PEAK-FLOW TRAVEL-TIME CHARACTERISTICS OF THE FRASER RIVER, BRITISH COLUMBIA

by

Anne Shuttleworth

B.Sc. (hons), University of Wales, 1979

THESIS SUBMITTED IN PARTIAL FULFILLMENT OF

THE REQUIREMENTS FOR THE DEGREE OF

MASTER OF SCIENCE

in the Department

of

Geography

SIMON FRASER UNIVERSITY

February, 1984

APPROVAL

Name: Anne Shuttleworth

Degree: Master of Science

Title of Thesis: Peak-Flow Travel-Time Characteristics of the Fraser River, British Columbia

Examining Committee:

Chairperson: R. Hayter

E.J. Hickin
Senior Supervisor

M.C. Roberts

R.B. Sagar

Olav Slaymaker
Professor
External Examiner
Department of Geography
University of British Columbia

Date Approved: February 17, 1984

PARTIAL COPYRIGHT LICENSE

Title of Thesis/Project/Extended Essay

Peakflow Travel-times of the Fraser River, British Columbia

Author: ______________________

(signature)

Anne Shuttleworth

(name)

June 5, 1984

(date)

ABSTRACT

A time-area model was developed to synthesise hydrographs for the Fraser River, British Columbia. These hydrographs reflect the size and shape of the tributary basins, and the effects of storage and attenuation are incorporated by the use of Muskingum flood routing technique. As each tributary is added to the Fraser River flow, its relative importance to the mainstream peak is assessed. The effects of the timing and magnitude of tributary inputs are shown to be critical in determining the time of absolute peak-flow.

Apparent travel times between gauging stations for rainfall events were examined for the period 1971 to 1980. Travel times were derived from hourly discharge data which were obtained directly from streamflow recorder charts. A negative relationship between discharge and travel time should exist where river flow mechanics dominate, but no decisive patterns were found, either on an annual or seasonal basis. Wave celerity appeared to be very fast in many cases, and in several others, the upstream peak occurred after the downstream peak. This behaviour is predicted by the time-area model. Therefore it is concluded that it is basin morphology and not translation that exerts the primary control upon runoff patterns in the Fraser Basin.

Real hydrograph changes in time and space can be interpreted using the model as a means to infer the source area of runoff. When data are grouped according to the spatial

distribution of precipitation, trends between discharge and travel time can be identified, for cases where substantial tributary inputs between two stations do not dominate the time-of-peak. It is concluded that the time-area model is a useful tool for interpreting travel time characteristics of the Fraser River flood waves.

DEDICATION

In memory of Mum

ACKNOWLEDGEMENTS

Sincere thanks are offered to Ted Hickin, my Senior Supervisor, for his guidance, patience and support. Also warm thanks go to Mike Roberts and Brian Sagar for their help and advice, and to Professor Olav Slaymaker for kindly acting as my external examiner.

I would like to say a special thank-you to Oliver Nagy of the Water Survey of Canada data section for making data acquisition a pleasure, and thanks also to Gary Schaeffer of the Atmospheric Environment Service.

Many thanks to everyone who has given me their valued friendship and support in the recent years, especially: Helen, Larry, Steve R, Chris, Barb, Rick, Mo, Brent, Careen, Joan, Ron, Pat, Uschi, Gary, Steve G, Geir, Cathy and Lorraine.
I thank you all.

Last, but not least, I wish to thank Michel for his encouragement, love and understanding.

TABLE OF CONTENTS

LIST OF TABLES

LIST OF FIGURES

I. Background to study

1.1 Introduction

Relatively few studies have addressed the problem of flow travel time variations within a watershed (Pilgrim, 1977). Studies have tended to polarize, either toward examining the effects of geomorphic characteristics upon response time (lag time), or considering changes in flood wave shape as it passes along a channel. In a large basin however, both aspects are important in determining peak flow travel times.

Travel time is defined by Linsley, Kohler and Paulhus (1949) as the elapsed time between flood wave crests at two stations. They noted the lack of constancy for a given reach but suggested travel time behaved in a regular fashion: being long at low stages, decreasing to a minimum at moderate stages, and increasing somewhat to stages above bankfull. Between reaches, travel time will vary depending on hydrologic and hydraulic differences. However, the regular relationship between travel time and discharge can be disrupted by major tributary inputs and a nonuniform precipitation distribution.

In this study, a time-area model of the Fraser basin is used to interpret peak flow travel time irregularities for rainfall events. The Fraser Basin is irregularly shaped, contains contrasting hydrometeorological regions and supplies numerous tributaries to the Fraser River. Therefore it permits an examination of the way in which spatially limited processes are incorporated into the overall basin response. Time-area concepts were used to simulate a series of contrasting precipitation inputs which were modified to Instantaneous Unit Hydrographs using the Muskingum flood routing technique. Peak flow travel times, derived from the model, then can be used to interpret real travel time irregularities. The model helps to identify where river flow mechanisms are dominant and where other factors need to be considered.

1.2 Hydrological Background

1.2.1 Drainage Basin Models and the Hydrograph

The land component of the hydrologic cycle is described by More (1969) as consisting of a precipitation input that is "distributed through a number of storages by a series of transfers leading to outputs of basin channel runoff, evapotranspiration and groundwater". A hydrograph represents the time distribution of that runoff from a drainage basin and Rogers (1972) noted that its peak corresponds to flow arriving from the maximum basin width. The hydrograph is therefore an integral of physiographic and climatic variables (Roberts and Klingeman, 1970). The relationships between basin features, climatic inputs and the hydrograph are documented or implicit in numerous papers, including those of Butler (1977), Howard and Smith (1977) and Weyman (1975).

Runoff reaches a river by numerous routes which Hewlett and Hibbert (1965) classified either as quick or delayed flow in order to avoid making process assumptions. The amounts translated by the differing means depend ultimately on basin characteristics, but quick flow generally corresponds to storm runoff. A hydrograph represents basin response to a storm and therefore hydrograph shape reflects basin storage and its influence on travel times. Numerous methods of hydrograph analysis are available (Chow, 1964) but the unit hydrograph of Sherman (1932) remains the most widely used technique. Sherman suggested that a unit hydrograph is specific to a given basin and that it can be used to understand individual basin processes. The theoretical bases of the unit hydrograph have been much debated. For example, Heerdegen (1973) questioned the rigidity of the assumptions. However, despite the problems, it still proves to be a valuable method.

Physical models of watersheds have been used to examine simple relationships between basin characteristics and runoff (for example Roberts and Klingeman, 1972; Black, 1970; and Black, 1972). Subsequent studies focussed on the similarities between models and basins. These models illustrate the relationships to be expected for ideal, reduced scale conditions and hence may be limited by problems of hydraulic similitude.

They are useful tools for understanding simple basin processes and are conceptually of great importance.

Simple mathematical modeling using correlation and regression techniques has had only limited success (Vorst and Bell, 1977). Correlations obtained are rarely applicable to a range of regions and possibly limited by poor mathematical representation of the relationships. Linear mathematical models do not explain several important features including the non-linear relationship between peak discharge rate and precipitation intensity (Wooding, 1965). Preliminary studies of non-linearity in laboratory channels were made by Amorocho (1963) and subsequently Kulandaiswamy, 1964; Snyder, Mills and Stephens, 1970; Reed et al, 1975; Pilgrim, 1976; and Singh and Woolhiser, 1976. A linear rainfall-runoff relationship is satisfactory for large basins but more deviations from linearity are observed in smaller basins (Wang et al,1981).

Computer technology has enabled advances in the synthesis of watershed processes: the Stanford Watershed Model IV (Crawford and Linsley, 1966) is a classic example. As models become more sophisticated, parameter optimisation makes them more specific to the catchment for which they were developed, as exemplified in models of the Meuse (Bultot and Dupriez, 1976) and the Fraser (Quick and Pipes, 1976a). Vorst and Bell (1977) cautioned that parameter values outside the range of streamflow data used may not be reliable.

The scale of a study is important, both spatially and temporally. For the micro-scale, an examination of the physical complexities of the watershed is necessary. At the macro-scale, general trends in parameters are investigated, allowing descriptions to be simplified. While it is well known that basin size influences runoff, an understanding of the implications of different scales is important for data transfer. Pilgrim and coworkers (1982) examined the effects of catchment size on hydrological relations and concluded general patterns between large and small catchments exist, but they are not simple and well defined.

Tools for predicting the hydrograph characteristics of a watershed abound, ranging from simple, empirical precipitation - runoff relations to complex routing techniques (Murphey et al, 1977). Until recently, spatial processes in hydrology have been relatively neglected in contrast to the multitude of studies of temporal stochastic processes (Ord and Rees, 1979). Many studies are limited because they consider hydrographs at only one point in the river system and assume that a single storm produces a simple flood hydrograph. The concept of multiple inputs however, is attracting more attention, for example, Mawdsley and Tagg (1981) and Huthman and Wilke (1982).

1.2.2 The Time-area concept

Geomorphology controls the time taken for translation of flow and ultimately influences hydrograph shape (Gray, 1965). One of the earliest attempts to quantify the relationships among flood hydrograph characteristics and the size, shape and gradient of a drainage system was that of Horton (1932). More recently, Vorst and Bell (1977) found that the distance travelled to the outlet is usually significant in predicting flood response time. Singh (1975) showed time of concentration to vary with the space-time distribution of rainfall. He reports Mulvany (1850) as possibly being the first to show clearly the relation between time of concentration and maximum runoff.

The time taken for an actual drop of water to travel through the system can be used to construct a time-area map of a basin. By assuming that flow velocity remains constant, isochrones[1] can be placed objectively at equal distances from the basin outlet. Laurenson (1964) proposed that the time interval between precipitation and its effect at the outlet was a more realistic interpretation than the actual travel time of a water droplet.

A time-area histogram may be used to simulate an

---

[1] lines of equal (travel) time

Instantaneous Unit Hydrograph[2] at the basin outlet (Clark,1945; Nash 1957; and Dooge, 1959). The concepts are well founded since area is undisputedly correlated with runoff volume and distance with the time crdinates of the hydrograph. In experimental work with physical models, Němec (1972) found full confirmation of time-area concepts. Clark (1945) claimed that the method allows the identification of elements in the hydrograph attributable to certain parts of the basin. The method involves applying flood routing techniques, designed to mathematically simulate flows using storage modifications. This assumes that it is possible to separate watershed translation and storage effects. The former allows pure translation (inflow curve = outflow curve) via the stream network to the outlet based solely on channel travel time. At the outlet the translated flow is damped by the incorporation of storage effects. This is achieved by routing the flow through a hypothetical linear reservoir[3] producing an outflow hydrograph for a basin.

The first attempts were based mainly on an assumed constant translation speed and uniform precipitation distribution. Clark (1945) obtained unit hydrographs which were not perfect, but within the accuracy usually attainable in runoff calculations.

---

[2] The hydrograph of runoff from an infinitesimally small period of precipitation that would result from an inch of water spread uniformily over an area and then allowed to runoff.

[3] Chow (1964) describes a linear reservoir as a fictitious reservoir in which storage is dependent on outflow, and a linear channel as a fictitious channel in which the time necessary for translation of a discharge through a given channel length is constant.

Pilgrim (1976) noted that detailed field work did not necessarily produce more accurate results for flood flows than measurement of a geomorphic parameter, such as area, from a map. Clark (1945) considered large drainage areas by deriving unit hydrographs for each subbasin, and then routing these to a main stream point. Kirkby (1978) suugested that output from a large drainage basin is damped relative to the input and therefore the prediction of storm hydrographs for large areas is successful because of the damping. In view of this, he also suggested that there were few advantages to be gained by including more detail than necessary.

Nash (1957) proposed that a drainage area can be simulated by a number of reservoirs in series. Studies prior to Nash, had implied this concept but not mentioned it explicitly (Chow, 1964). The Nash model however, may not yield satisfactory results in basins with a large delay time between precipitation excess and commencement of surface runoff. This is because the model is based on storage routing alone, and only makes indirect use of the translation property (Kulandaiswamy, 1964).

To incorporate the translation element, Dooge (1959) represented a basin as a series of alternating linear channels and reservoirs. In simple terms, a channel allows pure translation and a reservoir incorporates storage effects allowing hydrograph shape to change. In Dooge's model, subareas delimited by isochrones are represented by a linear channel in series with a linear reservoir. Outflow from the linear channel is determined from a time-area diagram and added to the outflow from the preceeding subarea. The sum then provides inflow to a linear reservoir. Unfortunately, the Instantaneous Unit Hydrograph solution that results is mathematically complex and not easily solved for practical purposes.

As computer facilities expanded, more complex mathematical functions were handled. For example, Manderville and O'Donnell (1973) introduced time variant versions of the linear channel and linear reservoir to produce hydrographs that are not governed by the harsh assumptions in Unit hydrograph theory. Ragan (1966) attempted to introduce local inflows but model oversensitivity caused many problems, making it unsuitable for a basin such as the Fraser.

Generally, a time-area model produces simple but reliable hydrographs. The Clark version has been used in many subsequent studies including the well-repected Stanford Watershed Model IV of Crawford and Linsley (1966). In 1973, both time-area concepts and Muskingum flood routing were used in the HEC I model of the U.S. Army Engineeering Corps (Viessman, 1977).

1.2.3. Flood Routing

This is a technique for predicting temporal and spatial variations of a flood wave as it traverses a river reach or reservoir (Viessman et al, 1977). Flood routing was originally defined as the determination of a hydrograph from an upstream hydrograph (Ding, 1974) and the first reference to what is considered to be a real application of routing along a river was by Graeff, in 1833 (Viessman et al, 1977). Most procedures have been developed since 1900; the majority being of a hydrological storage nature involving solutions based upon inflow and outflow hydrographs. Inflow hydrographs may be substituted by precipitation data or lagged input from time-area histograms. A second category of flood routing consists of hydraulic methods based on equations for unsteady flow in channels.

Hydrologic flood routing methods calculate the volume of water temporarily stored in a reach. Natural streams often have high channel resistance and storage capacity which modify a wave as it traverses the reach. Ideally, the rate of storage in the reach should equal the difference between ordinates of input and output hydrographs. A negligible loss or gain in total volume is assumed, and the equation of continuity applied.

The best known hydrological flood routing technique is the Muskingum method which views channel storage as having two constituents: prism storage depends on output and wedge storage reflects the difference between input and output. During flood wave advance, input is greater than output producing a wedge of storage. Conversely, output becomes greater than input during flood wave recession and a negative wedge results. The method assumes a unique relationship between stage and discharge, a storage constant (x) that is acceptable for all flows and a storage time constant (k) that is reasonably close to travel time within the reach. That is, k and x relate channel storage to input and output.

The kinematic wave theory (Lighthill and Whitham, 1955) offers an another approach. A kinematic wave has constant amplitude and only one velocity at each point in the wave. Convection-diffusion principles were originally defined by Hayami (1951) who recognised that irregularities in channel geometry affect flood waves and attempted to make allowances for them. However, this method has disadvantages which make it incompatible with a time-area model. For example, the problem of evaluating channel geometry irregularities is immense and the the method can be applied only to a limited range of discharges. Also, major tributaries cannot be included unless routing is from tributary to tributary and therefore problems arise in poorly gauged areas.

Resistance to flow, storage and channel irregularities govern outflow from a reach and can lengthen the time base and lower the crest of a storm hydrograph. This results in attenuation. The modeling of wave attenuation however, presents problems to both flood routing methods discussed here. For example, the prediction of attenuation by the Muskingum method is not easily reconciled with the assumption of a unique stage-discharge relation. The kinematic wave solution may not be appropriate either, because Henderson (1966) suggested that kinematic wave velocity may not occur during attenuation. Wave subsidence is usually more important in shallow sloped rivers than in steep mountain streams.

Simple mathematics are inadequate for nonuniform channels of complex section with varying slope and roughness. Even if these irregularities are assessed by extensive, costly surveys, they are sufficiently dynamic to limit the use of any data set. Also, the wave itself is usually an intermediate between pure translation and pondage and cannot be defined simply. Routing in natural channels is therefore highly complex, but assumptions can be made to limit the effect of channel irregularities.

The Muskingum method is one of the most satisfactory, and therefore popular, techniques of flood routing (National Environmental Research Council, 1975) and its relevance today is emphasised by Pilon and Cheng (1983). The model has an overall simplicity, both conceptually and practically. Chow (1964) suggested that although the method is not exact, it is adequate, especially for flood control plans and designs. Also, this method can include tributaries that are poorly gauged. Therefore, the Muskingum method was chosen for the Fraser Basin time-area model.

The Muskingum Method.

Developed by G.T.McCarthy (1938) through studies of the Muskingum Conservancy District by the U.S. Army Corps of Engineers, the Muskingum method is based on the continuity equation (1)

$$I-O=ds/dt \tag{1}$$

where I=inflow
O=outflow
ds/dt=changes in storage

Storage in a given channel reach depends on inflow, outflow and other hydraulic characteristics of the channel section. It can be expressed as:

$$S = \frac{b}{a}\left[xI^{\frac{m}{n}} + (1-x)\,O^{\frac{m}{n}}\right] \tag{2}$$

where S is storage,
a and n are constants reflecting stage-discharge at the end of the reach,
b and m represent stage-volume characteristics,
and x defines relative weights of I and O.

The Muskingum method assumes $m/n = 1$ and lets $b/a = k$. The value k is a storage constant for the reach and thus

$$S = K\left[xI+(1-x)O\right] \tag{3}$$

In application, equation (2) is expressed in the finite form

$$\frac{I_2+I_1}{2} - \frac{O_2+O_1}{2} = \frac{S_2-S_1}{t} \tag{4}$$

therefore from equation (3)

$$S_2-S_1=K\left[x(I_2-I_1)+(1-x)(O_2-O_1)\right] \tag{5}$$

which, when substituted in (4) becomes the Muskingum routing equation:

$$O_2=C_0I_2+C_1I_1+C_2I_1 \tag{6}$$

in which

$$C_0 = \frac{-Kx+0.5\Delta t}{K-Kx+0.5\Delta t}$$

$$C_1 = \frac{Kx+0.5\Delta t}{K-Kx+0.5\Delta t}$$

$$C_2 = \frac{K-Kx-0.5\Delta t}{K-Kx+0.5\Delta t}$$

and $C_0+C_1+C_2=1.0$

Routing involves solving equation (6) for successive time increments.

Determination of constants.

The Muskingum method assumes a unique stage-discharge relationship, so x should allow the volume stored to be the same for both rising and falling stages. Usually, x varies between 0.0 and 0.5. If storage is a function of outflow alone, as in a reservoir, x=0.0. When no storage occurs, x=0.5. In the latter case, inflow and outflow carry equal weight, and hydrograph shape is not changed (pure translation). Clark (1945) routed a cosine curve using x values between 0.0 and 1.0 and found x was responsible for changing peak magnitude. The constant k indicates the time required for the centre of mass of a flood wave to pass from an upstream point to a designated point downstream. It is therefore a function of reach length and wave velocity (Strupczewski and Kundzewicz, 1980) and usually approximates the travel time through a reach (Viessman et al, 1977).

If actual input and output hydrographs for a reach are available, k and x can be determined by trial and error. This involves plotting graphs from equation (3) for a series of x values. The best x value plots the narrowest hysterisis loop, and k equals the reciprocal of the slope through this loop. Unfortunately, the data available for the Fraser River do not permit this approach. There are many lateral inputs between continuous gauging stations, and although these inputs may be measured daily, they are too infrequent to be useful. Brakensiek (1963) calculated coefficients by using the section rating function and conveyance data for ungauged watersheds, but concluded that numerous assumptions limited their utility. By assuming the input hydrograph to be an isoceles triangle, Overton (1966) determined k and x from an analytical solution of the output hydrograph shape. Clearly this method is not practical for the large, irregularly shaped Fraser Basin. He found that k was related to the shape of the inflow hydrograph and the lag time between the inflow and outflow hydrograph peaks, whereas x was a function of k and attenuation. Also, because of the sensitivity of hydrograph shape in small catchments, he found that k and x varied for each storm.

The Muskingum method can be applied to many problems, when the model deficiencies are understood, and more importantly, where the deficiencies are not critical (Strupczewski and Kundzewicz, 1981). Singh and McCann (1980) found agreement among writers that the effectiveness of the Muskingum method depends on the accuracy with which the parameters are estimated. Although they found significant changes in k and x, depending on the method used, they concluded that no method had any particular advantages. Different investigators use different initial conditions to determine k and x (Singh and McCann, 1980). This hinders comparison of the methods and the utility of the results.

Dooge et al (1982) advocate the estimation of Muskingum parameters from geometric and hydraulic characteristics of the reach in the absence of input and output data. They attempted to establish relationships for any cross-sectional shape and any friction law. Although small differences in the values of k and x resulted, they suggested the real values would lie between the extremes. Methods using hydraulic characteristics can include nonlinearity because k and x can be varied with discharge. Ponce and Yevjevich (1978) prefer to vary k and x in time and space, depending on flow variability. However, they quote Kousis (1978) who suggested calculations are relatively insensitive to x and therefore it can be constant. Chow (1964) considered a variable x to be a refinement that is seldom necessary.

The nature of k is agreed upon by many hydrologists (Strupczewski and Kundzewicz, 1980) but the nature and values of x are more inconsistent. In some cases, a negative x may be calculated. Theoretically, this is debatable because classical hydrology views x as a measure of wedge storage, but Strupczewski and Kundzewicz (1980) conclude a negative x is admissible in mathematical modeling. In general, x increases with increasing input rate and decreasing output rate.

The routing time t is critical. Theoretically, t must be small relative to other time elements and sufficiently short to validate the assumption of a linear relationship between hydrograph ordinates. The routing period is normally assigned any convenient value between the limits k/3 and k (Viessman et al, 1977). However, errors introduced when t=k are insignificant compared with the inaccuracies of the basic storage assumption (Nash, 1957). Long reaches are often simulated unsatisfactorily and Strupczewski and Kundzewicz (1980) prefer a cascade of models. This concept was used in the present study.

Because k and x depend on the method of estimation, it is difficult to determine the best values. Also, in a natural system, constant values are suspect. Clark (1945) was criticised in the discussion of his classic paper for using uniform velocity to establish isochrones. He examined various refinements, but his results were not improved. Clark suggested this was caused simply by two wrongs making a right; the second "wrong" being the use of a single k value. While his answer may not be mathematically acceptable, it is an adequate solution for this method. Therefore average values of k and x seem reasonable when long rivers are simulated.

1.2.4. U.B.C. Model of the Fraser Basin

Funding from the British Columbia Disaster Relief fund supported developmental work on watershed and flow models of the Fraser basin by Quick and Pipes (1972, 1975, 1976a, 1977). The computer model consists of two major parts: the first is a watershed model, run on continuously updated meteorological information to combine both rainfall and snowmelt runoff parameters, and the second part routes flow through the channel network, including lakes and reservoirs.

The model has been in operation for several years now, and is a reasonably good predictor of runoff volume and timing for the system. Based on the total water balance, the model coefficients are held constant through the season and from year to year. Quick and Pipes considered this important for producing reliable short and long term forecasts. Indeed, the principle uses of the model are forecasting and planning.

Quick (1965) noted that it is difficult to quantify runoff due to the complexity of snowmelt. Therefore, much study of the phenomenon was necessary at the outset. Area-elevation bands were delimited to allow a pertinent distribution of temperature and precipitation elements. The soil moisture deficit is then used to apportion runoff to three different flow routes which are distinguished by speed of transfer. In physical terms these routes represent direct runoff, interflow and groundwater.

Routing of the fast component is based on the conceptual model of Nash (1957) with an additional linear storage reservoir for the medium component. A linear storage reservoir is used to accumulate the slow input which is subsequently released as recession flow.

The watershed model was designed to use a sparse data input, nevertheless, estimating inputs from the precipitation station network in the Fraser Basin is still difficult. As well, the numerous influences upon the water balance require constant revision. The flow model uses mean daily river discharges and therefore the Muskingum method is unsuitable because the travel time between stations is less than one day (the routing period). Therefore, a nonlinear kinematic wave method was developed utilising routing coefficients calibrated from stage-discharge and stage-velocity curve data for each section. This method of calibration avoided problems created by the addition of large amounts of ungauged lateral inflow.

The model is flexible because it allows travel time and channel storage to be independent. However, real travel time and time difference between two peaks at two successive stations may not be the same. The present study examines time differences between peaks at successive gauging stations and requires rain-produced floods with discrete inputs rather than snowmelt. A major problem of the U.B.C. model is the use of mean daily streamflow data. Hourly data should be more reliable since travel times between stations are usually less than one day.

1.2.5 Travel Times

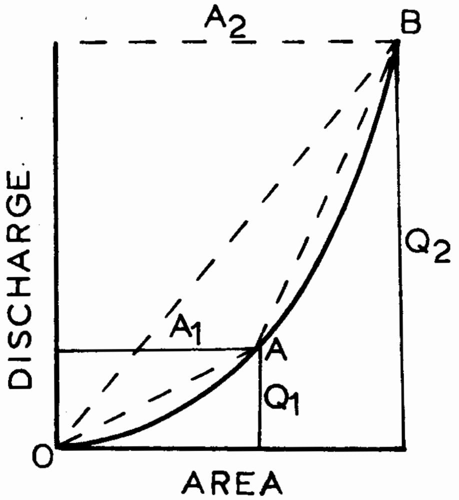
Velocity usually increases with stage, therefore area - discharge curves are usually concave upwards (Linsley et al, 1958) (figure 1.1). For a wave enclosed within the banks of a rectangular channel, it can be shown that the speed of the wave crest is greater than at its foot. Therefore, a hydrograph rising limb at a downstream point will be steeper than its counterpart upstream (N.E.R.C., 1975). The speed of flood wave movement is greatly influenced by slope, roughness, and hydraulic radius of the channel. Therefore, small inbank floods will travel faster than a flood that overtops the channel. Linsley et al (1949) suggest that minimum travel time occurs at moderate stages (figure 1.2).

Wave velocity or celerity may be calculated by Chezy or Manning formulae, but the resulting relationship with water velocity is contingent on the particular formula utilised, and channel shape. Therefore, neither supplies a finite solution. Alternatively, celerity (c) of a monoclinal rising flood wave can be expressed as:

$$C=\frac{dQ}{dA} \qquad (1)$$

where Q=discharge
A=area

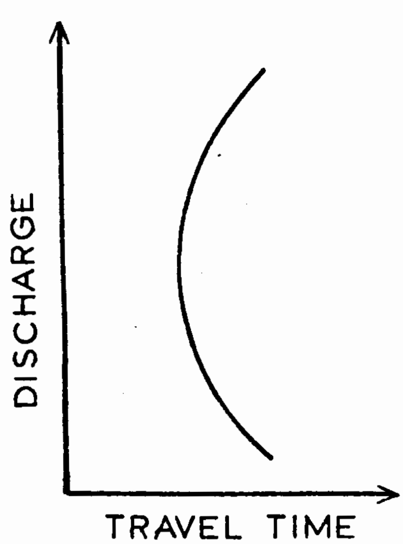
This is known as the Kleitz-Seddon law. It assumes that the wave has a permanent form and therefore there are many doubts as to its applicability. Quick and Pipes (1976b) however, found that



- slopes of secants OA and OB represent water velocities for sections 1 and 2
- slope of secant AB represents wave celerity

(after Linsley, Kohler and Paulus, 1958)

FIGURE 1.1 Typical area-discharge relationship for a stream and its influence on wave celerity



(after Linsley, Kohler and Paulus, 1949)

FIGURE 1.2 Stage - travel time relationship dQ/dA tends to overestimate wave speed.

Ideally, peak travel time is a direct function of head. In general use therefore, equation (2) can be applied.

$$C=v\pm\sqrt{gy} \qquad (2)$$

where C=wave celerity
v=water velocity
g=acceleration due to gravity
y=water head

Although this equation is good for a monoclinal wave, it is questionable when applied to a normal flood wave.

Observations suggest wave celerity is between 1.4 and 2.0 times greater than the water velocity (Linsley et al, 1949). Problems exist in the calculation of wave celerity in real rivers because of channel nonuniformity. Quick and Pipes (1976b) suggest gauging stations are often located above a contracted section and therefore reflect the velocities at these sections. Conversely, wave celerity is influenced more by the slower sections and thus, direct calculation of velocity is difficult.

During attenuation, the observed speed of the flood peak remains dependent on celerity, but it is also influenced by hydrograph shape (N.E.R.C., 1975). As peak curvature decreases, wave velocities may depart from kinematic expectations because outflow is no longer merely dependent on water depth. Now, neither discharge nor head remain constant. Subsequent calculations must consider the wave as having progressively decaying amplitude. Approximations usually yield adequate results, but crude simplification may hide important natural variations and care must be exercised.

The main methods of measurement of travel times are by timing coloured dye or radioactive tracers in the flow, and by interpreting water level records. Pilgrim (1976 and 1977) used isotope tracer methods to examine travel times and their relation to flood runoff and flood storage. Therefore, isochrone spacings are determined from the travel time of the water and not speed of flood wave movement. His studies revealed an initial positive increase of travel time to fairly constant values at medium to high discharges. Tracer analyses are usually restricted to small watersheds with limited tributary additions, and prior to this study, to low flows.

Stall and Hiestand (1969) estimated contaminant travel times in Illinois streams from hydraulic geometry relations with reasonable accuracy but generally found the computed travel time to be the minimum expected, especially at low flows. In Britain however, Brady and Johnson (1981) only found a good agreement between travel times measured by radioactive tracers with others derived theoretically, at relatively high flows.

Gergov (1971) proposed a method of determining travel time from the Chezy formula, utilising hydraulic stream expressions and requiring relatively uniform time - space precipitation distribution. Therefore the method is usually better for small homogeneous basins. However, he suggested that the influence of storm distribution and area of flood formation may have considerable effects on travel time values for a stream with major tributaries; this notion is central to the present study.

1.3 The Problem Defined.

The difference between time-of-peak at two consecutive stations can be attributed directly to flood wave velocity, and/or to basin and river network controls on generating maximum runoff. In the former case, a regular relationship to discharge should be expected on which to base prediction. In the latter case, a knowledge of basin geomorpholgy and of meteorological features are required to understand travel time irregularities. While engineering analyses of river systems revolve around successful prediction of flow, science seeks explanation of flow phenomena. By explaining and understanding the processes involved, more reliable predictions should be provided.

As basin heteorogeneity increases and the likelihood for complete storm coverage decreases, the strength of the relationship between precipitation and runoff diminishes. An additional problem in the Fraser Basin is that precipitation variability in both spatial and temporal frames at the higher levels is inadequately monitored by the sparse distribution of weather stations. Therefore simple input - output analyses are restricted. In contrast to the hydrologic approach, mathematical flood routing methods require input hydrographs to produce downstream outflow hydrographs. Such models can only be as good as the input data and the validity of the assumptions inherent in the methods.

Within a basin of the planimetric and vertical dimensions of the Fraser, many diverse hydrologic conditions exist. The basin heteorogeneity allows highflows to be generated by different causes in spatially distinct areas. Therefore the effects of tributary inputs, and the consequences of transfers through contrasting hydrometeorological regions, on a mainstream flood wave may be studied for a range of discharges. It should be noted that much precipitation falls as snow in the Fraser Basin and thus a high percentage of runoff is produced by spring melt. To model this process, detailed information of snowpack conditions and energy inputs over the basin are required. While snowmelt produces a general rise in river level, rainfall is typically associated with a flood wave. Therefore, it should be emphasised that only peaks caused by rainfall will be used in an analysis of travel times.

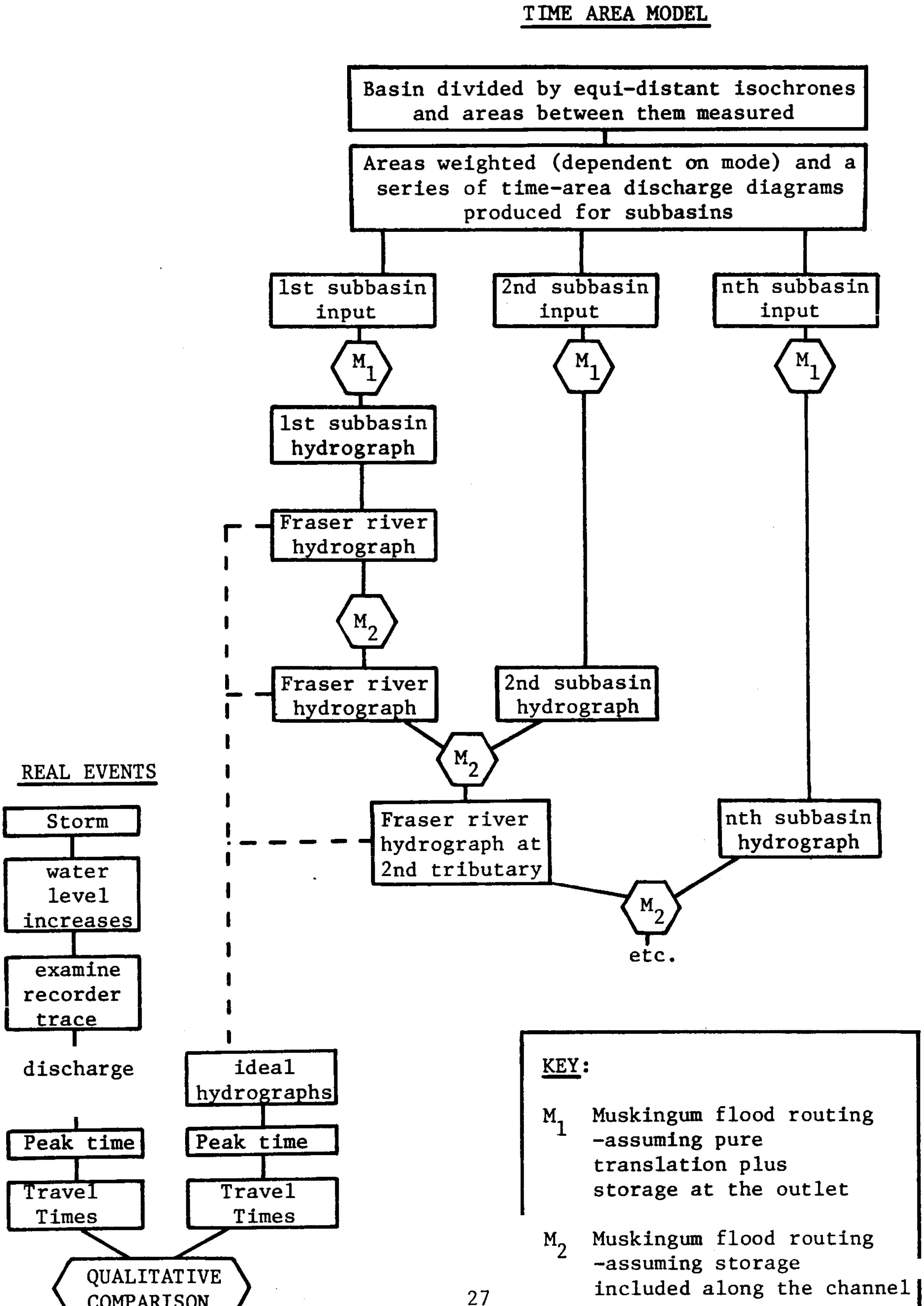
For a small, simple channel, a relationship between travel time and discharge might be expected, as described by Linsley et al (1958). However, given the different generating processes operating at any particular time in the large, complex Fraser Basin, it is reasonable to suggest that for a specific event, flow in the main channel will not behave in a simple mathematically defined manner. Moreover, travel time will be influenced by the contrasting hydrometeorological features of its tributaries. Thus it is suggested that the most significant variable in peak timing is a function of relative overland distances between sources and main channels. That is, the time of concentration for each spatial distribution of precipitation.

A model based on time-area considerations is developed in this study to simulate a series of contrasting inputs, in order to increase our understanding of the relative importance of the main tributaries and their effects upon mainstream hydrographs (figure 1.3). Data for real events were obtained from continuous recording gauge stations operated by the Water Survey of Canada on the Fraser River. This involved a detailed analysis of the recorder charts to obtain times and discharges of river flow on an hourly basis, together with an interpretation of meteorological conditions. Real hydrographs, in certain cases, exhibit characteristics that can be interpreted by the model. Therefore a series of hydrographs reflect the important tributaries for an event, and can be used to infer the region of precipitation source. This in turn can be verified by meteorological data, allowing a reclassification of events by source area of precipitation.

The main objective of this study is to model Fraser River hydrographs using the time-area concepts of Clark (1945) and Muskingum flood routing. The model hydrographs are then used as an aid to the interpretation of peak flow travel time irregularities between gauging stations. Within this central theme, there are several secondary objectives.

1. To examine the ways in which tributary inputs affect main channel flow volumes using simple time-area concepts.
2. To compare Fraser River hydrographs with hydrographs from

FIGURE 1.3 Flow Chart of the Problem.



the time-area model and identify characteristics attributable to localised storm inputs.

3. To assess the degree to which peak travel time is related to discharge in the manner suggested by Linsley et al (1949).
4. To determine if time differences between peaks at successive mainstream stations result predominately from flood wave translation or the integrated effect of geomorphic controls on runoff.

These objectives will be pursued in accordance with the following format.

1.4 Organisation of the Study.

This study is arranged in five chapters. Chapter 2 describes precipitation and runoff, based on a review of the pertinent hydrologic characteristics of the Fraser Basin. The third chapter introduces the basic time-area model of the Fraser Basin, its variations and implications for tributary contributions. In the light of this, Chapter 4 examines real peak events on the Fraser River. Conclusions drawn from the above are presented in the final chapter together with a discussion of the important controls on time-of-peak and the use of the time-area model for interpreting flood peak "travel time". The work is concluded with a discussion of several potentially useful research directions suggested by the present results.

II. Physical Environment of the Fraser Basin

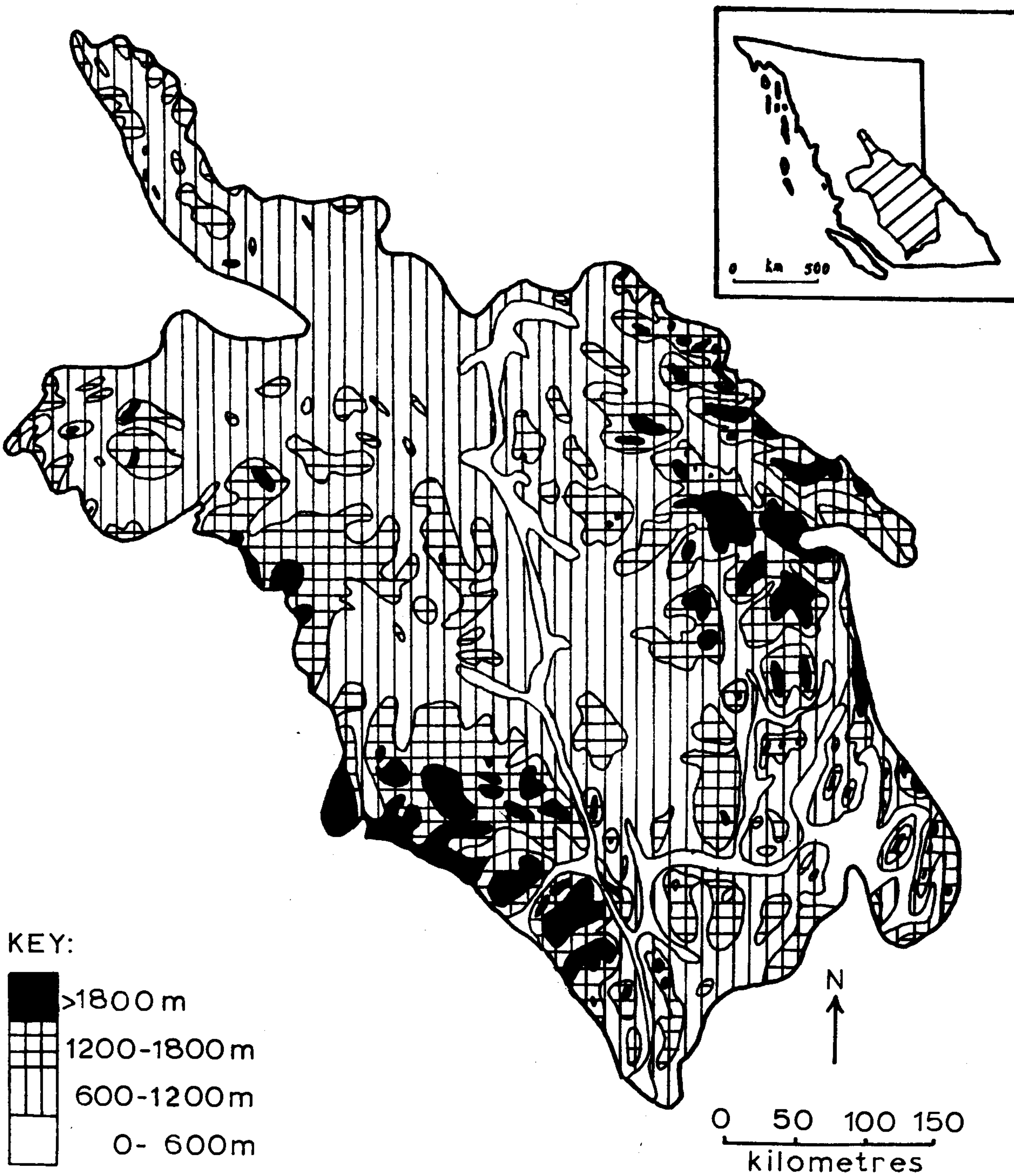
2.1 Physical Characteristics of the Fraser Basin above Hope

The Fraser Basin above Hope drains approximately one quarter of south - central British Columbia. This 218,000km^2 area extends from the United States border at 49° north, to 56° north, and it is bounded to the east by the Rocky Mountain divide and to the west by the Coast Mountains. The elevation range is quite striking, for example, ninety percent of the basin is higher than Prince George (665 metres above sea level (a.s.l)), and two percent lies at elevations higher than 2400 metres a.s.l. (Bruce, 1964) (figure 2.1). From Red Pass to Hope, the Fraser river is approximately 1200km long, and flows through regions with diverse relief, climate and vegetation.

2.1.1. Physiography

Physiographic regions are identified by similarities of relief and lithology, and fairly uniform land cover; therby unifying the controls on hydrological processes. Physiographic differentiation can be attributed to the evolution of the landscape at several scales. Ryder (1981) recognised three distinct temporal and spatial scales responsible for the

FIGURE 2.1 Elevation of the Fraser Basin



geomorphology of the southern part of the Coast Mountains of British Columbia. The oldest and most extensive landforms were produced by tectonic activity and subaerial denudation during the Tertiary. Subsequent landforms developed more locally within this framework, reflecting differences in climate and relief. During the Pleistocene, the existing landforms were modified by the effects of continental glaciation. Landscape details have been modified locally since the last extensive ice-cover (approximately 10,000 years B.P.) and therefore the landscape today is a composite form produced at all scales.

Reduced glaciers still exist in the highest parts of the basin, but present trends of advance and retreat are contradictory (Slaymaker and Macpherson, 1977). Most glaciers are now retreating, but a few show signs of advance, especially in the Coast Mountains. Generally, ablation is occurring more rapidly in the Rocky Mountains than the Coast Mountains (Slaymaker, 1972a).

The physiographic regions of British Columbia were definitively described by Holland (1964). Eight of his 18 regions are represented in the study area (figure 2.2) and the pertinent features of each region are summarised in table 2.1. Physiographic regions are clearly apparent in regionalisation of hydrometeorological features as will be shown later.

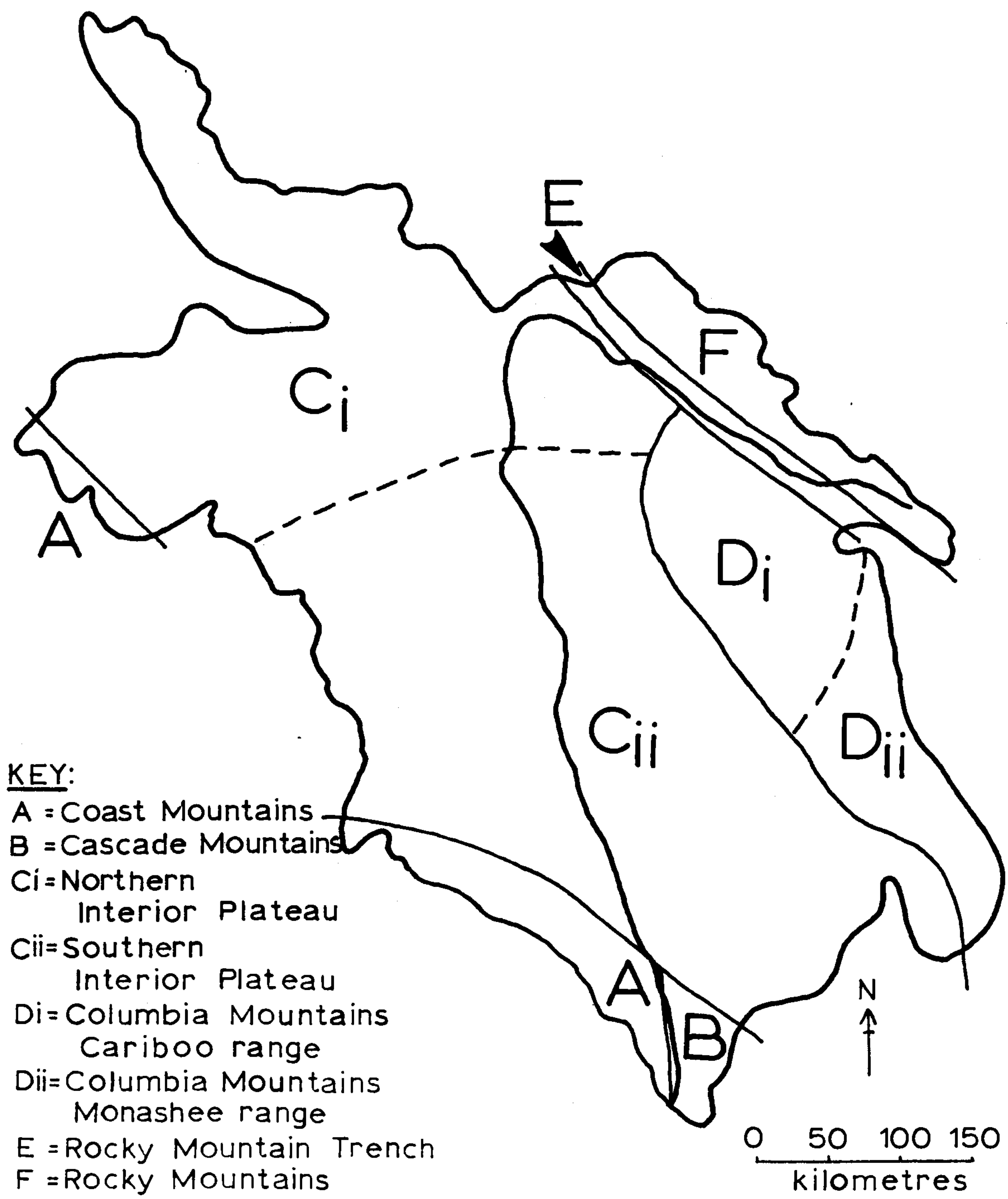


FIGURE 2.2 Physiographic Regions of the Fraser Basin

TABLE 2.1 PHYSIOGRAPHIC REGIONS

Name	elevation range (km)	glaciation	drainage	other features	discharge intensity cms/sq.km	% basin above Hope
Coast Mountains	0.5–0.4	icefields, icecaps and glaciers	main rivers eroded deep channels, glacier-fed	Pleistocene erosion sharpened higher peaks and softened lower ones. Mainly granitoid rocks and extensive metamorphism. Pleistocene and Recent vulcanism.	0.060	7
Cascade Mountains	1.5–2.5	a few small glaciers in Skagit range	rivers cut deep valleys	Separated from Coast Mountains by fault. Strongly folded and metamorphosed sediments plus volcanics and granite intrusions.	0.045	2
Skeena + Omineca Mountains	1.5–2.1	cirque remnants, modified valleys erosion features	major rivers cut through mountains	Very complex and tightly folded sedimentaries.	0.020	2
Rocky Mountain Trench	0.6–0.9	once occupied by ice, and form modified	drains north via Fraser and south via Columbia	Structurally controlled erosion feature. Width varies between 3 and 5 km. Absent where McGregor Plateau abuts Rocky Mountains.	0.008	0.5
Interior Plateau	0.6–0.9	drumlins, eskers and other drift covered features	large rivers and lakes	Highlands are transitional between plateau and mountains. Most runoff via Fraser river.	0.004	75
Columbia Mountains	0.6–2.0	many features of alpine glaciation e.g. cirques etc.	high relative relief	Very rugged. Metamorphic, granites and sedimentary in Cariboo range; quartzite and limestone in Selkirks.	0.035	9
Rocky Mountains	1.0–3.5	icefields and alpine features	developed in response to high input	Massive limestones, quartzites + argillites characterise region. Flat and dipping beds modified by glacial oversteepening.	0.045	4.5

After Holland (1964) and Slaymaker (1972a)

2.1.2 Basin Dimensions and configuration

The absolute size and shape of the basin govern the distance runoff must travel to the mainstream, and resistance to flow controls velocity. These parameters are generally accepted as inflencing time-of-peak and steepness of a flood hydrograph. Several studies have attempted to define the importance of basin size and shape to runoff, for example Renard and Keppel, 1966; Pilgrim et al, 1982; and Alexander, 1972. Alexander also investigated the relationship between area and time of concentration for a basin, and his results correspond well with Linsley et al (1958) and Hoyt and Langbein (1955). The Fraser Basin consists of several contrasting regions, and because the relationship between basin area and runoff, should be examined in areas of similar climate and physiography, each sub-basin should be treated separately.

Pilgrim et al (1982)found a correlation between basin size and the geomorphological variables describing its physical form. However, basin shape may influence the strength of this relationship. Murphey and his coworkers (1977) found a combination of basin shape and size to be one of the best geomorphic predictors of hydrograph features in the southwest United States. Unfortunately, it is difficult to quantify this highly irregular and markedly variable parameter. The effects of basin shape upon the timing of a flood peak are only qualitatively understood. Bell and Vorst (1981) suggest that shape is of most use in "between basin" studies and in multi-variate analyses where it is necessary to express the combined effects of many paramters. This quality is used in time-area studies where shape and size are the most important descriptors of each subbasin.

The general geometric shape of small watersheds is between ovoid and pear shape (Gray, 1961) but in the Fraser Basin, subbasins are rarely of the classic form. The ideal pear-shaped basin is restricted to small or first order tributaries like the Nahatlatch. The deformation and crenulation of the basin perimeter reflects locations of small tributaries within the basin (figure 2.3). As basin size increases, shape departs from the ideal. The Chilcotin and Quesnel basins for example, are broadest at their headwaters. Elongated basins are mostly large, for example the Fraser River to Hansard, the North Thompson (excluding the Clearwater) and the Stuart basin, but small basins like Williams Lake river, are also represented in this class (table 2.2). However, the shape variations perceived are dependent upon the scale of examimation.

The overall shape of the Fraser Basin is complicated by the sharp bend in the Fraser River near Prince George. This causes the north-east region of the basin to be folded back upon itself, and river length (excluding meandering) then exceeds the length dimension of the basin. Therefore, runoff is influenced by basin shape, not only because the whole basin is effectively elongated, but also because irregularly shaped tributary basins produce non-synchronised inputs to the main stream.

FIGURE 2.3 Subbasin locations and shapes

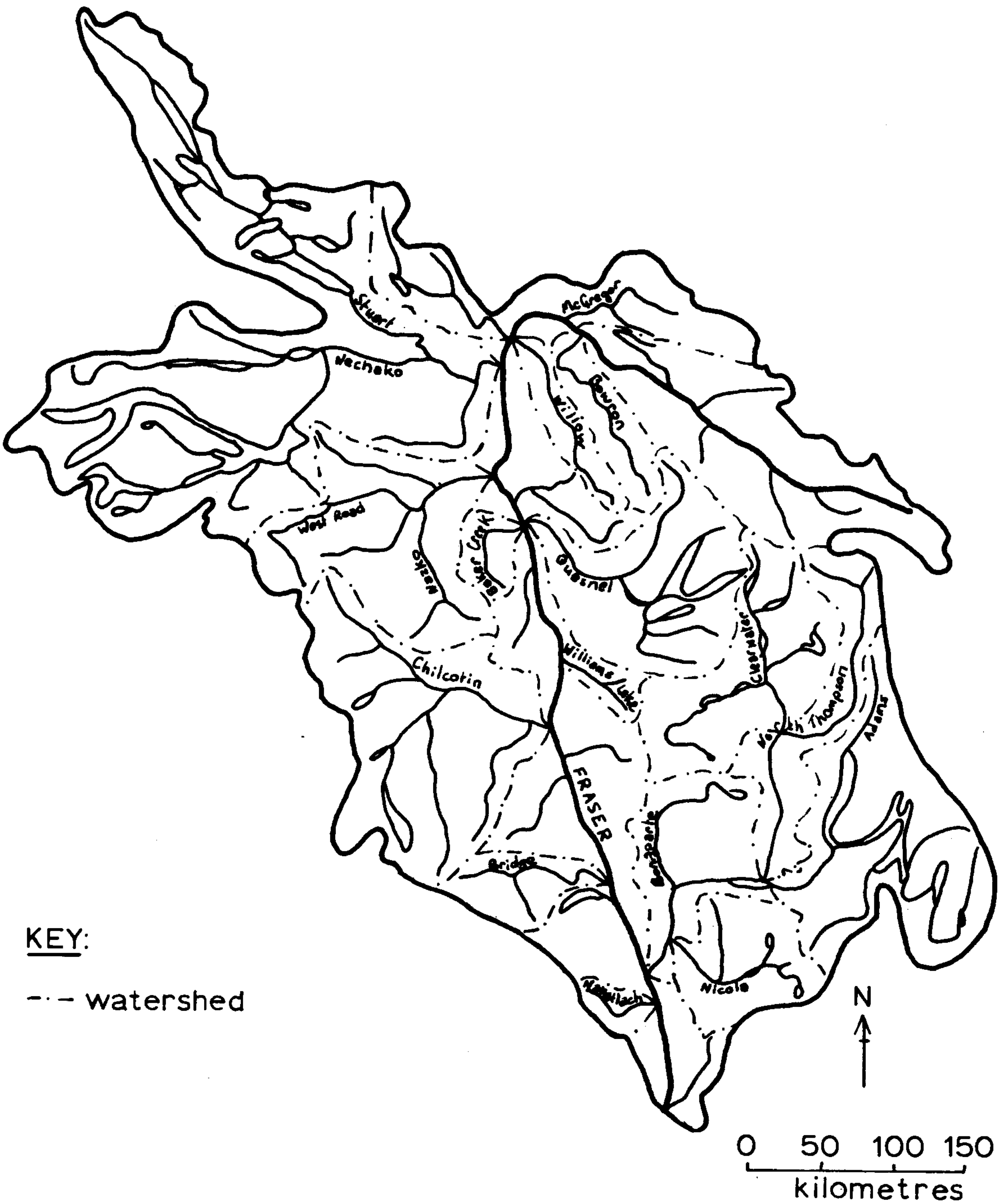


Table 2.2 Shape characteristics of selected subbasins

Shape	Name	Size sq.km	Tributary of	Region
Pear	Nahatlatch	1000	Fraser	S.I.P.
	Baker Cr.	1570	Fraser	N.I.P.
Offset-Pear	Bridge	4000	Fraser	Coast Mtns
Split Pear	Nechako	42800	Fraser	N.I.P.
Square	Clearwater	10200	N.Thompson	Columbia Mtns
Broad-headwaters	Chilcotin	19300	Fraser	N.I.P.
	Quesnel	11500	Fraser	N.I.P.
	Nazko	3350	West Road	N.I.P.
Elongated	Fraser to Hansard	18070	-	Columbia Mtns
	Adams	3100	S.Thompson	Columbia Mtns
	N.Thompson	10000	Thompson	Columbia Mtns
	Stuart	15000	Nechako	N.I.P.
	Williams-Lake	2300	Fraser	N.I.P.
	Willow	2870	Fraser	N.I.P.

note: N.I.P. is Northern Interior Plateau
S.I.P. is Southern Interior Plateau

In a similar manner, the density of tributary streams determines runoff speed. Channel spacing, or drainage density[1] develops in response to mean annual precipitation and influences the outflow of precipitation excess. Hence, drainage density is intrinsic in the time of concentration after a storm and in the

---

[1] term coined by Horton (1945) to describe the ratio of total length of streams within a basin to total basin area.

time taken to travel through a basin to the mainstream. A high drainage density allows faster removal of surface runoff and a corresponding peak discharge increase and lag-time decrease. Gregory and Walling (1968) suggested that the dynamic nature of drainage density, that is, its ability to expand and contract dependent on precipitation conditions, should be included in studies wherever possible.

The problems of determining drainage density are numerous and well-documented (McCoy, 1971; Gardiner, 1979; and Richards, 1979). Carlston and Langbein (1960) recognised that drainage density is the reciprocal of the mean orthogonal distance between channels. Therefore,

$$\text{Drainage Density} = 1.414 \frac{N}{L} \qquad (1)$$

where N is the number of intersections between a randomly orientated sampling net
L is the total length of sampling lines.

Mark (1974) examined the theory and empirical evidence, including samples from southern British Columbia, and concluded that equation (2) was more applicable.

$$\text{Drainage Density} = 1.571 \frac{N}{L} \qquad (2)$$

Drainage density within the Fraser Basin was estimated for this study using Mark's (1974) method. Samples were taken from a 1:1,000,000 scale map compatible with the time-area model, for each of the five hydrometeorological regions (table 2.3).

Table 2.3 Fraser Basin drainage densities

Region	Drainage density (km/km^2)
Southern Interior Plateau	0.15
Northern Interior Plateau	0.16
Mountains - Cariboo Range	0.18
Mountains - Monashee Range	0.18
Mountains - Coast	0.19

At this scale, the differences between regions are small, but nevertheless reflect a reasonable pattern. These figures may be interpreted in conjunction with table 2.1. The Coast Mountains have the highest drainage density, despite being sampled on the leeward side. They also have the highest discharge intensity. Drainage densities for the two eastern mountain samples are only slightly lower, as are their discharge intensities. Together, these mountain ranges cover 20.5% of the Fraser Basin. About 73% of the basin lies in the Interior Plateau which has a very low discharge intensity and numerous lakes. At the scale of 1:1,000,000 the drainage density measure does not reflect this contrast. This is possibly due in part to the inclusion of all intermittent streams, because it is assumed that the whole basin contributes to high flow conditions. However, the dynamic nature of the stream network and the incorporation of lakes into the drainage density index need to be examined further. These results represent the meso-scale of drainage densities (Gregory and Gardiner, 1975) because they reflect both mean annual precipitation and lithology/topography. Gardiner (1979) noted that biases introduced by scale and location are still probable.

The value of the drainage density index is widely acclaimed, but Black (1970) and Dingman (1978) indicate that it may not be sensitive enough to indicate runoff behaviour. Despite the problems, drainage density remains the most significant measurement of drainage basin morphology (Richards, 1979) and it is a useful integrative index reflecting physiography, soil development and vegetation.

Ruggedness measures the combination of drainage density and maximum relief and therefore expresses slopes within a basin (Melton, 1957). Black (1970) showed steeper slopes result in faster runoff, shorter time of concentration and higher maximum flows. In the Fraser Basin, differences between the drainage density indices for the Interior and the mountains would be increased by the ruggedness measure. Ruggedness is also interrelated with increased amounts of precipitation received in areas of higher relief.

2.2 Meteorological Conditions

Precipitation and temperature are both highly spatially variable at any time in the Fraser Basin. The hydrology of the basin is further influenced by the interaction of precipitation and temperature in winter and their effects upon snow distribution.

2.2.1 Precipitation

Most precipitation in the Fraser Basin is frontal in origin, bringing widespread precipitation especially in winter. Orographic effects tend to increase the intensity of frontal precipitation by reducing the speed of depression system movement. In summer, frontal activity is weak and less frequent, but Walker (1961) found a large proportion of summer precipitation still falls when fronts are present (table 2.4).

Table 2.4. Percentage of precipitation occurring with fronts present.

	Jan	Apr	Jul	Oct
Vancouver	99	83	92	94
Revelstoke	98	80	48	88
Prince George	100	75	41	95

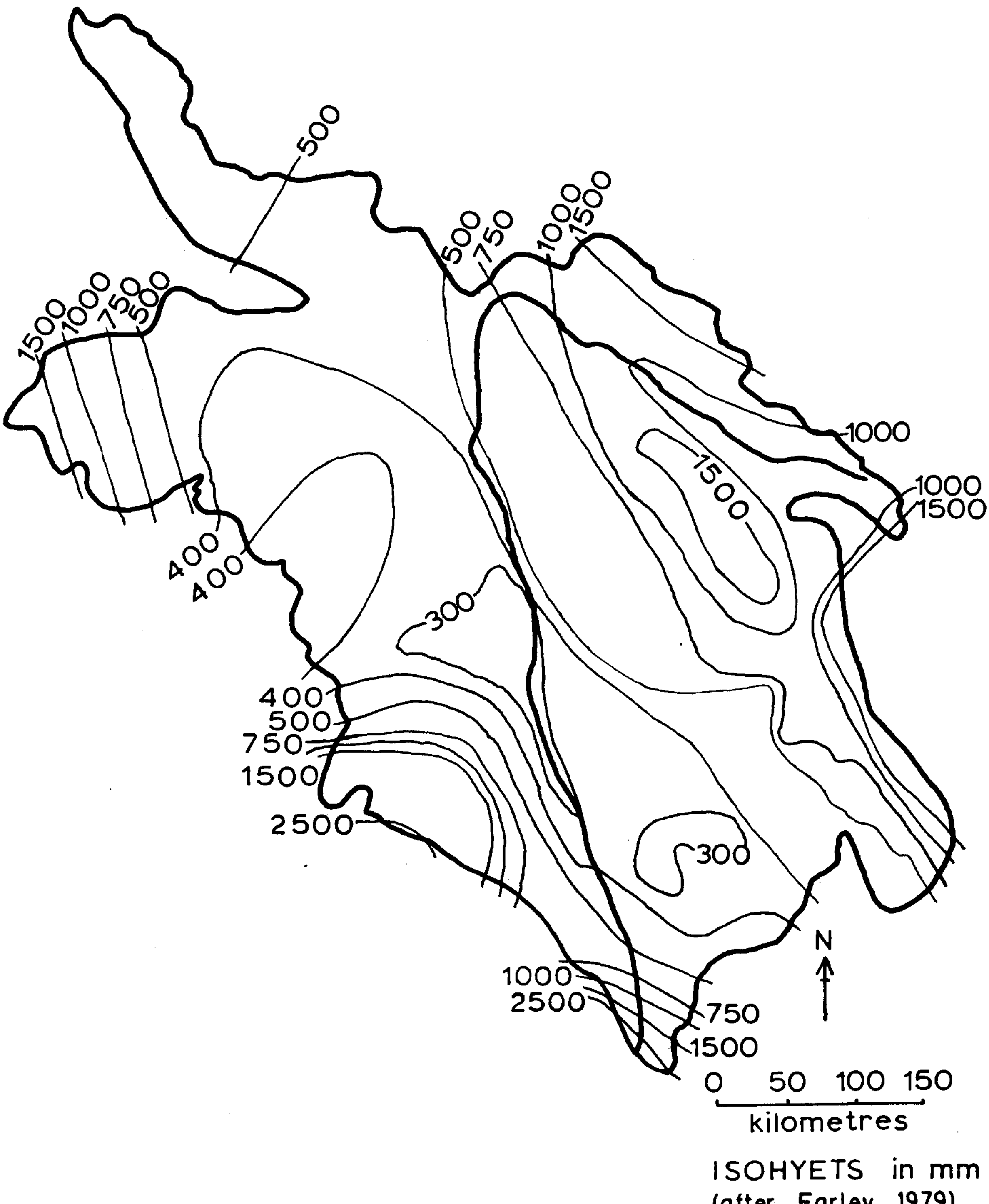
After Walker(1961)

Moist air predominately approaches the Fraser Basin from the southwest, and thus perpendicularly to the Coast Mountains resulting in high precipitation on the windward slopes and aridity to the lee side (figure 2.4). Moisture is reduced for a considerable distance downwind of the barrier in Washington and Oregon (Schermerhorn, 1969). This is also apparent in the Fraser Basin but the aridity of the Interior Plateau is interupted by the Columbia Mountains where precipitation is induced again.

The irregularity of a barrier is reflected in the lee side precipitation pattern. In the northern Interior Plateau, for example, generally higher precipitation is attributed to uplift of moist air that has passed over the lower section of the Coast Mountains between 53°N and 55°N (Ingledow, 1969). Also, gaps in the Coast Mountains, allow moist air to penetrate into the Nechako Valley (Wallis, 1963).

Precipitation usually increases with height (Barry and Chorley, 1974) but Riehl (1965) showed a steeper precipitation gradient on the leeward side of Pacific Northwest mountains. In the Fraser Basin, a "spillover" effect may occur and the Nechako, Lillooet and Bridge subbasins, for example, have steep precipitation gradients to maximum values at the Coast Mountain boundary. The Chilcotin basin is an exception because it does not drain the high precipitation areas west of the Coast Mountains. Otherwise, the lee effect is masked only where convection precipitation is important.

FIGURE 2.4 Average precipitation distribution in the Fraser Basin



Walker (1961) generalised the levels of maximum precipitation, but noted that terrain differences in British Columbia produced many local variations. The value of such generalisations is questioned by Glennie (1963) and March and coworkers (1979). Wilson, Valdes and Rodriguez-Iturbe (1979) examined the reliability of models subject to spatial variations of precipitation. They found the errors in a predicted hydrograph were greater for short, intense or localised storms than for those of frontal origin.

In winter, amounts of precipitation over the whole basin can be related to direction of upper air flow. This is not the case in summer when precipitation may be more localised. Orographic effects, for example, may trigger convective instability and convective thunderstorms which are associated with a variety of flow directions and amounts received (Walker, 1961). Also, the passage of a cold low in summer can induce local precipitation. Hage (reported in Walker, 1961) deduced that cold lows cause intense uplift and thus intense local precipitation within British Columbia. The resulting runoff from such events may be of local importance, but produce only a limited response in the mainstream Fraser. The more localised nature of storms helps explain precipitation variability in the Interior.

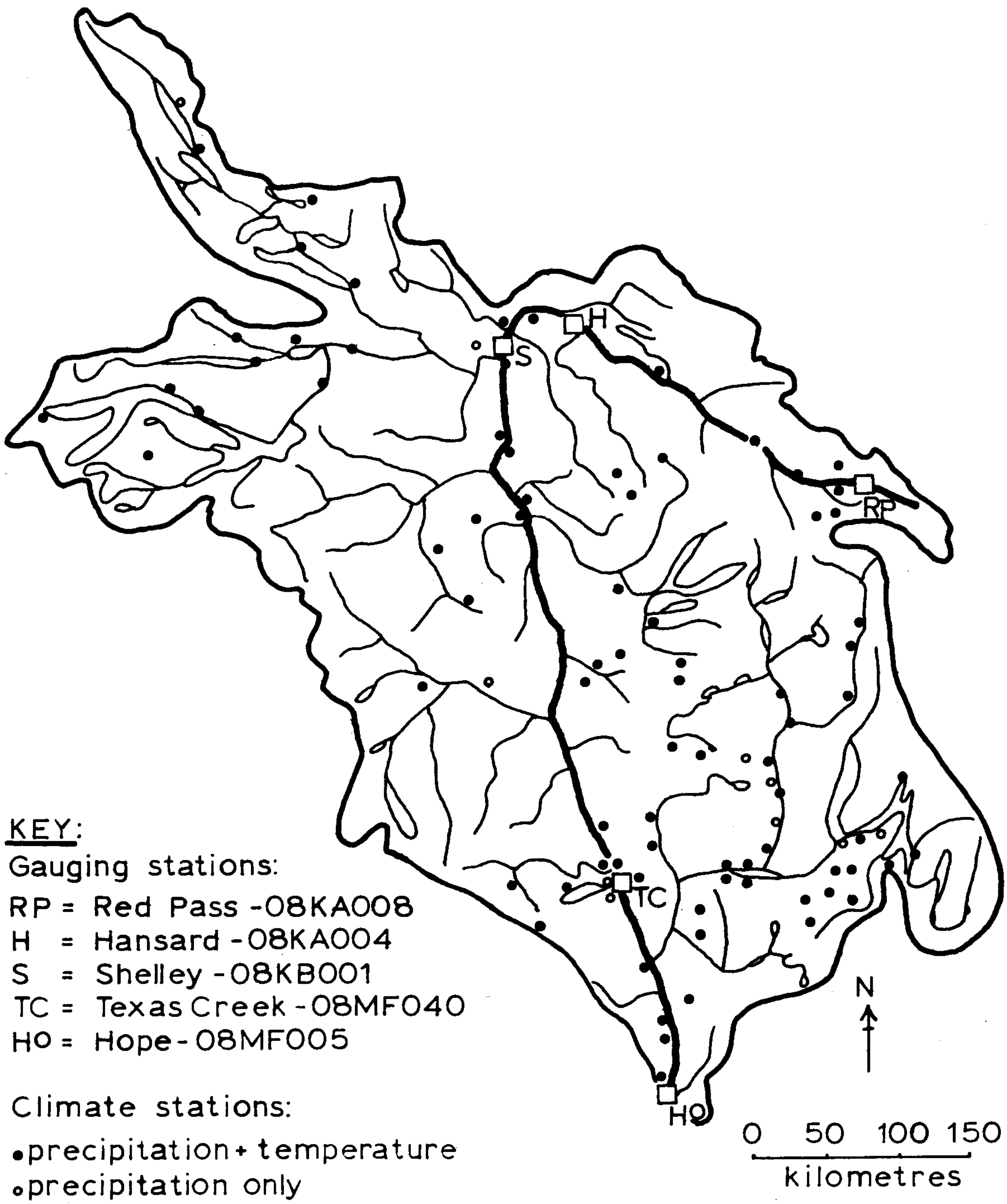
The annual pattern of precipitation distribution in the Fraser Basin is greatly influenced by the mountainous rim and topography within the basin. However, the use of mean annual precipitation figures is complicated when examining single events. This is because maximum precipitation is partly governed by the degree to which contrasting air masses penetrate the basin and therefore it does not occur everywhere at the same time (Wallis, 1963). The resulting precipitation distribution within the Fraser Basin is therefore highly variable. Unfortunately, climate stations located predominately in river valleys and the more accessible areas, provide an insufficient network for a thorough evaluation (figure 2.4).

2.2.2 Temperature

Temperatures in the Fraser Basin are primarily modified by the distribution of mountain chains, and the effects of continentality. Although, radiation receipt is highly variable in rugged topography, temperatures generally vary with altitude within the Fraser Basin. The direction of air flow influences seasonal values of daily maximum and minimum temperatures; an effect that is especially noticeable in winter when north and northeast winds lower the temperature. Easterly continental winds in summer and southerly maritime winds in the winter can increase temperatures. These are very general trends, however and micro-climatic effects cause much variability.

In the absence of climate station data, average lapse rates, typically about 6.5°C/1000 metres are used to interpolate temperatures between stations. The presence of an inversion complicates their use however, because large temperature changes

FIGURE 2.5 Locations of climate and gauging stations in the Fraser Basin



can occur when the height of an inversion layer fluctuates (Walker, 1961). This is relevant for evaluating potential snowmelt.

2.2.3 Snowfall and its influences on Fraser Basin runoff

During the winter months, snow covers the higher elevations of the mountainous Fraser Basin flanks and remains as snowpack until temperatures rise sufficiently to allow melt and subsequent runoff. The mountains of the Cariboo and Coast Mountains receive most of their snowfall in spring, but the remainder of the basin has a winter snow maximum (Wallis, 1963).

Snow accumulation tends to correlate with elevation but snow depth at any point depends on the amount of drifting. The lower limit of snowcover, or the snowline, is dynamic and constantly changing through winter. Snowline fluctuations affect the amount and timing of delayed runoff and therefore storage time is highly variable and difficult to evaluate.

As winter progresses, the snowpack compacts under its own weight and looses some water directly to the atmosphere by evaporation or sublimation. The nature of the snow varies and therefore potential runoff is best measured by its water equivalent. Loijens (1972) found elevation and slope acccounted for most variance of annual maximum snow-water equivalent in Banff National Park.

Heavy basin-wide snowmelt in the Fraser Basin is often triggered by the incursion of warm upper air flows from the southwest. Melt begins first in the south of the basin and then in successively more northern areas as arctic air is displaced by warmer maritme air. To the lee of the Cariboo mountains, significant localised melting can be induced by transverse upper winds producing a Chinook effect (Pollack and Bellows, 1972). On a diurnal basis, the amount of melting depends on the net heat exchange between the snowpack and its environment. Snowmelt during rainfall is a dominant hydrolcgic process in western Oregon (Harr, 1981) but further north within the Fraser Basin it is usually limited to the warmer southwest (Wallis, 1963) and the lower elevations (Woo and Slaymaker, 1975).

The timing and magnitude of the freshet is of great concern to the residents of the Fraser Basin, especially in the populous area below Hope. Glennie (1963) reported an old wives' tale that a high flood on the Fraser is preceeded by a fine, cold Indian Summer the previous year. The temperature and water content of the soil in the previous October can influence the available moisture in the spring, and thus freshet size. On the other hand, Bruce (1964) suggested the sequence of temperatures from April to June was the critical factor in the generation of a flood. The majority of snowmelt runoff originates in the Interior Plateau where snowfall is moderate and winter temperatures often fall to -35° C (Quick, 1965). In the north-east of the basin, as much as 80 percent of annual precipitation falls as snow producing the bulk of runoff in spring (Engineering Division, 1972).

Waylen and Woo (1983) examined highflows produced by snowmelt in three diverse subbasins of the Fraser. In a basin representing the southern Interior Plateau and in a subalpine basin of the Monashee mountains, the mean date of peak flows occurs near the end of May. In the basin representative of the Columbia mountains the date was almost a month later.

Snowcover is so heteorogenous that only estimates can be made of its extent, and the amount of melt at any time. Relationships between snow cover area and snowmelt runoff differ for each type of geomorphological catchment (Gupta et al, 1982). Several methods for estimating melt exist however. For example, temperature-index methods are considered to be the best in large, forested basins (Gray, 1970), together with 'phase-routing' which effectively slows the flow by varying storage constants. The U.B.C. Fraser Basin model (section 1.2.4) iccrporates these ideas.

In summary, snowmelt varies both temporally and spatially in the Fraser Basin. The contribution of meltwater to runoff is therefore very difficult to predict and impossible to evaluate precisely from meteorological records.

The above review of meteorological conditions in the Fraser Basin has highlighted regional differences and thus each tributary basin of the Fraser has contrasting hydrometeorological characteristics. These are presented in table 2.5.

Table 2.5 Climatic characteristics of the Fraser Basin

Name	Climatic group	Precipitation max	Precipitation min	%snow	max snow
Upper Fraser	alpine humid	even		90	winter-spring
Stuart	alpine arctic	summer	spring	40	autumn
Nechako	humid conti-nental	winter	spring		autumn
W.Road		summer	spring	N/A	N/A
Quesnel	dry and humid conti-nental	even	spring	30-50	spring
Chilcotin	dry conti-nental	summer	spring	33	autumn
Bridge	alpine maritime	winter	summer	71	autumn
North Thompson	alpine humid & dry con-tinental	winter summer	spring	40	autumn
South Thompson	as North	winter summer	spring	40	autumn
Thompson	alpine maritime & dry conti-nental	winter summer	spring	80-90	autumn

note: spring=March, April, May
Autumn=Sep, Oct, Nov

adapted from Wallis (1963)

2.3 Hydrometeorological Divisions and Highflow Regionalisation

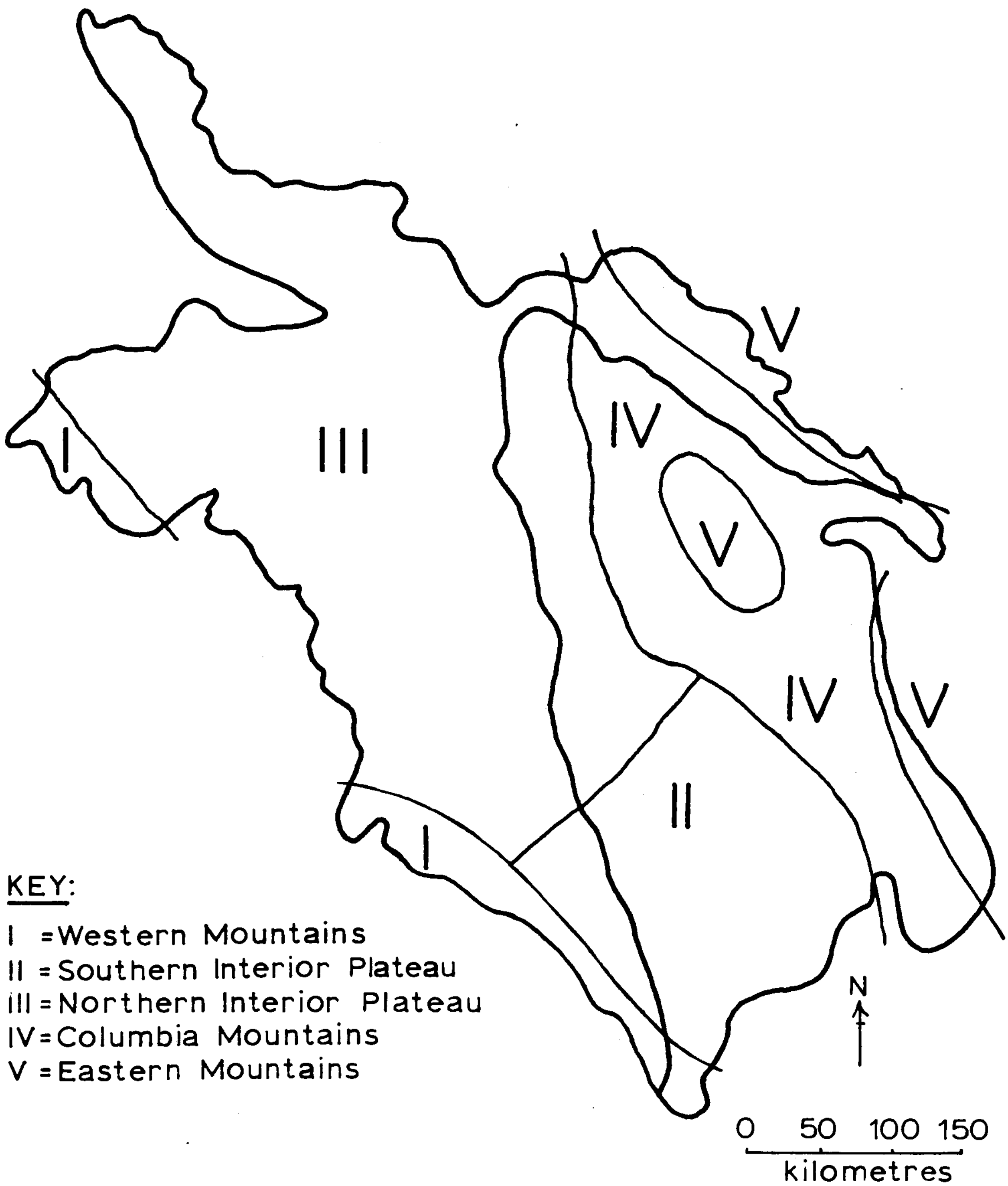
Many facets of the basin affect the volume and speed of runoff generation; some change daily, others remain more constant. Maps of mean precipitation, mean temperatures, and physiographic divisions often reveal common divisions which Waylen (1981) grouped into hydrometeorological regions (figure 2.6). Physiographic boundaries defined the three mountain zones, while vegetation indicators divided the Interior Plateau. Slaymaker (1972a) noted the control of physiography upon climate, vegetation and hydrology. Annual discharge hydrographs of Fraser River tributaries exhibit corresponding differences in the importance of seasonal peaks (table 2.6).

Based on observations that gauging stations in similar areas show corresponding annual discharge patterns, flow data can be regionalised to aid data transfer to ungauged areas. Some success in regionalisation has been achieved despite differences in basin characteristics and chance variations in

Table 2.6 Seasonal Runoff Peaks

Physiographic Region	Spring snowmelt	Glacial ablation	Autumn rain
Coast Mtns	Yes	Yes	Yes
Cascade Mtns W	Yes	-	Yes
Cascade Mtns E	Yes	-	-
Skeena Mtns	Yes	-	Yes
Omineca Mtns	-	-	Yes
Interior Plateau	Yes	-	-
Columbia Mtns	Yes	Yes	minor
Rocky Mtns	Yes	Yes	minor

FIGURE 2.6 Hydrometeorological Regions of the Fraser Basin



sampling. In British Columbia, Leith (1976) regressed ten variables with mean annual flood. He concluded the method was reasonably effective in large basins, but he was unable to identify important physiographic parameters. Leith used a computer data bank containing physiographic and hydrological data for 10km^2 grids. Data collection is described by Kreuder (1979), but this laborious task is not yet complete. The grid square method was also used in the South Thompson Basin by Obedkoff (1970) tc improve seasonal mean flow estimates by including evaporation and snow course data.

Waylen (1981) modeled the frequency characteristics of Fraser Basin highflows. He compared results from annual and partial duration series by using probability distributions to estimate parameters for a regrouping procedure. Parameter values were then regressed on basin and climatic data for each group. Waylen achieved partial success which he attributed to the inccrporation of both spatial and temporal measures.

Strong regional differences in the range of highflows within the Fraser Basin have been identified by the studies described above. The physiographic and climatic controls on these variations will also influence daily runoff patterns. The sub-basins will produce hydrographs reflecting contrasts in the quantity and timing of storm discharge. This, in turn, is important for Fraser river flood wave translation because inputs may not be synchronised. An understanding of regional differences therefore helps explain peak flow characteristics.

III. The Fraser River Time-Area Model

3.1 Introduction

The model developed in this study was designed to generate hydrograph time-of-peak data at four stations on the Fraser River for varied input conditions. The model is based on a series of time-area histograms which provide a series of lagged inflows to a flood routing equation. The result is an estimation of a complete hydrograph which can be compared to real hydrographs. The basic model was produced from a simple time-area map of the Fraser Basin. Subsequent variations were made to represent mean annual precipitation, snowmelt and three regional storms.

3.2 Mode I: The Basic Model

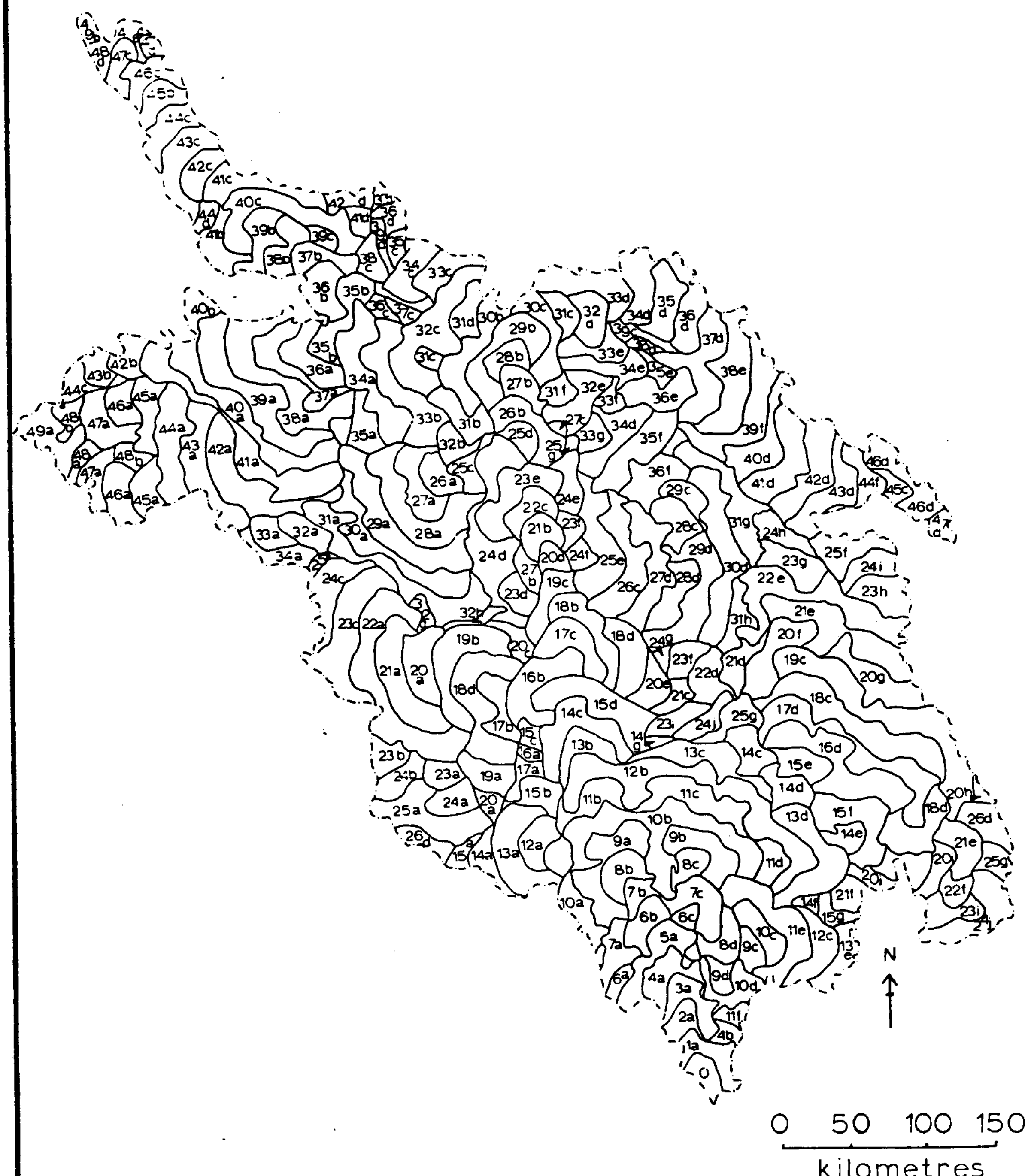
3.2.1. Description

To construct the time-area map at a scale of 1:1,000,000, dividers were used to place isochrones equal distances apart along the rivers. The isochrones delimit areas within which rainfall has sensibly equal travel distances to become runoff. A total of 49 areas were delimited, many being segmented into a number of Fraser tributary basins because of contrasting flow directions (figure 3.1). The isochrones were constructed on the basis of equivalent horizontal distance, and therefore water travel time, or velocity through each section can be varied conveniently. On the basis of Water Survey of Canada meter notes, the mean velocity of the Fraser River at Hope was found to be approximately twice that recorded at Upper Fraser river stations. Although velocity differences may reflect channel conditions adjacent to the gauge, it was considered necessary to increase velocities incrementally below Shelley.

The construction of isochrones allowed a detailed visual examination of the Fraser Basin stream network. Often, subbasin shape dominates the pattern of isochrones. In the Interior Plateau, lower drainage densities allow a simpler arrangement of isochrones, as in the West Road, Chilcotin and Bonaparte basins. In a few mountainous basins such as the McGregor and North Thompson, isochrones are arranged simply, but this contrasts with the majority of subbasins where topography and stream network complicate the patterns, for example, the Fraser Basin above Hansard and the western Nechako basin.

Isochrone positions are controlled more by drainage channel pattern than drainage density. The extent to which small streams affect isochrones depends on map scale and the chosen, horizontal spacing along the channels. The complexity of a time-area diagram increases when neighbouring streams flow in opposing directions, for example the headwater areas of Nechako,

FIGURE 3.1 TIME-AREA MAP OF THE FRASER BASIN



South Thompson and Stuart basins and also the Fraser Basin above Hansard. A pure, dendritic stream pattern will not displace isochrones significantly from an arc, but drainage density itself may affect the subsequent calculations. Unfortunately, network configuration and drainage density are too interrelated to allow examination of the quantitative impacts of one alone.

Basin drainage density is inherent in time-area modeling because it expresses the relative amount of time that runoff spends in channels rather than in other, much slower routes. Theoretically, isochrone spacings should be wider where drainage density is high because there, the potential runoff speed is faster (all other things being equal). Therefore, a similar drainage density throughout a basin is preferable for time-area modeling, as discussed in section 2.1.2.

At the scale of the time-area model described here, the inter-regional differences in drainage density are not especially high. Therefore, before giving drainage density more attention in a time-area model, quantitative studies are required to determine if it is proportionally more important than other topographical variables.

After construction of a time-area map, the areas between adjacent isochrones are measured by planimeter. These areas were used to construct time-area histograms at four points on the Fraser River, corresponding to Water Survey of Canada continuous recording gauge locations (figure 3.2). At this stage, basin storage is not included and the diagrams only represent travel

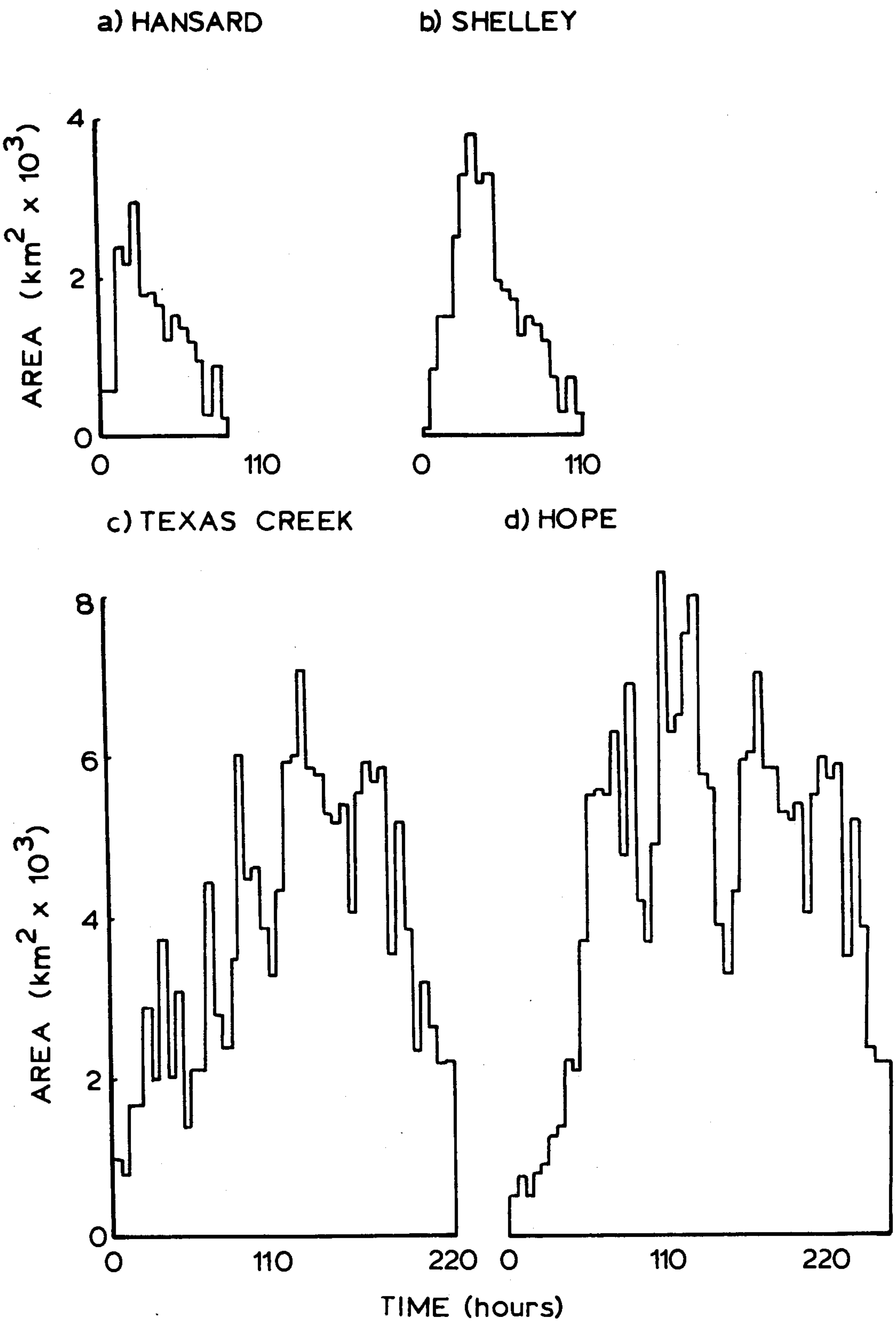


FIGURE 3.2 Simple time-area histograms for the Fraser Basin to the gauging stations

times / distances.

Hydrographs are produced by applying a mean value for basin precipitation[1] to the whole watershed. The lagged area - precipitation values are then input to the Muskingum flood routing procedure. Instantaneous unit hydrographs for the same four locations are shown in figure 3.3. While the resultant hydrographs reflect the overall basin conditions, they do not allow an interpretation of the relative importance of the contributing subbasins. Therefore a time-area histogram of each tributary was routed to a hypothetical linear reservoir situated at its Fraser River confluence (figures 3.4). The resultant hydrograph becomes inflow to the Fraser River and is then routed to the next major confluence or gauging station using Muskingum coefficients. This allows the effects of natural attenuation to be incorporated (figure 3.5). Each tributary hydrograph in turn, is added to the Fraser River flow at the corresponding time. See figure 1.3.

Determination and sensitivity of Muskingum constants.

As described in section 1.2.3, several methods of determining x and k exist, but all need either input and output hydrographs or field studies to determine hydraulic parameters. Each method results in slightly different values and is often based on different initial conditions. The problem remains, which is the most representative?

---

[1]assumed to be instantaneous (duration = 0 hours).

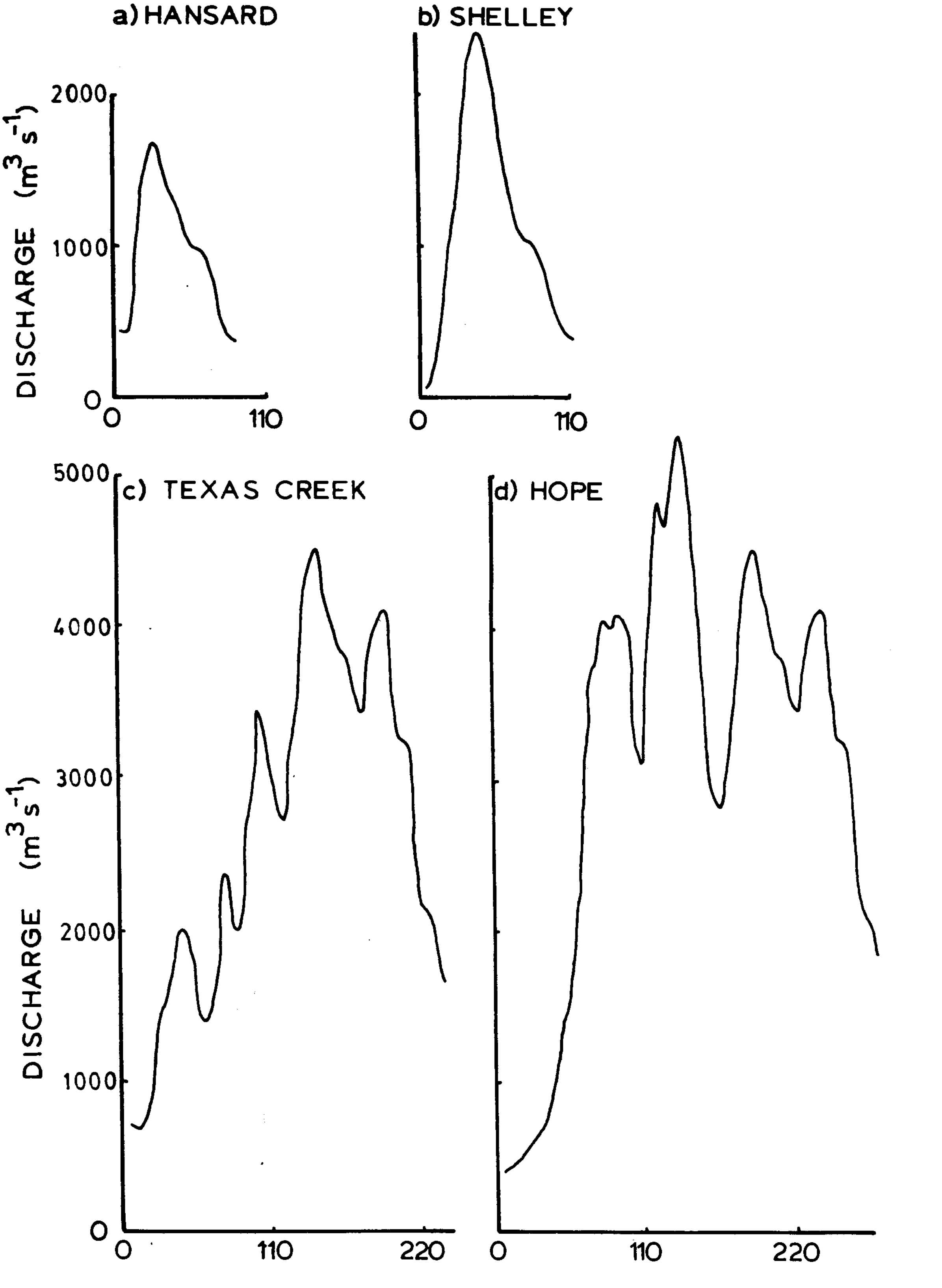


FIGURE 3.3 Instantaneous unit hydrographs for the Fraser Basin to the gauging stations

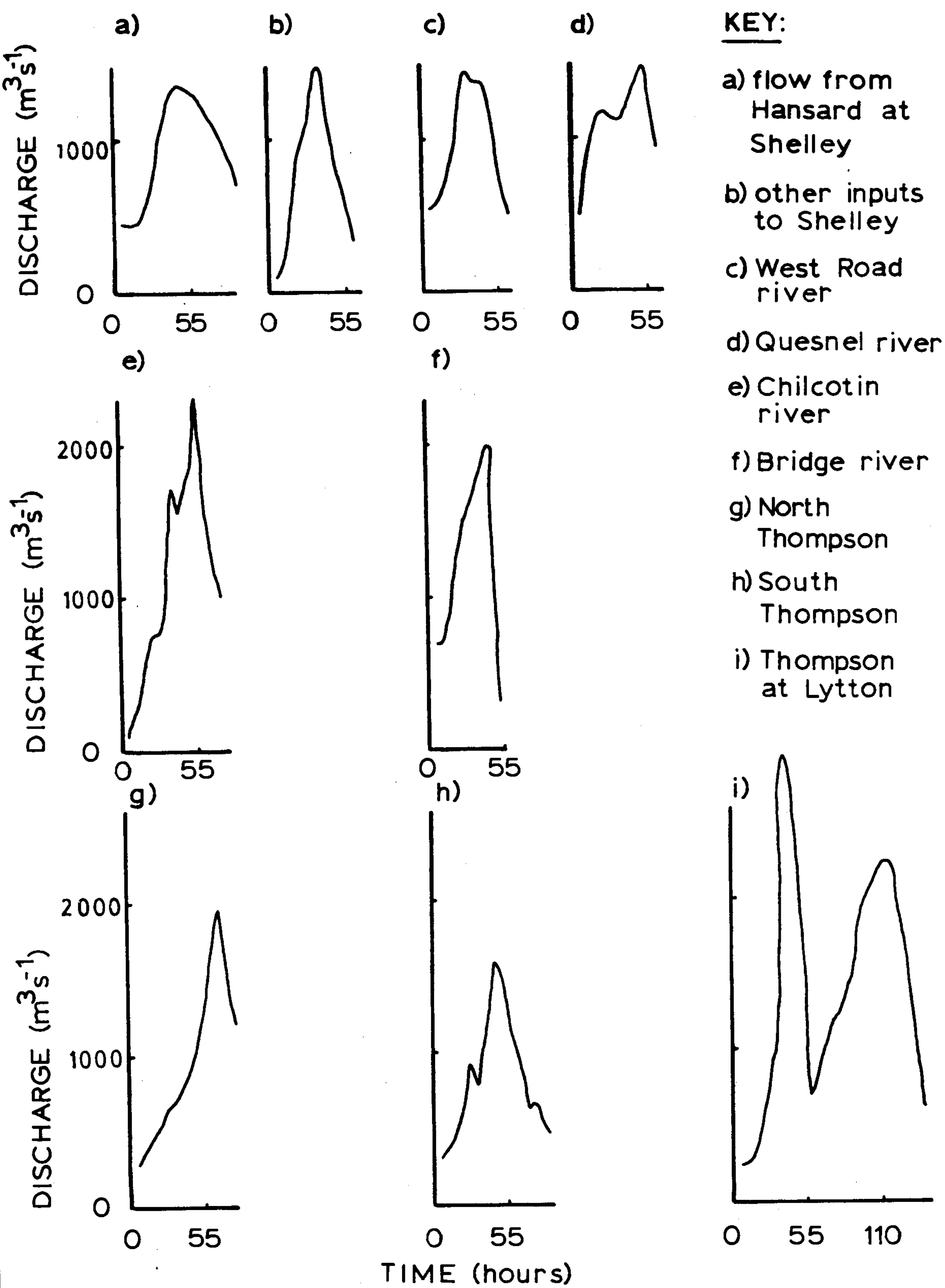


FIGURE 3.4 Hydrographs representing each subbasin

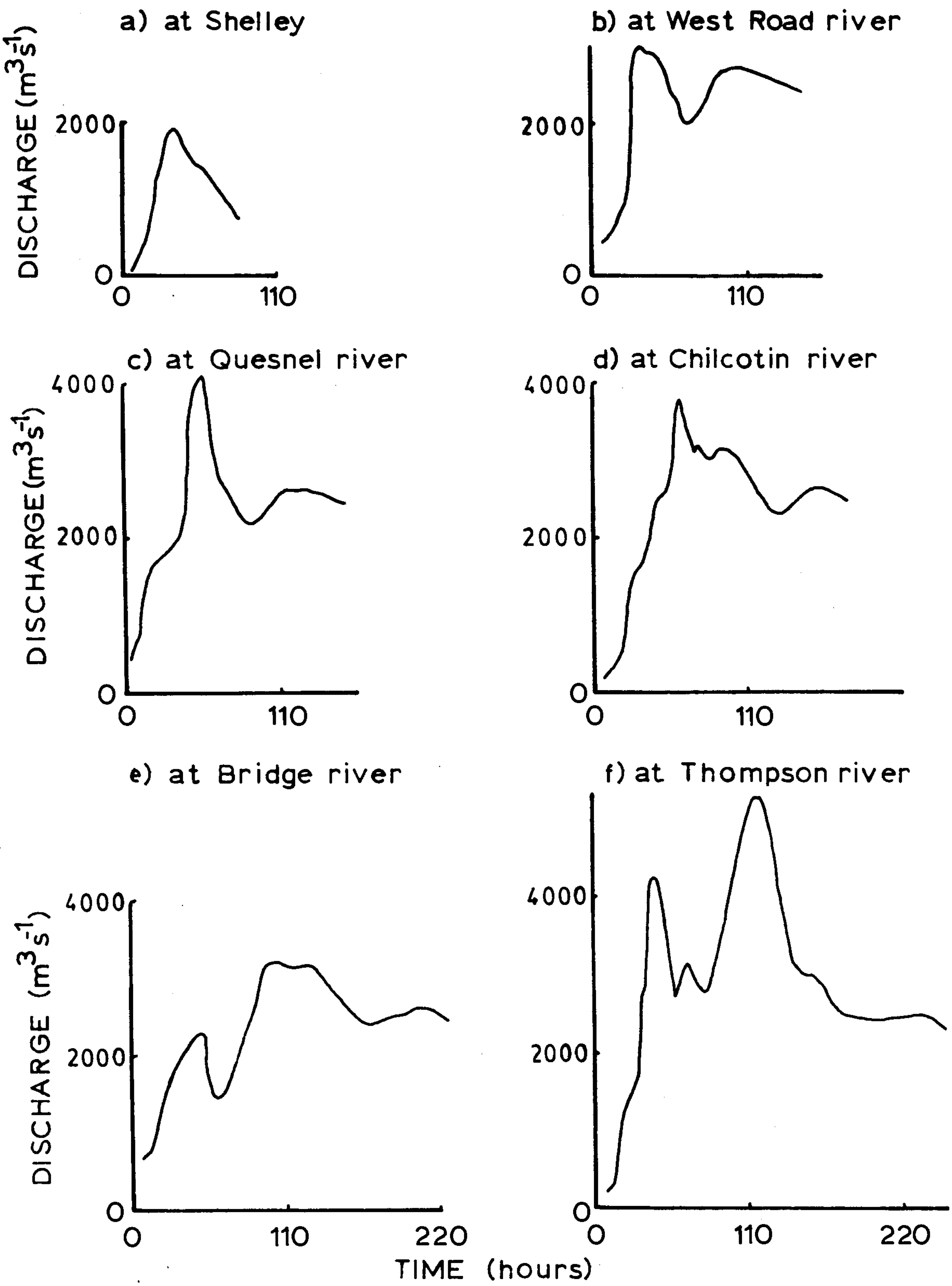


FIGURE 3.5 Fraser River hydrographs at successive confluences

The Fraser River data are not precise enough to estimate k and x from output, and adequate information regarding hydraulic characteristics are not available. Therefore, the empirical average of 0.2 was used for x and travel times based on average velocities used for k. The choice of mean values may counteract and balance natural variations along the Fraser River. Without extensive field work, the utility of "accurate" point values would be suspect and it is argued therefore that under these circumstances, and given the available data, averages for k and x are the appropriate choices.

Therefore, it was necessary to examine possible variations caused by changing k and x values. Overton (1966) found that inflow hydrographs varied for each storm in small catchments and k and x varied likewise. In the present study, k and x were considered indifferent to storm characteristics owing to the large size of the Fraser Basin. Overton also examined the effects of multiple routings on the parameters and found no significant changes in the routed hydrograph when k is less than the time to peak of the input hydrograph. Also, the time of peak of the outflow hydrograph did not change when x was between 0.0 and 0.5. This is relevant for the Fraser River Time-area model since peak timings are of utmost importance in the determination of apparent travel times.

Values of x were varied in one of the regional modes of the Fraser River time-area model to identify possible changes in the results. When tributary time-area inputs are routed to the

Fraser River, all the storage is incorporated at the confluence using x=0.0. Therefore variable x values can be incorporated only when routing on the Fraser River itself. As x was increased from 0.0 to 0.3, peak magnitude also increased (table 3.1).

Table 3.1
Peak Discharge for different values of x, expressed as a percentage of the value for x=0.2.

Fraser River Peak	x value				
	0.1	0.15	0.2	0.25	0.3
at Shelley	92	94	100	103	108
with Shelley	100	100	100	100	100
at Quesnel	97	98	100	103	105
with Quesnel	100	100	100	100	100
at Thompson	95	98	100	101	106
with Thompson	100	100	100	100	100
at Hope	99	100	100	100	101

However, the change appeared insufficient to change dominant peaks, and therefore apparent travel times. It is interesting to note that when another tributary is added, peak discharge remains the same regardless of the x value. This is because tributary inputs dominate and govern peak time, regardless of Fraser River discharge. Also, the degree of variation in discharge tends to become smaller toward Hope. This is possibly because the hydrograph becomes broader with each input, and therefore subject to less attenuation than a sharp peak.

Despite small peak discharge changes, time of peak is rarely altered (table 3.2) implying that although the hydrograph may be altered slightly, the time ordinates are preserved. This is of paramount importance to the Fraser River time area model.

Table 3.2 Time of peak with various x values

Fraser River Peak	x value 0.1	0.15	0.2	0.25	0.3
at Shelley	8	8	8	8	8
with Shelley	4	4	4	4	4
at Quesnel	11	11	11	11	11
with Quesnel	5	5	5	5	5
at Thompson	13	13	13	14	14
with Thompson	13	13	13	13	13
at Hope	15	15	15	15	15

The shape of hydrographs produced by each x value are very similar, and the most noticeable effect of x is the degree of attenuation. Sharp input hydrographs are attenuated most, especially by low values of x. However, once incorporated into the Fraser River flow, the differences are soon counteracted.

Attenuation between tributaries, and thus gauging stations, will be important in determining peak volumes. Long distances between tributaries enable considerable reduction of an upstream peak. In some cases, the effect may be sufficient to allow a smaller input downstream to dominate the overall hydrograph and therefore affect the timing of maximum flow. This method gives an immediate impression of the relative importance of each tributary input in producing a mainstream flood peak, with obvious implications for future monitoring.

3.2.2. Basic Model Interpretation

The input to Hansard takes the form of an almost regular hydrograph, except for a broad peak from its elongated basin (figure 3.4a). At Shelley, the shorter contributing area produces a sharp, symmetrical peak with a timebase half that of Hansard (figure 3.4b). When the translated flow from Hansard is added to the Shelley input, the Shelley waters form the dominant peak because they are combined with the Hansard rising limb (figure 3.5a). The Hansard peak becomes a blip on the Shelley recession limb. Depending on the relative importance of flows from each subbasin, the time-to-peak has the potential to change quite considerably (see section 3.6). The Nechako and Stuart river inputs were included as average values for all versions of the model after a visual examination of the recorder charts of the Nechako at Isle Pierre which is relatively near the Fraser confluence.

The West Road tributary basin is disproportionately broader in the middle. This produces a very steep and sharp hydrograph (figure 3.4c) which drains into the Fraser River before the arrival of the Upper Fraser (Hansard and Shelley). The two peaks are of almost equal volume and either the West Road tributary or the Upper Fraser is capable of altering the balance between peaks. That is, the maximum recorded peak on the Fraser at the West Road confluence may be associated with the West Road river alone, the Upper Fraser contribution alone or some combination of both.

The Quesnel river drains the windward Cariboo mountains and the eastern Interior Plateau. Its basin is broadest in the upper reaches but it is shorter than the West Road basin (figure 3.4d). Therefore when the Quesnel is added to the Fraser, the bulk of its discharge arrives in synchrony with the West Road peak. At this point on the Fraser, the first peak is accentuated and dominates the somewhat attenuated second peak from the Upper Fraser (figure 3.5c). Minor inputs from small tributaries can increase flows ahead of these main peaks, but may be insignificant in the overall hydrograph.

The next tributary of significance is the Chilcotin. Draining the western edge of the Coast Mountains, the basin is also widest at its headwaters. Basic time-area considerations produce a "reversed" input hydrograph (figure 3.4e) whose main input is delayed. Despite the delay, its waters are a supplememt to the rising limb of the first Fraser peak (figure 3.5d) because of the travel time from the Quesnel tributary. However, the desertified nature of the Chilcotin basin would considerably alter its real contribution and timing. In the basic model, it is suggested that the combination with the underestimated Quesnel allows a reasonable simulation of the whole basin.

The Bridge basin is also widest near its headwaters (figure 3.4f), but its relative shortness allows most of the flow to enter the Fraser before the peak flows described above (figure

3.5e). Because the headwaters are glacially-fed, the actual amount of input is difficult to determine with a time-area method. The generally high precipitation received in the basin should negate any reduced flow from the glacial areas, for the basic model.

The gauging station at Texas Creek which is a short distance below the Bridge confluence, receives a three-peaked flow according to the model. This is fewer peaks than were determined by routing the whole basin time-area inputs to Texas Creek. Attenuation however, merges inputs considerably. The absolute timing of maximum peak is flexible within a few hours but different ratios of inputs could alter the peak of importance and consequently time-to-peak data.

The North Thompson river has a broad headwater catchment draining the southern Cariboo. The upper reaches consist of lakes draining icefields, and so the real unit hydrograph from this basin probably does not reflect the broadness. The length of the basin causes the time-of-peak to be lagged (figure 3.4g). The South Thompson river drains the Monashee Mountains and its basin is also broadest at the headwaters but much shorter (figure 3.5f). The South Thompson peak appears as a blip on the North Thompson river hydrograph rising limb at their confluence. Lakes are of much importance in the Thompson basin, but the degree to which they reduce and delay peaks can not be assessed easily and incorporated within the simple principles of the model. Below the junction of the North and South Thompson rivers, the Bonaparte to the north and the Nicola to the south supply the bulk of local flows. Therefore the flow added by the whole Thompson basin to the Fraser River has two peaks arriving marginally before the peaks created in the Bridge catchment and middle Fraser area respectively.

The Thompson river supplies a major input to the Fraser. This causes backwater effects, which as noted by Clark (1945), alter the storage relationships in that reach. Theoretically, the Muskingum x component should be changed to accommodate backwater, but the average value of 0.2 is assumed throughout the model in the absence of contrary empirical evidence.

The first Thompson peak "drowns" the Fraser and the Bridge creates a small blip afterwards. The timing of the arrival of flow from the Bridge basin therrefore may be critical if flooding is imminent. The second Thompson peak supplements the main Fraser flow. Again, timing may be critical. After these main peaks, the steep falling limb of the hydrograph is interupted by a plateau caused by attenuated Upper Fraser flow. By the time the flood wave has reached the Hope station, the first peak has attenuated considerably and acts as a "warning" high before the main peak. Despite the relatively low level of the Upper Fraser waters, its role in prolonging high level conditions could be problematic in terms of overbank flooding.

3.3 Mode II: Rainfall Simulaticn

3.3.1. Description

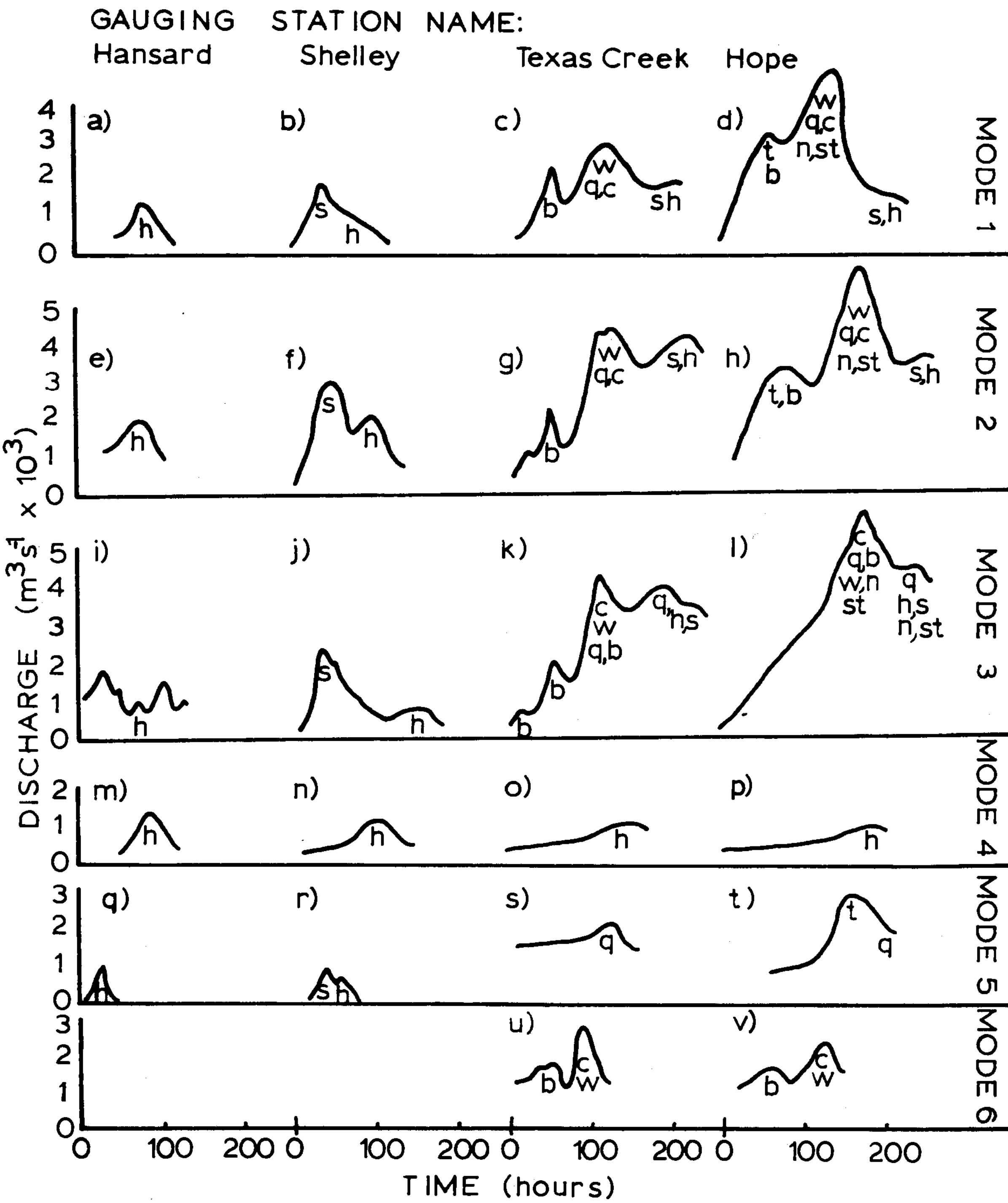
In many real situations, a model as described in the previous secticn, is inadequate. Therefore, the first adaptation of the model incorporates the spatial variation of precipitation. Meteorological data show an uneven distribution of amounts and timing of precipitation for most events. Mean annual precipitaticn values were used to weight precipitation amounts received by each isochrone-delimited area. The resultant hydrcgraph enables an interpretation of the relative tributary additions, by acccunting for the wetness or dryness of each subbasin. As such, the model remains relatively crude. The contributions of Interior Plateau tributaries are possibly still overestimated because the mean annual precipitation figures represent total precipitation, nct the proportion from a given precipitation event. This is noteworthy because most of the summer precipitation in the Interior Plateau is from local convectional storms, while elsewhere, frontal precipitation is most important.

A multitude of variations are possible, but the aim of this mode was to simulate basic hydrometeorological differences. Regions of 'highflcw, delimited on the basis of mean annual flood series similarity, have been shcwn to correspond closely to mean annual precipitation isohyets (Waylen, 1981). Weighting in terms of mean annual precipitation is therefore relevant to mean annual flood volumes and consequently, runoff yield. Also, vegetation responds to mean annual precipitation and therefore, runoff coefficients for a large area will be related to amount of precipitation in a secondary and less direct manner. A logical extension of this mode would be to assign precipitation contributions to the three seasons of interest: spring, summer and autumn.

3.3.2. Interpretation

The main features illustrated by this simulation were not significantly different from the basic model. When all the area above Shelley contributes to the Fraser River flow, the time of the peak reflects maximum inflow below Hansard (figure3.6f). The Hansard basin produces a larger runoff volume from increased precipitation in the mountains (figure 3.6e) and forms a more pronounced second peak in the flow at Shelley. The West Road tributary provides a lower peak volume preceeding the Upper Fraser flows. Small changes of the West Road river input volume were simulated. Although minor effects were visible on the mainstream Fraser, attenuation has made the changes practically insignificant when the flow arrives at Hope.

The height order of the first and second Fraser River peaks are reversed by high runoff from the Quesnel basin. An irregularly shaped input hydrograph is contributed from the arid



KEY TO CONTRIBUTING SUBBASINS:

h=Hansard	q=Quesnel	n= North Thompson
s= Shelley	c= Chilcotin	s= South Thompson
w=West Road	b= Bridge	t= Thompson

FIGURE 3.6 Comparative hydrographs for all modes at gauging stations

Chilcotin river basin because the driest region borders the Fraser and therefore most of its runoff arrives late. The Chilcotin addition coincides with the Quesnel rising limb and produces a twin-pronged initial peak on the Fraser. The balance between dominating peaks appears very delicate. Therefore, the recording of an instantaneous peak for example, could be altered by several hours depending on storm characteristics.

High temperatures and the lack of precipitation in the Chilcotin basin and the area around the Fraser / Thompson confluence cause secondary effects. Evaporation and water withdrawals for irrigation may cause some reduction of river discharge. Occasionally, water also will be required to fight forest fires in this very hot, dry region.

The Bridge river discharges into the Fraser River before the arrival of upstream peaks, despite having its highest runoff delayed. This produces a multi-peaked hydrograph at Texas Creek (figure 3.6g). In the Thompson basin, considerable contrasts exist between high volume runoff from the mountainous headwaters and barely significant runoff below Kamloops. The North Thompson river input reflects an increased supply from the mountains, but the South Thompson input is still more sensitive to the effects of basin shape. Combined, they produce one large, steep-limbed peak.

At the Thompson-Fraser confluence the inputs seem well synchronised: implying that the high runoff production areas lie similar travel times away in both basins. The combination of the

Bridge and lower Thompson flows produce a minor first peak. The main peak is produced by the Upper Thompson, Quesnel and West Road together. The volume and peak timing of the North Thompson river are of major significance in this mode. Because it drains higher elevations however, it is feasible to suggest some runoff may be held as snow occasionally. In this case, the peak may be reduced or delayed or a combination of both. At Hope, one major peak is clear and compared with the basic model, the Upper Fraser flow is more substantial (figure 3.6h).

3.4 Mode III: Snowmelt Adaptation

3.4.1. Description

The problem of evaluating snowmelt was discussed in chapter two. Clearly, any model using meteorological inputs can only hope to reasonably estimate the runoff component due to snowmelt. Snowmelt in the Fraser Basin is of prime importance in producing the annual peak flow and therefore, it should not be neglected. Much additional information would be necessary to analyse snowmelt but this is incompatible with the time-area model developed in this study.

Only one case of snowmelt is therefore modeled here. The appropriate scenario suggests that cold temperatures prevail at the time of a storm, resulting in snow only at the higher elevations. Temperatures rise after the storm enabling total melt of the new snow. In this mode, the snowmelt runoff is lagged accordingly and distributed over a twenty-four hour period to simulate a diurnal melting regime. Woo and Slaymaker (1975) found diurnal runoff cycles to be pronounced in a small catchment in the Southern Coastal Mountains of British Columbia.

Mode III is therefore concerned with minor delays caused by snowfall rather than general basin melt. This phenomenon is more likely to occur at either end of the winter season. Variations were considered but would probably not justifiably improve the model. Examples include a weighted increase in snow with elevation on an already weighted precipitation index, to account for the percentage of snow to rain; or routing some percentage of the runoff via greater storages and with longer travel times.

The summary of snow survey measurements from 1935 to 1975 (British Columbia Water Investigatios Branch, 1975) was inspected, but no easily definable pattern of snow (water equivalent) distribution could be determined. Monitoring amounts of snowfall is obviously a great advantage. A visual analysis of river gauge recorder charts, revealed that a textbook hydrograph shape only existed in late summer and autumn especially at Hope and Texas Creek. This illustrates that snowmelt is incrementing flow volume in the early part of the year. The events later in the year therefore, allow an examination of basin processes without snowmelt complications.

3.4.2. Interpretation

One of the salient differences to mode I is the sensitivity of the relatively small basin above Hansard to diurnal melt fluctuations (figure 3.6i). As volumes become higher downstream, the diurnal variation becomes negligible. The effect at Shelley is not so erratic because simulated snowmelt is less from the generally lower altitude drainage area (latitude effects were ignored). The combination of Hansard and Shelley reflects the time of peak from the area contributing to Shelley, but because of decreased volume in the Hansard rising limb, the Shelley peak occurs after the maximum at Hansard (figure 3.6j). Snowmelt in the region above Hansard produces a secondary, small rise on the falling limb.

The West Road river input shows very little difference to previous modes. Snowmelt arrives relatively late and is absorbed into the first peak from the Upper Fraser. Substantial snowmelt also arrives late from the Quesnel basin and joins the second peak flow. Obviously the interplay of factors would determine the dominant peak and thus, apparent travel time of peaks between two mainstream stations. Attenuation merges the peaks more closely near the Chilcotin river confluence. The Chilcotin adds most input late because its highest runoff is from the extreme west and also subject to small snowmelt delay. However, the resultant input joins the Fraser river before the upstream peaks. When the Chilcotin peak was varied in terms of arrival time, substantial differences occurred in the Fraser to the extent of altering a hypothetical instantaneous peak. The Bridge river input coincides with the early Chilcotin peak. Diurnal effects again seem to be important in this basin. The hydrograph of the Fraser at Texas Creek is dominated by highflow contributions from the Quesnel basin (figure 3.6k).

The snowmelt mode causes large changes in the Thompson input due to initial storage and subsequent release of snowmelt. The input has an extended timebase and a much lower peak, but it still coincides with the first peak on the Fraser. When all flows are summed at this point, the snowmelt acts to "fill-in-the-lows" in the rainfall mode hydrograph. Therefore at Hope, the hydrograph shows one major rise and fall, with a small recession limb plateau created by Upper Fraser runoff (figure 3.61). The hydrograph has a broader peak with the total volume distributed more evenly than in previous modes.

From the three modes described above, it is apparent that the highest peak at any point on the Fraser River usually corresponds to a particular input upstream. The hydrograph at Hope is the summation of all basin input: attenuated and summed. Therefore, to determine true peak travel time, it is necessary to identify the input to which the Hope peak corresponds. When only daily data are available, or the concern is with maximum discharge, an analysis of peak travel times will be unrealistic for the whole basin.

3.5 Modes IV,V,VI: Regional Adaptions

In a catchment with the size and hydrometeorological variety of the Fraser Basin, it is unlikely that precipitation is received everywhere and commences at the same time. Therefore, three storm centres for smaller areas were simulated, utilising weighted precipitation values.

Mode IV was concentrated on the Rocky Mountain Trench to examine attenuation and peak travel times when unaffected by inputs between Hansard and Hope. The broad hydrograph was only slightly attenuated, but compared with the normal volumes at Hope, its arrival may not be noticeable (figures 3.6m to 3.6p).

Mode V deposited precipitation on the west (windward) side of the Cariboo range. A small area upstream of Hansard, supplied a short, sharp, low peaked input (figure 3.6q). When a small input from above Shelley is added to the Hansard flow, its peak dominates. In this instance, time-of-peak at Shelley is the same as at Hansard (figure 3.6r). Therefore, apparent flood wave travel time is zero hours. A large, steep peaked input from the Quesnel basin joins the Fraser before the Upper Fraser flow and forms the dominant peak. The North Thompson river basin produces a sharp, but broad based peak. Although it is greatly attenuated at the Fraser confluence, its volume is still significant and its arrival coincides with the Quesnel flow (figure 3.6s). The Upper Fraser input is now lost within the recession limb and one large, broad peak is observed at Hope (figure 3.6t). Because the

Thompson river flows dominate, an analysis of travel times between Texas Creek and Hope would be incongruous.

Mode VI simulated a storm that was a supposed spill-over effect from the Coast Mountains, providing a series of inputs between Shelley and Texas Creek (figures 3.6u to 3.6v). The West Road river headwaters develop a very peaked input hydrograph which is almost completely attenuated on entering the Fraser. The Chilcotin basin also generates a steep peak whose arrival at the Fraser coincides with the West Road river input, producing a wide peak. The area of the Bridge basin affected by the storm creates substantial runoff. Its peak occurs before the arrival of upstream flow (figure 3.6u) but the relative balance with the second peak is probably governed by local storm characteristics. It is feasible that a storm moving from the northwest would reduce the separation between the two peaks. Conversely, a storm approaching from the southwest may separate them. However, storm movement is generally fast in relation to runoff speeds and hydrograph shape may not alter drastically. It is suggested that timing differences are substantial only when precipitation events occur in close succession.

The hydrographs from regional simulations (modes IV, V, VI) seem less sensitive to changes in tributary input timing than when the whole basin produces runoff. However, it is possible to envisage many scenarios depending on relative interactions.

3.6 Discussion of generated data

First, it should be restated that this model is dependent totally on the concept of isochrones and consistent proportions of runoff generation throughout the basin. Time-area concepts were tested in six modes as summarised in table 3.3.

Table 3.3. Model variations.

Mode no.	Model description
I	basic unadjusted model of whole basin
II	areas weighted by mean annual precipitation index for whole basin
III	snowfall at higher levels witheld from calculations, followed by later melt and release of runoff
IV	local storm centered over Rocky Mountain Trench
V	local storm on windward side of the Cariboo mountains
VI	overspill effect on lee of Coast mountains

Generally, hydrographs remain very similar for events involving the whole basin. Hydrographs at Texas Creek are the most varied in all but two cases, illustrating the importance of tributary inputs in the reach below Shelley. Addition of the Thompson river at Lytton tends to smooth the hydrograph and heighten the maximum peak.

The model illustrates the importance of tributary input timing for determining time-of-peak (table 3.4) Time-of-peak thus appears related to tributary input. Therefore, the term "travel times" is a misnomer for this system because basin

Table 3.4. Time of peak derived from model.

	Hansard	Shelley	Texas Creek	Hope
Mode I:	50	30	112	162
Mode II:	50	45	140	157
Mode III:	28	34	101	157
Mode IV:	73	78	129	168
Mode V:	28	39	129	134
Mode VI:	--	--	95	146

(in hours from commencement)

Table 3.5 Apparent travel times

	Hansard to Shelley	Shelley to Texas Creek	Texas Creek to Hope
Mode I:	-20	+82	+50
Mode II:	-15	+95	+17
Mode III:	+ 6	+67	+56
Mode IV:	+ 5	+51	+39
Mode V:	+11	+90	+ 5
Mode VI:			+51

(in hours from commencement)

characteristics totally swamp most flood wave movement (table 3.5). Undoubtedly, some element of wave progression will be experienced on the Fraser, but predictions using crest travel times are not reasonable. Peak time differences however will be referred to as travel times throughout this study, because these ideas are compatible with data collected and ideas of streamflow in general.

Differences between the six modes reflect the relative importance of the tributaries. The dominant source of peak at each station is presented in table 3.6.

Table 3.6 Source of main peak volume

	Fraser River at Shelley	Texas Creek	Hope
Mode I:	- S	- W Q C	- W Q - N St
Mode II:	- S	- - Q C	- - Q C N -
Mode III:	H S	- W Q C	- W Q - N St
Mode IV:	H -	H - - -	H - - - - -
Mode V:	H S	- - Q -	- - - - N St
Mode VI:	- -	- W - C	- W - C - -

Key: H=Fraser basin above Hansard, S=contribution added between Hansard and Shelley, W=West Road river, Q=Quesnel river, C=Chilcotin river, N=North Thompson river, St=South Thompson river.

The following can be concluded.

1. A negative travel time between Hansard and Shelley exists when the input between Hansard and Shelley is dominant. The peak at Shelley can only be later than Hansard when either its contributing area supplies an insignificant inflow, or the Hansard volume is delayed (snowmelt). It is unlikely that the peak at Shelley will occur after the peak at Hansard for widespread rainfall events.
2. A negative travel time between Hansard and Shelley allows apparently slower travel times between Shelley and Texas Creek. This is because Texas Creek shares the same overall

basin timebase, but inputs below Shelley are more important to peak timing than arrival of an upstream floodwave.

3. A slow travel time between Shelley and Texas Creek can occur also, if only the lower sections of the Hansard and Shelley subbasins provide runoff. Then, runoff has shorter distances to travel and the peaks at Hansard and Shelley are relatively early compared to peaks at Texas Creek.
4. Travel times decrease between Shelley and Texas Creek in mode III because the Texas Creek peak receives runoff from a large area that is not subject to substantial snowmelt delay (high elevation snowfall case) and therefore there is a relatively early peak at Texas Creek.
5. The shortest travel time between Shelley and Texas Creek is when a flood is produced only above Hansard. This suggests that tributary additions in this reach delay the peak. For mode IV (no additions), on other reaches travel times for each reach are central to the range of derived values.
6. Between Texas Creek and Hope, travel time is highly variable. The relative importance of the Thompson is paramount. The shortest travel time occurs when the Thompson input is much more important than the Fraser flow. This is because the Thompson is a shorter basin, its time-to-peak is earlier and therefore, most of its volume arrives at Hope relatively fast. This is especially clear for a storm centered over the Cariboo mountains because runoff generating areas of the Thompson are closer to Hope than

those of the Fraser. Slow travel times for this reach occur when the Thompson does not contribute.

7. When all the basin operates together, the slowest travel times are recorded. This suggests inputs from all subbasins are significant in the timing of the Hope peak.

These conclusions were deduced from relatively simple cases involving one storm. As the number of storms increases, the runoff pattern and resultant hydrographs become more complex. The areal extent of a storm is of prime interest because tributary hydrograph shapes reflect basin shape and size, and, the spatial characteristics of precipitation. Rogers (1972) however, found that a reasonable variation in precipitation amount and intensity within a general storm had little effect on the shape of the surface runoff hydrograph.

The arrival time of tributary peaks appear important in Fraser River hydrographs, therefore they were examined briefly. In most cases for the middle Fraser, each tributary arrives before the main upstream peak and accounts for a substantial proportion of Fraser volume at that time. Overall, small timing changes of one tributary will not drastically alter the resultant Fraser River hydrograph. Also, as the Fraser becomes larger, it becomes less sensitive to the smaller tributary inputs.

From tables 3.3 and 3.4, it is apparent that negative travel times between Hansard and Shelley occur when the Fraser

Basin above Hansard does not contribute to the main peak. Therefore, time of peak at Shelley is not dependent on the arrival of Hansard flow. As the Hansard flow becomes more important, travel times can become positive. The Shelley peak input coincides with the arrival of the rising limb of the Hansard hydrograph and this combination produces a large peak before the arrival of peak Hansard discharge. When the Hansard peak arrives, its effects are minimised because there is little additional runoff to Shelley at this time. Therefore, the Shelley input must be reduced for the Hansard peak to dominate at Shelley.

The relation between runoff volumes and Shelley peak timing was examined using mode II. When runoff is received from the whole basin, the volume below Hansard must be 40% of normal to allow the Hansard peak to dominate. This would require a storm across the centre of the Fraser Basin, that did not extend much north of Hansard. If only the western part of the area between Hansard and Shelley is contributing, a 50% reduction in runoff is required. This is due to the configuration of the subbasins. The McGregor river confluence with the Fraser River is below Hansard, but its basin lies parallel to the northern part of the Hansard basin. Here, high runoff is generated from the mountain slopes. Clearly, the phenomenon of negative travel times is acceptable, because time differences between peaks are not dependent on wave celerity.

The time-area model and its variations provide a useful tool for interpreting Fraser River hydrographs. By doing this, the model has achieved its potential. It is restricted by its simplicity; real world complexities need highly sophisticated models to enable predictions. It provides an invaluable aid, however, to understanding gauging station water level data used in the analysis of travel time.

IV. Peak Flow Times

4.1 Introduction

Based on an understanding of how hydrographs at selected points along the Fraser River are produced, streamflow data can be analysed with more insight. The time of occurrence of each peak was extracted from streamflow recorder chart data, and apparent travel times were derived from the time differences between peaks at two adjacent stations. It is postulated that travel times are governed by discharge if peak timing is determined by flood wave movement alone. In the absence of such a relationship, it is postulated that basin geomorphology is an important control on runoff response. This chapter commences with a review of the details of data collection methods. Peak flow data are then used to examine the nature of a relationship between discharge and travel time for the Fraser river. Finally, meteorological characteristics are integrated in an attempt to reduce "noise" and explain anomalies.

4.2 Data Collection

4.2.1 Streamflow data and gauging stations

Streamflow data were obtained from the Water Survey of Canada which currently operates 184[1] active stations within the Fraser Basin. This study requires a continuous record of stream levels to define hydrograph shape precisely, but unfortunately most stations supply only one reading per day. Hourly water levels therefore were obtained directly from the existing continuous recorder charts for Fraser River gauging stations between 1970 and 1980. These data were converted to discharges by use of the appropriate rating tables and ajustment factors.

The prerequisites described above reduced the number of useable stations to five (table 4.1). Subsequent analyses were performed for each reach between adjacent gauging stations. Station locations are illustrated in figure 2.4. Of these, the station near the Fraser headwaters is situated on Moose river but its contribution to the Fraser River, and its location near the source were considered acceptable for inclusion here.

The Texas Creek and Hope stations are classified as having regulated flow following the completion of the Kenney Dam on the Nechako river in 1952. The dam reduced the effective drainage area of the Fraser Basin by 14000 km^2 because Alcan diverted the

[1]Surface Water Data for British Columbia, 1982, published by the Inland Waters Directorate, Ottawa 1983

Table 4.1 Gauging Station details

station#	Stn. name	area (sq.km)	latitude longitude	Years of record (1)	G.S.C. datum (2)
08KA008	Moose river	460	55 55 12 118 48 00	58-80 nat	1039
08KA004	Fraser river at Hansard	18070	54 04 43 121 50 52	70-80 nat	599
08KB001	Fraser river at Shelley	32500	54 00 40 122 37 00	50-80 nat	567
08MF040	Fraser above Texas Creek	152360	50 36 50 121 51 10	51-80 reg	180
08MF005	Fraser river at Hope	217880	49 22 50 121 27 05	50-80 reg	38

(1) nat = natural flow, reg = regulated flow
(2) the Geological Survey of Canada datum in metres.

stored water to Kitimat. The records at Isle Pierre (08JC002), 40 km above the Fraser River confluence, were assessed and implied that this tributary had introduced no sizeable flood waves into the Fraser during the study period, despite periodic spillway openings at the dam. The yearly station discharge pattern shows a general increase to a late spring maximum and then a gradual decline to winter levels. This is probably due to the moderating influence of lakes, especially in the Stuart basin.

The Hope gauging station is often used as the key station to the Fraser. Its records, dating to 1912, are considered to be the most reliable on the Fraser, and it is situated sufficiently inland to be above the daily tidal fluctuations.

4.2.2 Meteorological data

These were provided by the Atmospheric Environment Service. The Fraser Basin has 105 meteorological stations (figure 2.4) supplying precipitation data on a daily basis. Precipitation is recorded as rain or snow after 1977, but prior to this, temperatures are required to estimate snowfall.

Published daily values provide a static impression of precipitation distribution. Therefore, for several days before and after each streamflow peak, frontal movements, cold low digressions and Upper Air flow patterns wer tracked on six-hour surface charts and twelve hour Upper air charts.

Slaymaker (1972b) found a trend of increasing Fraser River discharge at Hope from 1952-1969 despite an expected reduction due to retention behind the Kenney Dam. Although he found no statistically significant precipitation increase, he deduced that more precipitation, especially snowfall in the upper ungauged areas, caused the higher discharges. This could occur following a northward shift in the mean position of the Arctic Front, allowing moist Pacific air to exert a greater influence in central British Columbia. However, this is difficult to monitor because of the rarity of high elevation meteorological stations. The Barkerville station in the Cariboo Mountains is therefore important and its data are key inputs for flood predictions and the U.B.C. model.

4.3 Observed Peak flow Travel Time relationships

The theories discussed in section 1.2, indicate a negative, relationship between discharge and travel time should exist. Both travel times in hours and the velocities necessary for flood waves with corresponding travel times, were examined for a dependence upon actual discharge values. No decisive patterns could be determined between the two sets of variables, either on the basis of a whole year or when split by season. The seasonal division was made to differentiate between different runoff generating processes.

1. Spring events with a substantial snowmelt contribution.
2. Smaller scale summer events caused by localised storms and increased water losses during transfer.
3. Autumn rainfall-runoff events due to increased frontal activity.

At this stage in the study, only cases with positive travel times between two stations were considered. In most cases, only a qualitative analysis was attemped due partly to the lack of clear relationships exhibited by the data (figure 4.1). The implications from analysis are presented in table 4.2. This table highlights general trends but not firm relationships and additional observations subsequently will be discussed.

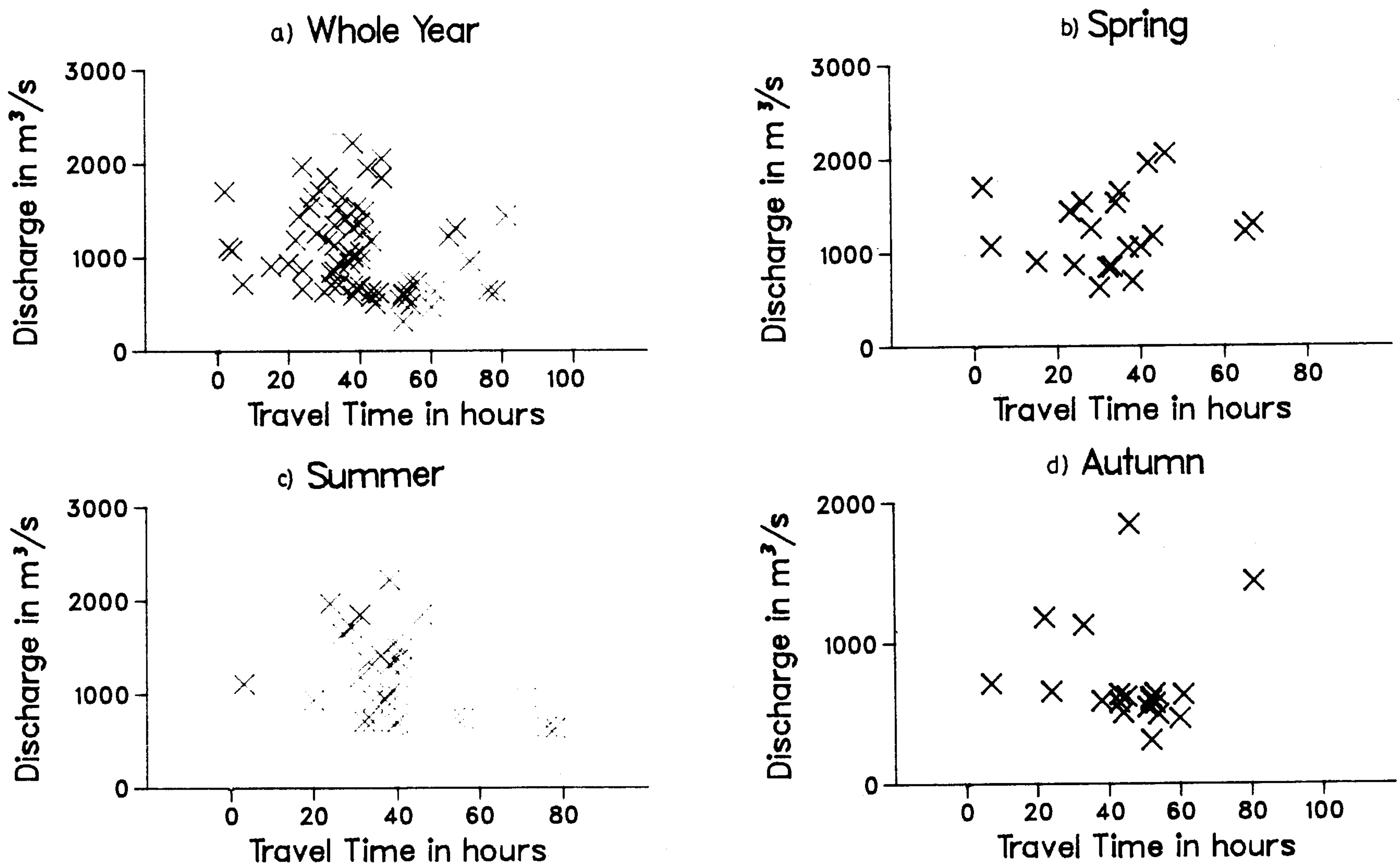


FIGURE 4.1.1 **Discharge–Travel Time relationships for Red Pass to Hansard**

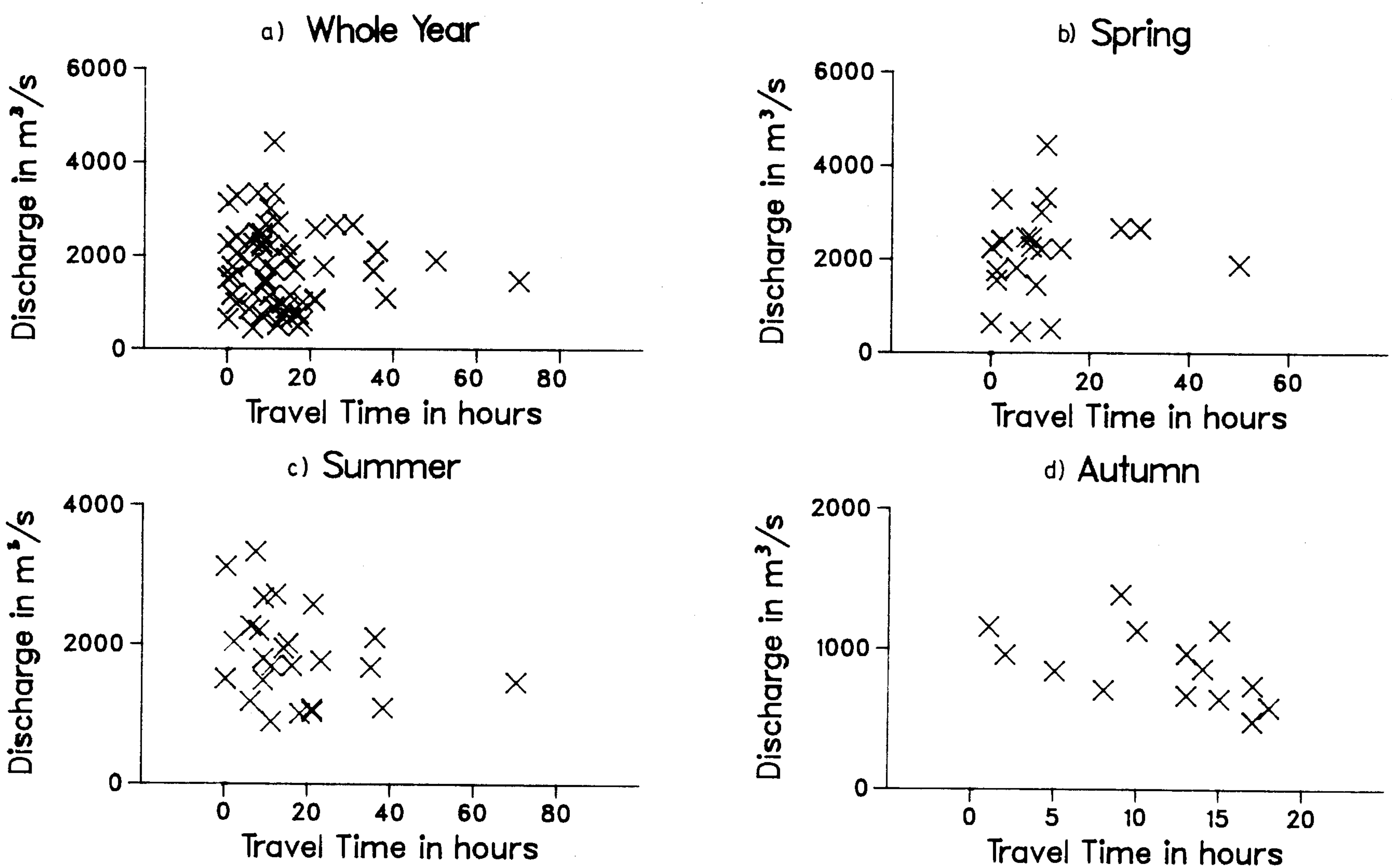


FIGURE 4.1.2 Discharge–Travel Time relationships for Hansard to Shelley

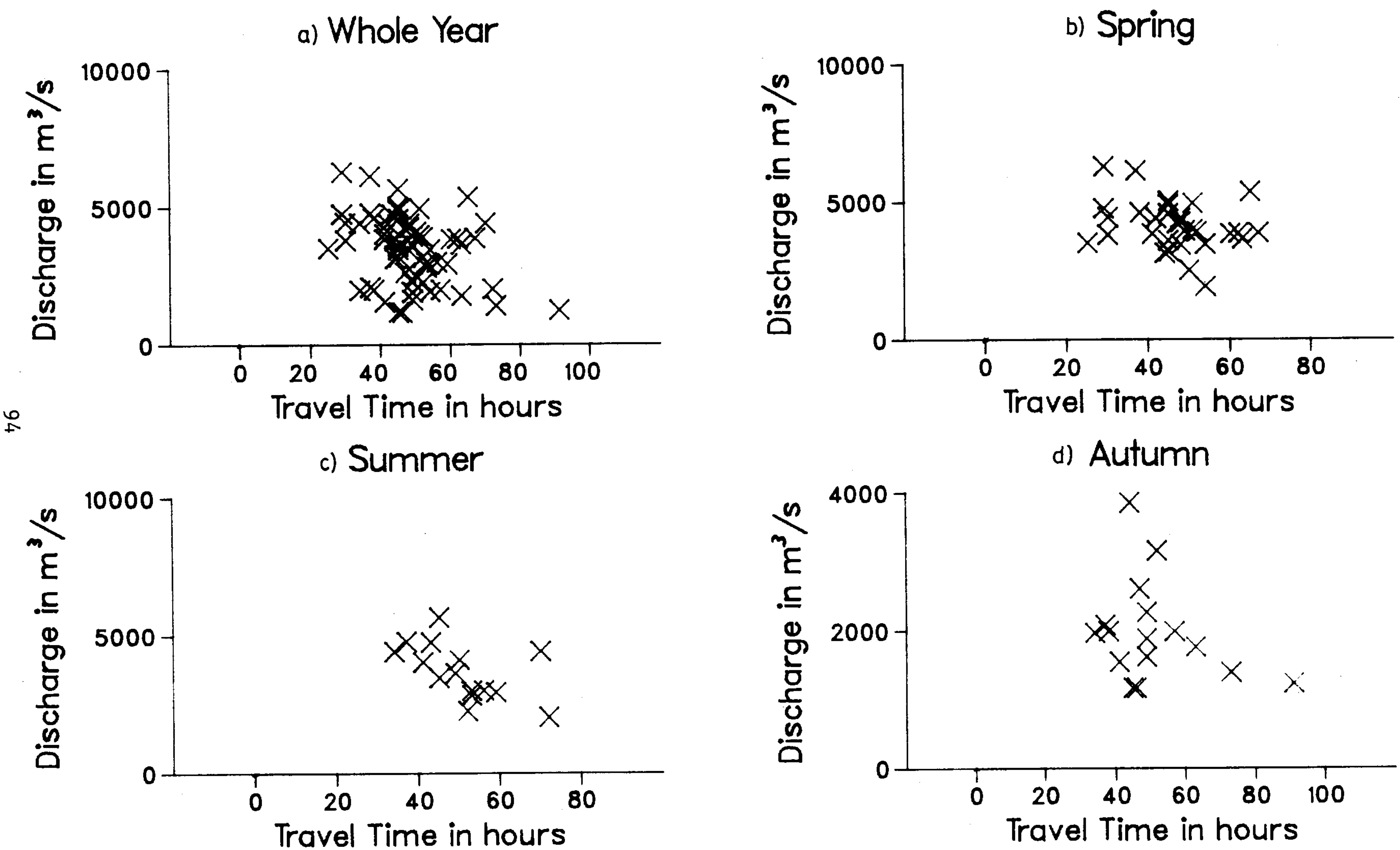


FIGURE 4.1.3 **Discharge – Travel Time relationships for Shelley to Texas Creek**

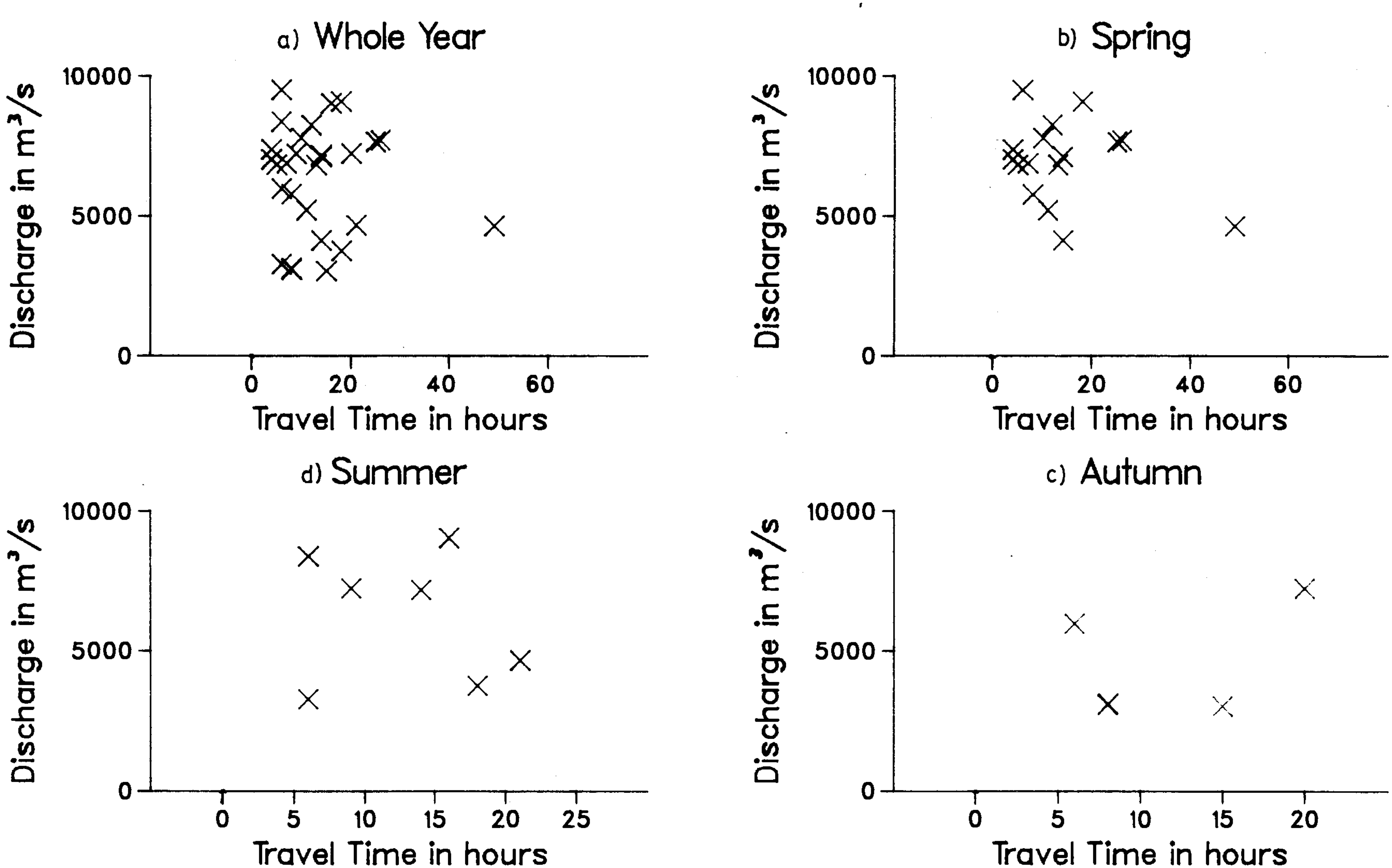


FIGURE 4.1.4 **Discharge — Travel Time relationships for Texas Creek to Hope**

Table 4.2 Discharge/travel time relationships

Reach	Whole Year	Spring	Summer	Autumn
RP-H	large Q range for medium TT	as annual	as annual	as annual
H-S	broad Q range with short TT	as annual	as annual	very broad negative trend, low slope
S-TC	broad Q range with short TT	large TT range for medium Q	negative trend, low slope, broad scatter	as summer
TC-Ho	broad Q range with short TT	as annual		

key: Q=discharge, TT=travel time, RP=Red Pass,
H=Hansard, S=Shelley, TC=Texas Creek, Ho=Hope

On an annual basis very little can be determined about the nature of an association between discharge and travel time, except with regard to the range of both variables. With the exception of annual data for the reach between Red Pass and Hansard, most discharge values can occur with relatively short travel times (figures 4.1.2a, 4.1.3a and 4.1.4a). The maximum discharge range on the reach between Red Pass and Hansard occurs central to the spread of travel time values (figure 4.1.1a). For the Hansard to Shelley reach it appears that the longest travel times occur at medium discharges, contrary to the predictions of Linsley et al (1949). Annual data suggest travel times are not representing wave celerities in most cases.

The freshet peaks are the highest in all reaches (figures 4.1.1b, 4.1.2b, 4.1.3b, 4.1.4b). While it is possible to comment on a few points on the spring plots, there are more exceptions than general rules. At this time of year the observed unpredictability of travel times is probably resulting from spatially and temporally uneven inputs.

In summer, the broadest range of travel times occurs at low discharges for the upper two reaches (figures 4.1.1c and 4.1.2c). For the Shelley to Texas Creek reach (figure 4.1.3c) a vague trend of shorter travel times with higher discharge is indicated, possibly responding to localised convectional storms. This conclusion is based on the nature of most summer precipitation in the Shelley to Texas Creek region, and observations from the model developed in the study. The model implies that travel time will not be dependent on discharge when there is basin-wide precipitation, but that a relationship can emerge for small, localised storms.

In autumn, low discharges appear to occur only with relatively long travel times for the middle two reaches (figures 4.1.2d and 4.1.3d), but the pattern is still confused (figures 4.1.1d and 4.1.4d). For the total river length between the extreme stations, travel time appears to be totally uncorrelated with discharge.

The importance of the Thompson river input to the Fraser River below Texas Creek led to an examination of discharge and travel time relationships between Spences Bridge and Hope. The patterns exhibited here were no better than those on the Fraser River itself. In general, the search to establish patterns between discharge and travel time seems fruitless.

Considering discharge at only one station may contribute to the paucity of results because travel times will respond to conditions along the whole section. Therefore, several tests were made to incorporate discharge values for more than one point. The first approach required the computation of peak difference, that is, the volume by which the peak has increased downstream. This measure should account for lateral inputs which may confuse the discharge/travel time pattern. Trends appear to be absent from annual data. Seasonally divided data were no clearer except for the two uppermost reaches in autumn, where a tenuous, negative, linear relationship appears. Often, between Hansard and Shelley, autumn peak differences are the lowest, signifying relatively smaller inputs downstream. The nature of autumn precipitation allows widespread receipt within the basin. Therefore, low peak differences may reflect a thirsty groundwater condition and influent streams in the area contributing to the Fraser River at Shelley. With the elimination of substantial input from this area, the Hansard peak can dominate and allow almost regular travel times on this reach. If the increases in discharge are expressed as percentages, the result is not improved.

In a second attempt to reduce "noise", simple discharges were separated into two groups according to the relative height of the preceeding peak; that is, higher or lower. This division should reduce problems created when measured events occur too close together in time and the system is carrying previous highwater conditions in its "memory". Again, typical broad scatters were produced in most cases. A scattered, positive relationship between discharge and travel time is indicated on the lower two reaches when highflows are smaller than their predecessor. Because river velocities are a function of discharge, the associated maximum velocity for the preceeding peak will also be high. Although velocity will decrease after the first peak because of increased baseflow, it may still be relatively high when the second peak is generated. Therefore, velocity is already quite fast. It is unlikely that a small peak preceded by a larger one will record a slow travel time, regardless of the complexities of unsynchronised inputs from the subbasins. In the opposite case where prior events are larger, high discharges appear to be restricted to relatively fast velocities.

A final examination was made of discharges and velocities averaged over the whole reach, instead of values from fixed points which may not be representative. No relationships could be determined and the overall conclusion from this section is that a simple correlation between discharge and travel time or apparent velocity does not seem to exist on the Fraser River.

4.4 Influence of Meteorological Conditions

It is hypothesised that storms from different directions may complicate the patterns due to contrasts in precipitation mechanisms, volumes produced and timing of runoff generation. Three meteorological features were considered important in governing direction of storm movement, namely, fronts, upper air flows and cold low centres. The predominant frontal approach is from the northwest, but speed of passage and upper air flows are variable for each storm. Upper air flows tend to steer storms and therefore the Geostrophic winds are a useful indicator for tracking (Gary Schaeffer, pers comm).

In spring, warming upper air from the southwest (Hawaii) may be more responsible for inducing snowmelt peaks. In winter, cold low centres are likely to dominate over the Northeast Pacific and may proceed northwards along the coast, allowing a series of fronts to pass over British Columbia. Conversely, in summer the lows themselves are more likely to move over British Columbia and produce instability showers. Therefore the changing relation between the Hawaiian High and the North Pacific lows is very important in determining the spatial extent of storm activity in British Columbia. In the following analysis each flood peak was classified according to direction of probable storm approach and upper air features.

Approximately one third of the spring peaks in the data set were associated with general instability and southwest upper air flows, implying that snowmelt is important. Snowmelt produces an inconsistent input to the Fraser River allowing the Hansard volume to dominate at Shelley. Further downstream between Texas Creek and Hope, snowmelt events exhibit much variation in the discharge/travel time relationship. In summer, the meteorological conditions associated with peaks are highly variable. While the variability does not allow conclusions about the effects of meteorological conditions on the discharge/velocity relationship, it does explain the confused patterns exhibited by the data. In autumn, northwest fronts may cross British Columbia in multiples, creating notable flood peaks between Red Pass and Hansard. For a series of storms, river inputs continue to increase after the first "peak", thereby extending the overall hydrograph timebase. The first storm may not produce the highest peak, and peaks at adjacent stations may not be caused by the same storm. The duration of each frontal passage does not appear to have a distinguishable effect.

The fastest travel times between Hansard and Shelley occur with a relatively low discharge produced by a front approaching from the northwest. This supports the idea that the time of peak at Shelley is independent of the arrival of a flood wave from Hansard. The slowest travel times for this reach are associated with westerly fronts and may result from relatively late precipitation receipt in the basin above Hansard. This, allows a time separation between the input from the area contributing more immediately to Shelley and the arrival of a flood wave from upstream. Therefore, "slow" travel times possibly illustrate the time difference actually produced by flood wave movement. For all sections of the Fraser River, the greatest range and variability of travel times is in summer. Additionally, travel times for each reach seem to behave without reference to neighbouring sections.

Average values of both discharge and velocity were considered also for each reach. Results were inconclusive because responses were not differentiated by direction of either frontal approach or upper air flow in any reach. Unstable conditions tend to be associated with higher discharges for a given velocity on the middle sections. Possibly travel times during unstable conditions are more representative of real travel time because the whole basin may not be contributing.

Attempts to model the direction of frontal approach proved unsuccessful because the time of passage is relatively short compared to basin dimensions and surface processes. Therefore, the main benefit of a classification based on direction, is to distinguish storm characteristics accrued from the source area rather than time differences introduced by actual movement direction across the basin.

The overall conclusion is therefore, that the "noise" interfering with supposed generalised trends cannot be fully explained by large scale meteorological differences. The most outstanding feature is the presence of unstable conditions and southwest upper air flows, enabling snowmelt in spring. Weather conditions associated with a peak in summer are more variable, implying that antecedent conditions exert an influence. Certain groups of meteorological factors indicate tentative reasons for erratic relationships between discharge and travel time, but not general trends. Therefore a range of discharges can be associated with most travel times for the upstream reach.

Until this point, only positive travel times were analysed. Positive travel times are defined as events in which the upstream peak occurs before its downstream counterpart. There are a significant number of cases where the reverse occurs, implying that travel times are negative. Most negative travel times were observed on the Hansard to Shelley section and fewest between Shelley and Texas Creek. They exhibited no preference for season or meteorological conditions (see table 4.3). This facilitates the conclusion that basin response, or the time of concentration [2] for a subbasin, is dominating time-of-peak at a station. Obviously in these instances, peak generation is a function of local basin factors and not wave progression!

Time of concentration was examined briefly to assess the importance of subbasins in determining time-of-peak. In almost

----

[2]The U.S. Department of Agriculture defines time of concentration as "the time required for runoff to travel from the hydraulically most distant part of the storm area to the watershed outlet or some other point of reference" (Viessman et al, 1977)

Table 4.3 Travel times and meteorological conditions

	% of positive times per reach				% of negative times per reach			
	R-H	H-S	S-Tc	Tc-Ho	R-H	H-S	S-Tc	Tc-Ho
unstable	30	32	30	20	17	9	25	33
NE front	3	3	9	14		18		
N front	4	3	5	5		9		
NW front	36	37	25	20	66	27	50	33
W front	21	20	25	30	1	32	25	33
SW front	6	5	6	11		5		
total +ve/-ve	93.4	78.1	95.6	84.4	6.6	21.9	4.4	15.6

all cases of negative travel times, upstream time of concentration is longer than that downstream. This allows the upstream peak to occur later, resulting in a negative "travel time". Time of concentration is highly variable between reaches and also between events over the whole system. The largest time of concentration occurs in spring due to incremental increases from snowmelt.

Cases where upstream time of concentration was longer than downstream can occur in each of the three seasons but mainly in summer between Red Pass and Hansard. This suggests that localised storm effects are important. Often, long upstream time of concentrations occur in conjunction with changing meteorological conditions, and sometimes these cases were anomalous in the discharge/travel time relationships. Now, the travel time irregularities on the section between Hansard and Shelley are well explained. In spring, despite a range of travel times and discharges, a long time of concentration upstream predominately accounts for very short travel times with low discharges. Many of the first summer peaks also have a long upstream time of concentration, which again explains the majority of extraneous points. For example, a very high head with slow travel time, and the very fastest travel times occur under these conditions. Exceptional autumn cases are also attributed to time of concentration irregularities. While the effects are not so explicit between Shelley and Texas Creek, between Texas Creek and Hope a long time of concentration upstream is associated with the faster travel times.

In conclusion, an examination of time of concentration explains much about apparently erratic travel times, and in many cases, time of concentration irregularities can be related to meteorological conditions. Once again the importance of subbasin contributions to time of peak at a station is highlighted. More specifically, the time taken for maximum runoff generation at a point, seems to be of most significance in determining time of peak anywhere along the Fraser River. The translation of a flood wave appears to be of minor importance because, for the majority of events, a negative travel time is exhibited between one or more stations.

4.5 Observed and model generated velocities

When the time differences between two peaks at neighbouring stations are converted to the velocities that are necessary for the same flood wave to be present in both peaks, some unrealistic velocities are implicated. This section deals more fully with real and apparent velocities from streamflow and model generated data.

It was presupposed that a flood wave traversing the system produced one peak which was observed at each successive station. Thus the times between peaks, derived from streamflow data, can be expressed as wave travel times (table 4.4.) When compared to generated travel times (table 3.5) the real data suggest wave celerity is extremely high. This could be due to a shortcoming of the model or a total disregard of flood wave translation in real peak timing. The dichotomy can be more easily resolved when actual water velocities and probable wave celerities are considered.

Observations suggest wave celerity to be 1.4 to 2.0 times greater than water velocity (Linsley et al, 1949). Velocity and discharge data, measured in the field three or four times per annum, were obtained for each station from Water Survey of Canada meter notes and used to calculate an expected range for wave celerity (table 4.5). Quick and Pipes (1976a) suggest that the best estimate of wave celerity for the Fraser River is 1.5

Table 4.4. Travel time statistics.

	Positive times #	mean	S.D.	Negative times #	mean	S.D.	Range
RP to H	69	38	16	5	39	24	-76 +76
H to S	70	11	8	28	12	9	-35 +38
S to Tc	63	48	11	7	17	8	-27 +73
Tc to Ho	31	14	9	8	9	6	-19 +49

(time from commencement in hours)

Key: RP=Red Pass, H=Hansard, S=Shelley,
Tc=Texas Creek and Ho=Hope

Table 4.5. Wave Velocity comparisons.

	Velocities from travel times (m/s)			Velocities from field data(1) (m/s)				
	#	mean	S.D.	#	mean	S.D.	range(3)	1.5(4)
RP to H	77	4.0	7.2	98	1.0	0.3	1.4-2.0	1.5
H to S(2)	65	5.4	6.4	90	1.5	0.4	2.1-3.0	2.25
S to Tc	82	4.6	4.8	78	2.7	0.3	3.8-5.4	4.05
Tc to Ho	39	3.8	2.3	54	2.3	0.7	3.2-4.6	3.45

(1) Derived from discharge-velocity curve at a station
(2) Only calculated when travel time more than zero hours.
(3) After Linsley et al (1949)
(4) After Quick and Pipes (1976a)

times the gauging station velocities. Clearly, the velocities derived from travel times have averages greater than is normally observed for wave translation. In addition, the velocities implicated by the travel times for the upper-most two reaches are higher than the expected range. The real velocity-discharge curves show that significant cross-sectional changes did not occur either as discharge increases, or for the ten years of study. Therefore, the lack of good relationships is not due to fundamental channel changes.

The reach-by-reach analysis of velocities was supplemented by a comparison of velocities from travel times between adjacent reaches. Spearman's "r" was very low for all pairs, suggesting independence of travel times: a phenomenon that is physically impossible. Meteorological conditions typical of erratic velocities from Red Pass through to Hope could not be discerned for these cases. Analysis of adjacent reaches is also hindered because few events were recorded at all stations. This may be due to data loss caused by problems at the gauging station, or simply no trace of a peak. The latter may be caused by attenuation or "drowning" of relatively small tributary inputs by larger Fraser River flows.

All the conclusions reached in this chapter point to discrete responses of the river at each gauging station. Rather than providing evidence of one flood wave passing through the system, it appears that "travel time" is not an acceptable, descriptive term for the time differences between peaks. Travel time traditionally refers to the maximum peak, but the model generated hydrographs suggest most highflow events are multi-peaked.

4.6 Time-area model hydrographs and Fraser River hydrographs compared

Hydrographs for late summer and autumn events were reconstructed from W.S.C. recorder chart data for the ten years of study. For this, the hourly data were invaluable. Hydrograph shape variations between stations per event, and between storms, were readily apparent. The hydrographs were divided into three types. The first showed a regular progression of hydrograph shape between all stations and positive travel times. The second group exhibited one or more "negative" travel times and the third exhibited changes in hydrograph shape. By comparing the changes in shape, volume and timing with the model, it was possible to infer precipitation source areas or storm centres.

The Fraser River hydrographs were classified by the apparent source region of precipitation. They were then checked with real precipitation distribution data which were available only on a daily basis. Precipitation will not commence everywhere at the same time in a basin as large as the Fraser, but the time lag introduced by runoff generation may be sufficient to ignore these aberrations (at this scale). Therefore, only the presence of precipitation was recorded, enabling an approximation of its spatial extent. Most classifications using hydrograph shape alone were successful. Although it is difficult to distinguish precipitation source areas precisely from the sparse network of meteorological stations, interpretations from the hydrographs were mutually supportive in 58% of the cases. Another 20% of the cases were considered to be half correct. Model hydrographs will not replicate Fraser River hydrographs exactly, because the model represents just one case of each storm. However, general hydrograph form does seem to be simulated well by the model.

Promising associations between hydrograph form and precipitation distribution were found for the majority of hydrographs. Hydrographs whose precipitation sources were incorrectly located occurred mainly in summer, with Arctic fronts or high pressure conditions. These meteorological conditions are associated with an uneven distribution of precipitation which may be causing the unpredictability. A common factor in extraneous autumn cases, was upper air flow from the southwest. This warm air flow possibly induces melt of early snow at the higher elevations, disrupting regular precipitation patterns. Generally, the inclusion of frontal activity and upper air characteristics did not offer any further explanations. Broad scale meteorological variables do not seem to be sensitive enough to account for runoff variations in a basin with the diversity of the Fraser.

The shape of the Fraser River hydrograph at Hansard may be exhibited occasionally at all stations downstream. This requires a limited duration, localised input in the north of the basin. Such a storm often produces a low, rounded hydrograph whose peak volume may not be a noticeable addition to the flow at Hope.

Examples are shown in figures 4.2c and 4.2d, and should be compared with figure 3.6. The Hansard peak may not be apparent at Hope, as in these two examples. This may occur if water is influent to the groundwater system in the middle Fraser, negating the increased volume from upstream, or the wave has attenuated completely. Volume increases downstream are explained by baseflow contributions. Although baseflow is excluded from the model hydrographs, flood peak volume changes are of the same order of magnitude as the model. The shape of the hydrograph at Hansard may pass through the system with few alterations during the summer, when precipitation in arid areas may not necessarily result in appreciable volumes of runoff. Storms in the north have the same effect. They are associated with fronts from the north, northeast or northwest, but the last also can be responsible for a range of spatial precipitation patterns.

An examination of shape changes in Fraser River hydrographs for events passing downstream revealed that pre-peak changes are more noticeable than changes after the peak. A rise or plateau on the hydrograph recession limb resulting from upper Fraser flows was rarely visible. Infact, attenuation is possibly more prevalent than expressed in the model. The next peak follows closely in many cases, obscuring the profile of the recession limb. Another facet of the system highlighted by examining real hydrographs, was the fairly even, less erratic shaped hydrographs of the Thompson river at Spences Bridge compared with the Fraser river at Texas Creek. The smoother shape

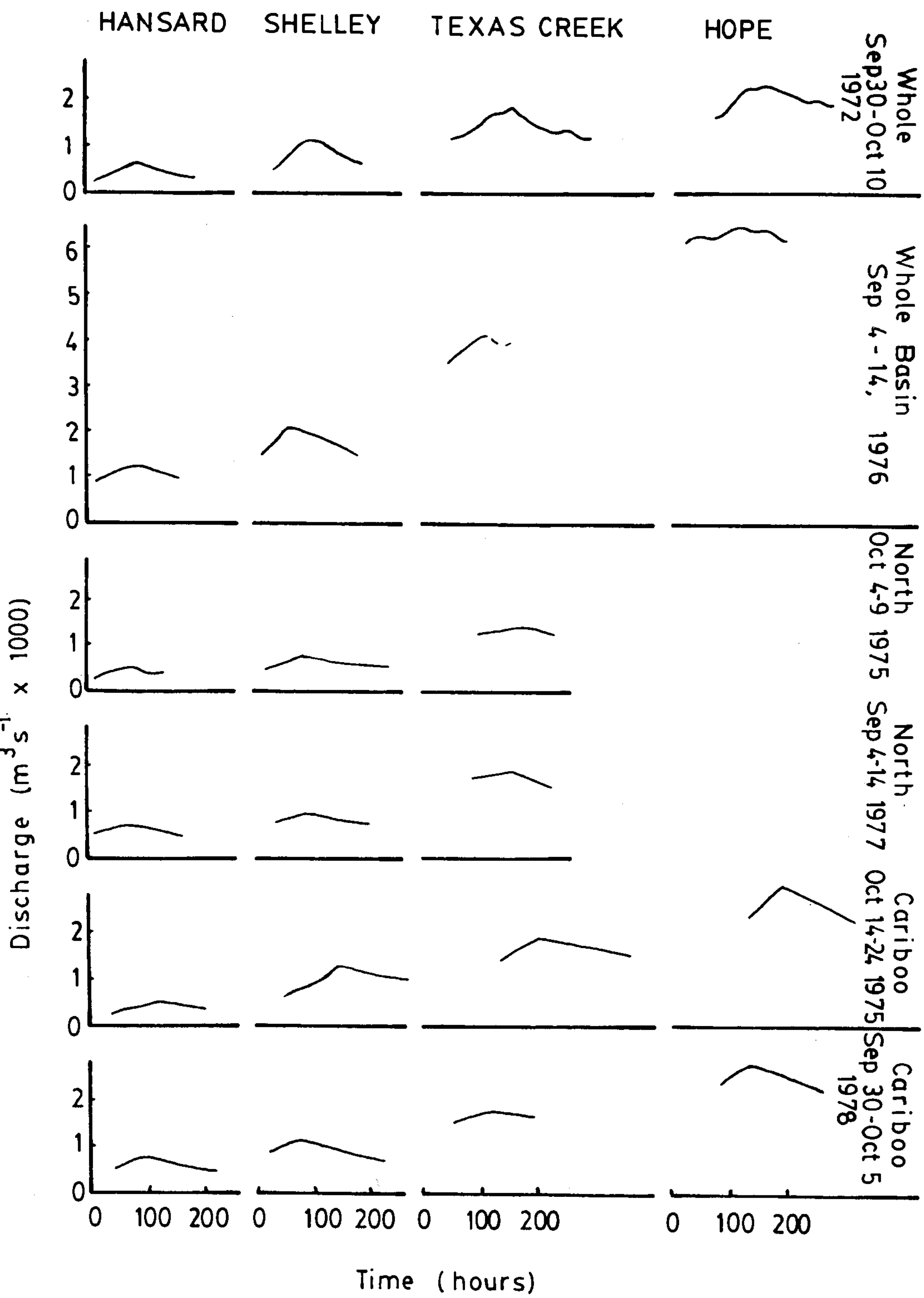


FIGURE 4.2 Fraser River hydrographs

reflects the regulating effect of lake storage on discharge fluctuations. Multi-peaked hydrographs are common for many events on the Fraser River (figure 4.2a and 4.2b). This illustrates the need to identify past and future discharge changes above baseflow, to assess which peaks are discrete.

The Fraser River hydrographs for precipitation over the whole basin showed good similarites to the model hydrographs. For example, the increased volume and broadening of the hydrograph shape from Hansard to Shelley (figure 4.2a and 4.2b), and the earlier time-of-peak in the latter case. The hydrograph for mode I at Texas Creek is very similar to that in figure 4.2a. The model suggests the first peak relates to Bridge river flows, the second peak to inputs along the middle Fraser River and the third blip to upper Fraser flows. The broadening of the main peak at Hope is due to the Thompson and the decreased slope of the upper Fraser flows due to attenuation (figure 4.2a). These three peaks are also visible at Hope in figure 4.2b. In this case the Thompson input was more substantial as evidenced by the larger volume increase.

Fraser River hydrographs from storms over the Cariboo Mountains (figures 4.2e and 4.2f) also followed the model closely. For example at Texas Creek the volume has increased but the peak has experienced attenuation. By Hope, the volume has increased and the shape is more pronounced. The differences between figures 4.2e and 4.2f are understandable since the extent of a Cariboo storm is difficult to determine and the degree of influence on the upper Fraser and Thompson will produce different hydrographs. In all cases however, the model should only be considered as relative.

Continuous discharge data allowed hydrograph shapes to be portrayed accurately, and a more precise evaluation of peak time. Many travel times between adjacent stations are less than 24 hours and therefore daily data do not possess enough sensitivity to determine the time of main peak. It is even more difficult to identify secondary peaks from daily data. Therefore much information is lost at the daily time scale.

4.7. Discharge/Travel Time relationships for selected precipitation source areas

After verifying the zone of precipitation, the basic relations between travel time and discharge were analysed again. Now that the data were split according to precipitation source region, the results were generally good, and implied faster travel times with larger discharges. Often the trends exhibited the curvi-linear form predicted by Linsley, Kohler and Paulhus (1949). For example, the Hansard to Shelley reach and the Shelley to Texas Creek reach with storm centres over the Cariboo mountains and in the north (figures 4.3.2a, 4.3.2b, 4.3.3a and 4.3.3b). This trend can also be seen for Hansard to Shelley with precipitation over the whole or centre of the basin, and to some extent between Texas Creek and Hope with the latter

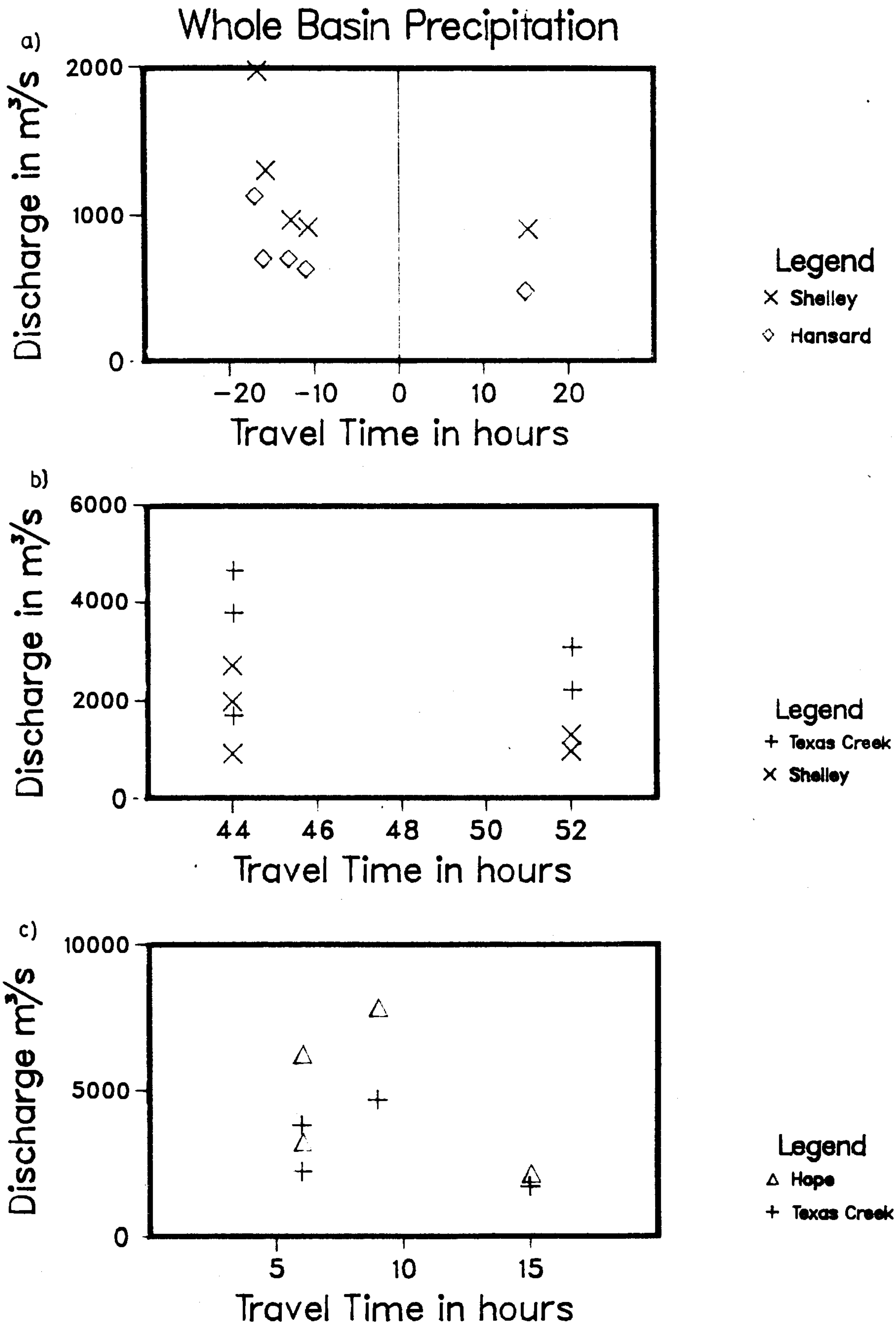


FIGURE 4.3.1 Discharge - travel time relationships by source region of precipitation

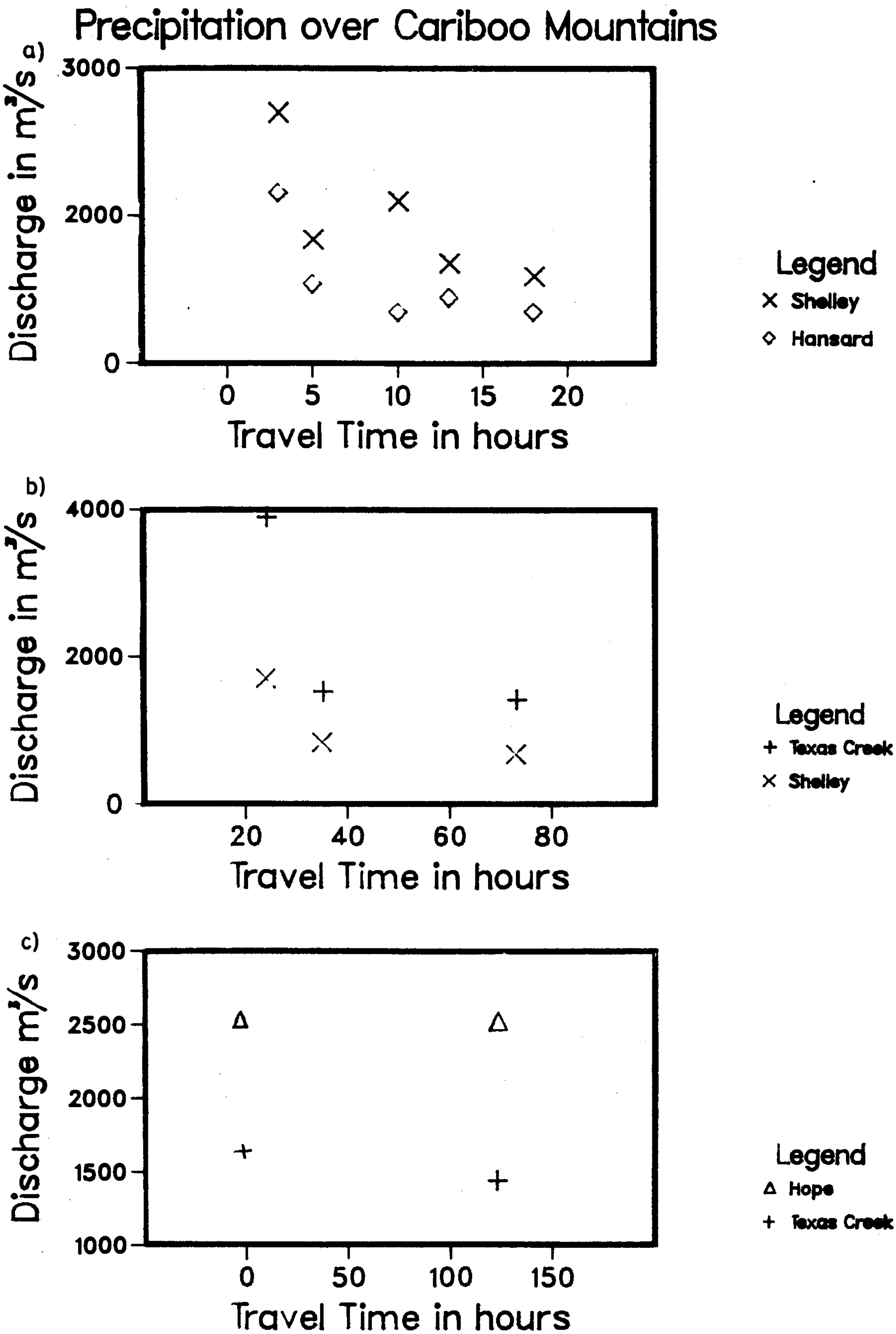


FIGURE 4.3.2 Discharge-travel time relationships by source region of precipitation

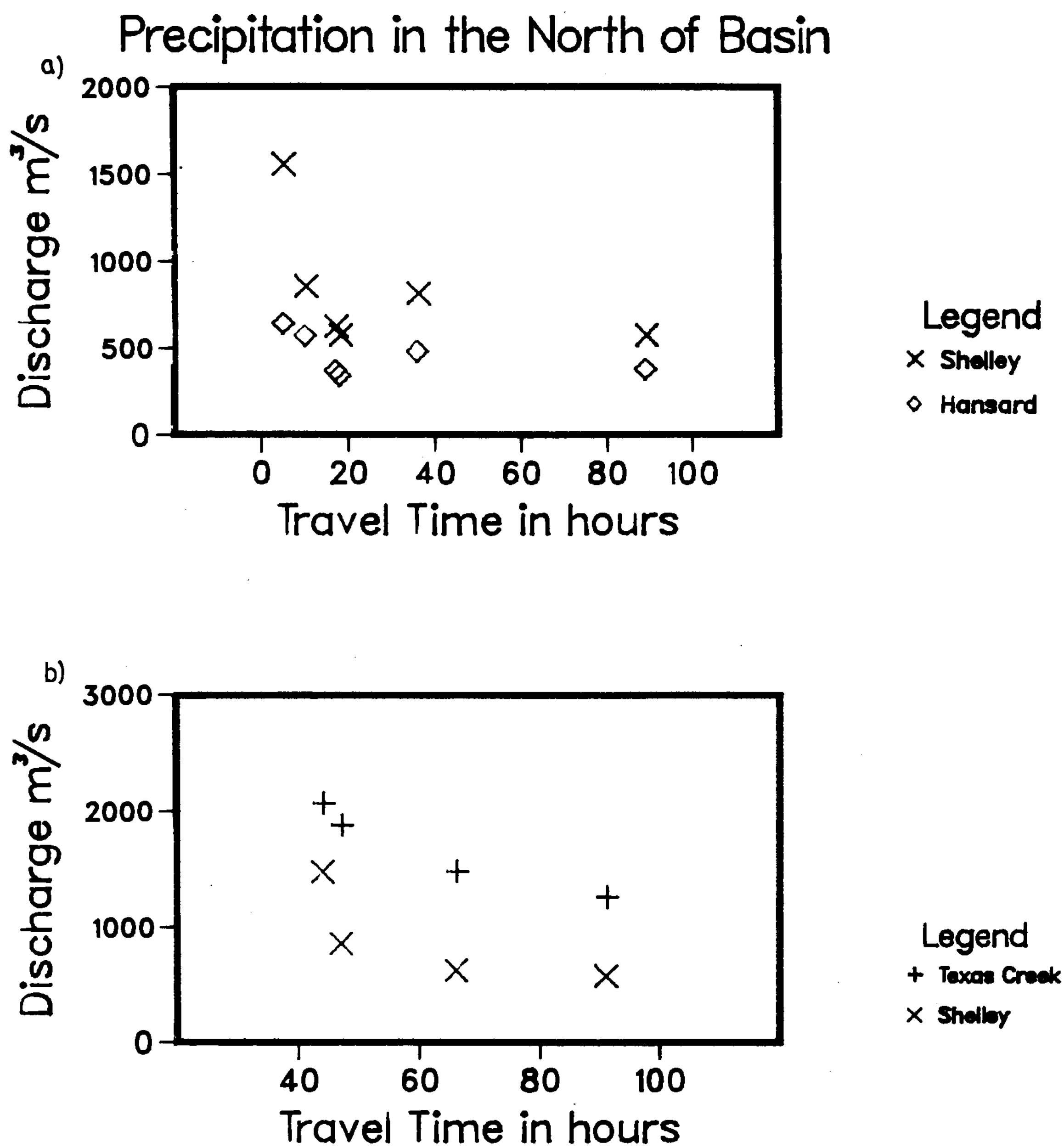


FIGURE 4.3.3 Discharge-travel time relationships by source region of precipitation

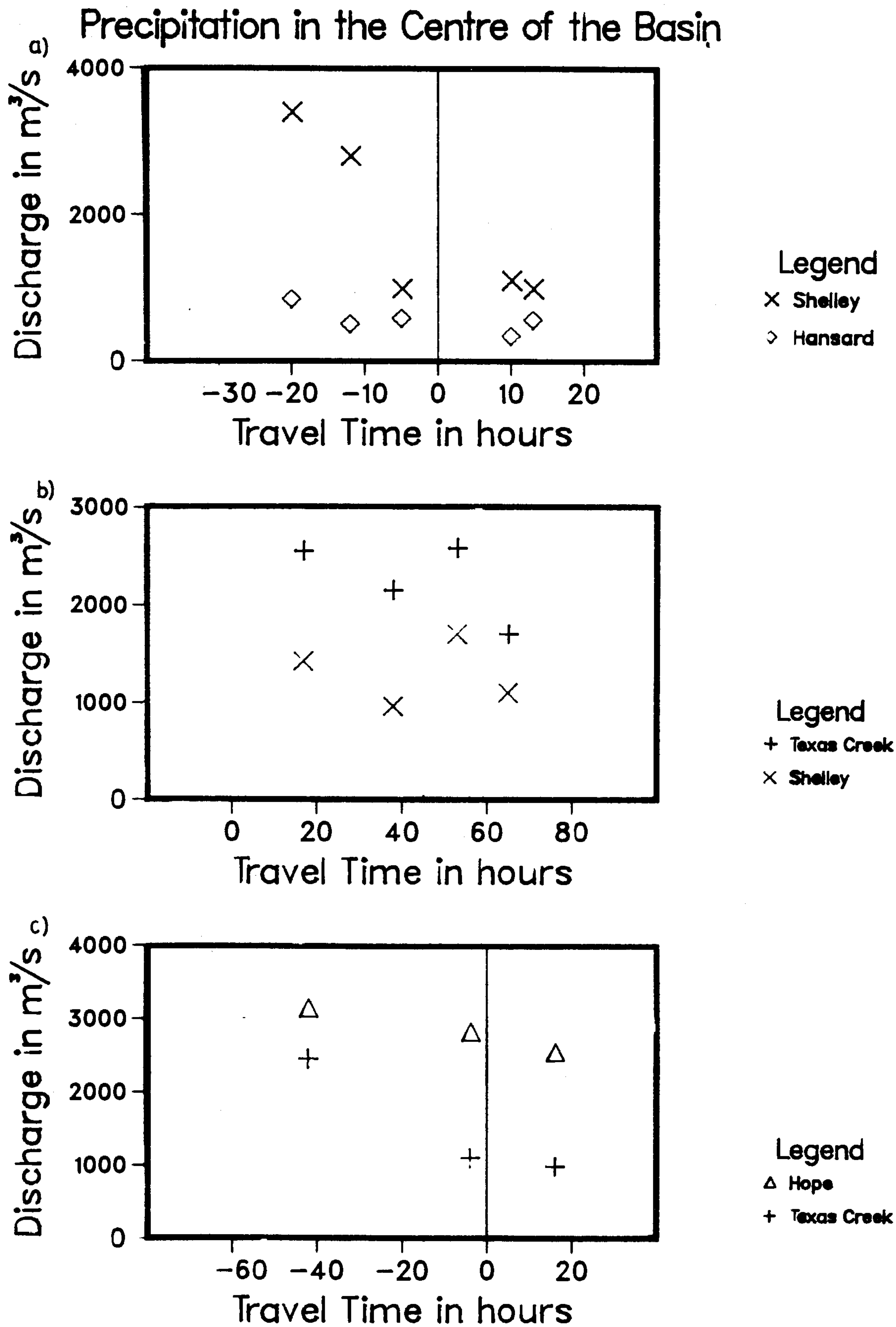


FIGURE 4.3.4 Discharge - travel time relationships by source region of precipitation

precipitation distribution (figures 4.3.1a, 4.3.4a and 4.3.4c). These trends however include "negative" travel times and may be purely coincidental.

When discharges at both upstream and downstream stations on a reach are compared, there seems to be a negative relationship between travel time and the difference between the two discharges. That is, discharge volumes are more likely to be similar volumes when the travel time is long. This is particularly noticeable on the Hansard to Shelley reach and the Shelley and Texas Creek reach (figures 4.3.2a, 4.3.3a, 4.3.4a). However, for precipitation in the centre of the basin, where this trend is most marked (figure 4.3.4a), the curvi-linear trend is also least apparent. Clearly the large additions between Hansard and Shelley allow "negative" travel times and disrupt the discharge/travel time relationship.

For cases where the negative curvi-linear trend is not apparent, the time-area model provides explanations. For example between Texas Creek and Hope with precipitation over the whole or centre of the basin (figure 4.3.1c and 4.3.4c), the relative importance of the Thompson confuses the pattern. "Negative" travel times occur when the Thompson supplies a more important input than the Fraser. Precipitation in the centre of the basin also confuses the discharge/travel time relationship between Shelley and Texas Creek (figure 4.3.4b), due to the relative importance of several tributaries in this reach. It is interesting also that basin-wide precipitation only provided two travel times for the Shelley to Texas Creek reach (figure 4.3.1b). This again suggests travel time is independent of discharge. The cases that do not fit the expected trend provide more evidence that wave progression is interupted when many tributaries are generating inputs. At these times, actual tributary input volumes and the balance between tributaries appear most important in determining absolute peak magnitude and timing.

In contrast to the timing complexities revealed in the original analysis, the most predictable reach is Hansard to Shelley. After eliminating the "noise" created by different spatial distributions of precipitation, the relationship between travel time and discharge for Hansard and Shelley is improved. This can possibly be attributed to both the shortness of the reach and the limited number of tributary inputs. Predictability between Shelley and Texas Creek varies, with some evidence of negative trends between discharge and travel time when a storm is present in the Cariboo mountains or the north of the basin (figure 4.3.2b and 4.3.3b). There are insufficient cases for the Texas Creek to Hope reach to draw good conclusions, but the addition of the Thompson river probably disturbs the shape of the Fraser hydrograph and thus expected travel times.

This section has demonstrated how the timing of tributary contributions can complicate an overall trend between discharge and travel time. Similar mainstream hydrographs are produced for each storm centered in a given area and a relationship between travel time and discharge can be expected, except where substantial tributary inputs can alter the time of peak.

The model developed in chapter 3 is successful in allowing observed hydrograph changes to be interpreted in terms of precipitation receipt/runoff source area. For certain precipitation source regions and reaches, trends between discharge and travel time can be shown to exist. While events with basin-wide precipitation do not readily conform to expected trends, the changes in hydrograph shape through the system allow an assessment of relative tributary inputs. In these cases, a more appropriate analysis would be that of highflow duration: a parameter of equivalent importance in flood damage calculations.

V. Discussion

5.1 Important Controls on Time of Peak

As discussed in chapter three, time of peak is dependent mainly on the time taken by water from the most productive basin zone to reach the main channel via the drainage net. Therefore, the governing factors can be split into two groups.

1. Hydraulics of overland flow.
2. Hydrologic and storm characteristics.

The multitude of interrelations between basin, storm and channel characteristics allow many permutations of the net effect. Langbein et al (1947) made several significant observations in this regard. They surmised that river systems have differing efficiencies as agents for the collection and conduction of water. Further to this, that variations in the supply are more sensitively perceived in the main channel as the hydrograph timebase diminishes.

The time taken for water to reach the basin mouth will be influenced by drainage density. Basins with a high drainage density can respond faster and attain greater flood peak volumes than a basin of equivalent size but lower drainage density. Heerdegen (1973) found these features were exhibited in Unit Hydrographs of contrasting Pennsylvania State rivers. He showed that both high peak discharges and unit hydrographs with a short baselength reflect basins with a combination of small area, short overland travel times, low drainage density and reduced channel sinuosity. Basin shape is inherent in all of these expressions. Five groups of topographical features which influence the hydrograph were listed by Sherman (1932). Shape, together with area and size, was in the primary grouping. After an examination of the similarities between models and the real world however, Black (1972) suggested that watershed eccentricity, rather than shape per se, was a useful expression of basin control on maximum peak flows and certain time parameters of the hydrograph.

Unfortunately, the absolute size and diversity of the Fraser Basin hinders the production of a simple unit hydrograph. For example, the observed hydrograph at Hope is the sum of attenuated inputs from many tributaries and therefore, it is poorly represented by a simple Unit hydrograph of the whole basin. However, unit hydrograph assumptions are more acceptable for individual tributary subbasins. Here, it is feasible to consider the characteristics of the subbasins as being homogeneous, especially as size decreases. In a small basin, a significant percentage of the time-to-peak is spent in overland flow and it is therefore influenced by topographic features. Conversely channel runoff predominates and stream hydraulic characteristics become more prominent in large basins.

At a large scale, studies identifying specific relationships between physiographic characteristics and hydrological parameters can only be fairly crude. For example, Heerdegen and Reich (1973) stated: "time of peak and peak discharge are dependent on area and other associated physiographic parameters". In order to be more specific, investigations must be very small scale. The main reasons for these difficulties is the interdependence between variables. Results from Langbein et al (1947) imply that none of the topographic parameters are unique in their influence on the hydrograph, and that each reflects a condition which also influences the others. Some of this variability may be reduced by using hydrometeorological regions.

Within a given hydrometeorological region, actual discharge is partly a function of basin size. Thereafter, the percentage of precipitation producing runoff indicates the relative importance of storage and pondage within the region. In arid zones with reduced slopes and streams influent to the groundwater system, time to peak is relatively long and peak discharge relatively low. Although great contrasts in climate and topography exist within the Fraser Basin, at the scale of a tributary subbasin, vegetation, climate and physiography are more uniform.

The very irregular shape of the Fraser Basin and numerous tributary inputs along the river, produce a complex hydrograph even before considering the spatial extent and characteristics of individual storms. The sheer size of the basin confounds the possibility of precipitation beginning concurrently throughout and it is plausible that precipitation will not be received everywhere, especially in summer. Also the amounts received at any one place are extremely variable. Therefore, it is feasible for all tributaries to add to the mainstream only under special conditions and at these times, they may be operating on separate timebases. Storm duration also affects the speed of runoff. Weyman (1975) postulated that the factors controlling infiltration are most noticeable for small storms and that multiple rain periods pose extra complications.

Mustonen (1966) cautioned that the quantification of climatic and basin characteristics only produces indices of the combined effects of several physical factors and therefore, it is potentially misleading to use complex statistics. While this is a justifiable assertion, quantitative models inevitably must be complex to allow reasonable results. The U.B.C. model of the Fraser is a good example of this. O'Donnell (1966) suggested that computers can assist in providing solutions but that the results are still limited by current knowledge and the understanding of the processes simulated.

Therefore, the model generated in this study was kept sufficiently simple to provide an approximation of time-of-peak at each major tributary junction and other subsequent points along the Fraser. The use of a time-area method for a precise peak prediction is not advisable however, because countless interactions and daily fluctuations in basin conditions affect travel times and therefore question rigid spacing of isochrones. Alternatively, a time-area model usefully provides an impression of basin responses, when the inherent raw assumptions are appreciated. In flood control it is also useful to be able to identify individual components of a hydrograph (Clark, 1945).

Studies utilising flood routing techniques often produce an accurate prediction of peak discharge but ignore hydrograph shape. The mathematical representation of peak time or peak volume by these methods may hide significant secondary peaks which can be important in determining total downstream responses. The multipeaked nature of the hydrographs generated for one event on the Fraser River clearly shows the importance of this omission. The highest number of peaks per event occur when input from the whole basin is lumped to produce a hydrograph at the lower Fraser River gauging points. While real hydrographs often exhibit more than one peak, they do not show the same number as in the model hydrographs described above. By routing each tributary to its Fraser River confluence and then routing the summation of flows, a more realistic model hydrograph is developed. Clark (1945) acknowledged that his basic method possibly exaggerates the influence of basin shape and the capacity of the basin to produce high peaks. With this constraint in mind, the use of the study model can be evaluated.

5.2 Use of the Time-Area Model to interpret Travel Times

In the Fraser Basin time-area model developed for this study, each subbasin is treated separately but small changes within them do not appear to be visible in hydrographs at Hope. The model allows the incorporation of a channel storage function and therefore includes attenuation. It also gives a reasonably realistic impression of how the tributary flows are synchronised in different circumstances. The model provides a useful representation of time-of-peak and spatial variations in basin response.

It appears that Fraser River peak flow times are independent of flood wave translation and are governed predominately by the time taken for runoff to accumulate in the channel. Climatic and physiographic characteristics are therefore of prime importance. It follows from this that differences between time of peak at two adjacent stations is rarely equal to flood wave velocity. Wave celerity is greater than water velocity (Linsley et al,1958), but this cannot account for the phenomenal velocities implicated in many cases on the Fraser River.

Travel time in the mainstream is related closely to the storage capacity of the open channel (Clark, 1945) contrasting to the hydrologic and basin characteristics governing water collection. The progression of a flood wave along the Fraser River can be observed when limited duration storms are centered in the north of the basin. When the input is more complicated than this, flood wave progression is disturbed. This explains the paucity of relationships found between travel time and discharge, even when classsified by frontal approach and upper air flow direction.

Subbasin contributions may be either synchronised for one storm throughout the basin, or may act independently. The time of arrival of peak flow from each tributary may or may not coincide with the arrival of peaks from the upstream Fraser River. This depends upon storm extent and duration. Variations in arrival times of peaks from a couple of tributaries can significantly alter the shape of the hydrograph at Hope, and consequently, instantaneous time of peak. Once again, the importance of tributary flow interactions dominates and completely masks correlations which could enable travel time predictions.

Sometimes, the estimation of travel times from real data is complicated by the absence of a peak at all stations. The time-area model enables several explanations:

1. Upper Fraser storms localised and effects attenuated at lower stations.
2. Lower localised storm, not experienced in the Upper Fraser.
3. Peak inputs delayed from the source in some regions, e.g. altitude effects upon snow retention.
4. Input from one or more tributaries synchronised with lower flows between peaks on the Fraser, may eliminate substantial

peaks at downstream stations.

5. Several storms over a few days obscure the effects of one. and also,

6. Recorder out of operation.

During other events, apparent negative travel times occur when the downstream peak is earlier than its upstream counterpart. These negative travel times do not appear to be correlated with either fast or slow flows on the adjacent reaches. Between Texas Creek and Hope negative travel times appear more dependent on the relative dominance of the Fraser or Thompson. Because the reason for a missing peak may not be clear, an examination of travel time irregularities using discharge data alone would be highly speculative. Therefore, the time-area model is an invaluable aid for the Fraser Basin.

The model has shown the necessity of understanding the complexity of a river system and also dealing with continuous data. For example, an apparent series of peaks on the recorder charts may be components of only one event, but distributed by time-area inputs. Familiarity with the model and real processes within the system allows an interpretation of the effects of one storm. These interpretretations in turn, permit explanations of the observed travel time irregularities.

5.3 Further Work

The model presented here assists an explanation of peak flow travel time irregularities and offers a better understanding of the Fraser Basin. Unfortunately, it cannot supply accurate predictions. The time-area model could possibly be improved by considering small but important variations. Examples of these include altering channel storage relations in the Fraser River to accomodate backwater effects, especially at the Thompson confluence, and also the inclusion of lateral inflow and minor tributaries. Field monitoring would be necessary to obtain a reasonable simulation, but the quality of the acquired data would then exceed that of the other inputs. Further refinements could assess the role of baseflow components and subsurface flows, but the end result probably would not justify the extra time involved.

An interesting addition could study model sensitivity to changing inputs in terms of timing, volume and extent of basin affected, especially with respect to Hansard and Shelley peak timing. It may be possible to identify several key precipitation stations and relate rainfall characteristics to hydrographs.

The statistical properties of travel times could also be examined. Anderson (1972) proposed that the Erlang distribution gives approximate peak inter-arrival times at stations, and this could be tested on a heterogenous basin like the Fraser. Weymen (1975) questioned if velocities consistently increase downstream as discharge increases. An answer for this could be incorporated usefully in a time-area model.

Precise input-output analysis however, is confounded by insurmountable problems of measurement: not only the determination of volumes in ungauged catchments, but also precipitation characteristics such as intensity and duration. Analysis of interrelationships between variables per se, is a thankless task and the results are unclear. Therefore, determining hydrologic response from geomorphic features is more appropriate. Existing models are deficient due to their linear and/or static nature. The ultimate model would be a process-based catchment model (Weyman, 1975) which would include topographic variables. Before this is attempted however, it is necessary to determine the most significant variables, for example, drainage density.

The relationship between drainage density and the time-area concept needs more thorough investigation. Drainage density is fundamental to the timing and magnitude of runoff, hence it is relevant to time-area ideas. Sokolov (1969) recognised drainage density as being the most important factor characterising conditions of flood flow formation. Current studies emphasise the dynamic nature of the stream network and its spatial variations. Collection of data concerning the dynamicity of the network poses many problems, but remote sensing and imagery should provide a useful source. Not only is a knowledge of network configuration important, but also the channel carrying capacity, that is, the amount of water that can be removed in a given time. This volume is dependent on relief, channel size and resistance, network shape and integration, among others. Gregory (1977) attempted to quantify these factors by producing an index of channel network volume, and subsequently (Gregory, 1979) an index embracing both channel velocity and basin relief.

Velocity is related closely to stream size and thus, a relation to stream order can also be expected. Gupta et al (1980) cited mathematical evidence linking an output hydrograph with channel network geometry. In a similar vein, Rogers (1972) proposed a new drainage basin parameter called Channel Length Frequency Distribution (C.L.F.D.). that has a qualitative relationship to the surface runoff hydrograph. C.L.F.D. is determined by measuring the distances from the head of each first order channel[1] to the basin outlet. Frequency histograms of these distances show similarities to surface runoff hydrographs of general storms that exhibited reasonable variations in precipitation amount and intensity. Although Rogers claimed that the correspondence between C.L.F.D. and hydrograph shape was too consistent to be coincidental, he achieved only partial success in quantifying the method. Nevertheless, similar studies could give more insight into mechanisms controlling peak timing and synchronisation of tributary inputs from contrasting hydrometeorological areas.

---

[1] after Strahler (1964)

The above studies make a first step to unify geomorphic principles with hydrologic response. Information relating these two continues to grow. It seems of particular advantage to pursue these studies in the Fraser Basin and results promise to be exciting. Preliminary studies to link the Instantaneous Unit Hydrograph with geomorphic parameters of a basin are presented by Rodriguez-Iturbe and Valdes (1979) Valdes et al (1979) and Rodriguez-Iturbe et al (1979). This work is based on the premise that the form and shape of a drainage net is governed by geomorphological laws. The concept is extended, suggesting hydrological response is also governed by basic themes which can be related back to geomorphological laws. To do this, general equations expressing the Instantaneous Unit Hydrograph as a function of Horton's numbers[2] are derived. Tests of the model in a real world situation produced very good results especially with regard to estimation of peak discharge and time-of-peak.

Gupta et al (1980) developed a model on similar lines to that of Rodriguez-Iturbe and Valdes. They achieved most success in the larger basins and subsequently questioned the validity of assuming a linear precipitation-runoff transformation. It seems plausible to test this type of model on the Fraser Basin. A slightly less mathematical route similar to Boyd (1978) and Boyd et al (1979) may be just as useful. They developed a relationship between lag time and drainage basin geomorphology using parameters originated by Horton and Strahler to allocate

---

[2]Horton, 1932 and 1945

storages. This eliminates the assumption implicit in the time-area model, that all storage elements are equal. It also removes the problem of interpolating isochrones from limited data. When compared to the real world, the Boyd model allowed reasonable calculations of lag time.

While these studies may at first appear somewhat removed from travel time determination, it is clear that basin geomorphology and its spatial variations can exert an important control upon time-of-peak and therefore apparent travel time. It is suggested that further studies should concentrate on a means of assessing the time distributions of discharge from all subbasins, including those that are ungauged. In particular, the role of drainage density should be assessed because of its direct influence on generation of discharge and peak flows. Drainage density can then be used as the basis for a routing model. The development of predictive relationships between geomorphic expressions, such as channel network geometry and hydrology are of fundamental significance for advancement in this field.

5.4 Conclusions

A simple time-area model based on Clark (1945) has been used to derive a series of hydrographs for the Fraser Basin. The model assumes peak flow to vary directly with maximum width of drainage area. The time-area values were then assigned weights which were derived from mean annual precipitation data. This allowed a representation of the timing of subbasin inputs, and the relative volumes to be expected under ideal conditions. Each tributary input was added to the Fraser River flow, and routed downstream incorporating the effects of attenuation and increased water velocity. The effects of timing and magnitude of tributary inputs relative to mainstream volume, were shown to be critical in determining the time of absolute peak.

The importance of subbasin additions to time of peak on the Fraser River is clear when peak times at two successive points are compared. In many cases, the time difference between peaks at the two stations does not necessarily reflect flood wave travel time. Therefore in a heteorogeneous, irregular-shaped basin, travel times do not behave in a manner contingent upon discharge. The inclusion of broad scale meteorological conditions and direction of storm approach do not clarify the situation. When data concerning the peak alone are extracted from the records, much information concerning hydrograph shape is masked. It is impossible to know if the peaks downstream correspond to peaks translated from upstream, or to large and/or unsynchronised inputs between the gauging stations. To distinguish between these it is necessary either to analyse a series of hydrographs for the same event at different gauging stations, or to determine the influence of storm extent upon Fraser River hydrographs.

The time of arrival of a highflow peak and the actual volume of flow received at this time are intrinsic in flood forecasting. Both variables are affected by the degree of synchronisation of subbasin inputs as shown by the present study. Because all the subbasins can be included, the model can be adapted to assess the effects of stream diversion projects such as the McGregor diversion proposed by the Fraser River Board (1963)[3]. The loss of the McGregor input would result in reduced flows in the Fraser River. Reid, Crowther and Partners (1978) estimated these reductions to be of the order of 28% at Shelley, 13% at Texas Creek and 9% at Hope. Consequently, the relative influence of the other tributaries will be changed. Not only would this affect peak timing and volume, but also thermal regime and the river's capacity to assimilate contaminants. The time-area model can be readily applied to this problem.

While the time-area method has been criticised for its time-consuming and laborious computations, it provides a useful

---

[3]The diversion of the McGregor river to the Arctic drainage was one of nine projects designated as System E by the Fraser River Board for flood protection and Hydro-electric production. Engineering studies on the project were suspended in 1978 by the B.C. Hydro and Power Authority.

explanation of travel time irregularities observed on the Fraser River. The method is not readily adaptable to flood forecasting studies, but it has highlighted the importance of basin geomorphology in determining time of peak, and the need for studies to quantify the spatial distribution and speed of runoff generation within each subbasin.

APPENDIX

Red Pass to Hansard

Date			Time	season	TT	Discharge ($m^3 s^{-1}$)
1971	Sept	8	0700	3	54	4885
	Sept	10	1800	3	45	6116
	Sept	20	1900	3	53	6343
	Oct	8	1500	3	44	4984
1973	May	8	0200	1	30	6230
	May	19	1500	1	35	16339
	May	26	1100	1	26	15291
	June	9	0200	1	42	19411
	June	26	0400	2	38	22115
	July	11	0600	2	3	11001
	July	17	1500	2	37	9330
	Sept	30	1600	3	22	11780
1974	May	27	2100	1	24	8566
	June	4	0600	1	28	12488
	July	2	2200	2	46	18377
	Aug	20	2400	2	78	6230
	Sept	8	0800	3	53	5663
	Sept	10	0800	3	43	6315
	Oct	4	1500	3	52	3058
	Oct	9	1200	3	7	7093
	Oct	19	1500	3	46	18377
1975	May	13	1700	1	37	10435
	June	6	0100	1	67	12983
	June	15	1300	1	34	15234
	June	22	1600	2	36	14045
	July	29	1800	2	38	9996
	Aug	12	1800	2	32	6966
	Aug	23	1800	2	20	9260
	Aug	31	0200	2	76	6343
	Sept	18	1400	3	52	6060
	Oct	19	0700	3	24	6513
1976	May	6	1000	1	4	10647
	May	12	1800	1	2	16962
	June	13	0100	1	65	12134
	June	21	0100	1	46	20445
	July	3	1100	2	31	18377
	July	10	0800	2	29	17032
	July	18	2300	2	40	13620
	Aug	8	1400	2	38	13054
	Aug	18	2100	2	41	14781
	Aug	28	0000	2	71	9458
	Sept	7	1200	3	33	11213
	Oct	2	1500	3	60	4587
	Oct	15	0900	3	38	5833
1977	May	12	1600	1	15	8948
	May	26	1000	1	43	11638
	June	3	1500	1	32	8297

Date			Time	season	TT	Discharge ($m^3 s^{-1}$)
	June	9	2100	1	40	10505
	June	23	1100	2	24	19567
	June	26	1200	2	31	18377
	July	3	1400	2	39	15248
	July	18	1600	2	33	13337
	July	31	0100	2	27	16339
	Aug	14	1800	2	41	12700
	Aug	26	0200	2	34	6966
	Sept	6	2000	3	43	5720
	May	23	1300	1	33	8410
	May	29	2300	1	38	6867
	June	6	2400	1	23	14300
	June	19	1300	2	40	9684
	July	2	1000	2	34	9203
	July	31	1700	2	31	11695
	July	25	1600	2	56	7214
	July	29	1900	2	54	7362
	Aug	16	2000	2	39	6569
	Aug	20	1500	2	40	6768
	Sept	7	0600	3	61	6272
	Sept	17	1200	3	51	5380
	Oct	3	0300	3	52	5876
	July	22	1000	3	81	14300
1975	Sep	7	0800	3	-16	6060
1976	May	26	1600	1	-36	9713
1978	Apr	30	1900	1	-48	4633
1979	June	7	0700	2	-37	24777
	June	18	1200	2	-13	14300

Hansard to Shelley

1973	June	9	0800	1	6	432
	June	26	1600	1	12	504
	July	12	1700	2	35	1654
	July	17	1500	2		1492
	Sept	13	1900	2	36	2085
	Oct	21	2100	3	15	1133
1974	May	28	0000	1	30	2652
	July	3	1000	2	12	2704
	Aug	21	1100	2	11	876
	Sept	12	0100	3	13	964
	Oct	5	0800	3	17	479
	Oct	9	2100	3	9	1384
1975	June	6	0300	1	2	2381
	June	15	2000	1	7	2458
	June	22	1700	1	14	2208
	June	28	2000	1	2	2383
	Aug	13	0000	2	6	1169
	Aug	31	2300	2	21	1058
	Sept	7	0900	3	1	1152

Date			Time	season	TT	Discharge ($m^3 s^{-1}$)
	Sept	19	0300	3	13	665
	Oct	7	0900	3	18	580
	Oct	19	1700	3	10	1124
1976	May	12	2000	1	2	3264
	June	13	1100	1	10	2197
	June	21	1200	1	11	3305
	July	3	1100	2		3107
	July	19	0100	2	2	2025
	Aug	19	0500	2	8	2184
	Aug	30	2200	2	70	1448
1977	May	5	0900	1	50	1873
	May	12	2400	1	8	2447
	May	26	1900	1	9	1427
	June	3	2000	1	5	1804
	June	10	0700	1	10	2993
	June	23	2000	2	9	2652
	June	26	1800	2	6	2246
	July	4	0400	2	14	1926
	July	19	1300	2	21	2560
	July	31	1600	2	15	1999
	Sept	7	1000	3	14	854
	Sept	28	1200	3	17	736
	Oct	4	0000	3	15	640
1978	May	23	1400	1	1	1534
	June	7	0000	1	0	2233
	June	24	0600	2	11	1672
	July	2	1900	2	9	1470
	July	13	0900	2	16	1679
	July	26	1300	2	21	1021
	July	30	1300	2	18	992
	Aug	8	1900	2	38	1077
	Sept	17	1700	3	5	835
	Oct	12	2200	3	8	701
	Oct	25	1200	3	2	954
1979	June	18	2000	1	8	2247
	July	2	1100	1	11	4419
	July	22	1700	2	7	3317
	Sept	12	0400	2	23	1750
1980	May	13	2100	1	0	624
	June	6	1200	1	1	1739
	June	26	0500	2	9	1780
1972	June	2	0600	1	- 9	4331
1973	Apr	28	0500	1	-16	992
	May	7	1600	1	-10	1611
	May	19	1100	1	- 4	1634
	Sep	30	2300	3	- 7	2085
1974	June	4	0300	1	- 3	2534
	June	24	1200	2	- 5	2356
	Oct	18	1000	3	-19	1362
1975	May	6	2000	1	- 3	1096
	May	13	1500	1	- 2	2321

Date			Time	season	TT	Discharge ($m^3 s^{-1}$)
	July	29	0400	2	-14	1623
1976	May	6	2000	1	-26	2650
	July	9	2200	2	-10	2717
	Aug	8	1300	2	- 1	2146
	Sep	6	0500	3	-31	1990
	Oct	1	0400	3	-35	687
	Oct	14	2000	3	-13	1209
1977	Aug	25	1300	2	-13	964
1978	Aug	16	0100	2	-19	1099
	Sep	6	1900	3	-11	908
	Oct	2	2200	2	- 5	983
1979	June	14	0100	2	-10	2692
1980	Apr	30	1600	1	- 2	2330
	May	3	1600	1	- 4	2300
	July	19	1100	2	- 6	2510
	July	24	1700	2	-19	1330
	Aug	18	2100	2	- 1	2559
	Aug	25	1300	2	-15	1410

Shelley to Texas Creek

1971	May	16	1700	1	45	4542
	May	30	0000	1	30	3772
	June	12	1700	1	38	4577
	June	18	0900	1	45	4866
	July	15	0000	1	51	3964
1973	May	20	1700	1	29	4733
	June	11	0500	1	45	3557
	June	28	1300	1	45	4998
	Oct	2	2200	3	47	2594
	Oct	23	2200	3	49	1608
1974	May	10	2300	1	51	4914
	May	29	1800	1	42	4372
	June	6	0300	1	48	4361
	July	4	2300	2	37	4746
	Sep	13	1500	3	38	1975
	Oct	11	1800	3	45	1152
1975	May	15	1500	1	48	3409
	June	8	1800	1	63	3613
	June	17	2400	1	52	3885
	June	30	1500	1	67	3840
	July	31	1500	2	59	2911
	Sep	9	1800	3	57	1975
	Sep	22	0400	3	73	1380
	Oct	11	0400	3	91	1218
	Oct	22	0800	3	63	1747
1976	May	8	0100	1	65	5324
	May	14	0100	1	29	6252
	June	14	1700	1	30	4409
	June	23	0100	1	37	6099
	July	5	0800	2	45	5627

Date			Time	season	TT	Discharge (m^3s^{-1})
	July	20	1100	2	34	4385
	Aug	10	0800	2	43	4722
	Aug	22	0300	2	70	4385
	Aug	23	2100	2	49	3591
	Aug	30	1900	3	44	3840
	Sep	8	0100	3	34	1954
	Oct	16	2100	3	49	2251
1977	May	14	1700	1	41	3806
	June	5	1600	1	44	3103
	June	12	0600	1	47	4180
	June	25	2000	1	48	4225
	July	6	0100	2	45	3432
	July	21	1500	2	50	4078
	Aug	27	1700	2	52	2230
	Sep	9	0900	3	49	1869
1978	May	25	2000	1	54	1900
	June	9	0600	1	54	3432
	June	26	1100	2	53	2956
	July	5	0000	2	53	2798
	July	15	1700	2	56	2973
	Aug	11	1900	2	72	1999
	Oct	4	1500	3	41	1529
1979	Apr	8	2000	1	62	3823
	May	20	1000	1	25	3470
	June	21	2100	1	49	3914
	July	24	1000	2	41	3995
1980	May	15	1800	1	45	3176
	June	8	1400	1	50	2478
	June	21	2300	1	60	3803
	Oct	4	1200	3	52	3145
1976	Aug	30	1900	2	- 3	3591
1977	May	7	0900	1	-14	3375
1978	May	2	1700	1	-22	1450
1979	June	24	0600	2	-27	3800
	Aug	5	0300	2	-20	1340
1980	May	2	1400	1	-26	3320

Texas Creek to Hope

1971	May	17	0300	1	10	7773
1973	May	10	0800	1	14	4106
	May	28	0800	1	14	7092
	June	11	0900	1	4	7357
	Oct	3	0600	3	8	3103
1974	May	11	0200	1	4	7006
	June	6	2300	3	20	7209
	July	5	0500	2	6	8353
1975	May	16	0400	1	13	6802
	June	19	0200	1	26	7685
	July	31	1200	2	21	4644
1976	May	9	0200	1	25	7626

Date			Time	season	TT	Discharge (m^3s^{-1})
	May	14	1900	1	18	9073
	June	23	0700	1	6	9480
	July	6	0000	2	16	9005
	Aug	10	1700	2	9	7209
	Aug	24	1100	2	14	7150
	Sep	8	0700	3	6	5948
	Oct	3	0500	3	15	3009
	Oct	17	0500	3	8	3056
1977	May	15	0100	1	8	5754
	June	12	1100	1	5	6802
	Aug	13	0400	2	18	3730
	Aug	27	2300	2	6	3248
1978	June	9	1300	1	7	6860
1979	May	9	0700	1	11	5182
	May	22	1100	1	49	4618
	June	9	1800	1	12	8232
1973	June	28	1200	1	- 1	5845
1974	May	5	1800	1	- 6	6286
1975	June	8	0300	1	-15	7269
1976	May	27	2000	1	- 7	7269
1977	June	25	1300	2	- 7	6626
1978	June	26	0900	2	- 2	5561
	July	15	1200	2	- 5	5009
1979	June	15	0500	2	-19	6456
	July	4	0400	2	- 4	6230

Key: season 1 = spring
season 2 = summer
season 3 = autumn
TT = travel time in hours

BIBLIOGRAPHY

Alexander, G.N. 1972. Effect of catchment area on flood magnitude. Journal of Hydrology, vol 16, pages 225-240.

Amorocho, J. 1963. Measures of linearity of hydrological systems. Journal of Geophysical Research, vol 68, no.8.

Andersson, M.G. 1976. Peak flow inter-arrival times. Journal of Hydrology, vol 30, pages 105-112.

Barry, R.G. and R.J. Chorley 1974. Atmosphere, weather and climate. Second Edition, 432pp, Methuen, London.

Bell, F.C. and P.C. Vorst, 1981. Geomorphic parameters of representative basins and their hydrological significance. Australian Resources Council. Department of National Develoment and Energy. Research project number 74/53. Technical paper number 58, 134pp.

Black, P.E. 1970. Runoff from watershed models. Water Resources Research, vol 6, part 2, pages 465-477.

Black, P.E. 1972. Hydrograph responses to geomorphic model watershed chararcteristics and precipitation variables. Journal of Hydrology, vol 17, pages 309-329.

Boyd, M.J. 1978. A storage routing model relating drainage basin hydrology and geomorphology. Water Resources Research, vol 14, no.5, pages 921-927.

Boyd, M.J., D.H. Pilgrim and I. Cordery 1979. A storage routing model based on catchment geomorphology. Journal of Hydrology, vol 42, pages 209-230.

Brady, J.A. and P.Johnson 1981. Predicting times of travel, dispersion and peak concentrations of pollution incidents in streams. Journal of Hydrology, vol 53, pages 135-150.

Brakensiek, D.L. 1963. Estimating cooeficients for storage routing. Journal of Geophysical Research, vol 68, pages 6471-6474.

British Columbia Water Investigations Branch. 1975. Summary of snow survey measurements 1935-1975. Water Resources Service. Department of Lands, Forests and Water Resources, Victoria, B.C.

Bruce, J.P. 1964. Large scale atmospheric circulations affecting floods on the Fraser River. Procedings western snow conference. Spokane, Washington. 32nd Annual Meeting, pages 44-51.

Bultot, F. and G.L.Dupriez 1976. Conceptual hydrological model for an average-sized catchment, I. Concepts and relationships. Journal of Hydrology, vol 29, pages 251-272.

Butler, S.S. 1979. Deterministic-systemic relationships in overland flow analysis. Water Resources Bulletin, vol 15, no.2, pages 519-529.

Carlston, C.W.and W.B.Langbein, 1960. Rapid approximation of drainage density: line intersection method. United States Geological Society, Water Resource Division bulletin, page 11. Unpublished information circular.

Chow, V.T. (ed) 1964. Handbook of applied hydrology. McGraw Hill, New York.

Clark, C.O. 1945. Storage and the unit hydrograph. Transactions American Society of Civil Engineers, vol 110, pages 1419-1488.

Crawford, N.H. and R.K. Linsley 1966. The Stanford Watershed Model IV:Stanford University, Department of Civil Engineering, technical report, 39 210pp.

Ding, J.Y. 1974. Variable unit hydrograph. Journal of Hydrology, vol 22, pages 53-69.

Dingman, S.L. 1978. Drainage density and streamflow: a closer look. Water Resources Research, vol 14, part 6, pages 1183-1187.

Dooge, J.C.L. 1959. A general theory of the unit hydrograph. Journal of Geophysical Research, vol 64, no. 1, pages 241-256.

Dooge, J.C.I., W.G. Strupczewski and J.J. Napiorkowski. 1982. Hydrodynamic derivation of storage parameters of the Muskingum model. Journal of Hydrology, volume 54, pages 371-387.

Engineering Division 1972. Spillway design floods for the Fraser River Basin. Water planning and management branch, Environment Canada, Ottawa.

Farley, A.L. 1979. Atlas of British Columbia. University of British Columbia Press, Vancouver, B.C.

Fraser River Board 1963. Final report of the Fraser River Board on flood control and hydro-electric power in the Fraser River Basin. Victoria, B.C.

Gardiner, V. 1979. Estimation of drainage density form topological variables. _Water Resources Research_, vol 15, pages 909-917.

Gergov, G 1971. Determination of the time of travel along the river network. _Journal of Hydrology_, vol 14, pages 293-306.

Glennie, J.F. 1963. _A hydrological study of the 1948 flood in the Stuart River Basin, British Columbia_. University of British Columbia.

Gray, D.M. 1961. Interrelationships of watershed characteristics. _Journal of Geophysical Research_, vol 66, no 4, pages 1215-1223.

Gray, D.M. 1965. Physiographic characteristics and the runoff pattern. in: _Research watersheds, proceedings of the hydrology symposium, 4_. National Research Council of Canada pages 147-164.

Gray, D.M. (ed) 1970. _Handbook on the principles of hydrology_. National Research Council of Canada

Gregory, K.J. 1977. Stream network volume: an index of channel morphometry. _Bulletin of the Geological Society of America_, volume 88, pages 1075-1080.

Gregory, K.J. 1979. Drainage network power. _Water Resources Research_, volume 15, pages 775-777.

Gregory, K.J. and V.Gardiner, 1975. Drainage density and climate. _Zeitschrift fur Geomorphologie_, vol 19, pages 287-298.

Gregory, K.J. and D.E.Walling, 1968. The variation of drainage density within a catchment. _International Association of Hydrological Sciences Bulletin_, volume 13, pages 61-68.

Gupta, V.K., E.Waymire and C.T. Wang. 1980. A representation of an instantaneous unit hydrograph from geomorphology. _Water Resources Research_, vol 16, no.5, pages 855-862.

Gupta, R.P., A.P.Duggal, S.V.Rao, G.Sankar and B.B.S.Singhal 1982. Snow cover area vs. snowmelt runoff relation and its dependence on geomorphology - a study from the Beas catchment (Himalayas, India). _Journal of Hydrology_, vol 58 pages 325-339.

Harr, R.D. 1981. Some characteristics of snowmelt during rainfall in Western Oregon. _Journal of Hydrology_, vol 53 pages 277-304.

Hayami, S. 1951. On the propagation of flood waves. Bulletin number 1, Disaster Prevention Research Institute, Kyoto University, Japan.

Heerdegen, R.G. 1973. The effects of physiography and climate on unit hydrographs. MSc Thesis. Pennsylvania State University , Dept. of Civil Engineering.

Heerdegen, R.G. and Reich, B.M. 1974. Unit hydrographs for catchments of different sizes and dissimilar regions. Journal of Hydrology, vol 22, pages 143-153.

Henderson, F.M. 1966. Open channel flow. Macmillan, New York.

Hewlett,J.D. and A.R.Hibbert, 1965. Factors affecting the response of small watersheds to precipitation in humid areas. International symposium on forest hydrology, pages 275-290, Pennsylvania State University. Pergamon Press, Oxford.

Holland, S.S. 1964. Landforms of British Columbia. British Columbia Department of Mines and Petroleum Resources, Bulletin 48.

Horton, R.E. 1932. Drainage basin characteristics. Transactions of the American Geophysical Union, vol 13, pages 350-361.

Horton, R.E. 1945. Erosional development of streams and their drainage basins: hydrophysical approach to quantitative morphology. Bulletin of the Geological Society of America, vol 56, pages 275-370.

Howard, C.D.D. and D.I. Smith 1977. Floods on unguaged watersheds. Canadian hydrology symposium. Edmonton, Alberta, Aug 29-31, 1977.

Hoyt, W,G, and W.B. Langbein 1955. Floods. Princeton University Press. Princeton, New Jersey.

Huthman, G. and K.Wilke 1982. Time and frequency domain analyses of hydrological systems. Journal of Hydrology, vol 55, pages 347-369.

Ingledow, T. and associates limited 1969. Hydrometric network study. 2 volumes. Government of Canada, Department of Energy, Mines and Resources.

Kirkby, M.J. 1978. Hillslope hydrology. Wiley Interscience Pub.

Kreuder,W.L. 1979. Development of hydrometric networks in British Columbia. Water Resources Research, vol 15, no. 6, pages 1677-1686.

Kousis, A. 1978. Theoretical estimation of flood routing parameters. Journal of the Hydraulics Division, American Society of Civil Engineers,, vol 104, no HY1, proceedings paper 13456, Jan. 1978, pages 109-115.

Kulandaiswamy, V.C. 1964. A basic study of the rainfall excess surface runoff relation in a basin system. PhD Thesis, Department of Civil Engineering, University of Illinois, Urbana, Illinois.

Langbein,W.B. 1947. Topographic characteristics of drainage basins. United States Geological Survey, Water supply paper, 986-C.

Laurenson, E.M. 1964. A catchment storage model for runoff Routing. Journal of Hydrology, vol 2, pages 141-163.

Leith, R.M. 1976. Streamflow regionalization in British Columbia number 2. regression of mean annual flows on physiographic parameters. Inland Waters Directorate report series 46, Environment Canada.

Lighthill, M.J. and G.B. Whitham, 1955. On kinematic waves: 1. Flood movement in long rivers Proceedings of the Royal Society of London, series A, vol 229, pages 281-316.

Linsley, R.K., M.A. Kohler and J.L.H. Paulhus 1949. Applied hydrology. McGraw-Hill, New York.

Linsley, R.K., M.A. Kohler and J.L.H. Paulhus 1958. Hydrology for Engineers. Second edition, McGraw-Hill, New York.

Loijens, H.S. 1972. Snow distribution in an alpine watershed of the Rocky Mountains, Canada. in: World Meteorological Organisation, Distribution of precipitation in mountainous areas. Geilo symposium, Norway, 31 Jul- 5 Aug. W.M.O. no. 326, vol 1, pages 175-183.

Manderville, A.N. and T.O'Donnell 1973. Introduction of time variance to linear conceptual catchment models. Water Resources Research, vol 9, pages 298-310.

March, W.J., J.R.Wallace and L.W.Swift 1979. An investigation into the effects of storm type on precipitation in a small mountain watershed. Water Resources Research, vol 15, part 2, pages 298-304.

Mark, D.M. 1974. Line intersection method for estimating drainage density. Geology, volume 2, pages 235-236.

Mawdsley, J.A. and A.F.Tagg 1979. Identification of unit hydrographs from multi-event analysis. Journal of Hydrology, vol 49, pages 315-327.

McCarthy, G.T. 1938. _The unit hydrograph and flood routing_. Unpublished manuscript presented at a conference of the North Atlantic Division United States Army Corps of Engineers, June 24.

McCoy, R.M. 1971. Rapid measurement of drainage density. _Bulletin of the Geological Society of America_, volume 82, pages 757-762.

Melton,M.A. 1957. _An analysis of the relations among elements of climate, surface properties and geomorphology_. Office of Naval Research, Geographical Branch, Project 389-042, Technical Report II, Columbia University.

More, R.J. 1969. The basin hydrological cycle. in: Chorley R.J.(ed) 1969. _Water, Earth and Man._ Methuen, London..

Murphey, J.B., D.E.Wallace and L.J. Lane 1977. Geomorphic parameters predict hydrograph characteristics in the Southwest. _Water Resources Bulletin_, vol 13, no.1, pages 25-38.

Mustonen, S.E. 1966. Effects of climatological and basin characteristics on annual runoff. _Water Resources Research_, vol 3, pages 123-130.

Nash, J.E. 1957. The form of the Instantaneous Unit Hydrograph. _Journal of Geophysical Research_, vol 64, no.1, pages 241-256.

National Environmental Research Council 1975. _Flood studies report_ in five volumes. Whitefriars Press ltd. England.

Němec, J. 1972, _Engineering hydrology_. McGraw-Hill.

Obedkoff, W. 1970. _A study of the grid square method for estimating annual runoff_. M.A.Sc Thesis. Department of Civil Engineering, University of British Columbia.

O'Donnell, T. 1966. Computer evaluation of catchment behaviour and parameters significant in hydrology. _River flood hydrology proceedings of the symposium of the Institute of Civil Engineers. Session C_. Part 5. March 18, 1965, London.

Ord, K. and M.Rees 1979. Spatial processes: recent developments with applications to hydrology. in: Lloyd, E.H., T.O'Donnell and J.C.Wilkinson _The mathematics of hydrology and water resources_. 138 pages. Academic Press, London.

Overton, D.E. 1966. Muskingum flood routing of upland streamflow. _Journal of Hydrology_, vol 4, pages 185-200.

Pilgrim, D.H. 1976. Travel times and nonlinearity of flood Runoff from tracer measurements in a small watershed. Water Resources Research, vol 12, no.3, pages 487-496.

Pilgrim, D.H. 1977. Isochrones of travel time and distribution of flood storage from a tracer study on a small watershed. Water Resources Research, vcl 13. no.3, pages 587-595.

Pilgrim, D.E., I. Cordery and E.C.Baron 1982. Effects of catchment size on runoff relationships. Journal of Hydrology, vol 58, pages 205-221.

Pilon, P.J. and L.C. Cheng. 1983. Documentation of flow estimation, analysis and routing package version I. FEAR I. Water Resources Branch, Inland Waters Directorate, Environment Canada, Ottawa.

Pollack, D.M. and D.Bellows 1972 Meteorological conditions associated with extreme flows in the Upper Fraser River. Atmospheric Environment Service, Canada.

Ponce, V.M. and V.Yevjevich. 1978. Muskingum-Cunge method with variable parameters. Journal of the Hydraulics Division, American Society of Civil Engineers, volume 104, HY12, pages 1663-1667.

Quick, M.C. 1965. A Hydrometeorological study of the Fraser Basin. University of British Columbia.

Quick, M.C. and A. Pipes 1972. Daily and seasonal forecasting with a water budget model. in: Role of snow and ice in hydrology. Proceedings of UNESCO/WMO/IAHS symposium, Banff, Sep. 1972. IAHS publication no. 106, pages 1017-1034.

Quick, M.C. and A. Pipes 1975. Nonlinear channel routing by computer. American Society of Civil Engineers, Journal of the Hydraulics Division, vol 101, no.HY6, pages 651-664.

Quick, M.C. and A. Pipes 1976a. A combined snowmelt and runoff model. Canadian Journal of Civil Engineering vol 3, no.3, pages 449-460.

Quick, M.C. and A.Pipes 1976b. Nonlinear channel routing by computer. (Discussion closure). American Society of Civil Engineers, Journal of the Hydraulics Division, vol 102, no.HY11, pages 1699-1701.

Quick, M.C. and A. Pipes 1977. University of British Columbia watershed model. Hydrological Sciences Bulletin vol 22, no.1, pages 153-161.

Ragan, R.M. 1966. Laboratory evaluation of a numerical flood routing technique for channels subject to lateral inflows. Water Resources Research, vol 2 no.1, pages 111-121.

Reed, D.W. P.Johnson and J.M.Firth 1975. A non-linear rainfall-runoff model; providing for variable lag time. Journal of Hydrology, vol 25, pages 295-305.

Reid, Crowther and Partners Ltd., 1978. Report on environmental studies for the McGregor diversion projects. 4 volumes for British Columbia Hydro and Power Authority, Vancouver, B.C.

Renard, K.G. and R.V. Keppel 1966. Hydrology of ephemeral streams in the Southwest. American Society of Civil Engineers, Journal of the Hydraulics Division, vol 92, no. HY2, pages 597-646.

Richards, K.S. 1979. Prediction of drainage density from surrogate Measures. Water Resources Research, volume 15, number 2, pages 435-442.

Riehl, H. 1965. Introduction to the atmosphere, 365 pages, McGraw Hill, New York.

Roberts, M.C. and P.C. Klingeman 1970. The influence of landform and precipitation parameters on flood hydrographs. Journal of Hydrology, vol 11, pages 393-411.

Rodda, J.C. 1969. The flood hydrograph. Chorley R.J. (ed) 1969. Water, Earth and Man. Methuen, London. pages 405-418.

Rodriguez-Iturbe, I and J.B. Valdes, 1979. The geomorphologic structure of the hydrologic response. Water Resources Research, vol 15, no.6, pages 1409-1420.

Rodriguez-Iturbe, I G. Devoto and J.B. Valdes, 1979. Discharge response analysis and hydrologic simililarity: The interrelation between the geomporphological Instantaneous Unit Hydrograph and the storm characteristics. Water Resources Research, vol 15, no.6, pages 1435-1444.

Rogers, W.F. 1972. New concept in hydrograph analysis. Water Resources Research, vol 8, no.4, pages 973-981.

Ryder, J.M. 1981. Geomorphology of the southern part of the Coast Mountains of British Columbia. Zeitschrift fur Geomorphologie, vol 37, pages 120-147.

Schermerhorn, V.P. 1969. Relations between topography and annual precipitation in Western Oregon and Washington. Water Resources Research, vol 3, part 3, pages 707-711.

Sherman, L.K.1932. Streamflow from rainfall and the unit hydrograph method. Engineering News Record, vol 108, pages 501-505.

Singh, V.P. 1975. Derivation of time of concentration. Journal of Hydrology, vol 30, pages 147-165.

Singh, V.P. and D.A.Woolhiser 1976. Sensitivity of linear surface runoff models to input errors. Journal of Hydrology, vol 29 pages 243-249.

Singh, V.P. and R.C.McCann 1980. Some notes on Muskingum method of flood routing. Journal of Hydrology, vol 48, pages 343-361.

Slaymaker, H.O. 1972a. Physiography and hydrology of six river basins, in: J. Lewis Robinson (ed) British Columbia, pages 32-68.

Slaymaker, H.O. 1972b. Recent fluctuations of the Fraser River, in: Leigh (ed) Contemporary geographic research trends. British Columbia Geographical Series, no.16, pages 3-13.

Slaymaker H.O. and H.J.MacPherson 1977. An overview of geomorphic processes in the Canadian Cordillera. Zeitschrift fur Geomorphologie, vol 21, part 2, pages 169-186.

Snyder, W.M., W.C.Mills and J.C.Stephens 1970. A method of derivation of nonconstant watershed response functions. Water Resources Research, vol 6, part 1, pages 261-274.

Sokolov, A.A. 1969. Interrelationships between geomorphological characteristics of a drainage basin and the stream. Soviet Hydrology , Selected papers 1, pages 16-22.

Stall, J.B. and D.W.Hiestand 1969. Provisional time-of-travel for Illinois streams. State of Illinois, Department of Registration and Education, report of investigation, 63. Urbana, Illinois.

Strahler, A.N. 1964. Quantitative geomorphology of drainage basins and channel networks, in: Chow, V.T. (ed) Handbook of applied hydrology 1964, pages 4.39-4.76.

Strupczewski, W. and Z.Kundzewicz 1980. Muskingum method revisited. Journal of Hydrology, vol 48, pages 327-342.

Strupczewski, W and Z. Kundzewicz. 1981. Translatory characteristics of the Muskingum method of flood routing - comments to reply. Journal of Hydrology, vol 52, pages 177-179.

Valdes, J.B., Y.Fiallo and I.Rodriguez-Iturbe 1979. A rainfall-runoff analysis of the geomcrphological Instantaneous Unit Hydrograph. Water Resources Research, vol 15, no.6, pages 1421-1434.

Viessman, W.jr., J.W. Knapp, G.L. Lewis and T.E. Harbaugh 1977. Introduction to hydrology. Secend edition, Harper and Row, New York.

Vorst, P and F.C.Bell 1977. Catchment geomorphology and its hydrological significance: a review. The Australian representative basins program report series. Report no. 2. Australian Water Resources council.

Walker, E.R. 1961. A synoptic climatology for parts of the Western Cordillera. Publication in meteorology no.35, science report no.8, Arctic Meteorological Research Group, McGill University.

Wallis, J.H. 1963 Precipitaticn of the Fraser River Basin: A descriptive study. M.A. Thesis, Department of Geography, University of British Columbia.

Wang, C.T., V.K.Gupta and E.Waymire 1981. A geomcrphologic synthesis of nonlinearity in surface runoff. Water Resources Research, vol 17, part 3, pages 545-554.

Waylen, P.R. 1981. Analysis of highflows in the Fraser River catchment, British Columbia. Unpublished PhD Thesis, McMaster University.

Waylen, P.R. and M.K.Woo 1983. Stochastic analysis of high flows in some central British Columbia Rivers, Canadian Journal of Civil Engineers, vol 10, part 2, pages 205-213.

Weymen, D.W. 1975. Runoff processes and streamflow modelling. Theory and practice in geography series. Oxford University Press.

Wilson, C.B., J.B.Valdes and I.Rodriguez-Iturbe 1979. On the influence of the spatial distribution of rainfall in storm runoff. Water Resources Research, vol 15, part 2, pages 321-328.

Woo, M.K. and H.O.Slaymaker 1975. Alpine streamflow response to variable snowpack thickness and extent. Geographiska Annaler 57, Series A, pages 201-211.

Wooding, R.A. 1965. A Hydraulic model for the catchment-stream problem Journal of Hydrolcgy, vol 3, pages 254-267.