanism of subsurface pipeflow which has been recognized as very similar to that during the generation of overland flow over a previous surface. This is the reason why the watershed model utilises the linear storage routing technique. The model separates the rainfall input into slow and fast flow components of runoff with the use of an exponential rainfall separation equation, but for higher runoff events evaluated in this study, a simple linear separation was sufficient.

Application of the model to the study watershed indicated that it performed well. It has been argued that the reason for this is the good physical representation of the physical processes and the use of highly accurate and representative data for both rainfall and streamflow.

The model application to the study watershed showed that the model parameters were reasonably consistent with their physical representation.
Chapter 7. CONCLUSIONS AND RECOMMENDATIONS

Statistical analysis of the variation of the model parameters for twenty six events proved that this variation was small except for four of the rainfall-runoff events. These events happened during intense summer storms and very dry soil moisture conditions. Furthermore, for all the rainfall-runoff events the rainfall separation equation took a linear form.

The above two findings showed a reasonably linear response of the watershed for larger rainfall events. It was argued that the linearity in the hydrologic response is due to the presence of vertical and lateral soil pipes. The vertical soil pipes are prefered infiltration routes which do not show the non-linear response of the vertical infiltration process through the soil matrix. Non-linearities in the watershed response are observed during intense rainfall events and dry soil conditions, and small rainfall events because of air entrainment in the soil and water movement through soil matrix, respectively.

The above results have not been validated for other watersheds, but at least give and indication that previous beliefs regarding the non-linear response of small, steep watersheds may not be valid. These findings have a direct effect on the rainfall-runoff simulation for flood prediction and engineering projects, since it has been proven that rainfall-runoff events in a small mountainous watershed can be simulated with a linear model.

Comparison has been made between the most common and widely used method of hydrograph separation in Forest Hydrology (Hewlett method) with the watershed model results, and the results showed that the quick flow volume obtained by the two methods is very similar. This result suggests that the Hewlett method can be applied to this particular watershed and perhaps in other similar basins, as well.

The second main objective of this work was to understand the physical processes during the generation of rainfall in areas with humid climates, analyse the rainfall distribution in the study watershed and examine the effects of rainfall representativeness on
runoff simulation.

Review of the literature showed that the distribution of the precipitation is affected by many meteorological and physiographic parameters. From the present study and other studies in areas with a humid climate the rainfall does not continue to increase with elevation, but only up to a certain elevation, and then it levels off or decreases with additional height.

Regression analysis of the rainfall depth per event in the study watershed showed the same trend. The rainfall depth increased up to the mid-elevation and then it decreased at upper elevations. On the other hand, the hourly rainfall intensities decreased with elevation indicating larger duration of the rainfall events at the upper watershed.

This trend of rainfall distribution is consistent with the decrease of temperature and humidity with elevation, and maximum precipitation may correlate well with the height of maximum cloud development.

Regression analysis of the rainfall depth between the rain gauges located in the same elevation but at the opposite hillslopes revealed that the larger hillslope facing northeast receives about 30% more rainfall per event than the opposite hillslope. This finding is of great importance for the simulation of rainfall-runoff and thus, for the study of the watershed response, because the watershed model is very sensitive to the representativeness of the rainfall input.

The large variation of the rainfall, even in the small study watershed, has large effects on the simulation of the runoff, proving that rainfall and its representativeness are the most important input factors to the model. Rainfall data from one rain gauge located in the lower watershed may not be representative of the rainfall distribution, and this has a large effect on the estimation of the runoff. But if the rainfall input to the model is highly accurate and representative, then the model performs well and the variation of the model parameters decreases indicating, once more, the linearity of the watershed
Chapter 7. CONCLUSIONS AND RECOMMENDATIONS

response.

The representativeness of rainfall data from one rain gauge may increase if predictor equations developed from hourly rainfall data from five rain station locations, are used.

7.2 RECOMMENDATIONS

Further work is necessary to improve the basis of the analysis, the watershed model, and extend the applicability of the findings to other watersheds in the Pacific Northwest. It is not argued that the findings of this thesis are applicable to other watersheds but at least, the present work addresses key questions which challenge some of the common beliefs about the hydrologic response of small mountainous watersheds and the rainfall distribution in mountainous areas. The next points are considered as important points for further research.

- The simplified watershed model could be improved if the parameters could be estimated from the physiographic features of the watershed. This estimation should incorporate the effects of scale, and the physical processes of the generation of pipeflow. Evaluation of the model parameters from physiographic and physical features could lead to a more physically representative model.

- The analysis presented should be extended in other watersheds in the Pacific Northwest which have similar characteristics to those of Jamieson Creek watershed. Also, the analysis should be extended to larger rainfall-runoff events, since this is very important for the estimation of peak flow and/or runoff volume for engineering design. The suggested improvement of the watershed model, the extension of the analysis to other watersheds and the validation of the findings, particularly the linearity of the watershed response, may allow the use of such a model for the prediction of runoff from mountainous watersheds in the region. Also, the results
of such a model may be used as input to submodels for the estimation of the production of sediments and pollutants.

- From the forest logging perspective, in the Pacific Northwest, it is necessary that the effects of logging on the storm hydrograph and the production of sediments and pollutants be estimated accurately. At present there is no such model for the above region even though the effects of logging are very important. Study of the logging effects on the runoff generation processes and derivation of relationships between logging practices and values of the parameters can help to derive an accurate and representative model for the estimation of logging effects on water quantity and quality.

- There are two main mechanisms for the production of flood runoff from mountainous slopes in the Pacific Northwest: a) large rainfall events, and b) rain on snow events. Incorporation of a snowmelt subroutine in the model, and study of rain on snow events can widen the applicability of the model.

- Larger scale analysis of the distribution of precipitation is necessary for the validation of the trend observed in the Jamieson Creek watershed. Equations for the prediction of the precipitation distribution can be developed for use in ungauged mountainous watersheds, so that the estimation of the runoff would be more accurate and reliable. Further work is necessary in order to examine the different distribution pattern of rainfall and snowfall and the effects of precipitation distribution on runoff simulation.
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Appendix A

SIMULATION RESULTS
Fig. A1. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 18-25/9/1972).
Fig. A.2. Comparison of simulated flows with best-fit and average values of the parameters with observed flow. (Event 23-25/6/1973).
Fig. A3. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 16-19/8/1973).
Fig. A4. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 19-22/9/1973).
Fig. A5. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 17-19/10/1973).
Fig.A6. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 1-3/10/1974).
Fig. A7. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 26-28/6/1974).
Fig. A8. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 4-6/11/1974).
Fig. A9. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 22-24/8/1975).
Fig. A10. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 25-26/8/1975).
Fig. A11. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 27-28/8/1975).
Fig.A12. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 2-5/10/1975).
Fig. A13. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 4-6/10/1975).
Fig. A14. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 14-16/10/1975).
Fig. A15. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 15-20/10/1975).
Fig. A16. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 15-17/8/1976).
Fig. A17. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 26-30/8/1976).
Fig. A18. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 4-8/9/1976).
Fig. A19. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 9-12/10/1976).
Fig. A20. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 23-26/10/1976).
Fig. A21. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 7-10/7/1977).
Fig. A22. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 16-17/7/1977).
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Fig. A23. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 23-25/8/1977).
Appendix B. SIMULATION RESULTS

Fig. A24. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 24-26/8/1977).
Appendix B. SIMULATION RESULTS

Fig. A25. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 18-21/9/1977).
Appendix B. SIMULATION RESULTS

Fig. A26. Comparison of simulated flow with best-fit and average values of the parameters with observed flow. (Event 22-26/9/1977).
Appendix C

PREDICTED AND OBSERVED AVERAGE RAINFALL
Fig. B1. Observed and predicted average rainfall using the predictor equations.

$\text{Observed average rainfall}$

$\text{Predicted average rainfall}$

$r^2 = 0.77$
Fig. B2. Observed and predicted average rainfall using the predictor equations.
(Event 15-17/8/1976)
Fig. B3. Observed and predicted average rainfall using the predictor equations.
(Event 20-21/8/1976)

\[ r^2 = 0.83 \]
Fig. B4. Observed and predicted average rainfall using predictor equations. (Event 26-28/8/1976)
Fig. B5. Observed and predicted average rainfall using the predictor equations.
(Event 4-6/9/1976)

- - - - Observed average rainfall
- - - - Predicted average rainfall

\[ r^2 = 0.85 \]
Fig. B6. Observed and predicted average rainfall using the predictor equations. (Event 9-10/10/1976)
Fig. B7. Observed and predicted average rainfall using the predictor equations.
(Event 7-9/7/1977)
Observed average rainfall

Predicted average rainfall

$r^2 = 0.98$

Fig. B8. Observed and predicted average rainfall using the predictor equations.
(Event 16/7/1977)
Fig. B9. Observed and predicted average rainfall using the predictor equations. (Event 23/8/1977)

\[ r^2 = 0.99 \]