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Stream Water and Chloride Generation in a Small Forested Catchment in South Western Australia

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STREAM WATER AND CHLORIDE GENERATION

IN A SMALL FORESTED CATCHMENT

IN SOUTH WESTERN AUSTRALIA

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ABSTRACT

Stream water and salt generation were investigated in a 82 ha, eucalypt forested catchment (Salmon) of deeply weathered, kaolinitic subsoils in the higher rainfall zone (1150 mm yr^{-1}) of the Darling Range in south-western Australia. Streamflow averaged only 11% of rainfall over ten years of below average rainfall. Evapotranspiration was approximately 90% of the catchment water balance of which interception was estimated to be 16% of rainfall. There was a net loss of water from the catchment groundwater over the ten years.

Less than 5% of streamflow is produced by direct rainfall-runoff (overland flow), averaged over ten years. Streamflow generation in Salmon catchment is therefore subsurface and predominantly as discharge from seasonal groundwater developed as perched aquifers in the duplex soils. Sources of streamflow are located in or close to the streamzone. The seasonal groundwaters produce saturated surface soils which function as source areas for direct rainfall-runoff and generate peak stream discharge during storm-period runoff. A maximum observed source area of 6% of the catchment was considered to be of a relatively infrequent occurrence. Discharge from the shallow, seasonal groundwaters was estimated to contribute more than 60% of streamflow volume during storm runoff. Rapid and large increases of water levels in shallow bores, combined with numerous soil 'pipes' enabling quick transmission through the shallow soils, was suggested as an explanation for this result.

The catchment chloride budget indicated a net loss of chloride in streamflow over ten years relative to input measured in gauges in well exposed clearings in the forest. Chloride fall in gauges installed under forest canopy was on average 25 kg ha^{-1} yr^{-1} greater and was sufficient to balance the catchment chloride budget. The source of the higher under-canopy chloride yield was not determined but is likely to be a process of cycling of chloride by the vegetation and therefore does not represent a net, additional input.

A permanent, more saline groundwater, deeper in the kaolinitic subsoils, was calculated to contribute little, if any, water to streamflow but to contribute 60 to 85 kg ha^{-1} yr^{-1} of chloride.

The sub-catchment water and chloride yields and the location of streamflow sources were found to correspond with hydrologic zones identified on the basis of topography, soils and vegetation mapping.

Models of sources of stream water and chloride were assessed. A model based on an exponential leaching of chloride was found to reproduce the yearly catchment water and chloride responses better than a linear-sum of source components model.

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1. INTRODUCTION

1.1 Streamflow Salinity

Salinization of previously arable land and consequently of streamflow is now a major environmental problem in Southern Australia. Approximately $5 \times 10^9\ m^3\ yr^{-1}$ of streamflow in Southern Australia now has an average salinity (expressed as total soluble salts, TSS) in excess of 1000 milligrams per litre (mgL^{-1}) (Peck et al., 1983).

Sadler and Williams (1981) have estimated that one third of the potential water yield from surface sources in the south west of Western Australia is already brackish or saline (greater than 1000 mgL^{-1} TSS). More significantly, a large fraction of the resource in the fresh to marginal quality (approximately 48% of the total resource) is at risk of salinization as a result of land use changes.

1.2 Land Use and Streamflow Salinization

Extensive clearing of indigenous eucalypt vegetation for dryland agriculture is now largely acknowledged to be responsible for the significant increases in streamflow salinization in South Western Australia (Sharma and Williamson, 1984). Clearing alters the water and salt balances across the landscape, resulting in increased groundwater discharge and mobilization of large stores of soil salts to streamflow.

1.3 Prediction and Management

The development of an understanding of the land and stream salinization problem in the south west has proceeded through identification and quantification of the important mechanisms

(Sharma and Williamson, 1984). Predictions of the effects of clearing forests for agriculture on stream water quality and quantity at a regional, water resource scale (e.g. Loh and Stokes, 1981) have been largely based on this knowledge. In turn, broadscale catchment management strategies have been developed (Sadler and Williams, 1981) to protect water resources with the result that legislative controls on clearing, and catchment rehabilitation by partial reforestation, have been implemented.

Improved management, particularly of more subtle hydrologic disturbances, such as silvicultural thinning, will require better predictive methods of hydrologic effects. The development of such methods (models) will require a more detailed knowledge of hydrologic processes at smaller time and space scales within catchments (Beven and O'Connell, 1982).

1.4 Study Background

Catchment water and salt budgets are being studied in the south west of Western Australia (e.g. Williamson and Bettenay, 1979) as part of a research effort to understand landscape influences on hydrology. More detailed observations of components (e.g. Sharma, 1983) and mechanisms (e.g. Johnson, et al., 1983) of the hydrological system have been made. Parts of the system are now well understood, at least conceptually (Sharma and Williamson, 1984); most notably in the context of the likely consequences of clearing for agriculture.

Relatively few models have been developed based on the understanding of this hydrological system. Peck (1976), Peck et al., (1977) and Loh and Stokes (1981) have developed fairly broad scale models. In contrast Smith and Hebbert (1983) formulated a detailed, mathematical model for simulating hillslope processes to which a salinity model was later added by Edgeloe (1982). A conceptual, catchment water and salt balance model, based on similar principles to the Sacramento model (Burnash et al., 1973) was developed by Mauger et al., 1985 (G. Mauger, pers. com.) for application to the Darling Range area of Western Australia.

In many of the experimental studies (e.g. Batini et al., 1977) and certainly in the modelling, there is an awareness of the important role of the seasonal, perched groundwater which develops in the near-surface soils. However there is little information on the characteristics of this system, particularly the interaction of the shallow groundwater with source areas and surface runoff, the source of salt from the deeper groundwater and the characteristics of stream water and salt yields.

1.5 Study Aims

The major objective of the study is to identify and characterise the important processes in the generation of stream water and solute for a small, forested catchment in the Darling Range of Western Australia.

A conceptual model of the sources of water and solute in streamflow is adopted. This model visualises the catchment as a representative hillslope section with three interacting sources of streamflow:-

i) direct rainfall-runoff
ii) seasonal, shallow (near surface) groundwater
iii) permanent, deeper, more saline groundwater.

The specific aims of the study are to:-

i) assess the spatial and temporal source variations of water and solute

ii) determine the magnitude and dynamics of sources of streamflow

iii) test the conceptual, hillslope model and the applicability of models of stream water and salt sources.

1.6 Study Outline

A review of forested catchment streamflow generation mechanisms and catchment solute budgets are presented in Chapter 2. Some simple solute leaching theory and the separation of composite stream hydrographs using chemical and isotope techniques are outlined. Finally in Chapter 2, some catchment hydrology and process studies in the Darling Range are reviewed and a conceptual hillslope streamflow source model is presented.

The physical, environmental and instrumentation details of the study site, Salmon Catchment, are described in Chapter 3.

In Chapter 4, ten years of catchment water and chloride input and output data are analysed. The important water and chloride components and budgets are calculated and discussed.

The within-catchment sources of stream water and chloride are presented in Chapter 5. Variations in the type, magnitude and variability of these sources are described as a guide to subsequent, more intensive investigations.

An analysis of some information from the deeper groundwaters in Salmon Catchment is undertaken in Chapter 6 to develop a simple groundwater flow system as a guide to the possible contribution of this groundwater to streamflow.

In Chapter 7 some results of a study of part of the shallow groundwaters in the catchment headwater are presented. The occurrence of these groundwaters and their relationships to areas of saturated surface soil conditions (source areas) and stream discharge are discussed.

The significance of direct rainfall runoff is assessed in Chapter 8 by separating stream hydrographs using a simple geometric method and a chemical method. The seasonal and yearly water and chloride responses are then presented and two source models of stream water and chloride are analysed. Finally, the approximate proportions of catchment source contributions to streamflow are estimated.

The study results are summarized in Chapter 9, conclusions drawn and recommendations for field investigations and modelling are made.

2. REVIEW OF CATCHMENT HYDROLOGY

2.1 Outline

Some of the important streamflow generation mechanisms are described in this Chapter and a review of the evidence for their significance in forested catchments is presented. In particular the importance of subsurface processes in the generation of streamflow and the role of preferred pathways through soils is noted.

A review of some basic catchment input to output salt balances is undertaken and a simple model for salt leaching is presented. The separation of composite stream hydrographs by chemical and isotopes is discussed.

Finally a conceptual hillslope hydrological model for some forested catchments in the south west is introduced and a review of catchment hydrology and process studies in the south west is presented.

2.2 Streamflow Concepts

The generation of streamflow from a catchment is a complex process involving different processes and, as noted by Pilgrim et al (1978),: "several ... processes or groups of processes predominate on different watersheds..." and "... it is also probable that different processes are dominant on a given watershed at various times". Andersson (1984) observed that: "The spatial variation or structure of a catchment is enormous, which in combination with climatic patterns give rise to an infinite variation in the pathways and residence time of water through and in the soil". This complexity results in variations in streamflow chemistry.

Sklash and Farvolden (1979) attempted to clarify some of the terminology in relation to stream chemistry by developing categories of streamflow generation. These are shown in Figure 2.1. The time aspects relate to whether water is 'new' (added during rainfall) or old (existing in soil water before rainfall). An alternative is to describe water as being from groundwater, the unsaturated zone, direct rainfall-runoff or a combination of these in a historical context. Sklash and Farvolden also presented a list of six ultimate delivery mechanisms of water to streams. Some of these are described in the next section as they may apply to streamflow generation in the south west of Australia.

2.2.1 Mechanisms

There are basically three mechanisms involved in the generation of streamflow:-

(i) Hortonian Overland Flow: as described by Horton (1933) involves surface ponding when the rainfall intensity exceeds the infiltration rate.

(ii) Saturated Overland Flow: occurs when near surface soils become saturated by rainfall and/or lateral subsurface inflow. Surface or overland runoff then results from rainfall on the area of saturated soil (a source area) and also by seepage from the soil to the surface (also termed return flow or exfiltration). Saturation of the soil typically occurs where the hydraulic conductivity of a subsoil is substantially less than the overlying soil so that there is an impedance to the vertical flow of water into the subsoil. This impedance may result in the development of a perched groundwater, particularly in duplex and multi-layered soils.

STREAMFLOW GENERATION

TIME ASPECTS	ULTIMATE DELIVERY MECHANISMS	HISTORICAL ASPECTS
1. Water added by specific events (event or new water)	1. Hortonian overland flow	1. Groundwater
2. Water present before event (old water)	2. Variable source area - subsurface storm flow	2. Unsaturated water
	3. Partial area - overland flow	3. Direct rainfall
	4. Variable source area - overland flow	4. Combination
	5. Channel interception	
	6. Groundwater Flow	

Figure 2.1 Classification of Streamflow Generation-Terminology (After Sklash and Farvolden, 1979)

(iii) Subsurface Flow: this mechanism occurs as lateral, near surface (within top metre or so) flow above a layer of lower hydraulic conductivity. During storm runoff this mechanism may contribute significantly to the stream by flow through macropores or preferred pathways in the soil.

A more sustained subsurface flow, associated with the groundwaters in the deeper profile may also contribute to streams, producing what is usually referred to as baseflow.

2.2.2 Streamflow Source-Area Concept

The spatial and temporal variation of the mechanisms of streamflow generation have been studied extensively in the field (see review by Dunne, 1978) and by modelling (Freeze, 1974). These and other studies have illustrated the importance of such factors as topography and hydraulic conductivity in determining the occurrence of particular mechanisms.

A notable contribution was the development of the concept of variable source areas of streamflow generation. Hewlett and Hibbert (1967) introduced the variable source area model for application to deep forested soils in humid environments. These source areas are located adjacent to stream channels and expand as the upslope soils saturate. Overland flow then occurs from larger areas and may consist of direct rainfall-runoff and subsurface flow which has appeared at the surface as overland flow.

Bonell et al (1984) further classified streamflow variable source areas into those producing predominantly surface or overland flow and those producing predominantly subsurface storm flow.

Computer simulations by Freeze (1972 a,b) and others have indicated the importance of topography (convergence/divergence), depth to an impeding layer and variations in hydraulic conductivity in the relative production of surface or subsurface flows. Concave slopes were found to be likely (variable) source areas for the generation of saturation overland flow. The expansion of the area of saturation was considered to be more likely a result of rainfall infiltration than influx of subsurface flows in producing saturation. Subsurface flow was considered to be important in maintaining saturated conditions between storms.

2.3 Observations of Streamflow Generation

Surface flow was found by Dunne and Black (1970 a,b) to predominate on grassed hillslopes of convergent, divergent and straight form in Vermont.

Saturation overland flow and return flow (velocities 500 times that of subsurface flow) were generated from saturated areas when a shallow water table rose to the surface during rainfall. Subsurface flows were found to be too small in magnitude and too slow in response time to be of any significance to stream stormflow. Instead the direct precipitation onto the stream channel and nearby wetlands was the source of the storm runoff. This formed a partial area, within a larger, variable source area. Dunne and Black also observed that the storage and translatory effects of the stream channel obscured the individual source area flood peaks downstream.

Betson and Marius (1969) produced evidence for runoff from relatively constant areas of hillslopes such as roads, rock outcrops, shallow soils and soils with limited infiltration capacity relative to rainfall intensities. Such areas were called partial areas by Betson (1964) and are a subset of the more general source area concept.

Weyman (1970) studied the relationships between the hillslope hydrograph and the stream hydrograph. He found that, although subsurface storm flow did occur on the hillslope, it was not significant in the stream hydrograph. Weyman observed that most streamflow was generated from a convergent headwater area upstream of the monitored hillslope. Here an extensive system of natural pipes, some 50mm in diameter, occurred at the base of a peat layer. These were thought to be a source of rapid discharge of water to the stream.

Jones (1971) observed that the preferred locations of such pipes was just above or within a soil layer of lower permeability. On lower slopes the development of the pipes appeared to require a steeper gully wall or stream channel bank. A concentration of pipes was found towards the base of the B horizon (0.3-0.6m below the soil surface) on a steep hillside in Japan (Tsukamoto et al., 1982). Some of the pipes, which averaged 140mm in diameter, could be traced for 5m into the hillside. Approximately 95% of the discharge from the 1.3m deep soils on the monitored hillslope came from the pipes. The pipes functioned when the level of saturation reached that of the pipes. Tanaka (1982) observed a similar predominance of subsurface discharge from soil pipes, uprooted and decayed tree stumps elsewhere in Japan.

Dominant subsurface flow in macropores was reported by Mosely (1979, 1982) from pit interception studies using tracers. Flow rates in the macropores, in the shallow (average depth 0.55m) soils on steep, forested slopes in the South Island, New Zealand, were 1000 times those of the surrounding matrix.

The important function of forests in providing channels or macropores for subsurface flow was noted by Beasley (1976). Subsurface flow was found by Beasley to be of sufficient magnitude and fast enough response on the hillslope to contribute significantly to the stream storm period flow.

From a detailed hillslope-stream study, Harr (1977) showed that subsurface flow contributed 96-98% of the storm flow hydrograph which varied from 23-51% of the storm rainfall. Direct rainfall-runoff from the stream channel and other source areas amounted to only 2-4% of rainfall.

More recently, O'Brien (1980) found subsurface flow to be 50-60% of storm flow and 92-93% of the annual flow from a 3.2 km^2 catchment in Massachusetts. Most of the subsurface discharge was considered to be from natural pipes located at the interface between "silty peat and muck". Discharge from the pipes occurred when the water levels in nearby bores were above the level of the pipes.

A rapid increase in the level of groundwater in bores has often been observed during storms in studies of the relationship between hillslope and stream hydrographs. O'Brien (1982) attributed the rapid level increase to the conversion of the capillary fringe of the tension saturated zone to a fully saturated state. The theoretical and experimental aspects of this process were reported by Gillham (1984) and Abdul and Gillham (1984). They showed that very little additional water (rainfall) is required under these conditions to fully saturate the profile above the level indicated in a bore.

This supports the groundwater ridging hypothesis (Sklash and Farvolden, 1979) in which groundwaters near the soil surface in lower landscape positions respond quickly to rainfall before those further upslope. Upslope water levels respond later because of the greater depth to the level of saturation. A groundwater mound or ridge will therefore develop on the lower slope and provide the source and driving gradient for the more rapid discharge of groundwater to streams. The relative significance of such subsurface discharge may be increased if it is delivered via macropores.

In a review of the nature and significance of macropore flow, Beven and Germann (1982) noted that, "it is likely that a variable zone of saturation at the base of the soil profile, or above a relatively impermeable horizon, will dominate lateral macropore flows through unsaturated soil in generating subsurface streamflows". On the likely response time they said that "if macropores are effective in transmitting flow to the stream channel, they may do so at velocities of the same order as overland flow".

2.4 Catchment Salt Processes

2.4.1 Salt Budget

The salt budget of a catchment is determined by measuring the atmospheric input of salt and the output of salt in streamflow. The difference between, or ratio of, output (O) and input (I) of salt, often a specific ion such as chloride, then provides additional information on the hydrology of the catchment.

Inputs are determined by measuring the total amount of water and salt in storage gauges distributed across a catchment. Output in streamflow is obtained by recording the discharge of water and concentration of salt in the water. The flow rate of water and flux rate of salt is then integrated to give the total water and salt output over a period (Barrett and Loh, 1982).

Catchment salt mass balances have been calculated for many regions, particularly in eastern and western North America and Western Europe. Feller and Kimmins (1979) summarized 20 such balances for forested catchments around the world with annual precipitations of between 550 and 3700 mm yr^{-1}. Of the 14 balances where chloride (Cl) was determined the differences between output and input averaged -3.2 kg ha^{-1} yr^{-1} with a range from -36 to 36 kg ha yr^{-1}. In their study of two catchments (23 and 68 ha) 60km east of Vancouver Feller and Kimmins found that the Cl O/I ratio ranged from 0.82 to 1.27 with a maximum output of 13.7 kg ha^{-1} yr^{-1}. These forested catchments were underlain by acid gneissic rock in an average rainfall of 2200-2700 mm yr^{-1}. A seasonal groundwater in the shallow podzolic soils was observed, with evidence of substantial subsurface flow and macrochannels in the profile.

At Hubbard Brook in New Hampshire, Juang and Johnson (1967) and Johnson et al (1969) measured yearly chloride output to input ratios of 2.2 to 2.4. The imbalance between output and input was considered by Juang and Johnston to be due to an additional (unmeasured) input. This imbalance which has also been observed by Van Denburgh and Freth (1965) in the western U.S.A. and Baldwin (1971) in California has been attributed to the impaction of particulates or aerosols on the forest vegetation which are subsequently washed-off by precipitation (McColl and Bush, 1978; Galloway, 1978; Reid et al, 1981, Sharma et al, 1983).

Therefore although bulk precipitation gauges are considered to measure salt input in rainfall and as dry fall (Galloway, 1978) there is the possibility that forest vegetation may be an additional mechanism for salt input to the catchment soils.

Variations in the magnitude and distribution of salt input will have consequences on the chemical characteristics of streamflow.

2.4.2 Hydrosalinity Models

Mathematical and computer simulation modelling of catchment hydrosalinity has been reviewed by Tanji (1981) for sub-catchment and catchment spatial scales in the western U.S.A. Tanji found that although there had been considerable progress over two decades in the understanding of the complexity of water and salt flows, most models are site specific and do not have wider applicability. In particular, Tanji concluded that "much more work is needed in the areas of field verification of models and increasing their utility for management and policy-making decisions".

At a more detailed level, Bresler (1981) and Raats (1981) reviewed the processes of transport and residence times of salt and water through soil systems. Raats considered the input-output relationships for a two-dimensional (in the vertical) saturated soil system and found an exponential distribution of arrival times. In Raats' words this was for a fairly simple system of "convective transport of solute distributed uniformly over the input surface, and not subject to diffusion, dispersion, to adsorption and to production and decay".

The exponential form of leaching was used by Peck (1973) for a laboratory analogy and by Hall and Gorgens (1979) for a large catchment in South Africa. The development (after Peck, 1973, 1976) is for the application of a recharge R (LT^{-1}) to a mass of solute M (ML^{-2}) on a soil surface through a storage W (L^3L^{-2}). The instantaneous leachate concentration C (ML^{-3}) is then:-

$$C = C_o \exp(-R.t/W) \qquad 2.1$$

where t is time since t_o
and C_o is initial concentration

The cumulative solute load L (ML^{-2}), by integration is:-

$$L = C_o W (1 - \exp(-R.t/W)) \qquad 2.2$$

(and $M = C_o W$)

Hall and Gorgens (1979) applied a similar model for the generation of monthly salt loads in streamflow, with additional terms for the continual generation of salt in a soil store.

Duffy (1984) developed conceptual models of salt loadings to streams from aquifers for the Upper Colorado River catchment. For steady state conditions (recharge equals discharge) an expression for the average concentration of aquifer outflow was developed:-

$$C = C_R (1 - \exp(-t/t_c)) + C_o \exp(-t/t_c) \qquad 2.3$$

Where C_R is the concentration of the steady state recharge, C_o is the initial concentration of the aquifer and t_c is the average aquifer residence time (equivalent to W/R in equation 2.1).

Integrating for the cumulative solute load L gives:-

$$L = W(C_o - C_R)(1 - \exp(-Rt/W)) + R.C_R.t \qquad 2.4$$

Note that equation 2.4 is equivalent to 2.2 when C_R is zero.

2.5 Separation of Streamflow Sources by Chemical Characteristics

2.5.1 Use

The separation of composite streamflow hydrographs into source components is considered to be one specific adaption of chemical methods in hydrology (Schwartz and Milne-Home, 1982). The authors' state that "... it is an approach to hydrograph separation that can be related to actual physical approaches operating in the basin". This is somewhat at variance with Pilgrim et al., (1979) who stated that such separations "yield information only on the proportions of old and new water and do not give direct information on runoff sources". The problem, according to Appelo et al., (1983) is that "the interpretation of single station data for the whole catchment area has to be supplemented and supported by knowledge of the actual spatial variations of water quality in the whole upstream reach". This view was emphasized by Walling and Webb (1980) for large catchments where the effects of aggregation of solute responses and channel routing may make interpretations of stream solute changes difficult. They did not however, comment on the likely significance of these effects for small catchments.

2.5.2 Mass Balance Equations

The steady and non-steady equations for the spatial and temporal distribution of conservative solutes in streamflow were presented by O'Connor (1976). From the conservation of mass of sources of solute in a single mixing volume:-

$$C_s \cdot \Delta V = C_s \cdot Q_s \cdot \Delta t + \sum_{i=1}^{n} C_i \cdot \Delta Q_i \cdot \Delta t$$

$$- \left(C_s + \frac{\partial C_s}{\partial x} \Delta x\right) \cdot \left(Q_s + \frac{\partial Q_s}{\partial x} \Delta x\right) \cdot \Delta t \qquad 2.5$$

Where ΔV : stream element volume (L^3)

Δt : time interval (T)

C : concentration of sources (i) and stream (s) (ML^{-3})

Q : flow of sources (i) and stream (s) (L^3T^{-1})

Δx : length of stream element (L)

n : number of sources

From equation 2.5 the general mass flux relationship is:-

$$\frac{\partial}{\partial t}(A \cdot C_s) = -\frac{\partial}{\partial x}(Q_s \cdot C_s) + \sum_{i=1}^{n} C_i \cdot \frac{\partial Q_i}{\partial x} \qquad 2.6$$

Where A : is area of flow (L^2)

By assuming steady state conditions (sufficiently large Δt for example), and integrating 2.6 yields:-

$$C_s \cdot Q_s = \sum_{i=1}^{n} C_i \, Q_i \qquad 2.7$$

and obviously $Q_s = \sum_{i=1}^{n} Q_i$ 2.8

Additional equations can be obtained, by assigning a characteristic concentration for any ionic species, to each of the inflows. Designate these concentrations as $C_i^{\ j}$, where i refers to the source of flow (i=1,n) and j refers to the ion (j=1,m species).

Therefore by mass balance:-

$$\sum_{i=1}^{n} C_i^{\ j}.Q_i = C_i^{\ j}.Q_s \qquad 2.9$$

and there are j = 1,m such equations.

Clearly for solution for the Qi unknowns, given the $C_i^{\ j}$ and Qs, there should ideally be m = n equations.

This condition is not always possible, and even where it is there have been problems in obtaining meaningful solutions (e.g. Woolhiser et al., 1982; Schwartz and Milne-Home, 1982). This difficulty arises because of the assumption of constant concentration ($C_i^{\ j}$), which, as noted by Pilgrim et al. (1979) is probably not valid. However there are difficulties in establishing how the C^j vary through time for each of the sources.

In applying this model to separate streamflow components, the $C_i^{\ j}$ are often assumed not to vary, and i = 2, so that there are two equations in two unknowns (Q_1 and Q_2). These components are often assumed to be surface runoff and groundwater inflow to the stream.

2.5.3 Results of Hydrograph Separations

Bernier (1985) summarized the results of nine studies reported in the literature of hydrograph separations using natural traces oxygen -18 and tritium. The contribution of pre-event or 'old' water to peak discharge and to total storm flow was typically 50-80%. Bernier concluded that "these findings strongly support the view held in the variable source area concept that subsurface flow plays a major role in storm-flow generation in vegetated basins".

Major ion chemistry and isotopes have been used to separate storm hydrographs (Fritz et al., 1976, Sklash and Farvolden, 1979). Groundwater (or pre-event or old water) was the source of more than 60% of peak discharge and total storm flow volumes.

The apparent rapid response of the groundwater during rainfall was attributed to groundwater "ridging" near the stream which results in increased depth and hydraulic gradient. Pinder and Jones (1969) invoked high hydraulic conductivities as a mechanism for the 32 - 42% contribution of groundwater to peak discharge.

In contrast to these, Gburek and Heald (1970) found that 100% of the storm runoff volume could be attributed to rainfall on a 'wetted' area on and immediately around the stream draining a 42 ha agricultural catchment in Pennsylvannia. Fritz et al. (1976) also noted the importance of rapid-rainfall-runoff from the channel and environs in generating most of the 'new' water in storm flow.

An increase in the leaching of solutes was observed by Gburek and Heald (1970) and Burt (1979) and was considered to be the result of the removal of solutes from previously unsaturated soil during rainfall infiltration. This variation in concentration is a problem in the application of the steady state (constant concentration) source component equations.

Fritz et al. (1976) determined that 'contamination' of the direct (surface) runoff component occurred in their study. This they indicated would produce errors in the separation, leading to over-estimation of the groundwater component. The cause of the increase in the concentration of the surface rainfall runoff was thought to have been relatively fast chemical reactions between the soil and the rainwater.

This variation in the concentrations of both surface and subsurface flows was found by Pilgrim et al., (1979). Concentrations of 'new' water added by rainfall were found to increase with increasing contact time with the soil. Their analysis indicated that if this variation in concentration was taken into consideration in the separation process, less 'old' water contributed during runoff. That is, time invariant concentrations will result in overestimates of the old water (or groundwater) component.

2.6 Catchment Hydrology in South Western Australia

2.6.1 A Hydrological System

A conceptual hydrological system shown in Figure 2.2 represents a typical hillslope section in the steeper, higher rainfall areas of the Darling Range of Western Australia.

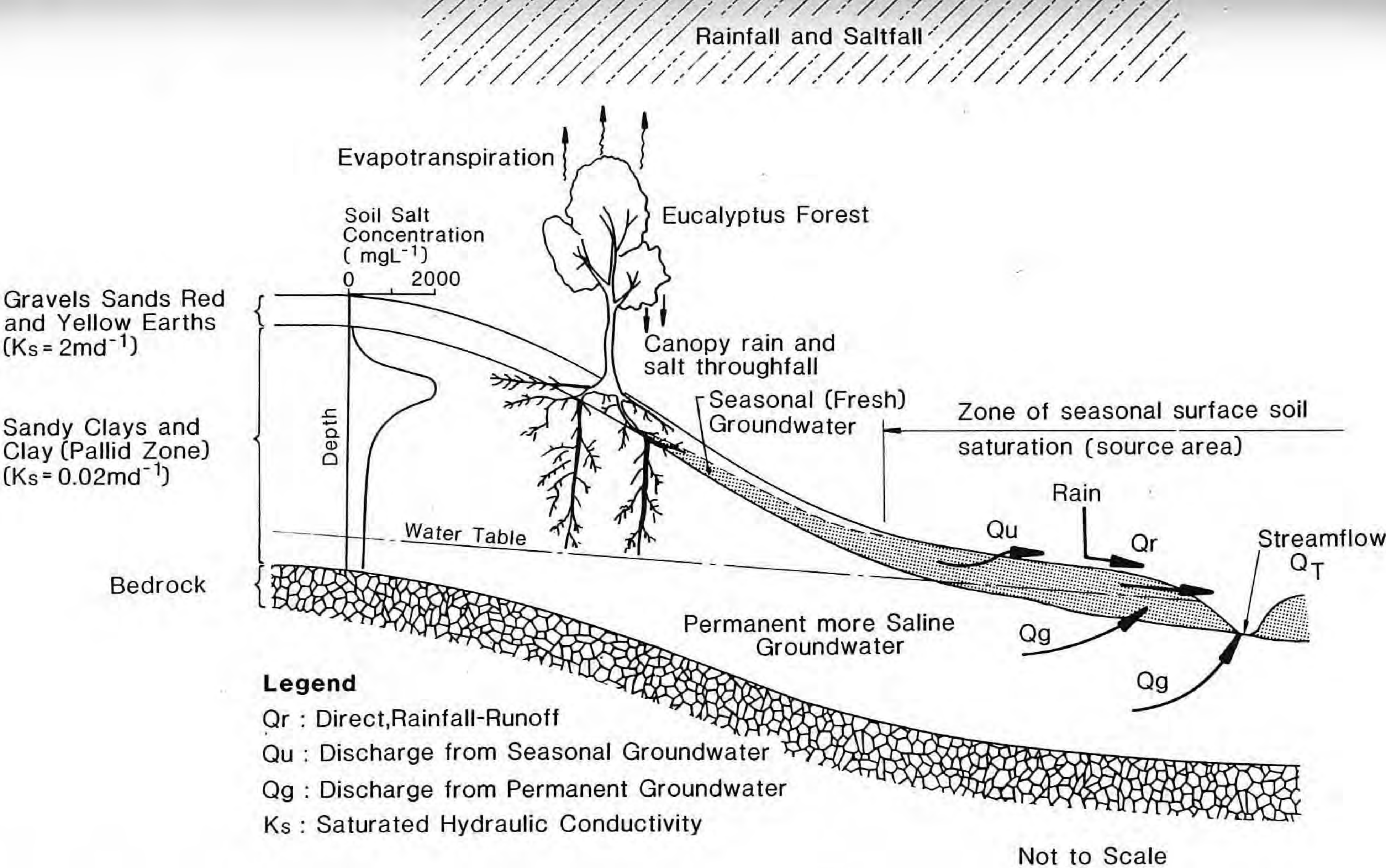


Figure 2.2

HILLSLOPE SOURCE MODEL OF STREAMFLOW
(Adapted from Sharma and Williamson, 1984)

The profile is characterised by a shallow (0 - 5m) surface soil overlying a deeper (0 - 50m) kaolinitic subsoil derived largely from insitu weathering of the basement rock. The saturated hydraulic conductivities of the surface and sub-soil materials differ by about two orders of magnitude (Sharma et al, 1980b; Peck et al., 1980).

Saturated hydraulic conductivities of $2m \cdot d^{-1}$ (Sharma et al., 1980b) in the sands, gravels and earths of the near surface soils mean that surface runoff is rare over most of the areas of forested catchments in the south west as rainfall intensities are generally much less.

The contrast in soil hydraulic properties is reflected in the distribution of tree roots and soil water. Carbon et al. (1980) found most root material in 1 - 2m deep sandy, gravelly surface soils. The density of roots decreased by 10 - 100 times in the underlying sandy loams and clays.

The deep-rooted, evergreen, eucalypt vegetation effectively exploits a large soil volume (Dell et al., 1983), transpiring water from deep into the profile and often to the water table (Carbon et al., 1980). Salts of oceanic origin are precipitated in rainfall and as dryfall (Hingston and Gailitis, 1976) and accumulate in the kaolinitic subsoils because of the very low hydraulic conductivities and the dominance of evapotranspiration in the water balance.

In the system depicted, seasonal, perched groundwaters (Qu) develop in the relatively low salinity surface soils above a permanent, deeper groundwater (Qg) in more saline (note the distribution of soil salt storage in Figure 2.2) kaolinitic subsoils. Water and salt from the deeper soils are transported into the shallow soils and then into the stream. Direct rainfall-runoff may occur from impervious areas such as rock outcrops and areas of saturated surface soil produced by the shallow and/or deeper groundwater.

2.6.2 Elements of the Water Balance

Sharma (1983) calculated six-monthly (October-March and April-September) components of the water balance equation for Salmon catchment (See Chapter 3) for the period 1973 to 1978. The equation was:-

$$ET = P - (RO + \Delta W + \Delta G) \qquad 2.10$$

where ET is evapotranspiration, P precipitation, RO streamflow, Δ W change in soil water storage and Δ G change in deeper groundwater storage. In this exercise, the results of which are reproduced in Table 2.1, Sharma was primarily interested in estimating evapotranspiration and its distribution through seasons.

The average soil water storage change was 212mm over winter and -217mm over summer. During summer months water was extracted at rates approaching 90mm per month. Sharma also noted that soil water deficits for a 1.2m deep profile did not exceed 100mm but were as high as 450mm for a freely draining site under a forest elsewhere in the south west of Australia.

Groundwater storage changes averaged 4mm and -17mm respectively over the two six monthly periods between 1974 and 1978. Only during the well above average rainfalls of 1974 was there a positive change in groundwater storage (32mm). Storages may have been positive in the other years at times earlier than the end of September analysis used by Sharma.

TABLE 2.1 SIX MONTHLY WATER BALANCE COMPONENTS FOR SALMON CATCHMENT

(From Sharma, 1983)

Year	Month	Days	P	RO	ΔW	ΔG	ΔW+ΔG+RO	ET	ET/Day	EO/Day	ET/EO
1974	O–M	182	171	12	-196	-5	-189	360	2.0	6.4	0.31
	A–S	183	1306	354	225	32	611	895	3.8	2.4	1.56
1975	O–M	182	108	6	-239	-12	-245	353	1.9	6.6	0.30
	A–S	183	806	77	152	-2	227	579	3.2	2.6	1.21
1976	O–M	183	273	2	-198	-22	-218	492	2.7	6.2	0.43
	A–S	183	610	13	149	-11	156	455	2.5	2.6	0.97
1977	O–M	182	208	20	-185	-12	-177	335	2.1	6.4	0.33
	A–S	183	775	55	234	0	289	486	2.7	2.5	1.05
1978	O–M	182	146	18	-270	-18	-270	417	2.3	6.3	0.36
	A–S	183	754	59	300	-1	358	396	2.2	2.3	0.94
	Mean										
	O–M		181 ± 28	11 ± 3	-217 ± 16	-14 ± 3	-220 ± 17	401 ± 25	2.2 ± 0.1	6.4 ± 0.1	0.35 ± 0.02
	A–S		850 ± 119	113 ± 61	212 ± 28	4 ± 7	328 ± 78	522 ± 52	2.9 ± 0.3	2.5 ± 0.1	1.15 ± 0.11

(Here P is precipitation, RO is streamflow, ΔW is change in soil water storage, ΔG is change in groundwater storage, ET is actual evapotranspiration and EO is pan evaporation.)

In the major conclusion, Sharma estimated evapotranspiration to be 70% of rainfall over the five years. During August 1977 evapotranspiration was calculated to have been 40% of rainfall. The average ratio of estimated ET to pan evaporation was 0.35 over summer and 1.15 over winter with a peak of 1.56 for the very wet 1974 winter.

On this evidence Sharma concluded that interception of rainfall and transpiration of infiltrated soil water were the most important elements of the water balance.

Streamflow from Salmon catchment averaged 123mm between 1974 and 1978 or about 12% of rainfall. Stokes and Loh (1982) calculated the average streamflow to be 10% of rainfall between 1974 and 1980 with most occuring between May and October.

2.6.3 Soil Water Distribution

The seasonal distribution of soil water has been studied by Sharma et al., (1982), some results of which are shown in Figures 2.3 and 2.4. Mean relative soil water storage for the top 6m of the profile for Salmon catchment (Figure 2.3) indicates seasonal maxima in August - September and minima in March. The minima over the four years, relative to 23/4/74, decreased by less than 50mm and can therefore be considered as reasonably constant from year to year. Maxima are more variable, reflecting the variation in winter rainfalls. Soil water storage changes of upwards of 300mm are indicated.

The largest variations of soil water with depth occurred between 1m and 5m for a site on Wights, a catchment adjacent to Salmon. Sharma et al., (1982) noted that there was relatively little seasonal change

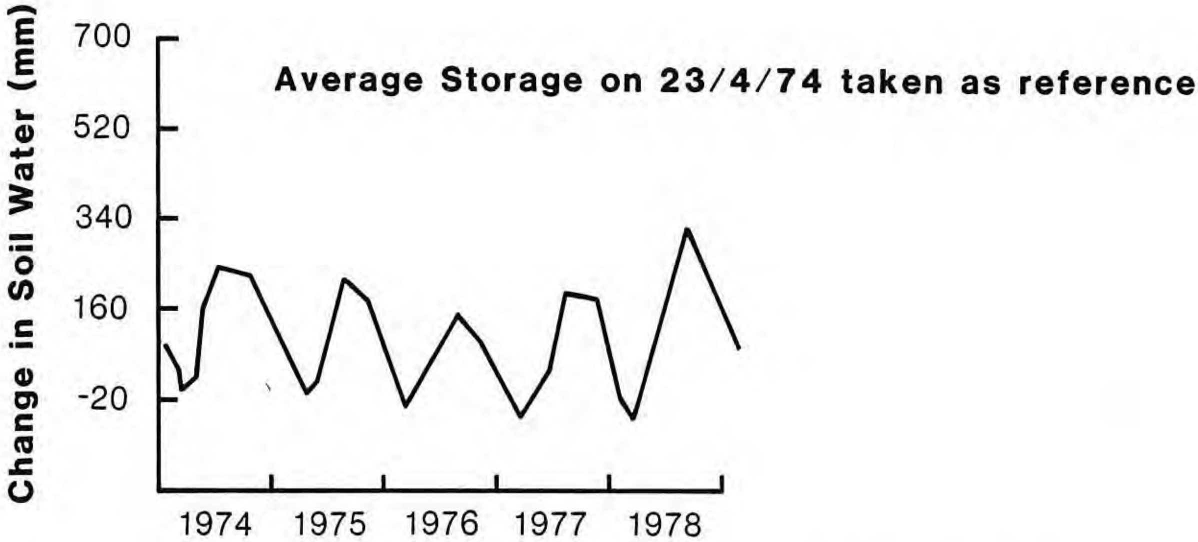


Figure 2.3 Average Soil Water Storage in top 6m of Salmon Catchment
(from Sharma et al, 1982)

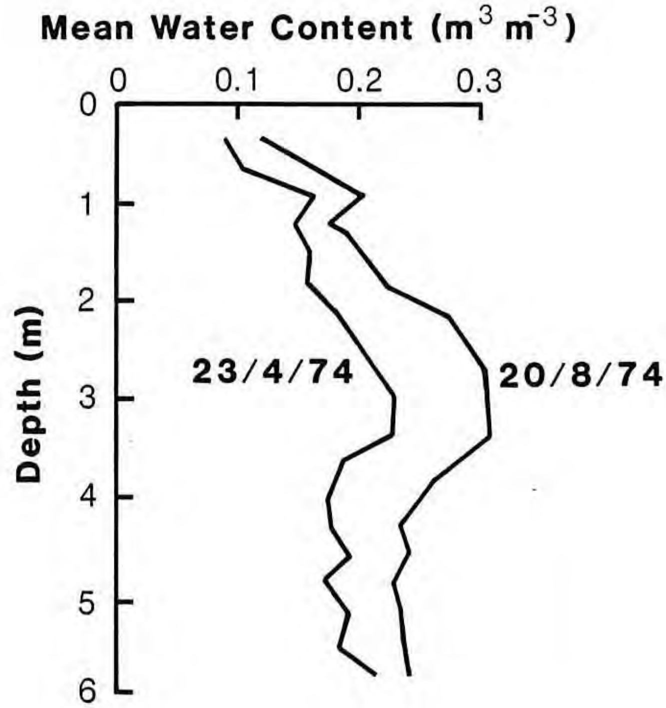


Figure 2.4 Typical Maximum and Minimum Mean Soil Water Storage at Site 5 Wights Catchment
(from Sharma et al, 1982)

at the bottom of the profile at 5.7m. This was attributed to the deep-rooted vegetation exploiting the soil water store down to this depth. By inference only small quantities of water were withdrawn below this depth although this was not measured.

A substantial storage of soil water may develop during winter resulting in saturation of parts of the profile. In lower landscape positions surface soil saturation may occur resulting in the formation of source areas for overland flow during subsequent rainfall. The development of saturated soil in the upper profile may also enhance recharge to deeper parts of the profile.

2.6.4 Recharge to Deeper Groundwaters

Johnston (1983) reported that only 17% and 14% of annual rainfall infiltrated deeper than 2m at two sites on Salmon catchment. At 6m, vertical water flux was 0.5% and 3% of rainfall at the same sites. Johnston indicated that there was evidence for more rapid recharge deeper into the profile, probably by preferred pathways such as root channels. From 7 bores in Salmon catchment Johnston estimated groundwater recharge to be 25mm yr^{-1} with an upper bound of 50mm yr^{-1}. The mean seasonal rise in the water level at one bore in the catchment was calculated to be equivalent to 32mm of water.

The mechanisms of recharge, from the shallow soil horizons to the deeper groundwater, were studied in some detail by Johnston et al. (1983). A $10m^2$ area was carefully excavated to about 8m depth after a winter period during which tracers had been applied in rainfall equivalent depths. In the authors' words this study was the first test of the hypothesis that:

"water moved from the ground surface to the saturated zone via preferred pathways or channels in the soil matrix".

Johnston et al. found a shallow, perched aquifer of from 0 - 1.3m thickness which developed for relatively short periods in the near surface profile. Tracer was recovered in this shallow profile 22m downslope, 9m along the contour and 3m upslope of the application area. The horizontal movement of the tracer occurred as distinct streams rather than as continuous sheet flow.

Water movement into the underlying clays occurred in well-defined, approximately cylindrical, vertical channels of coarse textured material with channel diameters of 20 to 100mm. These were thought to conduct water under conditions of saturation in the overlying soils (perched aquifer).

The occurrence of these channels, often filled with living and dead tree roots (Dell et al., 1983), is probably a common feature of the profiles in the forested areas of the south west. Saturation of the shallow soils overlying the deep clay profiles may therefore be a very important mechanism for recharge to the deeper groundwaters via preferred pathways.

2.6.5 Salt Budgets

Annual average salt budgets, of streamflow output to precipitation input, for forested catchments of $4km^2$ to $380km^2$ in the south west were reported by Peck and Hurle (1973). These ranged from 1.1:1 to 1.6:1 for forested areas whereas catchments with some clearing had salt output to input ratios of more than 20:1. Recent, more accurate salt balances by Stokes and Loh

(1982) and DCE (1984) have shown that salt balances can vary between 0.1:1 and 2.5:1 for small ($< 5km^2$) forested catchments.

Therefore, forested catchments in the south west are either accumulating salt or are exporting salt (a net catchment loss) through streamflow. Peck and Hurle (1973) considered that it was unlikely that errors in the measurement of salt input could be responsible for this apparent loss of salt for agricultural catchments but that this may have been the case for forested areas.

Chemical mass balances were used by Peck and Hurle to determine the amount of discharge of the more saline discharge of streams. This technique has been used subsequently to estimate the relative contributions of surface runoff and groundwater to streamflow.

2.6.6 Streamflow Source Components

Sharma et al. (1980a) and Stokes and Loh (1982) have used the conceptual hillslope model (Figure 2.2) to determine the relative contributions of surface runoff (direct rainfall), shallow groundwater and deeper groundwaters to streamflow from Salmon catchment. The steady-state mass balance equations (developed from 2.7 and 2.8) used were:-

$$Q_T = Q_r + Q_u + Q_g \qquad 2.11$$

$$F_T = F_r + F_u + F_g \qquad 2.12$$

where:- Q : flow ($m^3\ s^{-1}$)

F : chloride flux (kg s^{-1})

and subscripts:-

T : total at gauging station
r : direct rainfall-runoff
u : shallow groundwater
g : deeper groundwater

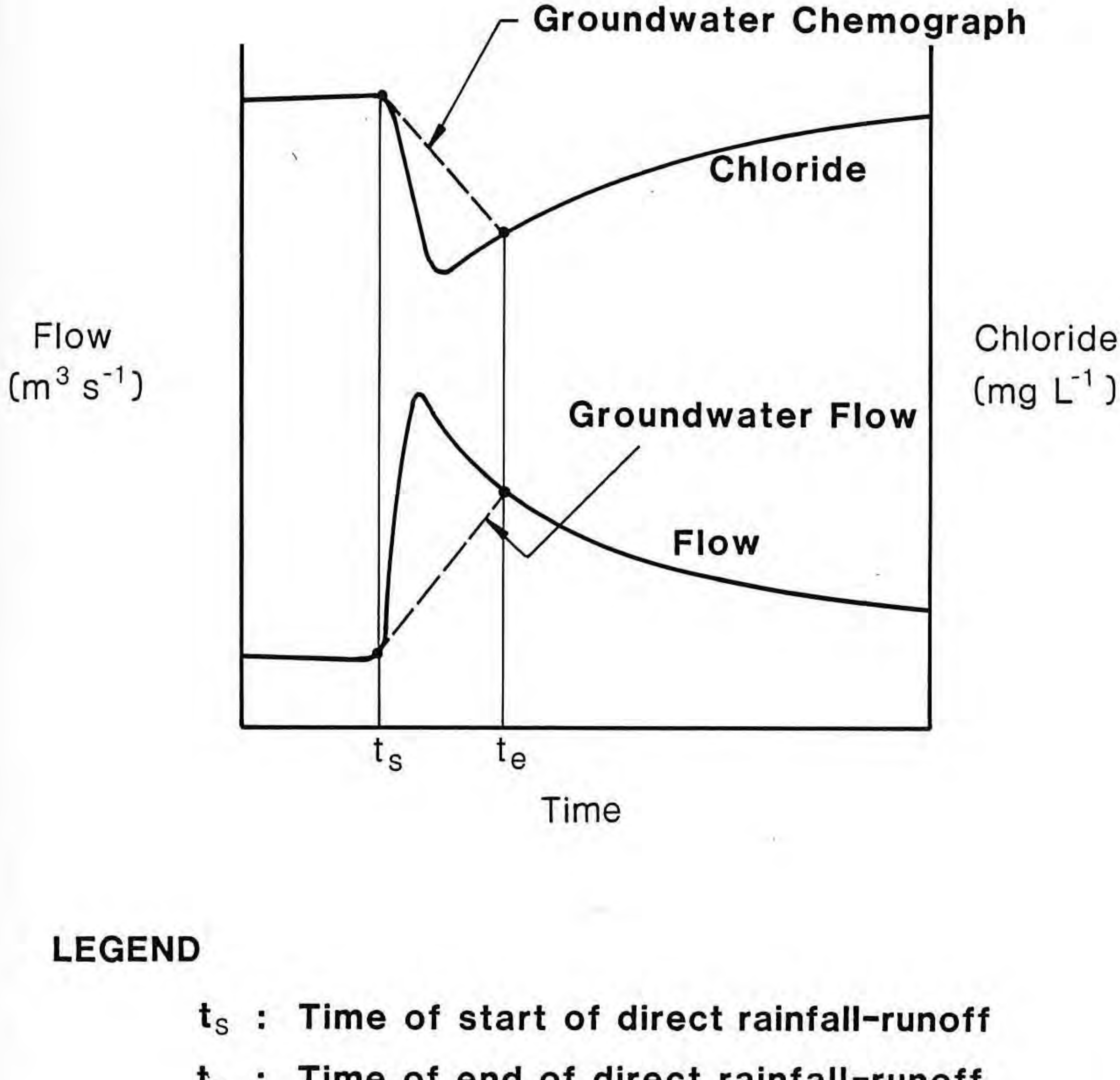
Here $F_i = C_i \times Q_i \times 0.001$ 2.13

where C_i : concentration ($mg\ l^{-1}$)
i : T, r, u and g

Sharma et al. (1980a) applied these equations to annual data for Wights catchment adjacent to Salmon and set the surface runoff component (Q_r) to zero in order to solve the two equations. Characteristic concentrations (C_i) were estimated from analyses of shallow and deep groundwater samples. Their results indicated that about 80% of streamflow was from the shallow groundwater.

The important restrictions on this approach were the use of annual data, constant characteristic concentrations and the necessity to set surface runoff to zero in order to solve the equations.

In attempting to improve the procedure, Stokes and Loh (1982), using continuous records of discharge and concentration, developed a simple hydrograph separation procedure to estimate the surface runoff (Q_r) component. The separation procedure is shown in Figure 2.5. Surface runoff is assumed to occur between the time of the hydrograph rise and a time set as a fixed period of time after the end of rainfall. This procedure assumes that the baseflow (combined shallow and deeper groundwater) component responds in a linear



LEGEND

t_s : **Time of start of direct rainfall-runoff**

t_e : **Time of end of direct rainfall-runoff**

Hydrograph and Chemograph Separation

Figure 2.5

way and that all surface runoff will have flowed past the gauging station in a fixed time since the end of rainfall.

This procedure was applied to Salmon catchment for the 1980/81 year. Shallow groundwater was found to contribute 91% of water and 38% of chloride to streamflow. Direct, surface runoff contributed less than 2% of water and chloride. The deeper groundwater was calculated to have contributed 7% of water and 60% of chloride. These were based on C_u and C_g values of 35 mg l^{-1} and 750 mg l^{-1} respectively as used by Sharma et al. (1980a).

Peck (1976) developed equations 2.11 and 2.12 to estimate the streamflow salinity effects of bauxite mining and agricultural clearing. Peck suggested that the variation of the deeper groundwater concentration (Cg) could be estimated from:-

$$Cg = (S/W) \exp(-G.t/W) \qquad 2.14$$

where S($M\ L^{-2}$) is the mass of solute, W ($L^3\ L^{-2}$) is the volume of water stored in the soil below unit area of land, G($L^3\ L^{-2}\ T^{-1}$) is the rate of groundwater discharge per unit of catchment area, and t (T) is the leaching time (see also equation 2.1).

Equation 2.14 was used with data from Wights catchment (adjacent to the study area) with S = 15 kg m^{-2}, W = 4300 mm and G = 430 mm to produce a Cg of 3500 mg l^{-1}. This result was thought to be too high and subsequent analysis (Peck, 1983) indicates that G is much less. Therefore groundwater contribution would be significantly smaller.

A development of these approaches will be made in later Chapters for application to the stream water and chloride response for Salmon catchment.

2.7 Discussion and Summary

Subsurface water has been shown to be a major contributor to streamflow from forested catchments around the world. Detailed plot, hillslope and catchment scale experiments have identified mechanisms whereby subsurface water can respond quickly and in large volume in the stream storm-period hydrograph. In particular, the phenomena of rapid, shallow groundwater ridging and preferred pathways or macropores for rapid water flow to streams have been high-lighted. These shallow groundwater are also responsible for areas of saturated surface soil which then function as dynamic source areas for direct runoff of rainfall.

Various natural and introduced tracers have been used to discriminate between sources of streamflow. Separation of hillslope and catchment hydrographs into source components have been made using such tracers. Subsurface flows have been identified as being significant proportions of storm flow hydrographs. In general these studies did not attempt to calculate the seasonal or yearly contributions of surface runoff and groundwater to streamflow.

In the south west of Australia, evapotranspiration is the dominant component of the water balance, particularly for forested catchments. Seasonal soil water storage variations of 200-300mm have been found in the top few metres of the generally duplex profiles of the region. Pathways or macropores down through the kaolinitic subsoils have been observed and these function as preferred water flow pathways for recharge to the deeper, permanent groundwater from the transient, perched aquifer in the surface soils.

Catchment scale salt balances have indicated a general accumulation of salt under forest with some catchments exporting salt. Chemical mass balance techniques have been used to calculate the approximate proportions of surface runoff and groundwater in streamflow. Shallow, seasonal, groundwaters have been found to be the primary source of water whereas the more saline, deeper groundwater is the primary source of salt in streamflow. However little is known about the extent and dynamics of the shallow and deeper groundwaters and how streamflow is generated.

It is an aim of this study to develop a clearer understanding of the spatial and temporal variability of streamflow generation in Salmon catchment, a small forested, experimental area in the south west and to test the applicability of simple models of stream water and chloride sources.

3. SITE AND EXPERIMENTAL DETAILS

3.1 Location

Two small experimental catchments were established near Wellington Reservoir (Figure 3.1), in 1973/74, to study the water and salt responses of clearing forested areas for agriculture. This is a joint project between the Water Authority of Western Australia (formerly the Public Works Department) and the C.S.I.R.O. Division of Groundwater Research (formerly Division of Land Resources Management).

3.2 Climate

The study area has a Mediterranean climate of cool, wet winters and warm to hot, dry summers. About 80% of the average annual rainfall of 1150mm (Hayes, personal communication) occurs between May and October (Figure 3.2). Class A pan evaporation is about 1600mm, with rainfall exceeding evaporation only during winter. Average maximum temperatures in the nearby town of Collie are 31.2^{o}C in January and 15.7^{o}C in July with average minima of 14^{o}C and 5^{o}C respectively.

3.3 Topography and Vegetation

The geology, soils and vegetation of Salmon catchment have been mapped and described in some detail by Bettenay et al. (1980). The catchment is broadly rectangular in shape, 81.8 ha in area and approximately 1500m in length. The elevation ranges from 302m (A.H.D.) in the south at the highest part of the divide to 190m at the weir (Figure 3.3). In general form the catchment consists of an 800m section upstream of the weir which has roughly parallel, relatively steep side slopes. Further upstream the catchment broadens in a convergent, headwater area with relatively lower slopes.

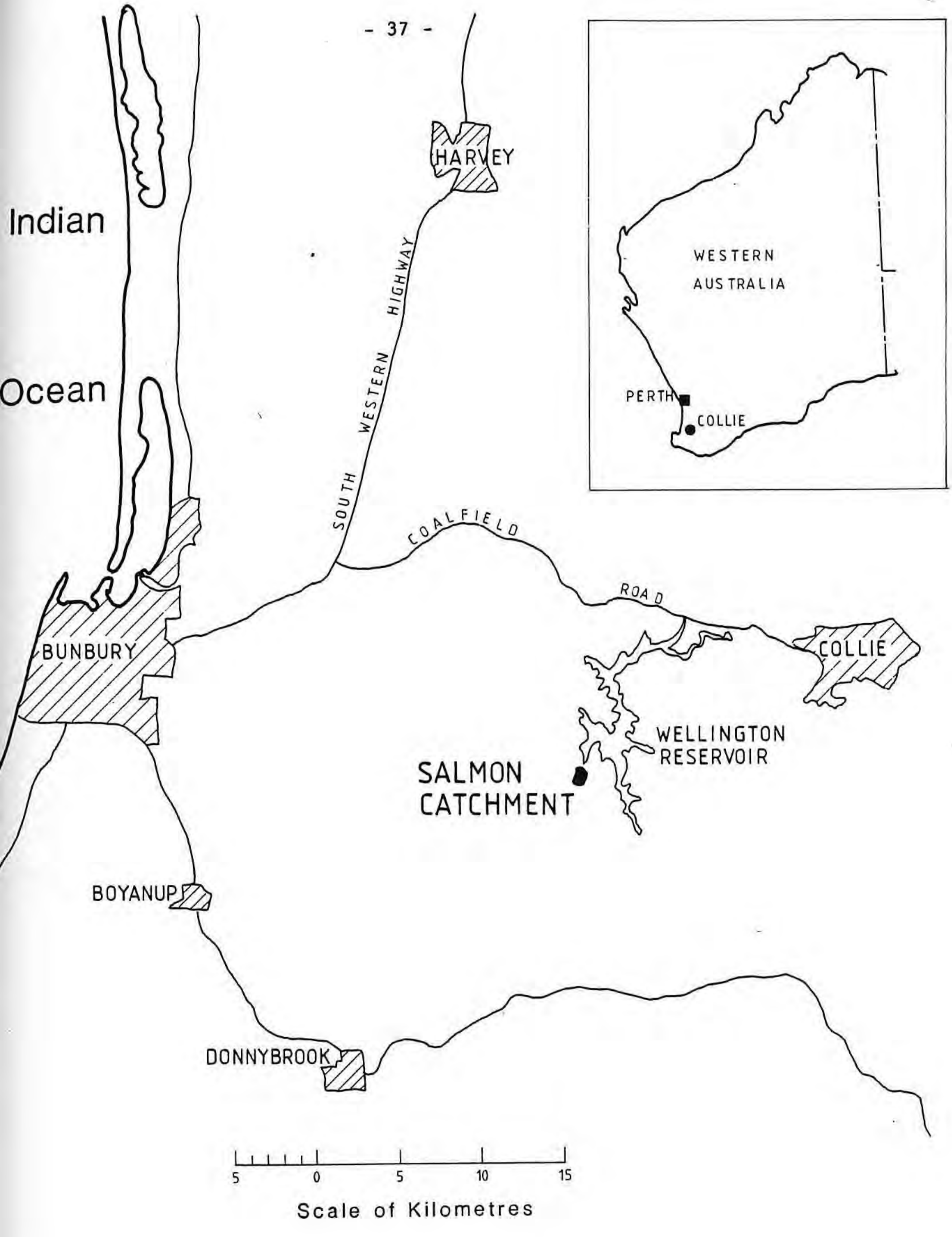


Figure 3.1
LOCALITY MAP

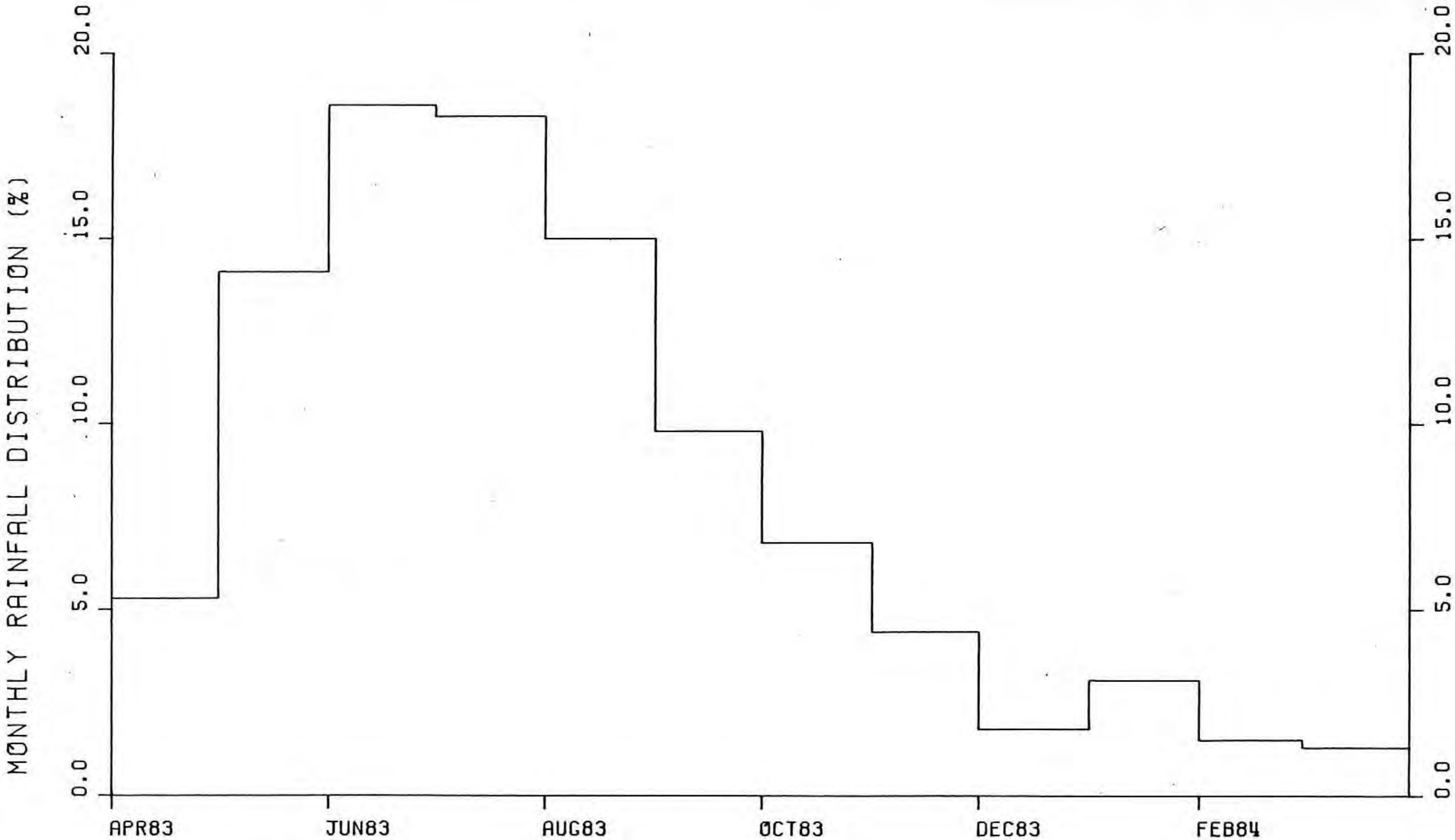


Figure 3.2
AVERAGE MONTHLY RAINFALL DISTRIBUTION

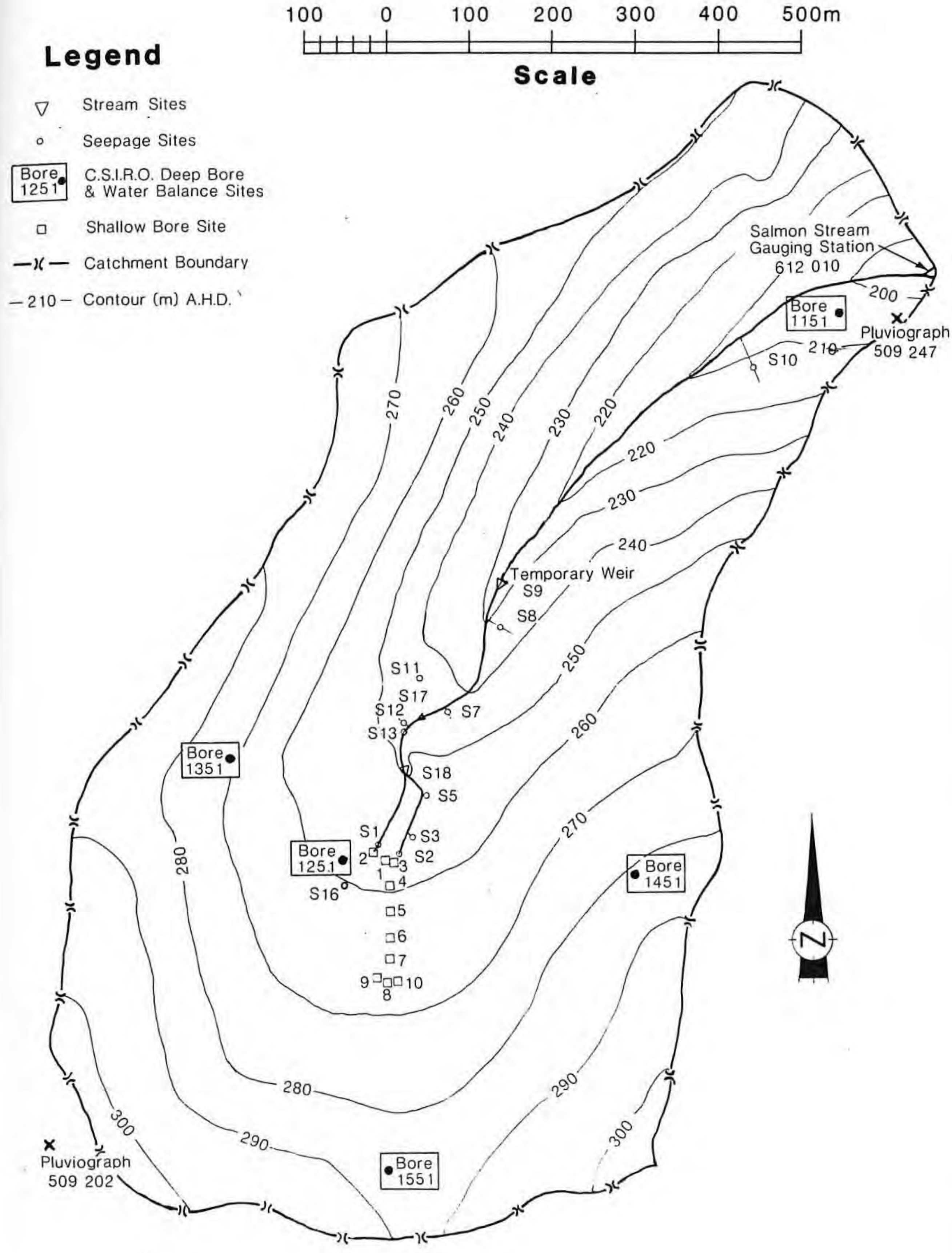


Figure 3.3
SALMON CATCHMENT

The stream rises as two small channels, each about 100m in length, draining the convergent, headwater area. These channels are 2 - 3m wide and incised 0.5 - 1.0m into the surrounding areas of red earths. Channel incision increases to 2 - 3m in the 50m reach downstream of the junction of the two channels. Here the 1 - 2m wide channel has incised through the earths to a gravelly, lateritic or mottled/pallid clay bed.

The 230m of channel downstream of the swamp area near S12 (Figure 3.3) to the site of the temporary weir at S9 passes through steeper side slopes, with often shallow soil over rock and in places exposed rock. An area of rock and shallow soils (0.5ha) intersects the channel from the east bank about 100 - 150m downstream of S9.

Over the lower 700m the channel, 0.5 - 1.5m wide and 0.5 - 1.0m deep, occasionally meanders across a 5 - 20m wide 'flood plain'.

There is a succession of pools, small cascades and a few small (less than 1m) waterfalls.

The depth of incision in the lower catchment, as evidenced by rock outcrops indicates the possibility of groundwaters intersecting the streamline in this area.

3.4 Geology

Basement rock is granite and gneiss with some amphibolities (Bettenay et al. 1980). Depths of weathering range from zero at rock outcrop to more than 30m in upslope locations.

A typical profile consists of medium to coarse textured gravelly soils of 1 - 5m thickness overlying kaolinitic subsoils of a much finer texture. These may extend 10 - 20m, varying from mottled clay to pallid clay before changing to a weathering zone transitional to the parent rock.

3.5 Soils and Vegetation

The soils map (Figure 3.4) has been reproduced from Bettenay et al. (1980). Soils range from gravels, pale sands and lateritic duricrust on the uplands through to red and yellow earths on the lowlands. Areas of stony red and yellow earths, often shallow to bedrock and outcrops of rock occur along the streamline lower in the catchment.

The vegetation (Figure 3.5, from Bettenay et al., 1980) is typical of the jarrah forest in the higher rainfall areas of the western Darling Range. The overstorey is predominantly E. marginata (jarrah) and E. calophylla (marri) with basal areas of 20 - 30 m^2 ha^{-1}. Carbon et al. (1979) estimated that this forest has a leaf area index of 1.8 (standard deviation 0.4).

3.6 Catchment Hydrologic Zones

Bettenay et al. (1980) grouped the various soil/vegetation types into 'hydrologic provinces' on the basis that these zones were considered: "... to be similar in general hydrologic properties , depth of weathering and vegetative cover". Four provinces were identified, and these are shown in Figure 3.6 and the areas and properties are listed in Table 3.1.

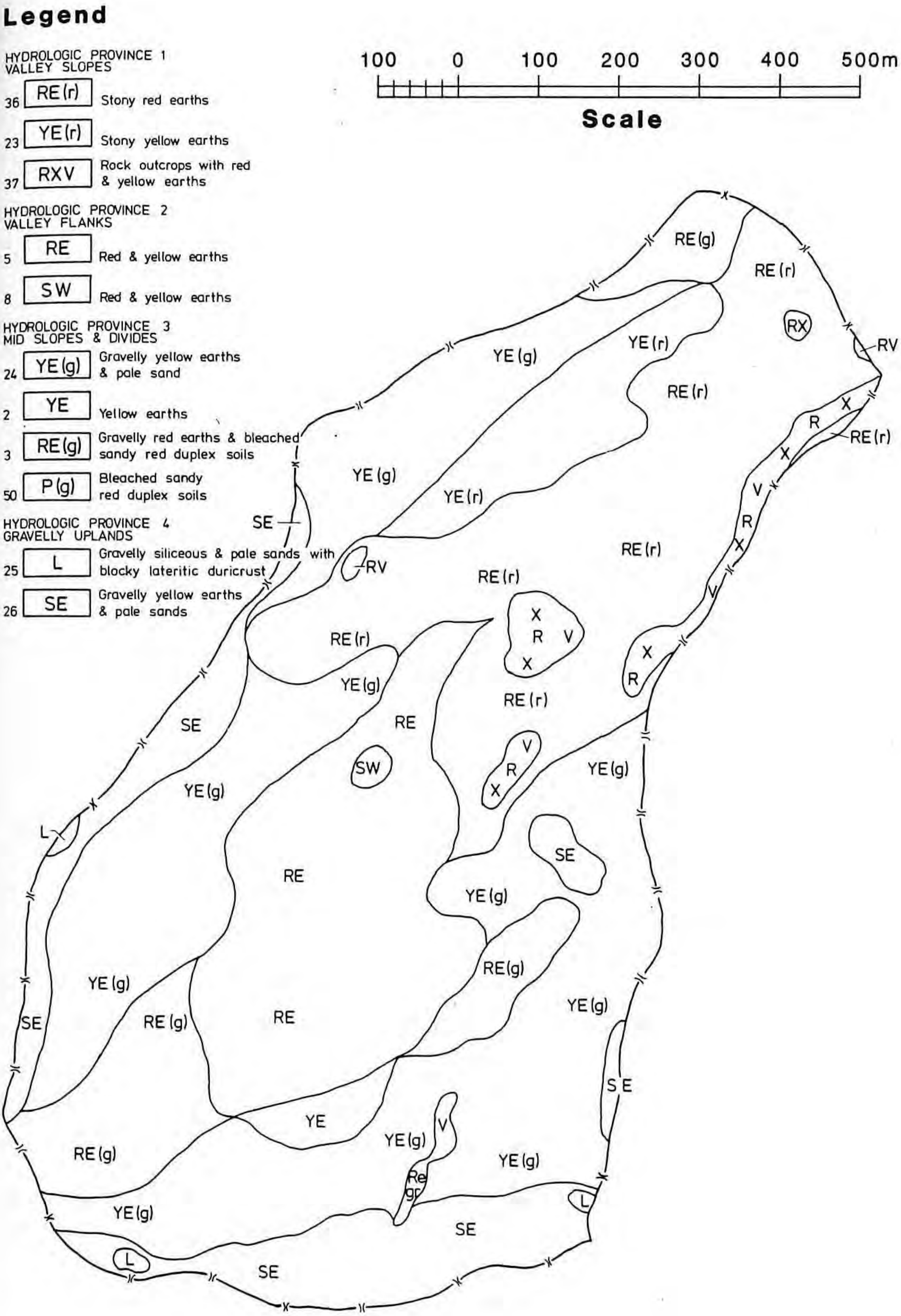


Figure 3.4
SOILS OF SALMON CATCHMENT
(From Bettenay et al., 1980)

Legend

TYPE T occurs on the crest and upper slopes of the steeply dissected valleys. The indicator species are Eucalyptus marginata, E. calophylla, Leucopogon verticillatus, Pteridium esculentum, Leucopogon capitellatus and Persoonia longifolia

TYPE T-S is separated from T by the presence of Banksia grandis and generally occurs on the more gradual ridges and upper slopes. Lateritic influence on the soil formation is more noticeable in this type.

TYPE U-T occupies the steep fertile lower slopes of the Salmon catchment. In addition to type T indicators Xanthorrhoea preissii and Macrozamia riedleii occur in moderate density.

TYPE Q occurs on the moist fertile lower slopes and flats and is indicated by the presence of Eucalyptus patens and heavy E. calophylla in the tree species and Hypocalymma angustifolia, Macrozamia riedleii and Xanthorrhoea preissii in the understory.

TYPE D occupies the broad moist valley heads of Salmon and Wight and occurs as a narrow transitional zone above Q on the less deeply incised eastern valley side of Wights. It differs from Q in the virtual absence of Eucalyptus patens and the dominance in the tree species of E. marginata and E. calophylla. It is indicated in the understorey species by the presence of Hypocalymma angustifolia, and absence of Leucopogon capitellatus.

TYPE C is a narrow belt fringing the streams occasionally broadening at the wider sections of the valley. It is indicated by the presence of Agonis linearifolia, Lepidosperma tetraquetrum and Astartea fasicularis. The tree species are E. patens and E. calophylla.

TYPE G-R is the granite and epidorite outcrop areas. It is typified by the virtual absence or low basal area of tree species and the presence of Grevillea bipinnatifida, Baeckea camphorosmae. Hakea lissocarpha, Hypocalymma angustifolia and Diplolaena drummondii.

TYPE O-F is a yellow sandy colluvium. It is typified by the presence of Hakea ruscifolia, Xanthorrhoea gracilis and Conospermum capitatum. The tree species are E. marginata, E. calophylla and Banksia grandis.

TYPE Q-T occurs in an upland valley situation of the Salmon catchment. It is indicated by the presence of E. calophylla, E. marginata and light E. patens in the tree species and Pteridium esculentum, Persoonia longifolia and Hibbertia sylvestris in the understorey.

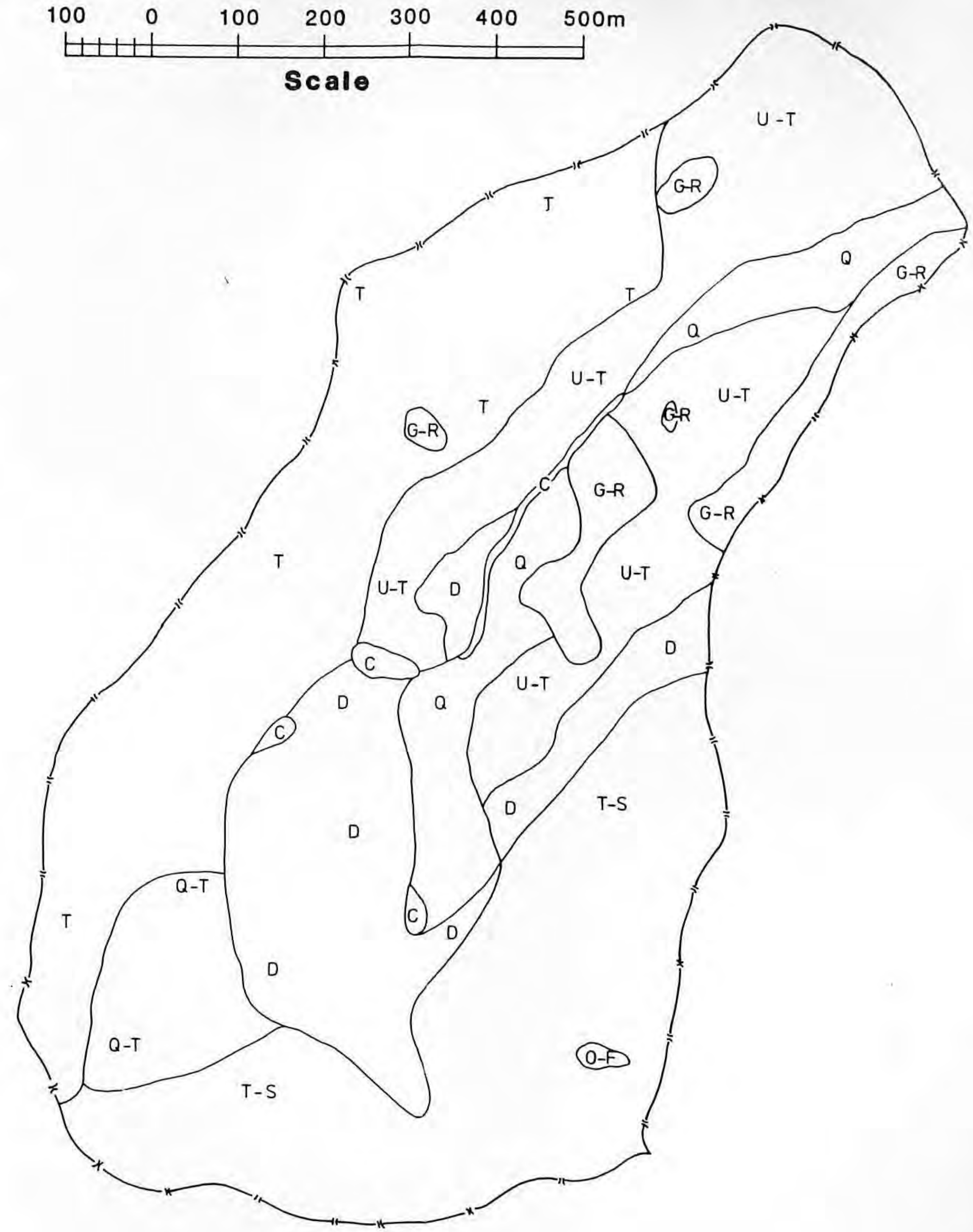
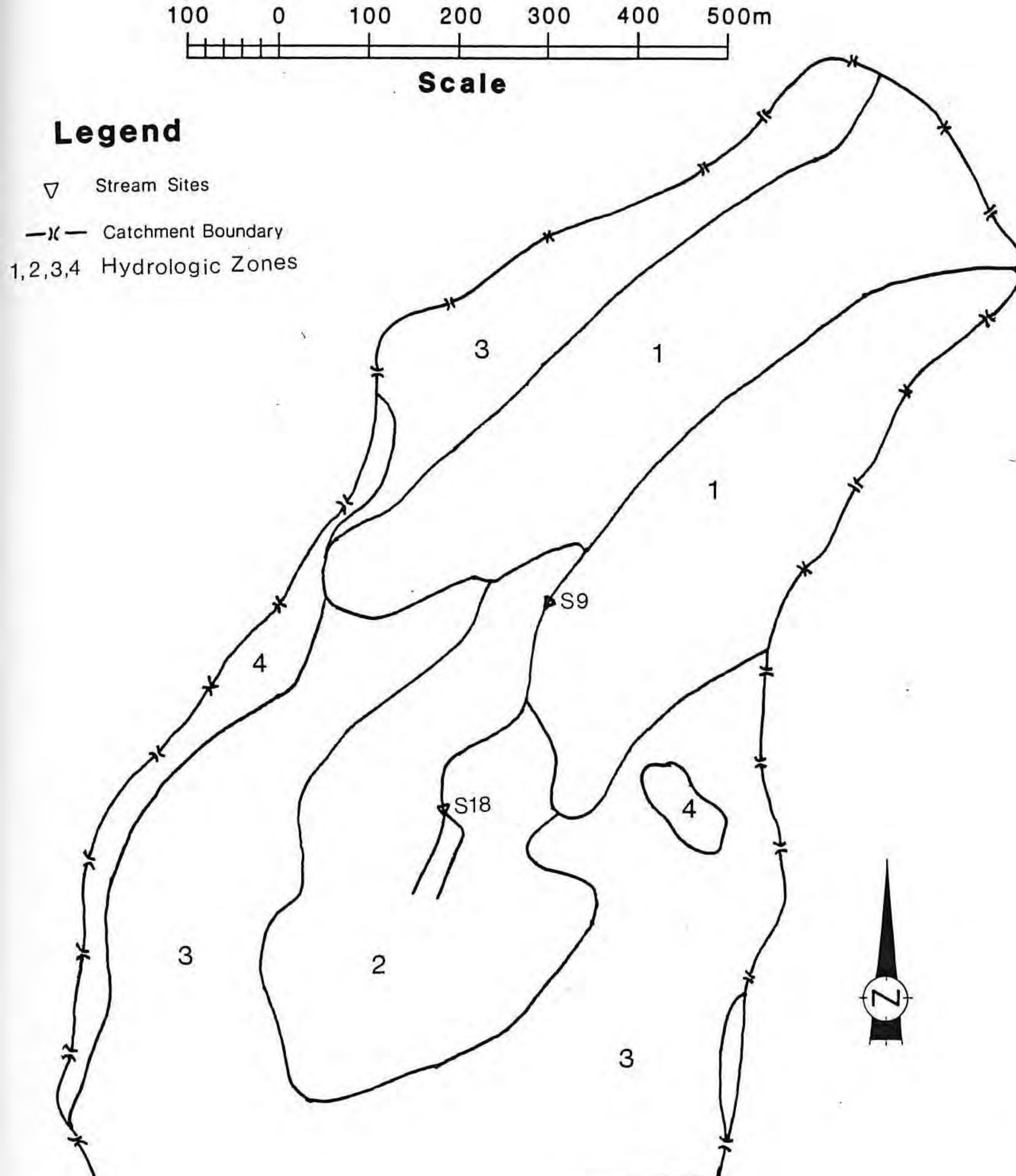


Figure 3.5
VEGETATION OF SALMON CATCHMENT
(From Bettenay et al., 1980)



HYDROLOGIC ZONES OF SALMON CATCHMENT

Figure 3.6

TABLE 3.1 DESCRIPTION OF SALMON CATCHMENT

Hydrologic Province 4 : Area 8.8ha (10.8%)	
Gravelly uplands: very gentle to moderate gradients; no run-off; jarrah-marri forest	
L	Upper slopes with gravelly pale, and siliceous, sands and lateritic duricrust
SE	Upper slopes and divides with gravelly yellow earths, and pale sands
Hydrologic Province 3 : Area 39.0 ha (47.6%)	
Mid-slopes and divides : gentle to moderate gradients; limited run-off; winter saturation and lateral water flow at lower ends; jarrah-marri forest	
Yeg	Upper and mid-slopes with gravelly yellow earths, and pale sands
YE	Upper and mid-slopes with yellow earths
REg	Upper and mid-slopes with gravelly red earths, and bleached sandy red duplex soils
Pg	Upper slopes with bleached sandy red duplex soils
Hydrologic Province 2 : Area 13.7ha (16.8%)	
Valley flanks: gentle to moderate slopes; winter saturation and lateral water flow; jarrah-marri forest	
RE	Valley sides and heads with red, and yellow, earths
Sw	Permanent and seasonal swamps with red, and yellow, earths
Hydrologic Province 1 : Area 20.3 ha (24.8%)	
Valley slopes: moderate gradients; winter saturation and lateral water flow; surface run-off from rock outcrops and shallow rocky soils; jarrah, marri and yarri forest, with some areas of low scrub	
REr	Valley slopes with stony red earths
YEr	Valley slopes with stony yellow earths
R,+	Major rock outcrops with limited red, and yellow, earths

(From Bettenay et al., 1980)

The shallow soils, rock outcrops and sandy loams of province 1 (25%) along the steeper hillsides flanking the lower streamline are potentially high rainfall-runoff areas. In the stream headwater area the duplex soils of province 2 (17%) were described by Bettenay et al. (1980) as a zone of saturation during winter with lateral subsurface flow. Upslope, some perching of shallow groundwater may occur in the sandy, gravelly, yellow and red earths of province 3 which overlie mottled and pallid clays at about 1m depths. Surface soil saturation in province 2 was thought to be sustained by lateral, subsurface flow from the perched groundwater province 3. No surface runoff and probably little lateral flow is likely from the deep, lateritic soils of province 4 on the uplands of the catchment.

3.7 Instrumentation

3.7.1 Rainfall

Two tipping-bucket rain gauges, recording on Leopold and Stevens graphical recorders, are located in and near the catchment (Figure 3.3) and operated by the Water Authority.

3.7.2 Streamflow and Salinity

A combination, sharp-crested, 90^{o}V notch weir with wing walls was constructed in April 1974 on a rock-bar at the catchment 'outlet'. Streamflow is recorded continuously by a float operated Leopold and Stevens graphical recorder with a Rimco punched-paper tape recorder for back-up. The theoretical stage-discharge rating has been checked by meterings and over ten years less than 0.2% of record has been lost.

Streamflow water quality has been obtained by an automatic pumping sampler. Samples have been routinely analysed for electrical conductivity, chloride concentration, temperature, turbidity and sediment and selected samples have been analyzed for major ions. Since 1980 a continuous record of electrical conductivity has been obtained by a toroidal cell recording onto punched paper tape.

In 1983 a small portable weir was installed at S9, 700m upstream of the main gauging site, and instrumented with a float-well and a graphical recorder. This site was selected to quantify the contribution of streamflow between S9 and the main station.

In 1984 the weir was moved upstream to a site downstream of the confluence of the two small channels at site S18 (Figure 3.3). This records streamflow from the saturated heatwater areas with a catchment of 33 ha.

3.7.3 Water Balance Sites

In 1973 five sites, each of about 0.1 ha, were selected by C.S.I.R.O. (Division of Groundwater Research) as water balance sites in Salmon catchment (Figure 3.3). At each site a storage rain gauge was installed to monitor rainfall and solute input (Peck et al., 1982b). Three neutron meter access tubes were also installed at each site to a depth of about 6m. A deep bore was drilled to bedrock (or auger 'refusal') with hollow augers and a 40mm diameter PVC pipe, slotted over the bottom 2-3m was inserted. These were designed to monitor the quality and potentiometric level of the deeper groundwater system. In addition a shallow bore was installed into the B horizon at each site, The soil salt storage (Johnston and Williamson, 1981), soil water (neutron count) and groundwater levels and quality (Peck et al., 1982a) were measured by C.S.I.R.O.

3.7.4 Shallow Bores

Bettenay et al. (1980) indicated that the sandy, red and yellow earths of "Hydrologic Province 2" of the convergent headwater area "receive water from upslope and so are frequently water-logged in late winter" and further that these areas have "lateral flow of perched water".

To investigate these groundwaters a transect of shallow ($<$ 3m) bores were installed in the headwater area (Figure 3.3) in 1983 with the aims of:-

(i) observing the development of perched groundwaters in relation to soil type and topography

(ii) relate the shallow bore water levels to the areas of surface soil saturation (variable source areas)

(iii) relate storm period bore water level response to streamflow response.

Cores of about 100mm diameter, in lengths of 900mm, were obtained at the six sites along the transect (where possible) to depths ranging from 1.5m to 3m from the surface. The cores were logged, on site, with particular note made of the structure, colour and the relative sand/gravel/clay content. This information was used to identify depths in the profile which were considered likely to function as perching layers or horizons of relative impedance to the vertical flow of water.

Several bores were installed to various depths at each site with the PVC tube slotted over the section of the profile considered likely to form a perched groundwater. The 44mm diameter PVC was slotted with a hacksaw and inserted into the augered hole with a cap glued to the bottom. A similarly slotted, 150mm diameter PVC pipe was installed at the bottom (Site 1) and top of the transect (Site 8).

At Sites 1 and 8 a continuous record of water level was obtained by float and clock-driven chart recorders (bores 6129321 and 6129342). The water levels in the remaining bores were manually read, at least every 3 weeks, and usually more frequently, except in September 1983. An estimate of peak water level was obtained by granulated cork which adhered to a rough-textured length of rope, weighted to hang down inside the bore.

3.8 Summary

Salmon catchment (81.8ha) is located in the higher rainfall (1150mm yr^{-1}) area of the Western Darling Range near Wellington Reservoir in the south west of Western Australia. The elevation ranges from 180m AHD at the gauging station to 302m at the divide, over a distance of about 1.5km. A first-order stream of approximately 1100m length originates in a broad, convergent headwater area of relatively flatter slopes, and becomes progressively more incised downstream.

Surface soils range from lateritic (duricrusts) on the divide, through gravelly (yellow) earths to loamy red earths and to shallow stony soils and areas of exposed basement in the mid and lower catchment.

The catchment vegetation is dominated by a relatively dense eucalypt forest overstorey of 20 - $30m^2$ ha^{-1} basal area with a leaf area index approaching 2.0. The soil/vegetation associations were grouped into four 'hydrologic zones' on the basis of general (expected) hydrological responses.

The catchment is (relatively) extensively instrumented for the measurement of streamflow, rainfall and saltfall. Less extensive measurements have been made of the deeper and shallower groundwaters and the (unsaturated) soil water.

4. CATCHMENT WATER AND CHLORIDE BUDGETS

4.1 Introduction

Yearly water and chloride inputs and stream outputs for Salmon catchment are presented in this chapter for the ten year period commencing in April 1974 (water years). These form the basic data set of information of the integrated response of the catchment as measured at the gauging station.

Analysis of these data will be made in order to develop a general understanding of the dominant hydrological processes of the catchment. Observations of processes within the catchment are presented in later chapters.

4.2 Catchment Water and Chloride Inputs

4.2.1 Measurements

Period rainfalls were measured in seven storage gauges located at the five water balance sites and the pluviometer sites (Figure 3.3). The five gauges at the water balance sites are located under forest canopy and therefore measure net rainfall (throughfall) whereas the two pluviometers are sited in well exposed, forest clearings. Saltfall was determined by measuring the chloride concentration of the water in the storage gauges (Peck et al., 1982b).

4.2.2 Rainfall

Catchment daily rainfall was obtained by Thiessen-weighting the daily rainfalls for the two pluviometers. Yearly rainfall was then calculated by summing over a water year commencing in April for the ten years 1974 to 1983 (Table 4.1).

TABLE 4.1 STORAGE GAUGE WATER AND CHLORIDE INPUTS

Water Year	Water (mm)			Chloride (kg ha^{-1})	
	Pluv.	Exposed	Canopy	Exposed	Canopy
1974	1493	1466	1330	91.6	127.7
1975	1001	1059	892	80.8	112.0
1976	845	867	748	44.5	71.5
1977	978	964	817	57.7	87.5
1978	986	1010	848	67.0	92.6
1979	831	848	681	41.4	67.7
1980	1284	1257	1048	77.5	111.0
1981	1473	1511	1180	107.2	132.8
1982	900	916	NA	63.0	NA
1983	1258	1264	NA	128.0	NA
Mean	1105	1116	943	75.9	100.4

Notes (1) Pluvio totals are for water years commencing April 1 whereas the exposed and canopy totals do not correspond exactly with the water year.

(2) NA is data not available.

The ten year average yearly rainfall was 1105mm which is about 45mm below the longer-term average (1926-1978) of 1150mm (Hayes, personal communication, 1980). Rainfall was above average in 1974, 1980, 1981 and 1983 and well below average in 1976 and 1979. The maximum variations from the longer-term average were +30% in 1974 to -28% in 1979.

Thiessen-weighted rainfalls for the two storage gauges in exposed positions and the five gauges under canopy are listed in Table 4.1. The differences in yearly rainfalls between the pluviometer and exposed gauges is due primarily to the periods of storage gauge measurement not coinciding with the water year from April 1 to March 31. The average exposed gauge rainfall was 1116mm, 11mm greater than the pluviometer average.

Under canopy rainfall which was only available for the eight years between 1974 and 1981 averaged 943mm (84% of the average for the exposed gauges). A linear regression between the exposed (Re) and canopy (Rc) gauge rainfalls produced:-

$$Rc = -1.1 + 0.84\ Re \qquad \text{with } r^2 = 0.945 \qquad 4.1$$

This indicates that net rainfall under canopy is 16% less than at the exposed sites. On average therefore, interception is 16% of gross rainfall.

4.2.3 Chloride

Bulk chloride inputs (wet and dry fall) were calculated for the fully-exposed and under canopy gauges using Thiessen-weightings (Table 4.1). The ten year average input as measured in the exposed gauges was 75.9 kg ha^{-1} with a range from 41.4 kg ha^{-1} in 1979 to

128 kg ha^{-1} in 1983. The eight year canopy input averaged 100 kg ha^{-1} which is 1.4 times the exposed gauge average over the same period.

A linear regression between canopy and exposed saltfall produced:-

$$Lc = 25.3 + 1.057\ Le \quad \text{with } r^2 = 0.976 \qquad 4.2$$

where Le and Lc are yearly exposed and canopy saltfall. The slope of the relationship (1.057) was found to be not significantly different from 1.0 at the 95% confidence level. Therefore it can be concluded that the chloride input as measured under the canopy is some 25kg ha^{-1} higher than that measured in the exposed gauges.

The additional chloride input under canopy may be a result of chloride cycling, within the system, by vegetation (see conceptual model by Johnston, 1984). That is the vegetation transports chloride from the soil to the leaves during transpiration where it is washed back to the surface by subsequent rainfall. Therefore there is no net input to the catchment, only a redistribution.

An alternative has been proposed (e.g. Juang and Johnson, 1967) in which aerosols impinge on the tall vegetation and are subsequently washed-off. In this case there is a net additional input of chloride to the catchment. Sharma et al. (1983) reported an increase in chloride under pines which was attributed to impingement.

The average chloride concentration for the exposed gauges was 6.7 mg l^{-1} (standard error 0.47) over the ten years in comparison with the average of 10.6 mg l^{-1} (0.31 standard error) for the under canopy gauges.

4.2.4 Adjustments for Water Year Period

The pluviometer water year rainfalls were adopted and the canopy rainfalls and exposed and canopy chloride were adjusted to correspond to a water year.

The canopy rainfalls were adjusted by:-

$$R_c^1 = (R_c/R_e).\text{Pluvio} \qquad 4.3$$

where R_c^1 is adjusted canopy, R_c canopy and R_e exposed rainfalls.

Exposed chloride inputs ($L_e{}^1$) were adjusted by:-

$$L_e{}^1 = (L_e/R_e)\ \text{Pluvio} \qquad 4.4$$

Canopy chloride inputs ($L_c{}^1$) were adjusted by:-

$$L_c^1 = (L_c/R_c)\ R_c^1 \qquad 4.5$$

These adjusted rainfall and chloride inputs are listed in Table 4.2. The canopy rain and chloride in 1982 and 1983 were determined from the regression equations 4.1 and 4.2 and then adjusted by the procedure noted above (equations 4.3 to 4.5).

TABLE 4.2 ADJUSTED WATER YEAR CATCHMENT WATER AND CHLORIDE INPUTS

Water Year	Water (mm)		Chloride (kg ha^{-1})		Chloride (mg l^{-1})	
	Exposed (3)	Canopy	Exposed	Canopy	Exposed	Canopy
1974	1493	1354	93.2	130.0	6.2	9.6
1975	1001	843	76.4	105.9	7.6	12.6
1976	845	729	43.4	69.7	5.1	9.6
1977	978	828	58.5	88.7	6.0	10.7
1978	986	827	65.4	90.3	6.6	10.9
1979	831	667	40.6	66.3	4.9	9.9
1980	1284	1070	79.2	113.3	6.2	10.6
1981	1473	1151	104.5	129.5	7.1	11.3
1982	900	756 (1)	61.9	90.3 (2)	6.9	11.9
1983	1258	1057 (1)	127.4	159.8 (2)	10.1	15.1
Mean	1105	928	75.1	104.4	6.7	11.2

Notes: (1) Predicted from equation 4.1 and then adjusted for water year.

(2) Predicted from equation 4.2 and then adjusted for water year.

(3) Exposed are adopted pluviometer rainfall.

4.3 Catchment Water and Chloride Outputs

4.3.1 Data Processing

A continuous record of stream discharge at the gauging station was available over the ten years and a record of stream water quality was obtained by analysis of pumped samples and an insitu continuous electrical conductivity device (refer Chapter 3). Stream water chloride concentration (Cl) was obtained by laboratory analysis of the pumped samples or by regression with the electrical conductivity (EC).

$$Cl = -13.3 + 2.97\ EC \qquad 4.6$$

with $r^2 = 0.97$

The stream water chloride load was obtained by interpolating a concentration for each discharge (q, m^3s^{-1}) between chloride samples; multiplying the discharge and concentration to obtain a chloride flux (f; kg s^{-1}) and integrating the discharge and flux to produce daily and then yearly flow volume and chloride load.

This procedure is accurate where there are enough samples to adequately define the variation of concentration with discharge as was the case from 1977 to 1983. Prior to 1977 samples were obtained daily and this may have resulted in an over estimation of the chloride load because the lower frequency of samples would bias towards higher concentrations during recession flows. Therefore a flow-weighted calculation was used to check this procedure.

In the flow-weighted procedure, all sample chlorides (Cl) are multiplied by the discharge (q) at time of sampling, summed and divided by the sum of the sample discharges:-

$$Cl_w = \sum_{i=1}^{n} Cl_i q_i \Big/ \sum_{i=1}^{n} q_i \qquad 4.7$$

where n is the number of samples.

This procedure can be expected to produce a good estimate of the flow-weighted concentration when the numbers of samples is sufficient to sample the bulk of the streamflow and most of the important variations in chlorides. This can be checked by subdividing the continuous discharge record into discharge intervals (on a logarithmic scale usually) and calculating the flow-weighted sample chloride concentration in each interval. The amount of discharge volume in each discharge interval is then multiplied by the flow-weighted sample chloride in each interval. The procedure effectively 'weights' the sampled chlorides and should reduce sample bias by adjusting for the amount of actual flow volume.

4.3.2 Results

The results of the three methods are listed in Table 4.3. The discrete sample flow-weighted methods are lower than the continuous integration method between 1974 and 1977 and generally higher after 1977. The continuous integration method is accurate from 1979 when a record of continuous electrical conductivity was available and was supplemented by the discrete, pumped samples. Therefore the continuous integration method results from 1979 onwards were adopted as the more accurate. Prior to 1979, the flow-volume weighted results were adopted.

TABLE 4.3 YEARLY FLOW-WEIGHTED CHLORIDE CONCENTRATION (mg l^{-1})

Water Year	Continuous Integration Method (1)	Discrete Sample Method (2)		
		Number of Samples	Flow-Weighted	
			Method (i)	Method (ii)
1974	62	170	52	52
1975	128	185	116	123
1976	NA	175	257	263
1977	120	119	110	111
1978	105	236	106	108
1979	207	157	205	206
1980	86	147	93	95
1981	73	282	76	81
1982	115	200	101	115
1983	58	236	58	63

NA : Not Available

Notes: (1) Integration method involves continuous interpolation between samples to produce file of continuous chloride flux, similar to discharge. Continuous electrical conductivity record available 1979 to 1983.

(2) Discrete Sample Method

(i) Sample flow-weighted

(ii) As for (i) and adjusted for actual flow volume in discrete discharge intervals

The water year stream outputs of water and chloride are listed in Table 4.4. Streamflow varied by a factor of more than twenty between the low of 17mm in 1979 to the highest of 367mm in 1974. The arithmetic and geometric mean yearly water yields are 125mm and 85mm respectively with corresponding chloride yields of 100kg ha^{-1} and 91 kg ha^{-1}.

Yearly flow-weighted chloride concentrations varied from 52 mg l^{-1} in 1974 to 263 mg l^{-1} in 1976 with an average of 80 mg l^{-1} over the ten years (total chloride divided by total water).

TABLE 4.4 CATCHMENT WATER YEAR WATER AND CHLORIDE OUTPUTS

Water Year	Water (mm)	Chloride	
		kg ha^{-1}	mg l^{-1}
1974	367	191	52
1975	84	103	123
1976	20	53	263
1977	74	82	111
1978	77	83	108
1979	17	35	207
1980	138	116	86
1981	173	127	73
1982	66	77	115
1983	230	134	58
Mean	125	100	

4.4 Catchment Water and Chloride Budgets

4.4.1 Water Balance

The water balance for a catchment with a non-leaky basement can be written as:-

$$P = Q + I + T + W + G \qquad 4.8$$

where P is rain, Q streamflow, I is interception, T is transpiration, W is change in soil water storage (generally unsaturated) and G is change in groundwater storage (saturated).

Using the totals (or means) over the ten years (Tables 4.2 and 4.4) equation 4.8 becomes:-

$$1105 = 125 + 0.16 \times 1105 + T + W + G \qquad 4.9$$

$$\text{i.e.} \qquad T + W + G = 803 \qquad 4.10$$

If W = 0 over the period, as was indicated to be generally the case from year to year by Sharma (1983) then:-

$$T + G = 803\text{mm} \qquad 4.11$$

Sharma (1983) calculated G to be -46mm over 1974 - 1978 (5 years) so if this was extrapolated to 10 years, to be -100mm, then ($G = -10\text{mm yr}^{-1}$) and

$$T = 813\text{mm yr}^{-1}$$

That is, transpiration is the major component of the water balance at about 74% with interception 16% and streamflow 10% over the ten years. There has been a net loss of groundwater.

The ratios of catchment water output to input are listed in Table 4.5 for both exposed and canopy input conditions. Streamflow averaged 10% of rainfall (almost 12% of under canopy fall) with a range from 2% in 1979 to 24.6% (27.1% of canopy) in 1974.

4.4.2 Chloride Balance

In contrast to the relatively low water output to input ratios, the ratios for chloride (Table 4.5) averaged 131% for exposed input and 93% for under canopy input with corresponding minima and maxima of 86% (53%) in 1979 to 205% (147%) in 1974. On average therefore Salmon catchment is either in approximate chloride balance as measured relative to under canopy input or is in negative balance (31% excess output) relative to exposed gauges. If the canopy input is from within the catchment then this implies that there is a net loss of chloride. That is:-

$$L_s = L_p - L_Q \tag{4.12}$$

where L is chloride load and subscripts P, Q and S are for precipitation, streamflow and storage change respectively. In addition, L_p might consist of:-

$$L_p = L_e + L_a \tag{4.13}$$

where e and a are subscripts denoting chloride load measured in exposed gauges and the apparent additional load measured under canopy.

That is:- $L_s = L_e + L_a - L_Q$ 4.14

for an approximate chloride balance or

$$L_s - L_a = L_e - L_Q \tag{4.15}$$

for a net loss of chloride.

TABLE 4.5 CATCHMENT WATER AND CHLORIDE BUDGETS

Water Year	Ratio : Output (O) to Input (I) (%)			
	Water		Chloride	
	Exposed (1)	Canopy (2)	Exposed	Canopy
1974	24.6	27.1	205	147
1975	8.4	10.0	135	97
1976	2.4	2.7	122	76
1977	7.6	8.9	140	92
1978	7.8	9.3	127	92
1979	2.0	2.5	86	53
1980	10.7	12.9	146	102
1981	11.7	15.0	122	98
1982	7.3	8.7	124	86
1983	18.3	21.8	105	83
Mean	10.0	11.9	131	93
Std Dev	6.9	7.8	31	24

Note (1) Based on fully exposed rain gauges.

(2) Based on rain gauges under forest canopy.

4.4.3 Water and Chloride Budgets

The yearly chloride and water balance ratios are plotted in Figure 4.1 for exposed and Figure 4.2 for under-canopy input conditions. These illustrate the differences between the exposed and canopy input conditions where the balances vary from a net export of chloride to a net gain respectively.

An important feature of the data is the variation between 1976 and 1979 and between 1983 and 1974. The difference in chloride balance between 1976 and 1979 is significant because there is very little difference in the water balance. This difference may be the result of output of chloride from the deeper, more saline, permanent groundwater.

The output of chloride in 1974 was also much greater than the input, in contrast to 1983 when the balance was much closer to one. However there is some doubt as to the accuracy of the 1983 chloride input because the average concentration was 10.1 mg l^{-1} (Table 4.2) which is significantly higher (five standard deviations) than the average of the other nine years. Applying the average concentration to the 1983 exposed rain input would produce an output to input of about 169% which would plot between the 1974 data and the other data.

Another possibility is that the output of chloride in streamflow in 1983 was less because the deeper groundwater did not contribute significant chloride. A discussion of the deeper groundwater system in Salmon catchment is presented in Chapter 6.

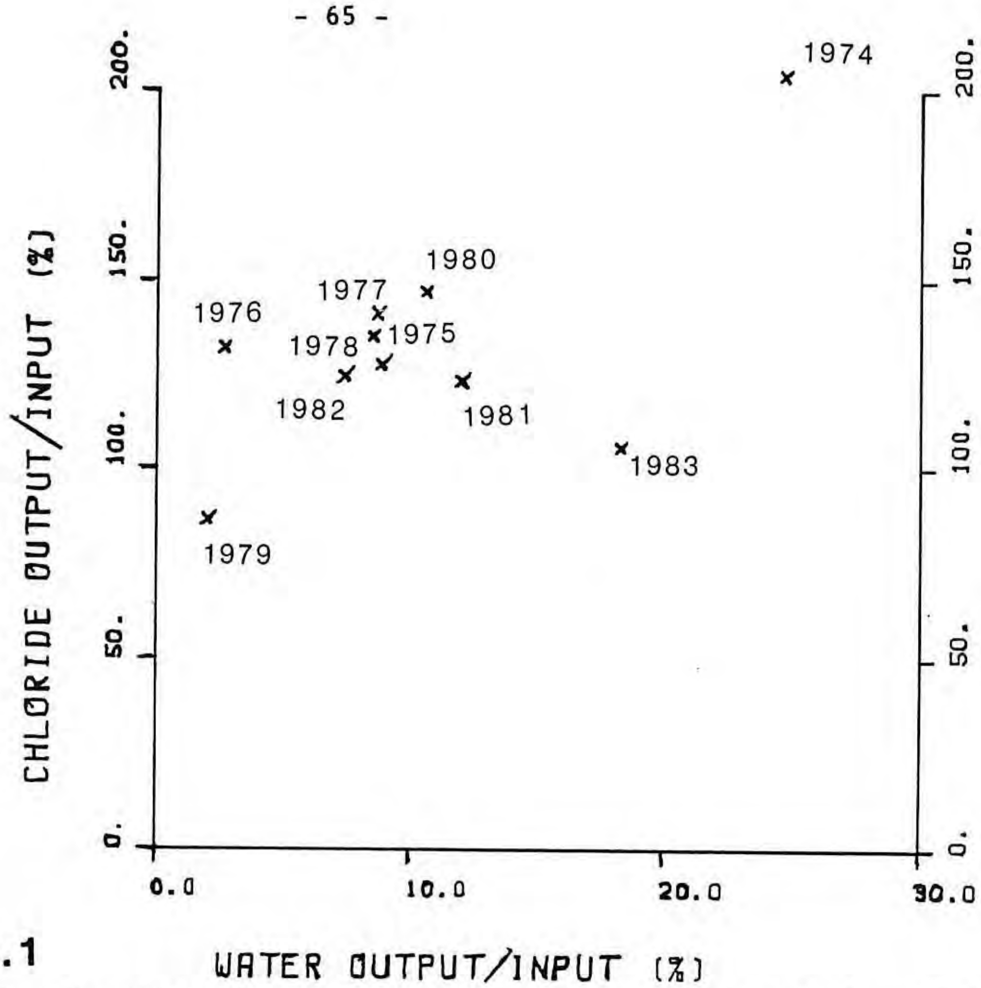


Figure 4.1

CATCHMENT CHLORIDE AND WATER BUDGETS : EXPOSED

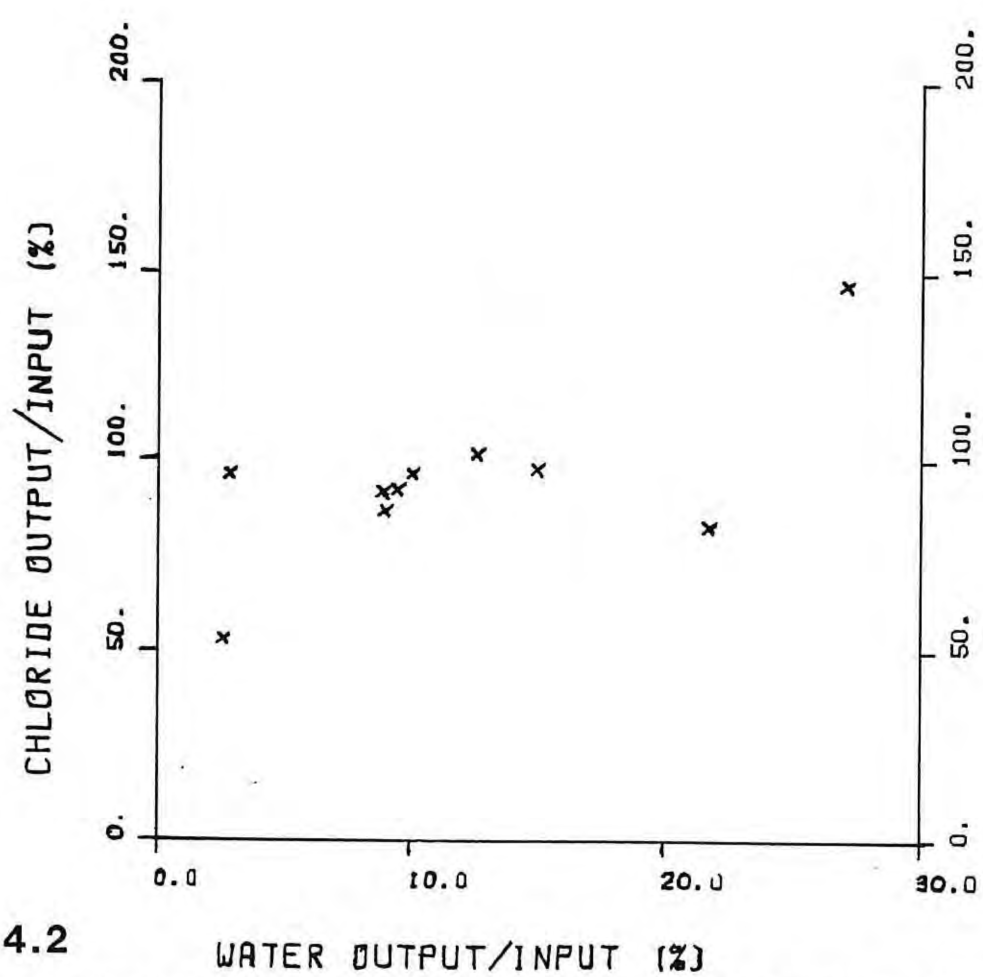


Figure 4.2

CATCHMENT CHLORIDE AND WATER BUDGET : CANOPY

The chloride output will be zero when there is no streamflow and therefore any relationships between the chloride and water balances must include the origin. The data indicate that the chloride balance should exceed 75% at a water balance of about 2.5% and greater than 100% by at least 5% water output to input for exposed input conditions. In contrast the chloride balance only exceeds 100% of canopy input when the water balance approaches 10%, double that for the exposed condition.

In general, Salmon catchment has a chloride output to input which is about ten times more efficient in the output of chloride than in the output of water.

4.5 Summary

Comparison of pluviometer rainfall and storage gauge rainfalls indicated a good relationship and the pluviometer data were adopted. Average yearly rainfall between 1974 and 1983 was 1105mm, about 45mm below the longer term average with only four of the years above the average.

Thiessen-weighted rainfall measured in gauges under the canopy was 84% of that measured in gauges in well-exposed forest clearings. Net interception was therefore about 16% of average rainfall, or 177mm yr^{-1}.

Catchment chloride input as measured under forest canopy was approximately 25 kg ha^{-1} greater than that measured in the exposed gauges.

Average streamflow over the ten years was 125mm with a range from 17mm (1979) to 367mm (1974). An approximate water balance indicated that transpiration was 74% of average rainfall, which when combined with the interception means that average evapotranspiration was about 90% (995mm) of average rainfall.

The ratio of catchment chloride output (streamflow) to input (precipitation) averaged 130% as measured in the exposed gauges and 93% by the under canopy gauges. If the additional chloride input under canopy is largely due to an internal cycling by the vegetation then Salmon catchment is producing more chloride in streamflow than is being input and therefore is in a state of negative chloride balance (output exceeds input).

The relationship between chloride output (O) to input (I) and water O/I is strongly non-linear. This and the negative chloride balance indicate that a mechanism is operating in the catchment to transport chloride from storage into streamflow. The mechanism is investigated in the following chapters.

5. OBSERVATIONS OF STREAMFLOW SOURCES

5.1 Location and Nature of Sources of Streamflow

An investigation of the nature and distribution of sources of stream water and chloride in Salmon catchment commenced in 1980 with a survey of the stream zone. Inspections were made before, during and following storm runoff and during periods of extended flow recession. Sites were established along the stream and where water was observed to flow from the ground. The location of these sites are shown in Figure 5.1 and some of their features are listed in Table 5.1.

Many of the sources of water were holes located in the stream bank (S13) or seepage areas in shallow depressions (S3, S5). These holes or 'soil pipes' were frequently located towards the base of the earthy, gravelly surface soils, above a clay, hardpan or other impeding layer. Relatively few discrete sources of water were observed along the stream below S17, principally because of the difficulty of access in the heavily vegetated, steep terrain. Three small streams (S7, S8 and S10) were found to discharge into the main stream from the relatively rocky, eastern slope of the catchment. These originate in seepage areas below rock outcrops and from shallow soils.

The seepage from site S11 was observed to infiltrate well short of the stream, some 30-40m downslope of S11. The area between S11 and the swamp (outflow to stream at S12) was seasonally saturated.

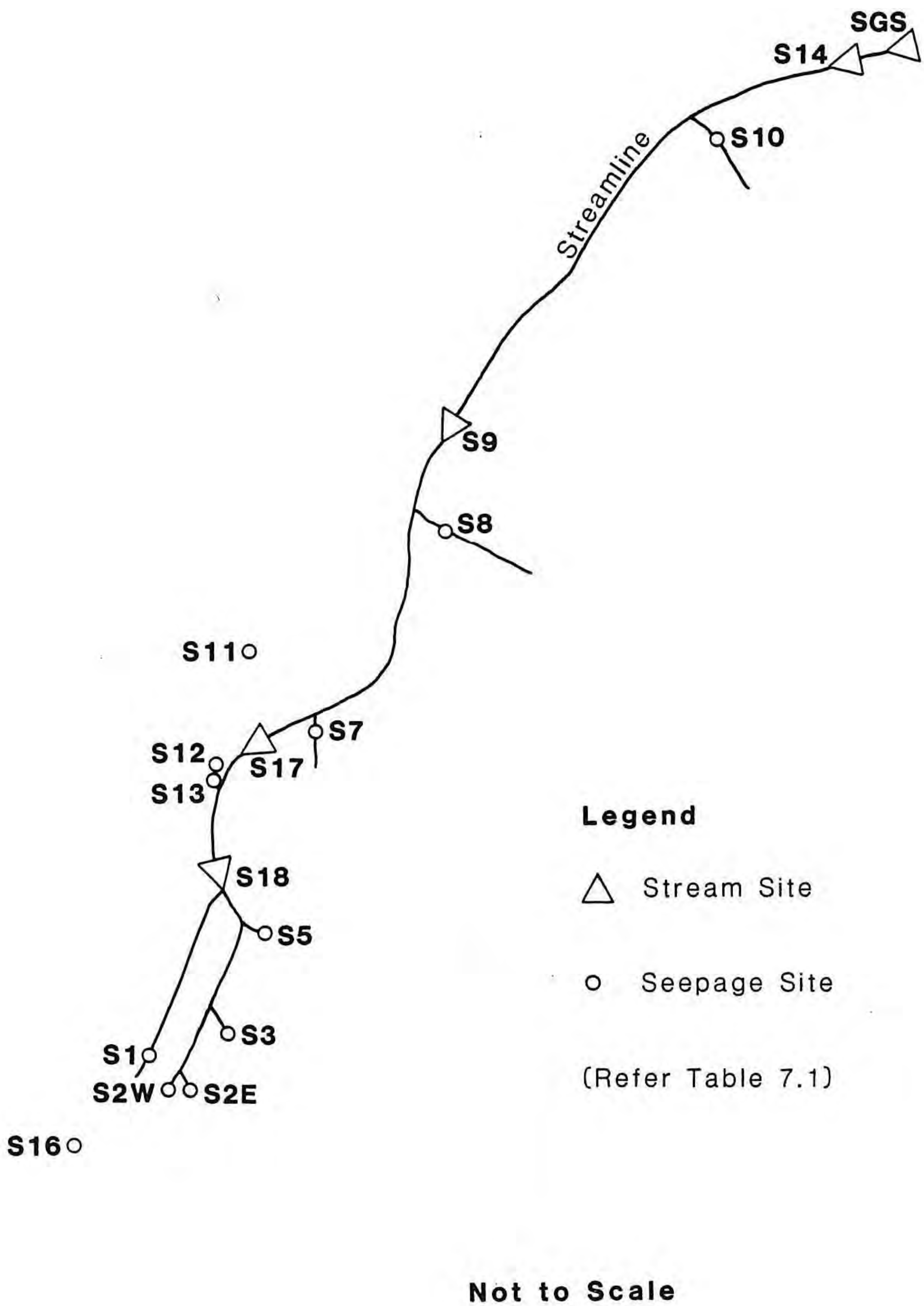


Figure 5.1
SAMPLING LOCATION

TABLE 5.1 SITE CHARACTERISTICS

SITE	CHARACTERISTICS
SGS	At gauging station, (612 011)
S1	Hole in stream bank, 14cm wide, 9cm high
S2W	Diffuse seepage areas, including many small,
S2E	vertical holes in headwater area
S3	Small seepage hollow, many small holes
S5	Hole, 7cm by 3cm, with many smaller holes in side channel seepage area
S7	Small, steep channel draining shallow soils
S8	Side stream, draining soils below rock outcrops
S9	Site on mainstream, 100cm wide, 50cm deep
S10	Very small side channel, source in gravels upslope
S11	Large hole in small channel draining swamp area
S12	Small, short, overland, overflow from swamp
S13	Large hole, 10cm diameter in streambank
S14	Rock bar on main stream
S16	Many small seepage holes in large, transient seepage area in variable source area
S17	Main stream
S18W	West channel of headwater stream
S18E	East channel of headwater stream
S18	Downstream of junction of two streams

5.2 Measurement of Source Water and Salinity

Insitu measurements of water quality were obtained during inspections by use of a conductivity meter and a mercury thermometer. The electrical conductivity was converted to chloride concentration by use of equation 4.6. These were the only measurements taken during 1980.

In 1981, estimates of discharge were obtained by:-

(i) current metering: a small current meter was used in relatively straight channel segments to estimate discharge by the area-velocity method

(ii) volumetric: a portable metal weir with a 'V' spout 15cm long, 10cm deep and with a crest width of 20cm was used for small discharges to direct water into a container which was timed to filling.

Seven sets of water quality measurements were obtained between July 10 and October 22 1980 and the spatial distribution are shown in Figure 5.2. Approximately 82% of water and 74% of chloride were produced between these dates. Eight sets of discharge and water quality were obtained between June 15 and October 29 1981 (Figure 5.3) during which time the yearly proportion of water and chloride yield for 1981/82 went from 2% to 93% and 5% to 89% respectively.

The distribution and variation of chloride concentrations between and within sites are very similar from 1980 to 1981. Concentrations above S18 are significantly less in magnitude and vary less (a range from 27 mg l^{-1} to 90 mg l^{-1}) than at some seepage sites and stream sites downstream.

The concentrations at the three small side streams S7, S8 and S10 are similar to those of some of the headwater sources such as S1 and S5. This indicates a similar type of source.

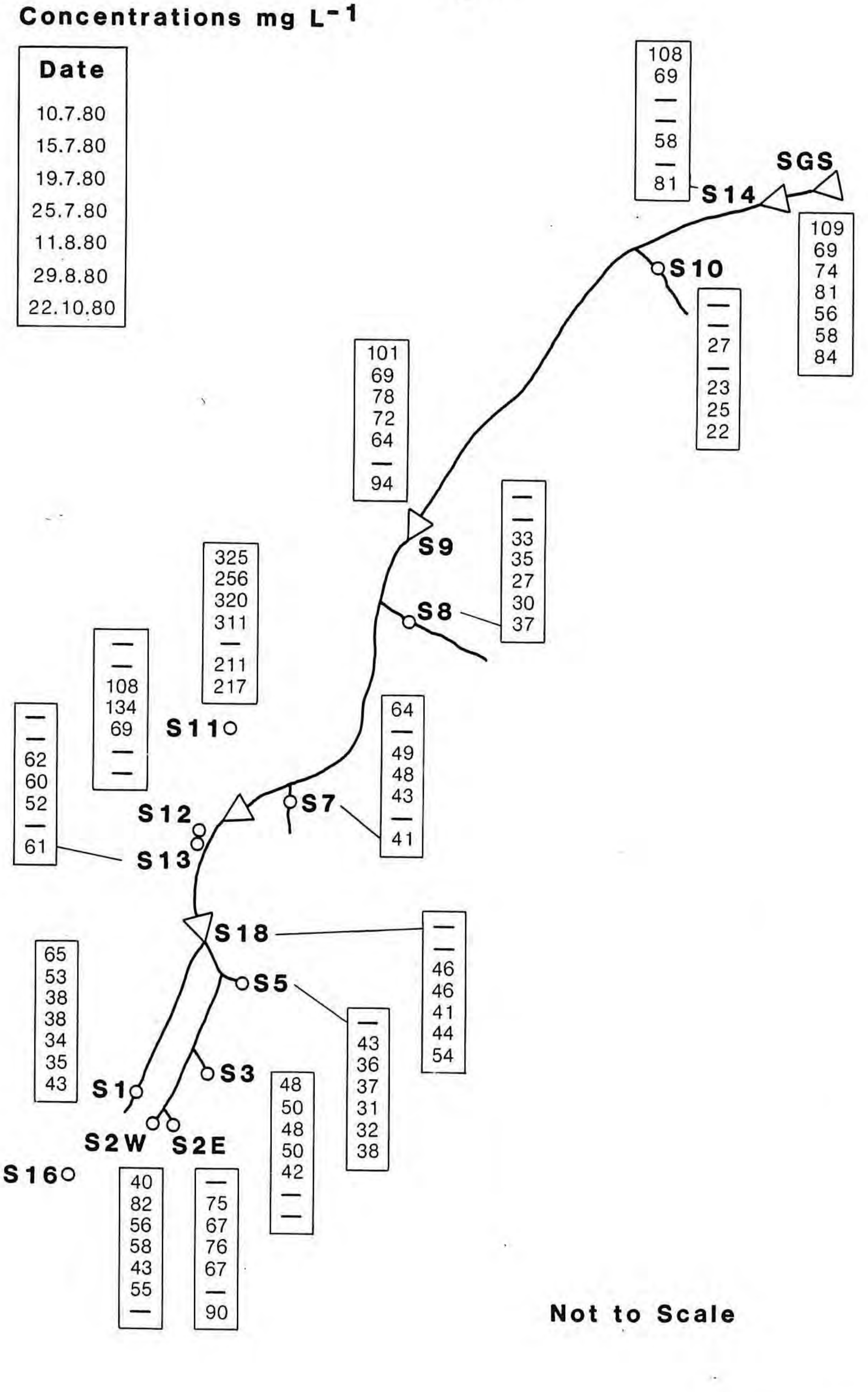


Figure 5.2
SOURCE CHLORIDE CONCENTRATIONS : 1980

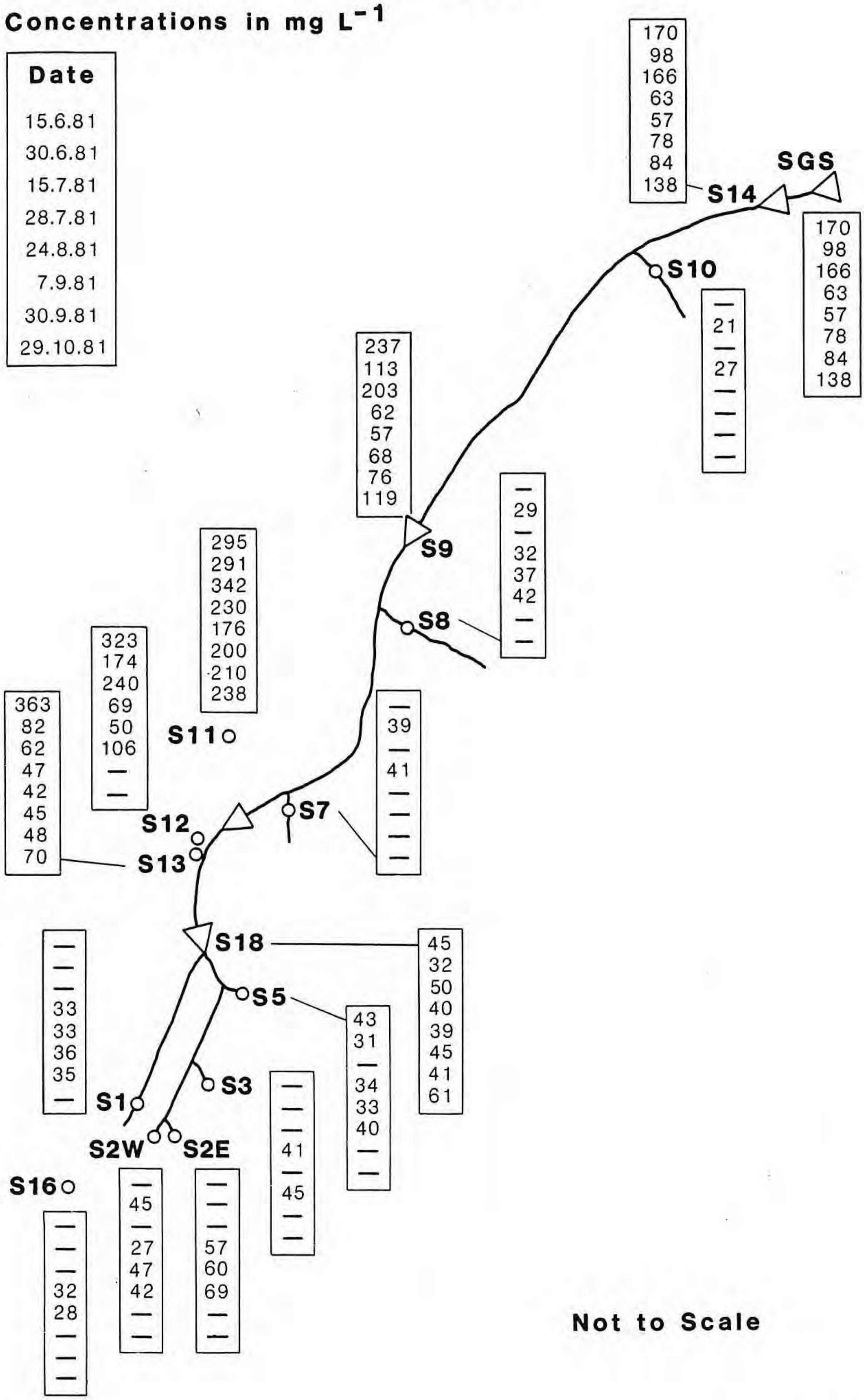


Figure 5.3
SOURCE CHLORIDE CONCENTRATIONS : 1981

In contrast, concentrations from around the swamp, at S12 and S11 are markedly higher, particularly during early and late sampling times. The high concentrations and the variation during the seasons is possibly an indication of the accumulation of soil salt and subsequent leaching.

Stream chloride concentrations at S9 and the main gauging station SGS also vary through the season in a similar manner as the concentrations at S11 and S12.

5.3 Stream Reach Water and Chloride Yields

The contribution of water and chloride at S18 and S9 as proportions of the total at SGS are shown through 1981 in Figure 5.4 and as proportions along the stream between S1 and SGS in Figures 5.5 and 5.6. The September 7 data are possibly misleading because these were taken during a period of substantial rainfall and steady flow recessions cannot be assumed.

Most of the season's water and chloride (93% and 86%) occurred after the inspection on July 15. After this time the proportion from upstream of S9 was more than 50% of SGS with a peak approaching 90% in late July through to late August. The relatively low proportions of 20-40% on June 15 and July 15 occurred before much rainfall had fallen.

From mid-July to late August about 40% of the water at SGS came from the headwater area upstream of S18 (Figure 5.5) with a contribution of chloride of between 20-30% (Figure 5.6). These remained fairly constant until October when the proportion of water and chloride decreased to about 20% and 10% as the catchment dried.

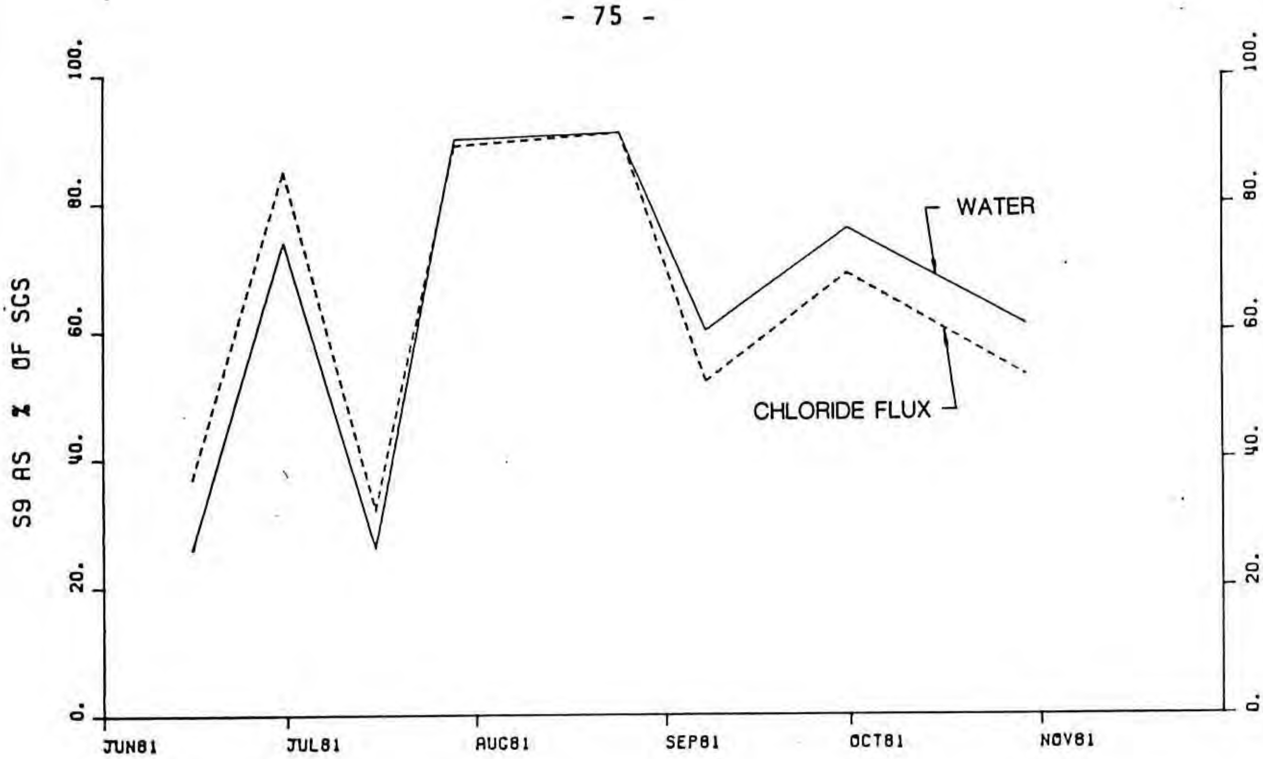


Figure 5.4
PROPORTION OF WATER AND CHLORIDE UPSTREAM OF S9

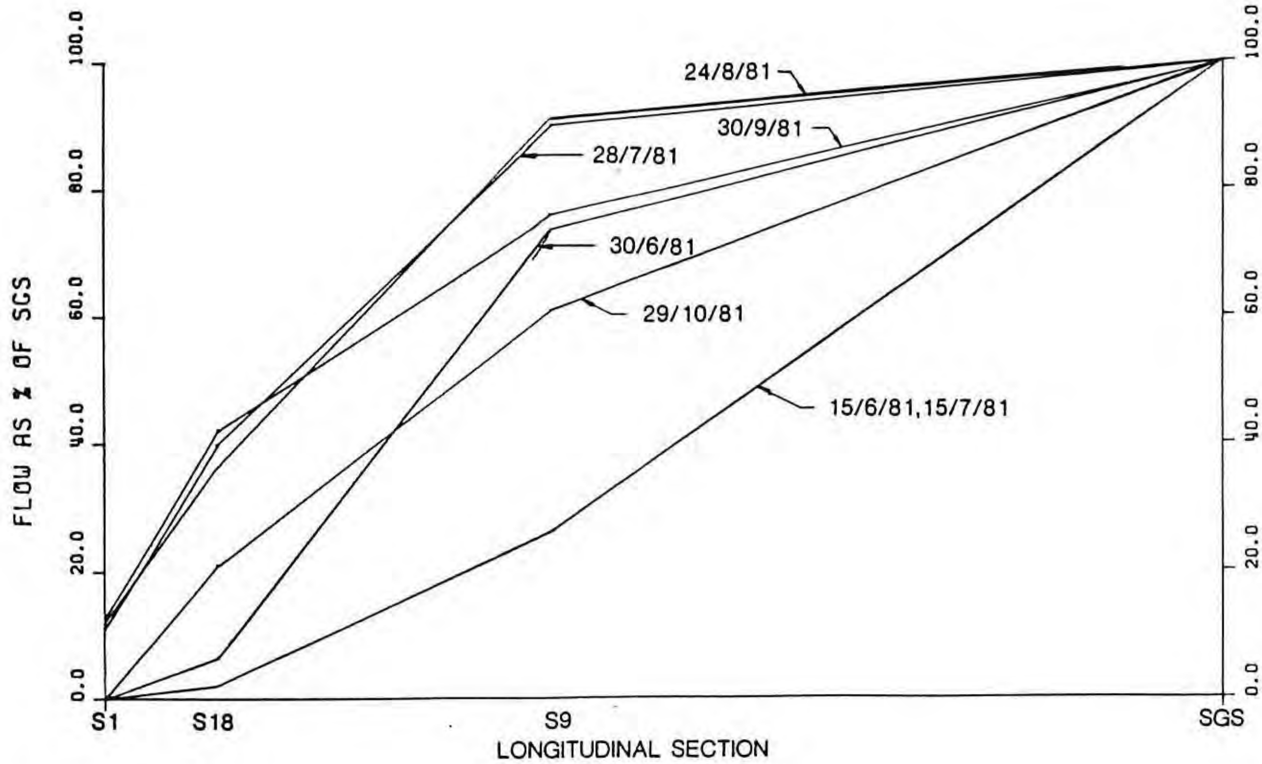


Figure 5.5
PROPORTIONS OF FLOW AT STREAM SITES : 1981

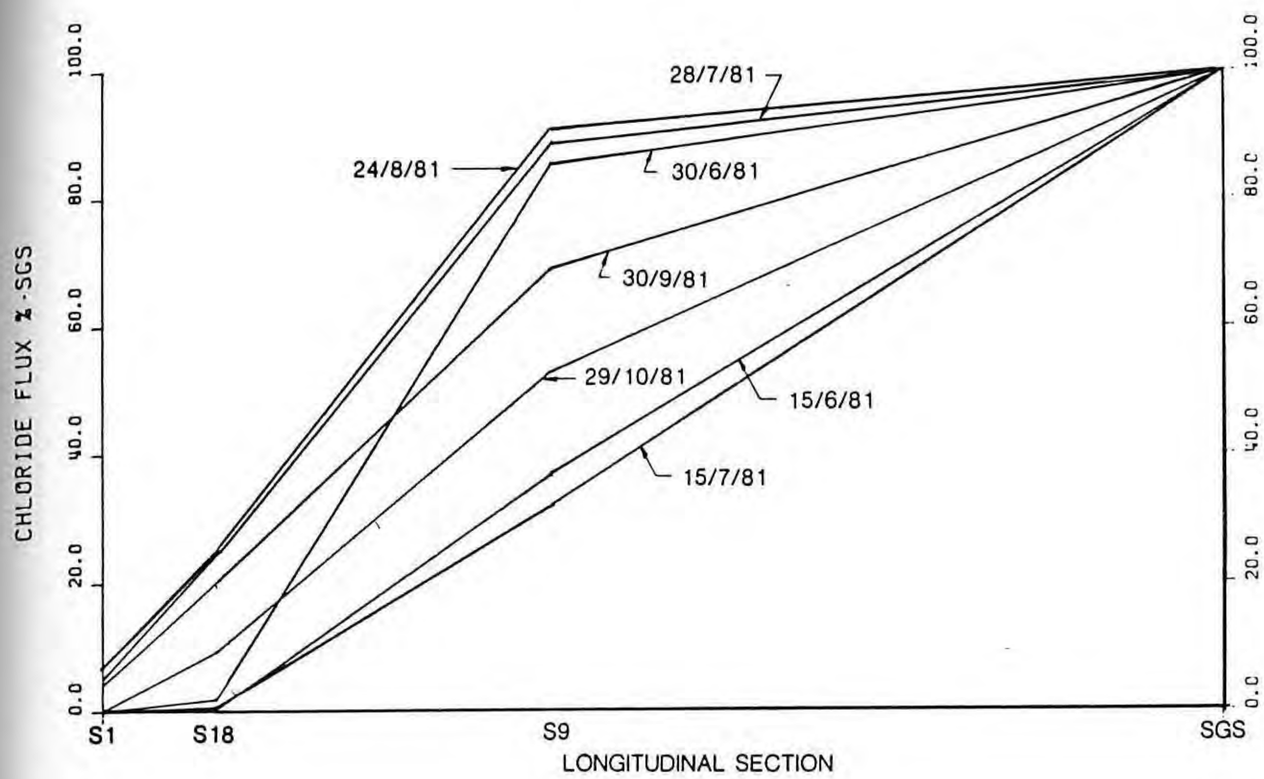


Figure 5.6
PROPORTIONS OF CHLORIDE FLUX : 1981

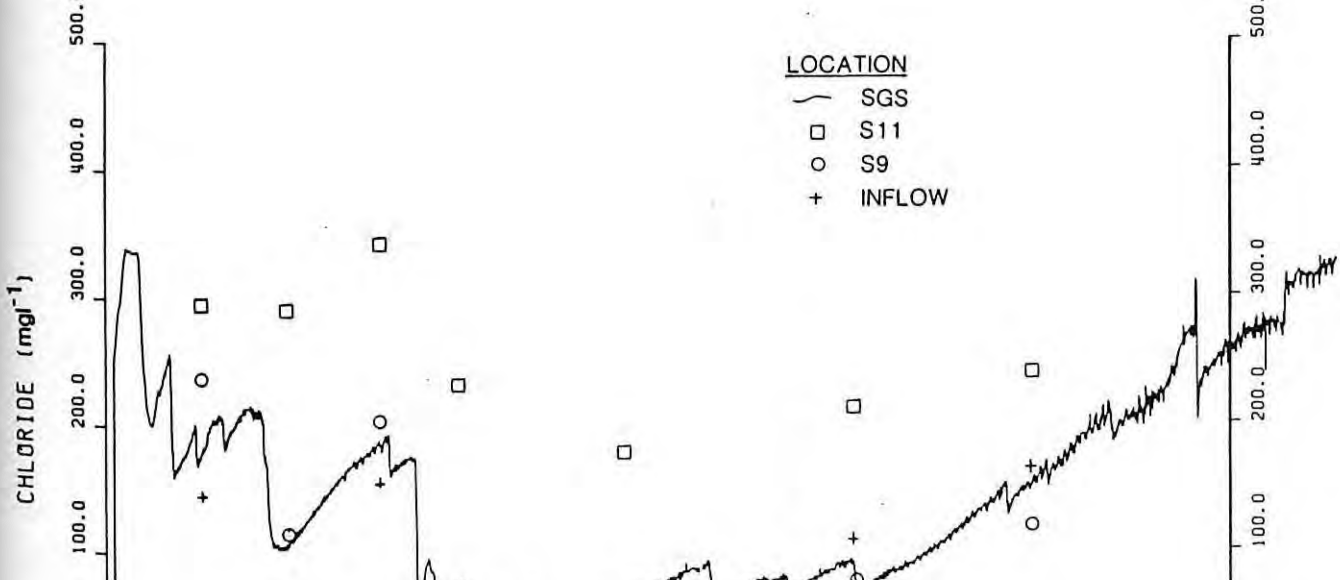


Figure 5.7
CHLORIDE IN REACH S9 TO SGS

The variation of chloride concentration at SGS, S18 and S9 through 1981 is shown in Figure 5.7. Concentrations at SGS decreased from 300 mg l^{-1} at the start of the season to about 50 mg l^{-1} during peak discharge during August. The concentration at S18 remained relatively constant at 40-50mg l^{-1} through the season whereas the chloride at S9 followed the same general trend as at SGS. However the concentrations at S9 were initially higher than those at SGS but decreased to be similar mid season and were generally fresher towards cease-to-flow. This variation in chloride along the streamline between S1 and SGS is illustrated in Figure 5.8 where the large salinity changes between S18 and S9 are emphasised.

The approximate proportions of water and chloride contributed by reaches SGS-S9, S9-S18 and upstream of S18 are summarised in Table 5.2. These are for the July 28 to September 30 period during which 60% of water and 50% of chloride for the year were output by the catchment. These data indicate that most water and chloride is produced upstream of S9 and in particular that most chloride is produced between S18 and S9. It must be emphasised that these data are for part of the total flow period and more importantly are based on just three samples during periods of recession.

TABLE 5.2 REACH PROPORTIONS OF WATER AND CHLORIDE 28/7 - 30/9/81

Reach	S18	S18-S9	S9-SGS
Water (%)	40	35 - 50	10 - 25
Chloride (%)	20 - 25	50 - 65	10 - 30

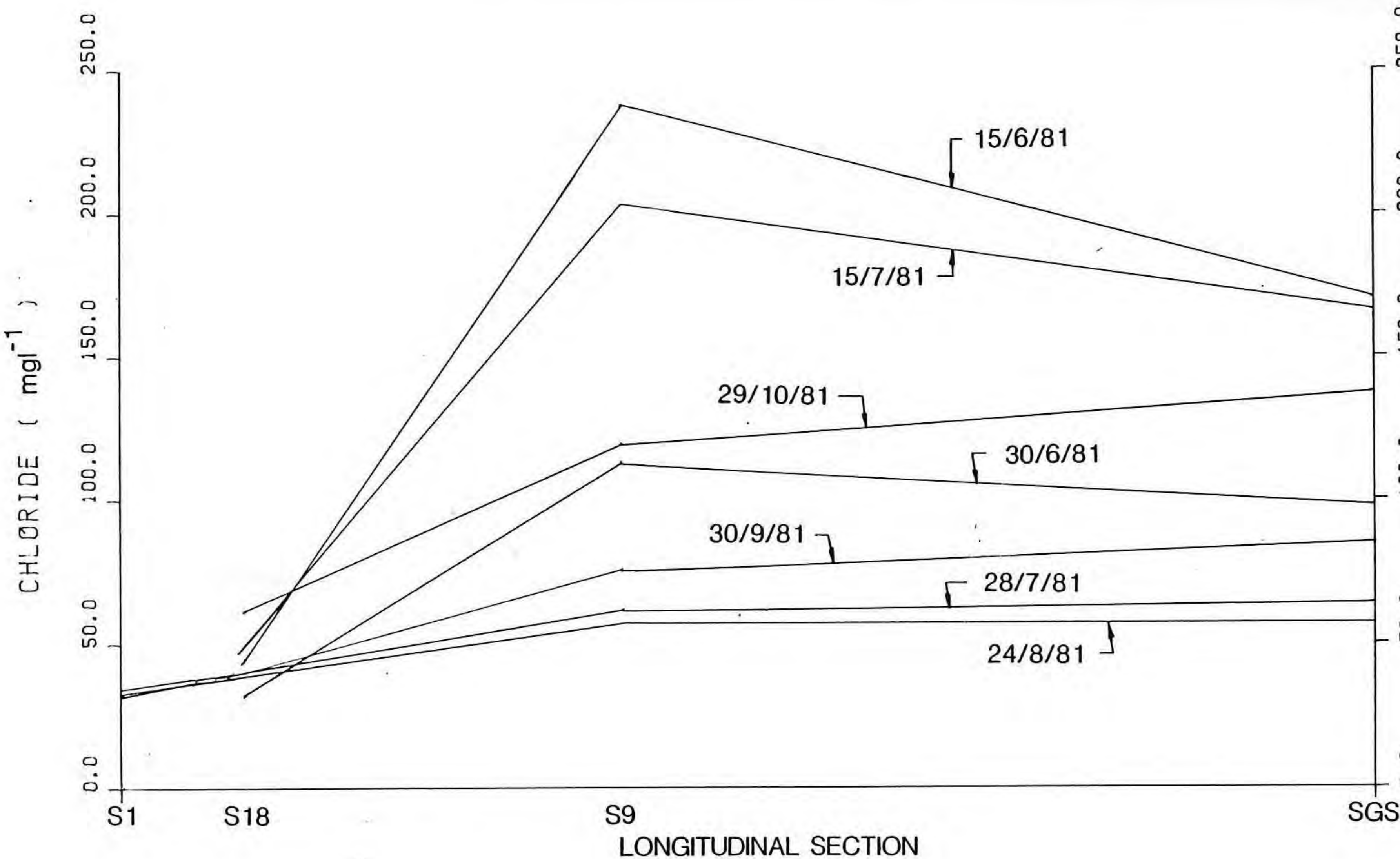


Figure 5.8
CHLORIDE AT STREAM SITES : 1981

5.4 Unmeasured Reach Inflows

The unmeasured inflows of water and chloride for the SGS - S9, S9 - S18 and upstream of S18 reach are listed in Tables 5.3, 5.4 and 5.5 respectively. These were calculated by the difference of measured inflows and measured outflows and are expressed as percentages of the measured outflow and as a chloride concentration.

Unmeasured inflows between SGS and S9 were generally small, as a percentage outflow at SGS, during higher discharges. The chloride concentration was lower than that at S9 or SGS early in the season, about the same mid winter and significantly higher towards the end of flow (Table 5.3). This may be evidence for a change in the relative contributions of a deeper, more saline groundwater and the seasonal fresher groundwater. As the shallow, seasonal groundwater decline in the last part of the season and the deeper groundwater reach their peak (see Chapter 6) the inflow concentration would be expected to increase.

The salinity of the unmeasured inflow between S9 and S18 (Table 5.4) is markedly higher than that at S9, particularly early and late in the 1981 flow season. The inflow salinity decreases from 321 mg l^{-1} chloride at the start of flow to 69 mg l^{-1} by mid-winter as the percentage of unmeasured water inflow decreases from 71% to 49%. However later in the season the proportion of unmeasured water decreases with an increase in the chloride concentration. This is probably a consequence of the relatively greater contribution of water from upstream of S18 and of a contribution of a more saline groundwater within the reach.

TABLE 5.3 INFLOW IN REACH SGS to S9

DATE	SGS FLOW (m^3s^{-1})	SGS CL (mgl^{-1})	UNMEASURED INFLOW WATER (%)	UNMEASURED INFLOW CHLORIDE (%)	UNMEASURED INFLOW CHLORIDE (mgl^{-1})
15/6	1.25	170	74	63	145
30/6	7.7	98	26	15	57
15/7	0.5	166	74	68	153
28/7	15.5	63	10	11	69
24/8	11.0	57	9	9	57
7/9	7.7	78	40	48	94
30/9	3.8	84	24	31	109
29/10	1.0	138	39	47	166

Note (i) all flows 10^{-3} m^3 s^{-1}

(ii) % Cl are of chloride flux in kg s^{-1}

TABLE 5.4 INFLOW IN REACH S9 to S18

DATE	S9 WATER (m^3s^{-1})	S9 CL (mgl^{-1})	UNMEASURED INFLOW WATER (%)	UNMEASURED INFLOW CHLORIDE (%)	UNMEASURED INFLOW CHLORIDE (mgl^{-1})
15/6	0.33	237	71	96	321
30/6	5.7	113	72	76	119
15/7	0.13	203	52	34	132
28/7	14.0	62	43	53	77
24/8	10.0	57	49	59	69
7/9	4.6	68	10	30	211
30/9	2.9	76	12	42	254
29/10	0.61	119	47	61	156

Note (i) and (ii) above

TABLE 5.5 INFLOW IN REACH UPSTREAM OF S18

DATE	S18 WATER (m^3s^{-1})	S18 CL (mgl^{-1})	UNMEASURED INFLOW WATER (%)	UNMEASURED INFLOW CHLORIDE (%)	UNMEASURED INFLOW CHLORIDE (mgl^{-1})
15/6	.006	45	83	83	45
30/6	0.51	32	57	57	33
15/7	0.012	50	-	-	-
28/7	6.1	40	56	48	34
24/8	4.0	39	58	39	41
7/9	2.9	45	13	11	46
30/9	1.6	41	71	75	43
29/10	0.21	61	-	-	-

Note (i) and (ii) above

Unmeasured inflows upstream of S18 (Table 5.5) have average chloride concentrations very similar to those measured at S18 even though the proportion of water changes significantly. The unmeasured concentration were similar to those of the sources and the low salinity and small seasonal variation indicates little if any contribution from a more saline groundwater in the reach upstream of S18 in 1981.

5.5 Contributions of Point Sources

5.5.1 Source S1

The large hole in the stream bank in the headwater area at S1 contributed 10% of the water and between 5% and 10% of the chloride recorded at SGS between late July and late September 1981. The chloride concentration was about 34 mg l^{-1}, about 5-10 mg l^{-1} less than at S18 and varied relatively little. Between 28% and 33% of the water at S18 came from S1.

5.5.2 Source at S13

This source occurs between the S18 and S9 sites and appears to have been a large root, 10cm in diameter, located a few metres upstream of the S12 outfall, on about a level with the stream channel invert. In June 1981 saline upwellings were noted from the bed of the stream, in-line with the S13 outlet but not from S13 itself. First flows from S13 were relatively saline (363 mgl^{-1}, Figure 5.3) but decreased rapidly to around 40 mgl^{-1}. In general salinities were only slightly higher than those in the stream above the outfall point.

For all measurements, except for the 'dry' ones of 15 June, 15 July and 29 October, discharge from S13 average 0.69 10^{-3} m^3 s^{-1} (range 0.5 - 0.85 x 10^{-3} m^3 s^{-1}). This is remarkably constant and possibly reflects discharge constrained by pipe hydraulics rather than the supply of water. As a proportion of SGS, S13 supplied between 4% and 21% of water and up to 12% of chloride.

5.6 Localised Saline Inflow

The stream reach between S18 (the junction of the two headwater channels) and the main stream site at S9 had been identified as having a significant change in salinity from the 1980 and 1981 data. A detailed stream water salinity and temperature survey was carried out in this reach on August 31 1983 in an attempt to localise the inflow source.

The chloride and temperature results of the general survey on August 31 1983 are shown in Figure 5.9. The more intensive survey results between S18 and S9 are shown in Figure 5.10. Downstream of S18 the salinity (as chloride) increased at the outfall from the swamp, just downstream of S13 and the temperature decreased. Salinity increased steadily downstream of S17. A small decrease in temperature of 0.1^{o}C was noted at the inflow from the fresh source at S7 although this may be a measurement error.

A sharp increase in salinity and decrease in temperature occurred between S8 and S9, over a distance of less than 100m. Inflow from the small sidestream source S8 was notably colder (13.3^{o}C) and may have accounted for the decrease in temperature of the water to S9. However the chloride concentration at S8 was 28 mg l^{-1}, less than that of the main stream upstream of the junction with S8. Therefore inflow from S8 could not be the cause of the increase in

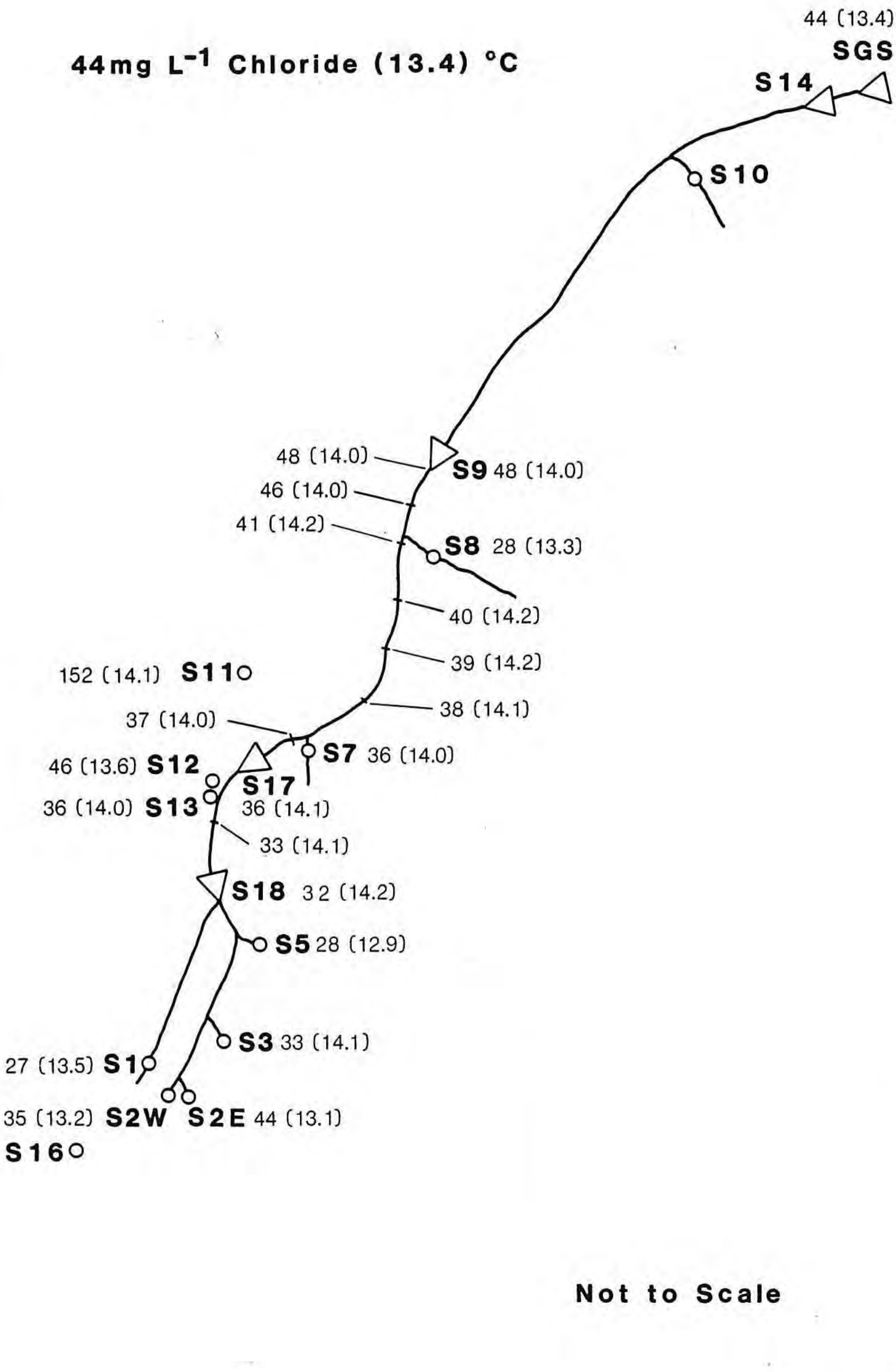


Figure 5.9

SOURCE CHLORIDE AND TEMPERATURE : 31.8.83

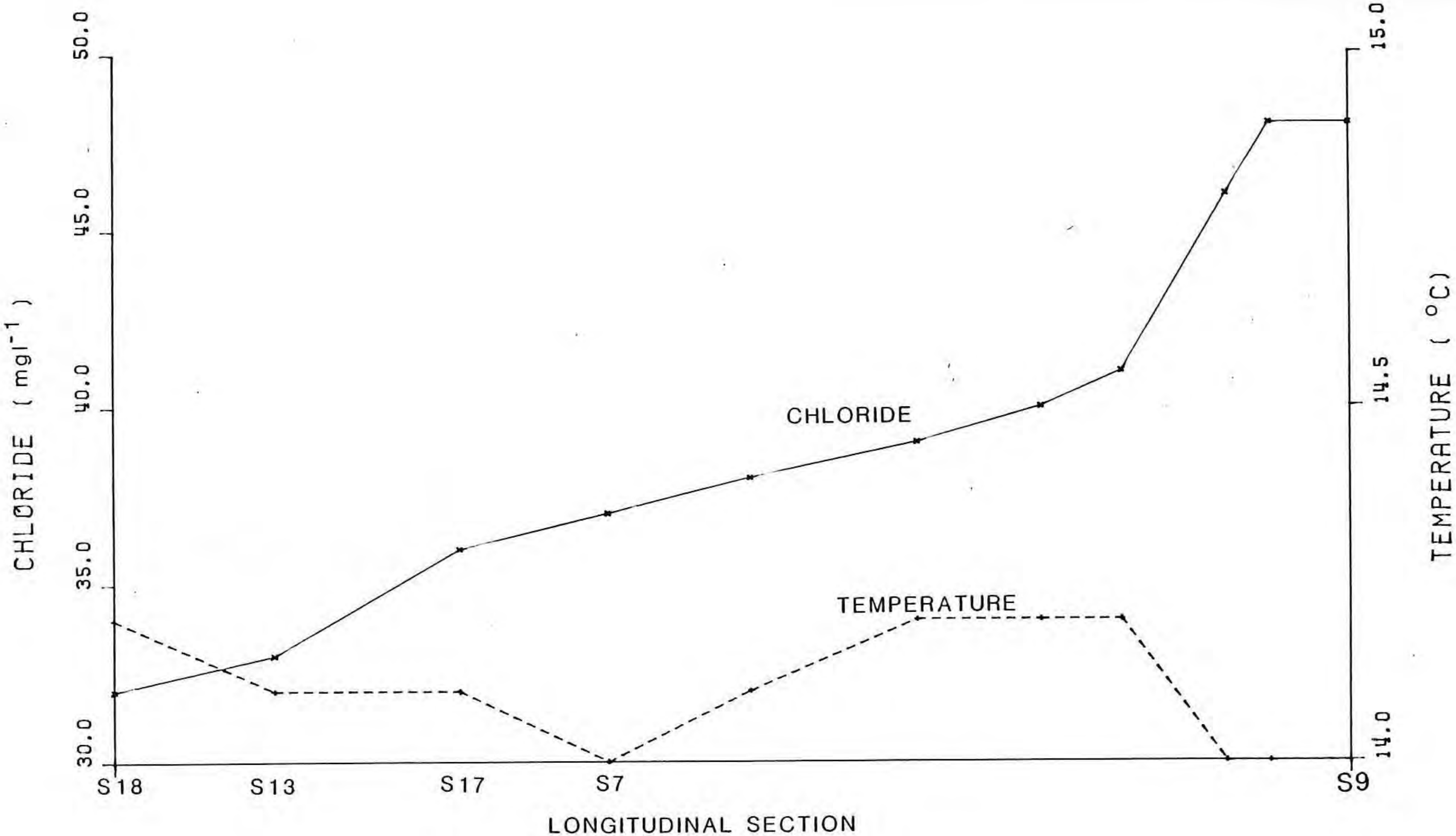


Figure 5.10
STREAM CHLORIDE AND TEMPERATURE : 31/8/83

salinity downstream. It must be inferred that a substantially more saline inflow was the cause of the 7 mg l^{-1} chloride increase to S9.

Water at S11 (152 mg l^{-1} Cl^{-1} and 14.1°C) is an expression of the more saline groundwater at the surface. However water from S11 re-infiltrates 30m downstream of S11 still some 20m short of the main stream channel. This water may enter the stream via a subsurface route downstream of S8.

Using the chloride concentrations of 41 and 48 mg l^{-1} near S8 and at S9 (Figure 5.9) there would only need to be an additional inflow of about 6% of the discharge at S9 (0.02 m^3 s^{-1}), of the quality at S11 (150 mg l^{-1}), to produce the observed 7 mg l^{-1} increase. With a higher inflow salinity, which is possible from the evidence of the deeper groundwater bore data, this would mean an inflow of less than 0.001m^3 s^{-1} on August 31.

5.7 Subcatchment Water and Chloride Budgets

A weir was installed temporarily near sampling point S9 for the 1983 winter and was moved upstream near S18 in 1984. Sampling of stream water quality at S9 in 1983 and S18 in 1984, although relatively infrequent, was used to calculate yearly chloride loads by the methods described in Chapter 4.

The yearly water and chloride yields for the subcatchments in 1983 and 1984 are listed in Table 5.6. Comparison between subcatchments between 1983 and 1984 is not possible firstly because S9 and S18 did not operate concurrently and secondly because 1983 was much wetter than 1984. However the general pattern is that significant runoff and chloride occurs downstream of S18.

TABLE 5.6 SUBCATCHMENT WATER AND CHLORIDE YIELDS

SITE	SGS		S9		SGS-S9		S18		SGS-S18	
Area(ha)	81.8		53.5		28.3		33.0		48.8	
Year	Water	Cl	Water	Cl	Water	Cl	Water	Cl	Water	Cl
1983 (1)	188	10.8	103	5.7	84.6	5.1	-	-	-	-
(2)	230	131	193	107	299	180	-	-	-	-
(3)		58		55		60				
1984 (1)	66.8	5.6	-	-	-	-	16.6	0.6	50.2	5.0
(2)	82	68	-	-	-	-	50	18.2	103	102
(3)		84						36		100

Note (1) Water in 10^3 m^3, chloride in tonnes

(2) Water in mm, chloride in kg ha^{-1}

(3) Average (flow-weighted) chloride in mg l^{-1}

TABLE 5.7 SUBCATCHMENT WATER AND CHLORIDE BUDGETS (%)

WATER Area(ha)	SGS 81.8		S9 53.5		SGS-S9 28.3		S18 33.0		SGS-S18 48.8	
Year	Water	Cl	Water	Cl	Water	Cl	Water	Cl	Water	Cl
1983 (1)	18.3	103	15.3	84.3	23.8	142	-	-	-	-
(2)	21.8	81.4	18.3	66.5	28.3	112	-	-	-	-
1984 (1)	7.2	88.3	-	-	-	-	4.4	23.4	9.0	133
(2)	8.5	64.2	-	-	-	-	5.2	17.0	10.7	96

Note (1) Water and chloride measured in exposed gauges

(2) Water and chloride measured under canopy

(3) 1984 inputs estimated using equations 4.1 and 4.2

In 1983 the 28.3 ha catchment between S9 and SGS produced 23% of rainfall as runoff in comparison to 15% upstream of S9 (Table 5.7). The SGS - S9 area also output between 112% and 142% of chloride input (canopy and exposed conditions respectively) in comparison to the 66.5% and 84.3% upstream of S9. Therefore the lower catchment is more efficient in producing water and chloride from 35% of the catchment area.

The output of water and chloride at SGS in 1984 were less than half those of 1983 (Table 5.6). Water and chloride output at S18 was 25% and 11% of SGS output in 1984 with corresponding output to input ratios of 4.4% and 5.2% and 23.4% to 17% (exposed and canopy) respectively. Most of the water and chloride was therefore produced downstream of S18 where in 1984 the chloride balance was 96% to 133% of input.

These results of the distribution of water and chloride are generally similar to those obtained by infrequent sampling in 1980 and 1981. The important difference is that the area downstream of S9 contributes significant amounts of water and chloride (45% and 47% respectively in 1983). This illustrates the difficulty of obtaining reliable spatial distributions of catchment yield from infrequent measurement.

The 1983 and 1984 results also indicate the relative efficiency of the distinct zones of the catchment in producing streamflow. Relatively little stream water and chloride were produced from the shallow slopes of the headwater area even though this area is 40% of the catchment. In contrast, the steeper slopes and shallow soils downstream of S18 produce significant runoff and notably, chloride output to input ratios of greater than 100%.

5.8 Summary

The location, water quality and water yields of sources of streamflow indicate that the catchment can be considered as three zones:-

(i) headwater area: streamflow generation in the shallower slopes of the topographically convergent upper catchment above the swamp originates at distinct holes in the stream bank and as diffuse seepage in depressions. Chloride concentrations are generally 20 - 40 mg l^{-1}.

(ii) middle catchment: the catchment area between the swamp outfall at S12 and the mainstream site at S9 produces significantly more water and particularly chloride than upstream of S18. In this zone the topography steepens, the stream is more incised and bedrock is exposed so that the deeper groundwater probably contributes to streamflow.

(iii) lower catchment: the 700m reach (35% of total catchment area) between S9 and SGS has steep side slopes and a narrow valley through which the stream wanders. This zone is the most efficient in producing runoff from rainfall (24% in 1983) although there was little evidence of significant surface runoff. Production of chloride was also relatively high in this zone, thus indicating the significance of subsurface flow to the stream.

6. DEEPER GROUNDWATERS

6.1 Source of Groundwater Data

Bores were installed by C.S.I.R.O. (Peck et al., 1982a) to monitor the deeper groundwater system in Salmon catchment. Selected portions of these data are used in this study to assess the role of the deeper groundwater in streamflow.

6.2 Areal Extent of Groundwater

The seven deep bores to bedrock are concentrated at the five water balance sites (Figure 3.3). Four of the sites are located in the upper half of the catchment and clearly there is limited information on the extent of the deeper groundwater, particularly in the lower half of the catchment.

All of the bores have permanent water levels above bedrock and in the upper half of the catchment the groundwater appears to extend over most of the area of the catchment surface divide. Depths to bedrock at bores 1251, 1351, 1451 and 1551 are approximately 25m with depths of saturation of 20m, 11-15m, 1-4m and 8-13m respectively. Although piezometric levels have been above ground at 1251, water content analysis of samples for the core (Johnston and Williamson, 1981) indicated saturated soil conditions between 4m and 5m below ground.

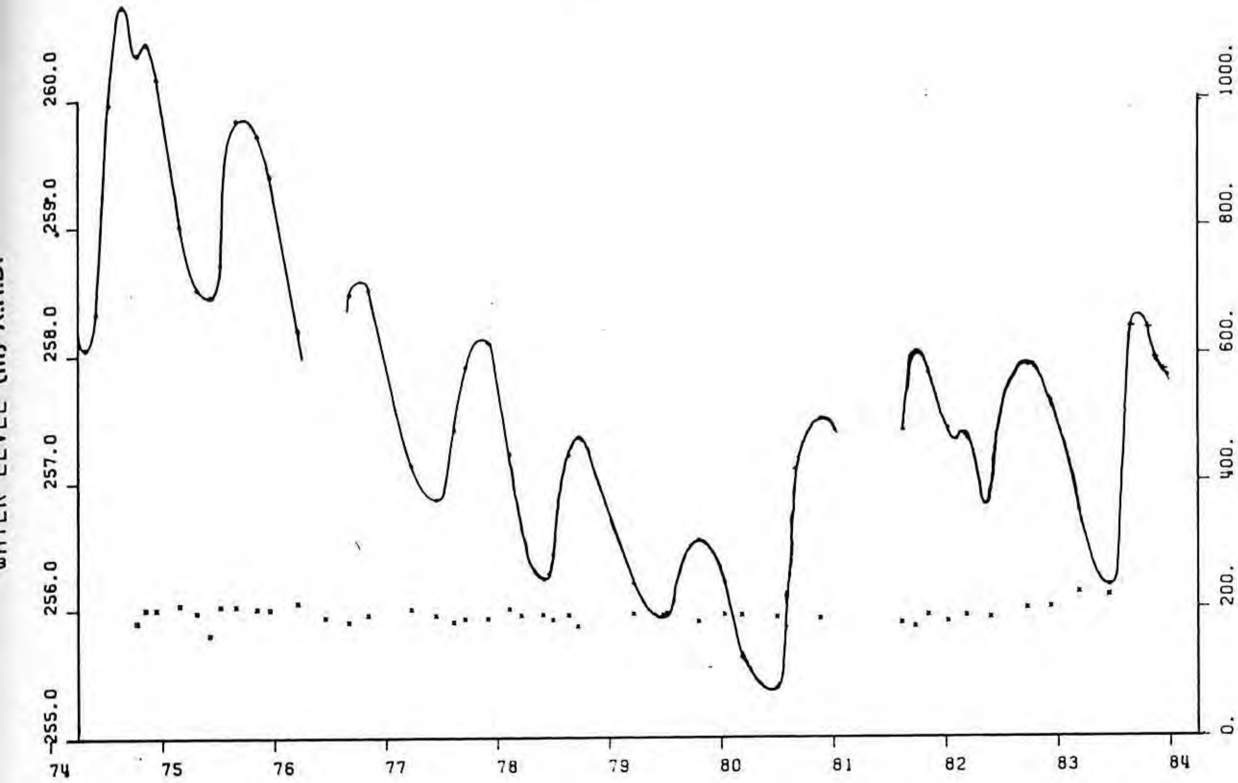
The areal extent of the deeper groundwater in the lower half is not known because there is only one bore (1151), located close to the gauging station. The water levels in this bore have always been below the invert of the stream downslope so that it is possible that the deeper groundwater at this site does not contribute to streamflow.

6.3 Bore Water Levels

Water levels in bores 1251 and 1351, (Figure 6.1, 6.2) show seasonal and longer-term trends. Levels have decreased from maxima in 1974, following the above average rains of 1973 and 1974, to minima in about 1980 as a result of the sequence of generally lower rainfalls in the late 1970's. The average long-term water level decline for bores 1251 to 1551 (measured at the yearly minima) were 2.4mm day^{-1} and 1.2mm day^{-1} for the 1974-1977 and 1978-1980 periods respectively. Water levels have increased since 1980 with a return to average or above average rainfalls.

Pronounced seasonal water level fluctuations occur in bores 1251 and 1351 with an increase of 2.7m in bore 1251 in 1974. Levels rise shortly after the start of rainfall in May-June and peak by October-November during which time about 80% of rain will have occurred (on average). Water levels decline over the following, relatively dry, months with minima by April-June. The typical 1.0-1.5m water level decrease in bores 1251 and 1351 represents 5.5 and 8.2mm day^{-1}.

The water level in bore 1251 has varied from almost 3m above ground level (in 1974) to 2.4m below ground level in 1980 (Figure 6.1). The seasonal minimum water level decreased below ground in 1976 and the seasonal peaks by 1978. Peak levels did not increase to, or above ground level, until 1981. As bore 1251 is located in the area of groundwater convergence, the variation of water levels above and below ground over the ten years may mean variations in the contribution to streamflow.



6.1
NDWATER LEVEL & CHLORIDE : 1251

6.2
NDWATER LEVEL AND CHLORIDE : 1351

6.4 Soil and Groundwater Chloride

Soil samples were obtained from cores to bedrock (where possible) and the soil solute content determined in the laboratory (Johnston and Williamson, 1981). The average soil chloride concentration for the five bores (numbers 1151, 1251, 1351, 1451 and 1551) was 1390 mg l^{-1}. However the cores from 1151 and particularly 1251 are probably unrepresentative of general profiles because only about 8m lengths of cores were obtained. The coring at 1151 ended at 8m due to rock and drilling was abandoned at 8m at 1251 due to problems with the stability of the hole. The average of the remaining bores (1351, 1451 and 1551) is 510 mg l^{-1} chloride.

An example of the distribution of soil chloride through the profile is shown in Figure 6.3 for bore 1351. A pronounced salt 'bulge' occurs between 3m and 8m below ground level with a peak of 1700 mg l^{-1} at about 5m depth. In the top few metres, soil chloride concentrations are generally less than 200 mg l^{-1} and mostly less than 50 mg l^{-1}. The gravelly, earthy surface soils are therefore low in salt accumulation relative to the underlying clays.

Below 10m depth, soil chloride concentrations are less than 200 mg l^{-1} except below 20m where concentrations increase to 500 mg l^{-1} at the base of the hole.

The quality of the water in the bores was determined from samples extracted and analysed by C.S.I.R.O. at approximately monthly intervals. The chloride concentrations of the bore water ranged from 200 mg^{-1} (bores 1251 and 1551) to 500 mg l^{-1} (bores 1351 and 1451).

The slotting intervals of these four bores (excluding 1151) are generally below the level of maximum soil chloride concentration as shown in Figure 6.3 for bore 1351. The bore water chloride concentrations reflect the soil chloride concentrations over the slotting intervals.

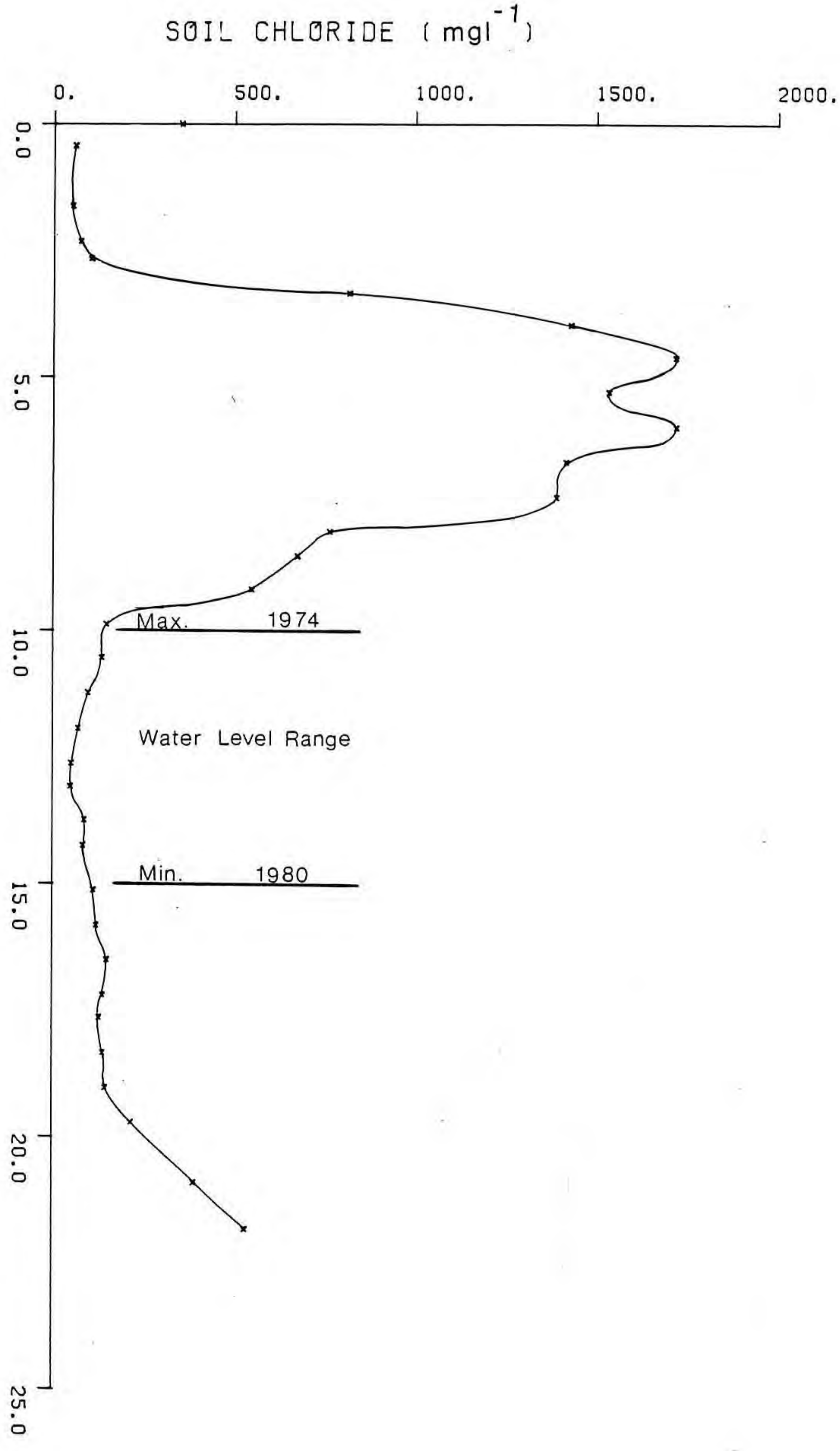


Figure 6.3
DISTRIBUTION OF SOIL CHLORIDE : CORE 1351

Over the ten year period bore water levels have varied by more than 5m and water levels have been above the top level of the slotting in all bores except 1451. Maximum water levels have been into or above the level of maximum soil chloride concentration and it is therefore possible that the bore water chlorides are less than those occurring in the groundwater system generally.

The small number of bores and the length of slotting of the casing of the existing bores provide only a small sample of the range and variation in time of groundwater chloride concentration. However there is almost certainly a spectrum of concentrations and these probably range from 200 mg l^{-1} to 500 mg l^{-1}.

6.5 Groundwater Flow System

Within the limited information on the extent of the deeper groundwater over Salmon catchment, a possible groundwater flow system has been developed for the upper catchment.

The groundwater flow system, shown in Figure 6.4, is based on a longitudinal section from near the top of the catchment at bore 1551 downslope past bore 1251 to the streamline below the stream sampling site at S9. There is frequent outcropping bedrock in and around the stream channel starting about 600m down the transect, below a swamp area. A cross-section through bores 1351 - 1251 - 1451 (Figure 6.5) indicates small hydraulic gradients along this line. Therefore the direction of groundwater flow is in general sympathy with the surface topography. That is, from the top of the surface divide, downslope into the surface topography convergent area below bore 1251.

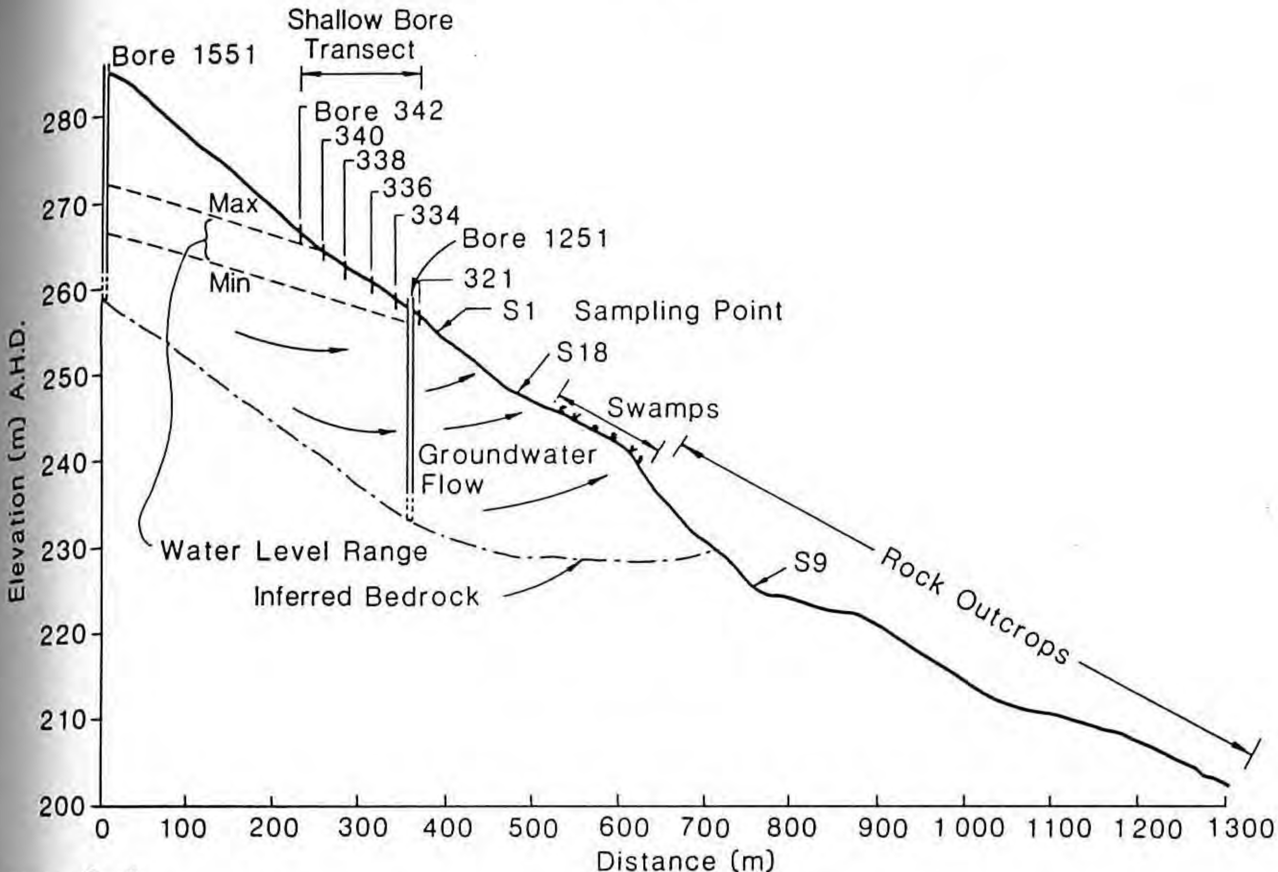


Figure **6.4**
LONGITUDINAL SECTION

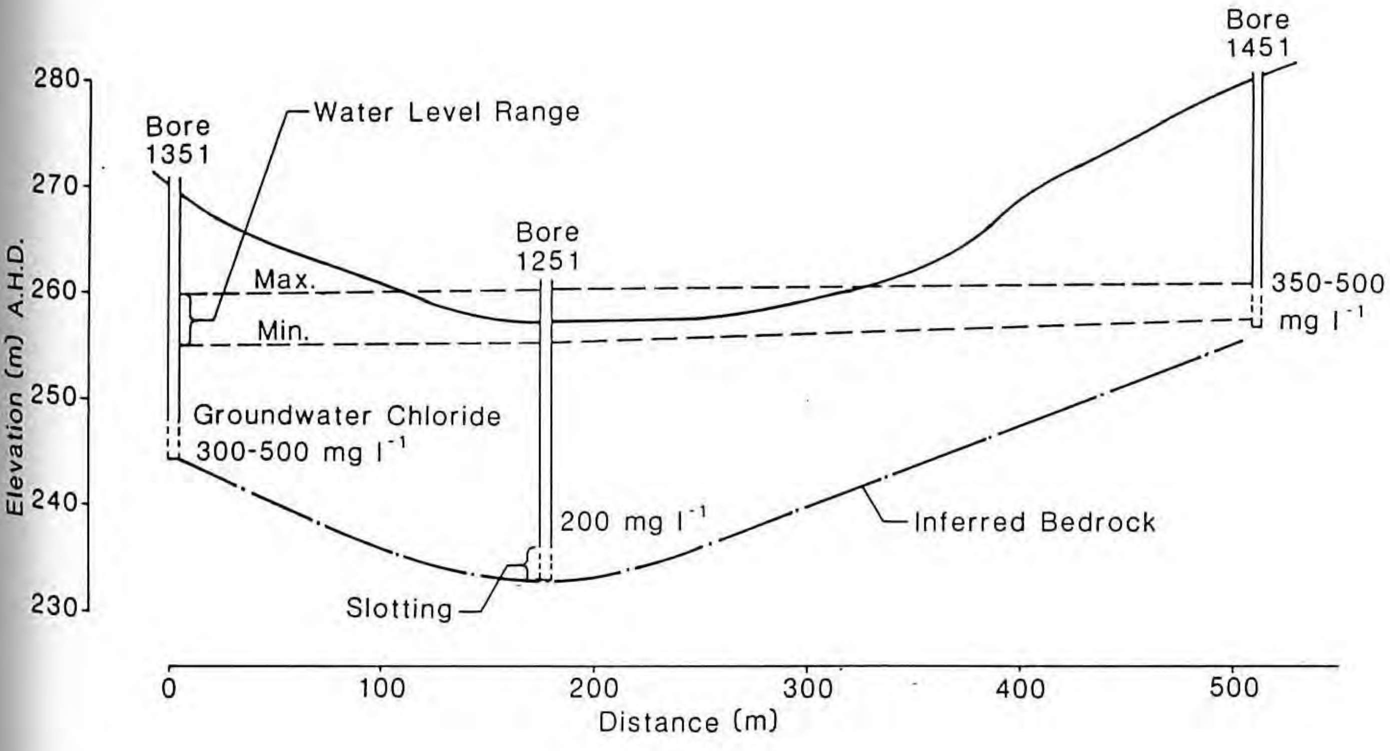


Figure **6.5**
WATER LEVELS : TRANSECT 1351-1251-1451

As noted, water levels in bore 1251 have varied from almost 3m above ground level in 1974 to 2m below ground level in 1980. The vertical hydraulic gradient at 1251 has been about 10% whereas the horizontal gradient been 1551 and 1251 has been about 3%. The relatively high vertical gradient at 1251 (and the above-ground potentiometric level) possibly indicate a semi-unconfined aquifer (Martin, 1982). That is, the aquifer is essentially unconfined but because the hydraulic conductivity is so small (Peck et al., 1980) vertical gradients can be developed in response to surface topography and a strong sink effect such as may occur through transpiring vegetation in the top few metres of the profile.

The presence of the swamp below bore 1251, the vertical hydraulic gradients in 1251, the bedrock outcrops below the swamp and variations of stream water quality through this area (see Figure 5.2) indicate a possible contribution of deeper groundwater to streamflow. The conceptual groundwater flow system shown in Figure 6.4 was developed on this basis.

6.6 Estimate of Groundwater Flux

Sharma et al. (1982) obtained an estimate of lateral groundwater flow by consideration of the one-dimensional horizontal water flow equation for an unconfined aquifer:-

$$S \frac{\partial h}{\partial t} = K \frac{\partial}{\partial x} \left(H \frac{\partial h}{\partial x} \right) + R \qquad 6.1$$

where S is specific yield (0.04), K is hydraulic conductivity (0.01 m day^{-1}), H is aquifer thickness (30m), R is a sink or source term (O), h is potentiometric head, t is time and x is distance. They calculated that for these parameters lateral flow over six months could cause a groundwater decline equivalent to about 14mm of water. The large observed **decreases (Figure 6.1) are the result of transpiring** vegetation. Over a 12 month period therefore lateral groundwater flux was estimated to be equivalent to just less than 30mm of water. For a groundwater chloride concentration of 200 mg l^{-1} (for example) this implies a flux of chloride equivalent to about 60 kg ha^{-1} for the whole catchment.

Johnston (1983) estimated that the annual recharge at bore 1351 was "much less than 50mm yr^{-1}" and that the mean seasonal rise was equivalent to 32mm of water. The average annual recharge from seven bores in Salmon was estimated by Johnston (1983) to be 25mm yr^{-1} but was considered unreliable because of the relatively small number of bores.

6.7 Summary

There is only limited information of the deeper groundwater in Salmon catchment because of the small number of deep bores and their location. Deeper groundwater apparently extends over most of the upper half of the catchment. However there are insufficient data for the lower half of the catchment.

Bore water level data indicate a significant decrease (5m) in levels from the peak in 1974 to a minimum in 1980 after which levels increased (about 2m) again although not to the levels of 1974.

Groundwater chloride concentrations measured in the deeper bores ranged from 200 to 500 mg l^{-1}. Prominent soil salt 'bulges' were found, 5-10m into the profile, but generally above the water levels.

A deeper groundwater flow system was constructed for the upper half of the catchment based on the observed hydrogeology and interpretations of stream salinities, rock outcrops and a swamp.

An estimate of the catchment flux of deeper groundwater of about 30mm yr^{-1} was used with a groundwater concentration of 200 mg l^{-1} to produce a chloride flux of 60 kg ha^{-1} yr^{-1}.

7. SHALLOW GROUNDWATERS

7.1 Shallow Bore Transect

A transect of shallow bores (< 3m) was installed over a distance of 150m in the headwater area of Salmon catchment (see Figure 3.3) in 1983 to observe the response of the seasonal groundwaters. The transect starts just upslope of the seepage sites S1 and S2 and extends upslope towards the C.S.I.R.O. water balance site 5 (bore 1551). Six sites were located on this transect, about 30m apart, at which shallow bores were installed. The average surface gradient between the lowest site, Site 1 (256.2m AHD) and the upper site, Site 8 (266.3m AHD) is approximately 6.7% (Figure 3.3).

A generalised soil profile along the transect, obtained by interpretation of 100mm diameter cores, is shown in Figure 7.1. Some points of note include:-

(i) the depth to the 'primary' perching material, the mottled to pallid clay, increases up the transect

(ii) a mottled, transitional clay material was relatively close to the ground surface at Sites 4 and 5

(iii) a dark, sandy-humic surface layer present at the lower sites was absent at the upper sites where surface gravels occurred.

(iv) red earths do not occur as part of the profile upslope of Site 6

(v) the surficial gravels upslope of Site 6 extend into the profile as gravels, often in a very sandy matrix which made core retrieval very difficult.

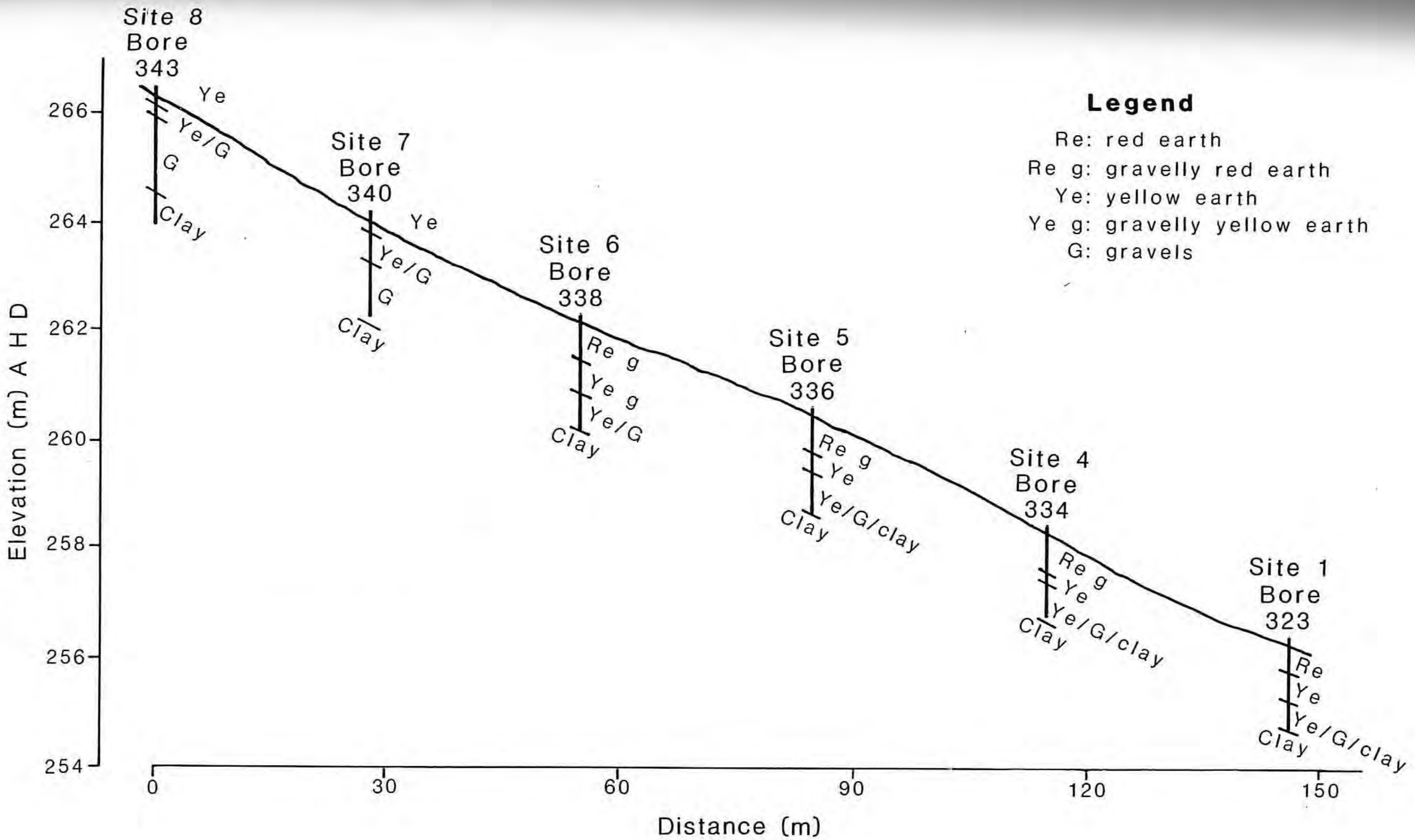


Figure 7.1
SHALLOW BORE LONGITUDINAL SECTION SOIL PROFILE

In general terms, the 150m transect rises vertically through about 10m, from sandy, red and yellow earths at the base of the slope to predominantly gravels at the top. Within this broad description there were several intermediate soil profile transitions such as red to yellow earth, yellow earth to clay which may function as perching layers.

7.2 Water Levels on Transect

In Figure 7.2 a - d water levels on the transect have been drawn representing four distinct seasonal periods in 1983:-

(a) before and following peak levels in early July
(b) the mid-winter period
(c) peak levels in early September
(d) final recession levels in late spring.

The position of the water level on 20/7/83 (Figure 7.2a) and through until 23/8/83 (Figure 7.2b) are very similar. These occur, in the main, at red to yellow earth and yellow earth to mottled-pallid clay transitions in the profile and may therefore be the effective perching layers on the hillslope. During the final stages of recession (Figure 7.2d), the bore water levels cluster around the mottled-pallid clay transition deeper in the profile.

Areas of surface soil saturation were observed downslope of Site 6 (bore 338) for a large part of the winter. The high levels (or ridging) around Site 5 (bore 336) is a notable feature which occurs through until the final recession period (Figure 7.2d). As noted earlier, the profiles at Sites 4 and 5 had less permeable mottled zones much closer to the ground surface. Downslope the red earth profile deepens, whereas upslope the gravel profile predominates.

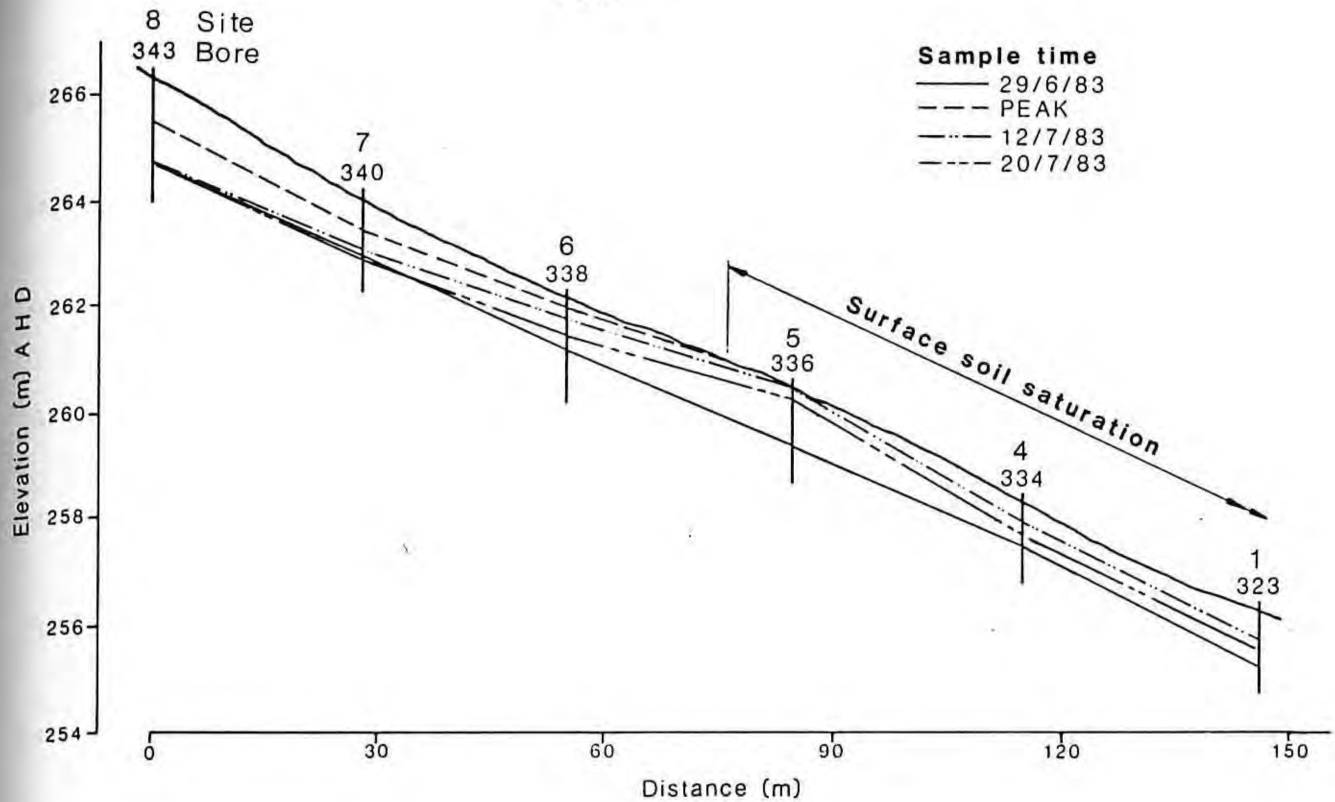


Figure 7.2a
SHALLOW BORE TRANSECT WATER LEVELS : 29/6/83 - 20/7/83

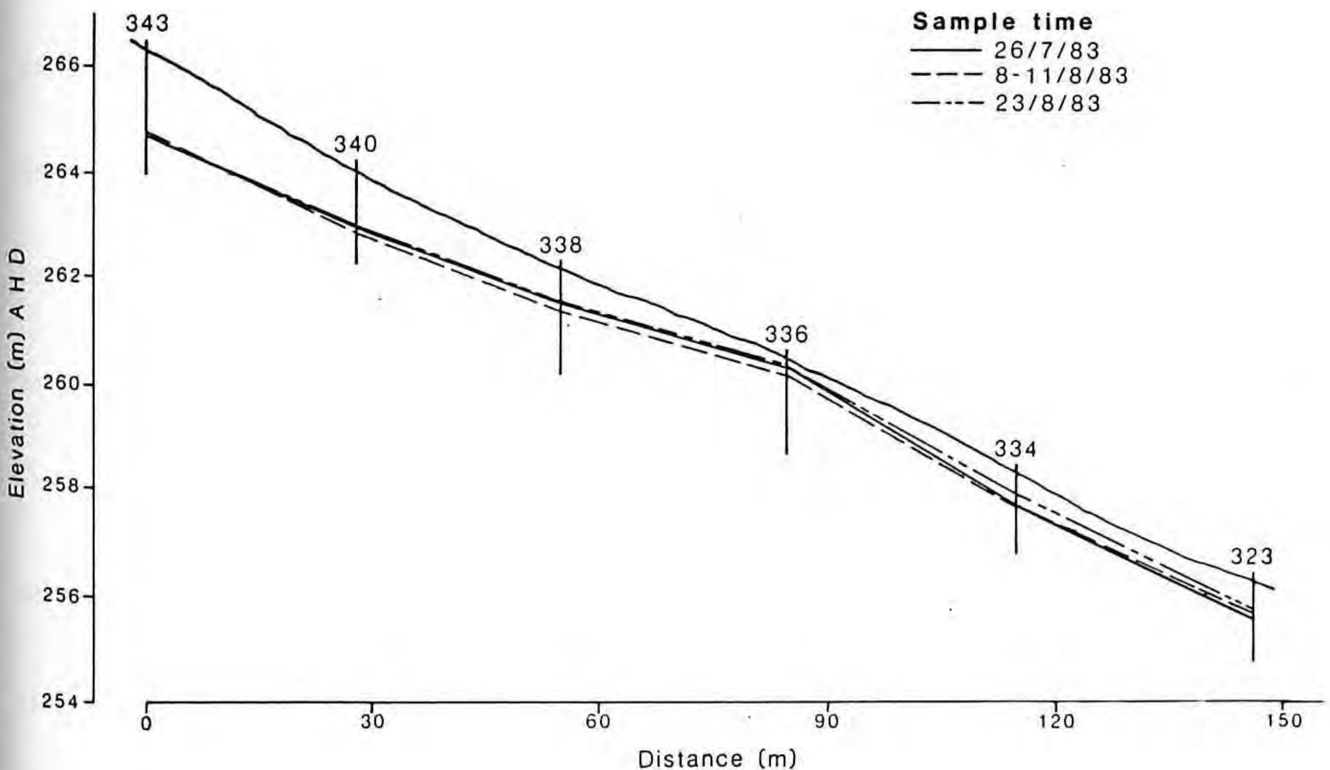


Figure 7.2b
SHALLOW BORE TRANSECT WATER LEVELS : 26/7/83 - 23/8/83

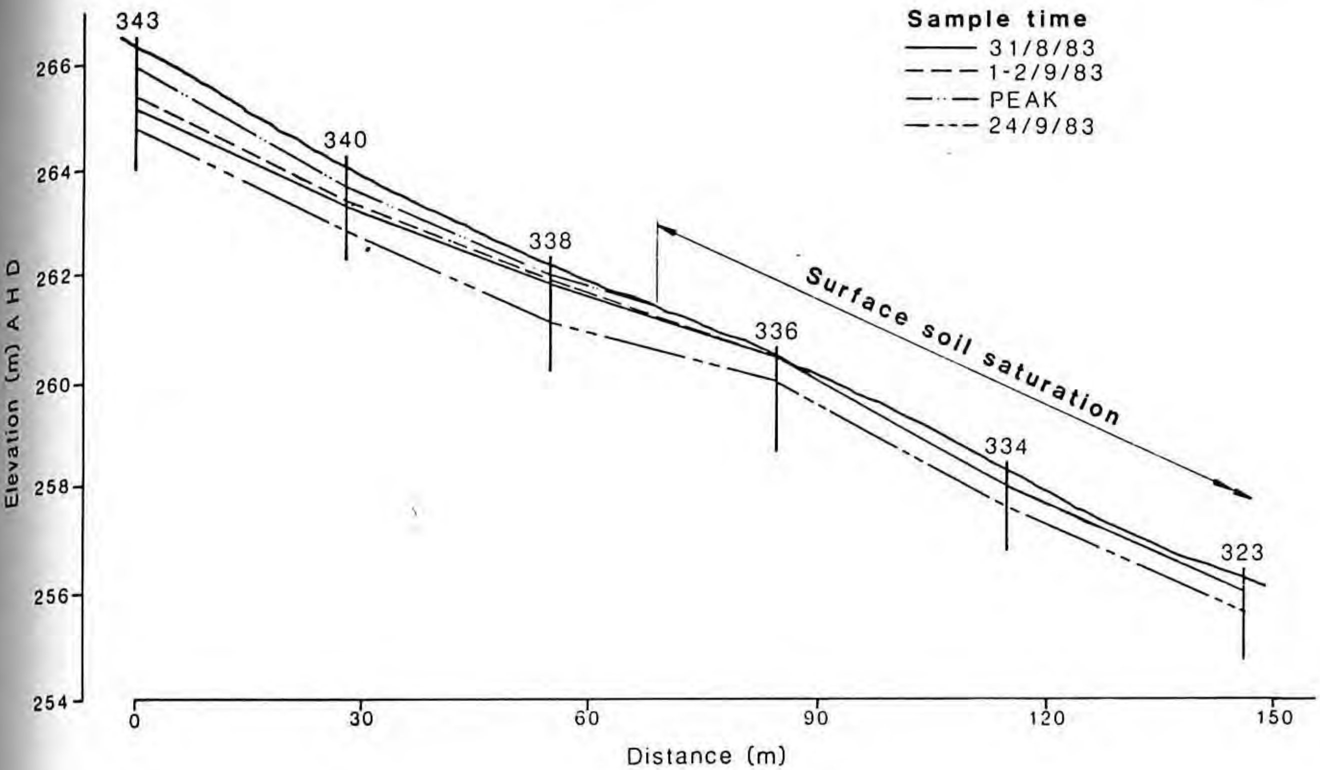


Figure 7.2c
SHALLOW BORE TRANSECT WATER LEVELS :31/8/83 - 24/9/83

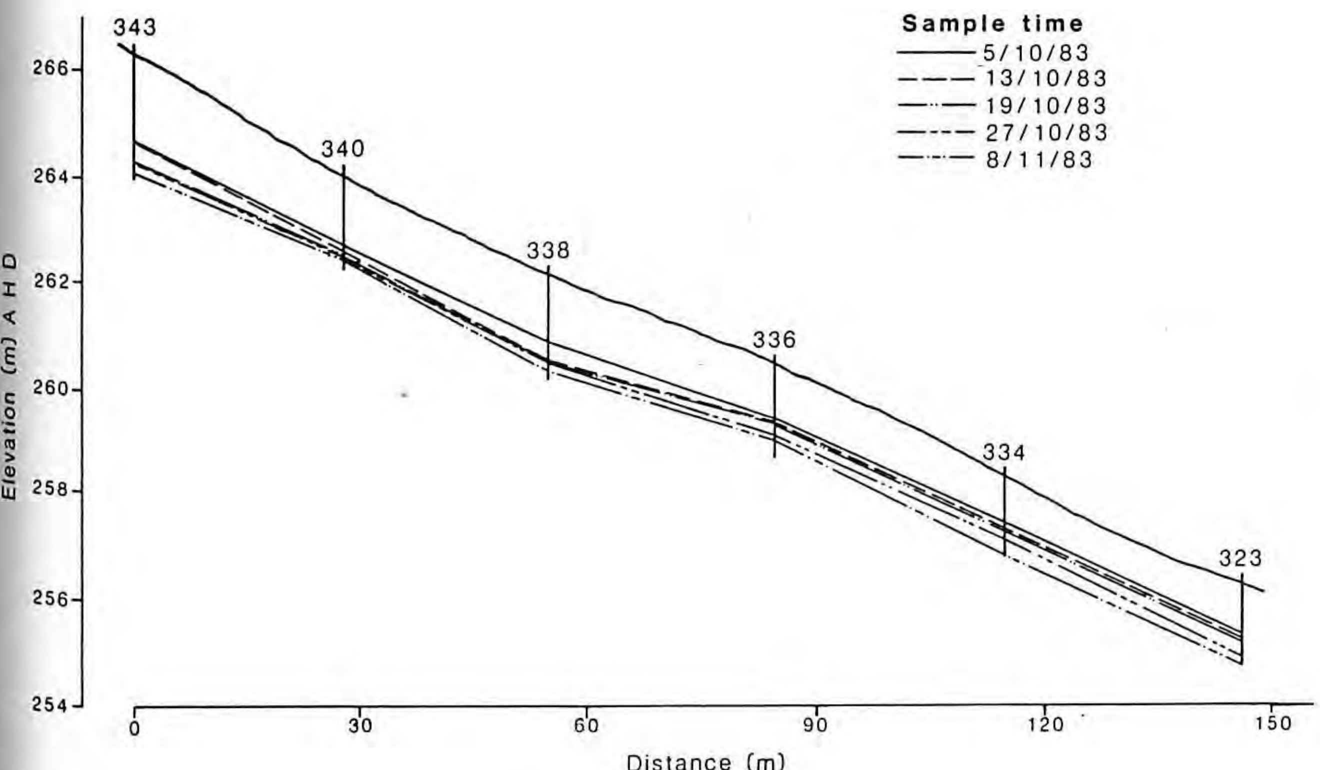


Figure 7.2d
SHALLOW BORE TRANSECT WATER LEVELS : 5/10/83 - 8/11/83

Lateral flow of water from upslope of Site 5 may have exceeded the capacity of the 'throttle' at Site 5, particularly during early July and September. On these occasions saturated depths of more than 1m occurred upslope of Site 5 and the reduction in the depth of flow at Site 5 may thus have resulted in the development of areas of surface soil saturation and seepage.

7.3 Saturated Areas

Inspections of Salmon catchment were made on five occasions in 1983 between July 13 and October 5. In general terms the spatial distributions of sources and salinity of streamflow were very similar to those of 1980 and 1981.

Approximately 740mm of rain fell in the period from the middle of June to August 31 when a detailed inspection of the catchment was made. The discharge of 0.035 m^3 s^{-1} at the main gauging station (SGS) was the highest at any visit between 1980 and 1983.

The area of surface soil saturation on August 31 1983 was the most extensive observed during the period of study (Figure 7.3). Saturation occurred from around S1, westwards through the C.S.I.R.O. water balance site (bore 1251), then south through near the shallow bore transect at Site 6 and then eastwards. From here the saturated area extended northwards, well to the east of S3 and S5. At the junction of the two headwater channels (S18), saturation was confined to the stream channel.

Further downstream, saturation occurred around the swamp, upslope of the saline seep S11, across the stream at a secondary swamp (S6) and in a band along the east bank from S7 to S8. Downstream of S9 the areas of saturation were confined to a 10-30m wide zone along the channel.

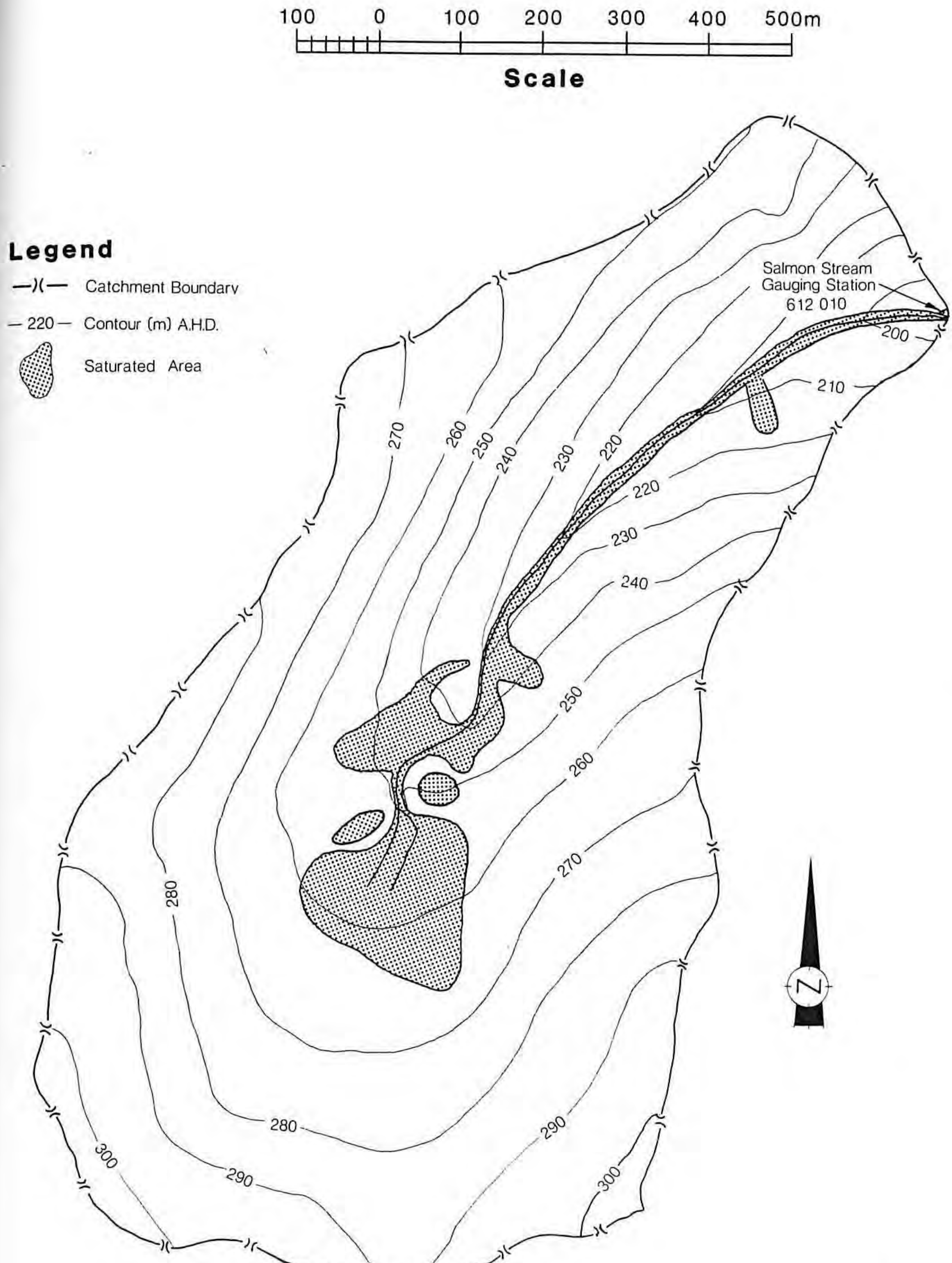


Figure 7.3
AREA OF SATURATED SOIL : 31/8/83

The estimated area of saturation upslope of S18 was 2ha. Downstream of S18 the extent and therefore area of saturation was difficult to determine but was probably of order 3ha. Of this total of 5ha the most variable area appears to be that in the convergent headwater area above S18. Downstream, the areas are confined to the swamp, rock outcrops and the zone immediately adjacent to and including the stream channel. These are what have been termed 'partial source areas' whereas the headwater area is more like the 'variable source area' concept as discussed in Chapter 2.

7.4 Shallow Groundwater Quality

Maximum and minimum chloride concentrations and temperature of water in the bores during 1983 are listed in Table 7.1. Water in bores with a bentonite plug generally had high chloride concentrations. In late August 1983 the bores were bailed-out and a salinity reading made shortly after the water level had recovered. The high chloride values were not repeated again thus suggesting some initial contamination, possibly from the bentonite.

Excluding the suspect values the general pattern of water quality is for lower chlorides at the top of the transect, higher values at mid-transect and lower chlorides again at the base of the transect. Values range between 25 and 40 mg l^{-1} with an occasional 50 mg l^{-1}. Concentrations at Site 2 are representative of those found for water issuing from the 'pipe' at seepage area S1. Similarly, water quality at Site 3 is very similar to that seeping out at S2, which is slightly more saline that at Site 2. The average concentration at Site 1 was about 33 mg l^{-1}.

The lower chlorides at the top of the transect is possibly a result of more effective leaching in the gravels. In contrast the higher values at mid transect provide additional evidence for less permeable soils.

Water temperatures were generally between 13-14^{o}C with a range over the season of 12-15^{o}C. A small warming trend was evident towards October.

TABLE 7.1 CHLORIDE AND TEMPERATURE IN SHALLOW BORES

SITE	BORE	CHLORIDE (mg l^{-1}) Min.	Max.	TEMPERATURE (^{o}C) Min.	Max.
1	321	-	-	-	-
	322	25	50	-	-
	323	26	36	13.1	14.4
	324	30	34	13.2	13.6
2	325	27	31	13.3	14.6
	326	27	40	12.9	13.8
	327	33	49	12.8	14.5 B
	328	23	28	12.6	14.3 B
	329	24	32	12.1	14.7
3	330	40	56	13.0	13.6
	331	35	50	13.0	14.8
	332	70	108*	13.1	13.7 B
	333	35	47	13.3	13.5
4	334	134*	152*	13.5	15.0 B
	335	32	41	12.8	13.5
5	336	58	169*	13.4	14.6 B
	337	33	37	13.8	14.5
6	338	57	147*	13.8	15.0 B
	339	26	46	13.6	14.7
7	340	21	21	13.8	15.6 B
	341	-	-	-	-
8	342	-	-	-	-
	343	-	-	14.9	14.9
	344	-	-	-	-
9	345	20	20	13.7	14.7
10	346	0	0	13.8	13.8

Notes (i) B denotes bore with bentonite seal
(ii) * suspect data

7.5 Continuous Shallow Bore Water Levels

The continuous water levels for bore 321 (Site 1) and bore 342 (Site 8) are shown in Figure 7.4 and 7.5 for June 1983 to November 1983. The record for bore 321 for the period 2-13 July is not available because of equipment malfunction. However the peak level in this period is approximately correct.

These plots of continuous water levels indicate significant differences in responses between and within bores. The steep hydrograph rises at the start of the record is indicative of the relatively impermeable material in which each bore is founded.

Water levels in bore 321 between 13 July and the middle of August rose and fell 0.1m about a 'base-level' approximately 0.7m below ground level. This level is roughly coincident with that of increased clay in the yellow earth. Above this level, in the red earths, hydrograph rise and falls were steeper, indicating higher lateral (and/or vertical) water flux in contrast to the underlying materials.

Peak water levels in early July and late August - early September were within 50mm of ground level. Saturation of the surface soils around this bore was observed on 31 August and from the water levels it is estimated that saturation probably occurred for about a week in early July and at least two weeks in late August - early September.

The hydrograph recession in bore 321 commences at 0.04m day^{-1} from the peak in early September in the yellow earths and decreases to 0.02 m day^{-1} when the water level falls to the level of the mottled zone by later September.

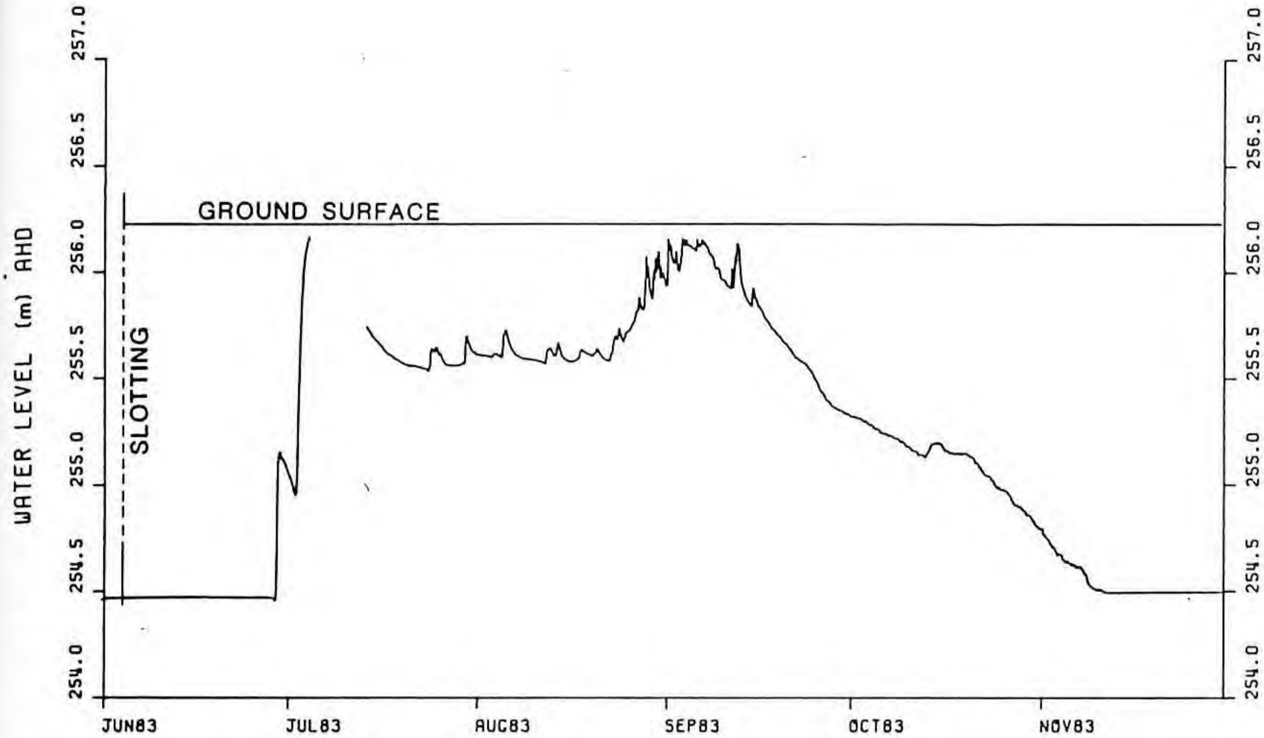


Figure 7.4
WATER LEVEL : BORE 321

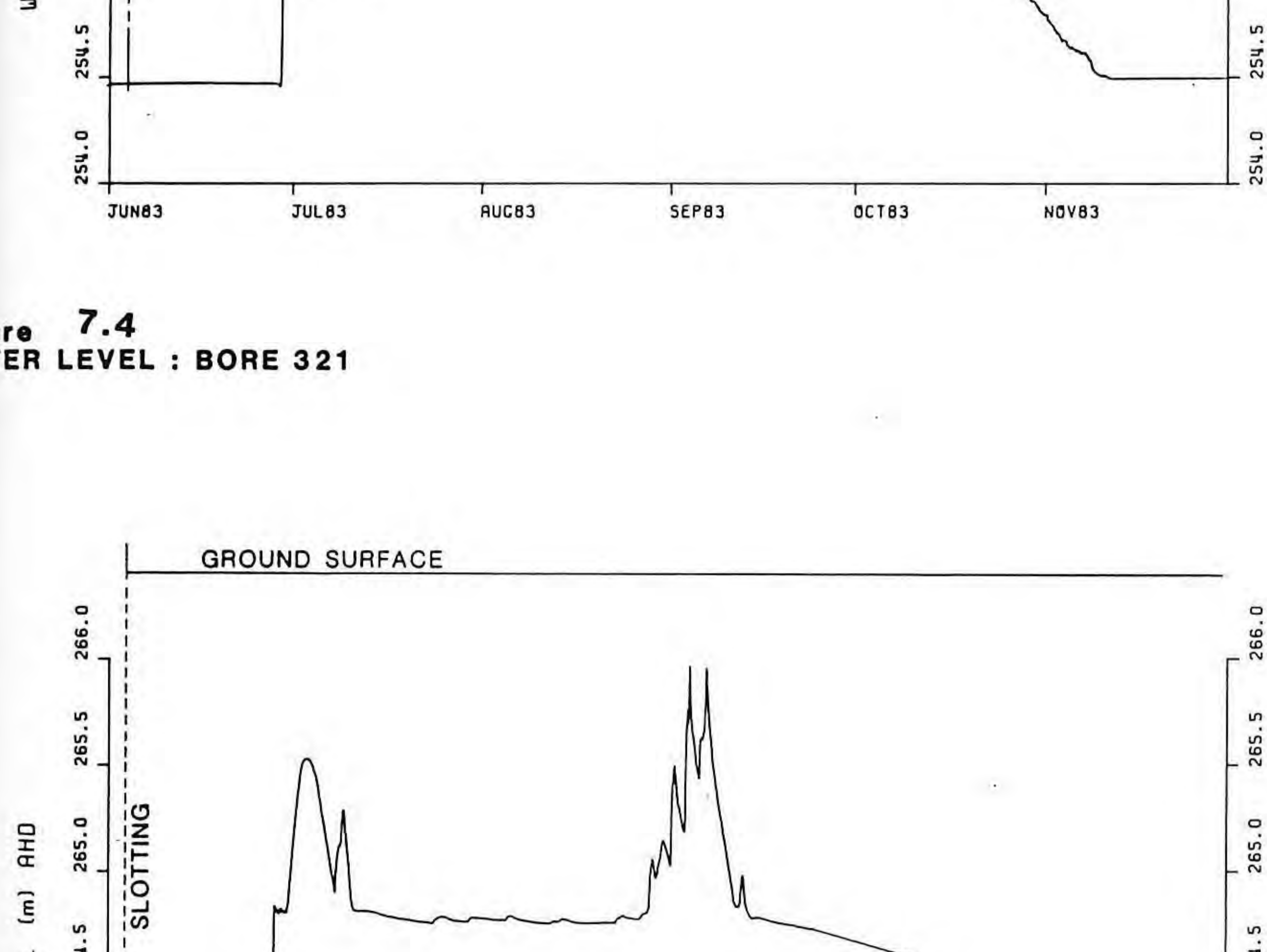


Figure 7.5
WATER LEVEL : BORE 342

The hydrograph response in bore 342 (Figure 7.5) is markedly different from that of bore 321. When drilling bore 342 it was not possible to obtain cores due to the gravels and the position of the effective perching layer was estimated to be at 2m. However water levels were virtually constant at about 1.6m from mid-July to mid-August which is more likely therefore to be the perching layer at the transition to clay.

Above this level, in the gravels, the water level response is rapid (relative to bore 321) with sharp rises and falls. Significant periods of perched water occurred in late June - early July and late August - early September when peak levels were within 0.4m and 0.85m of ground levels respectively which represent saturated depths for the perched aquifer of about 0.7m and 1.2m. Water level recession rates in the gravels in bore 342 were between 0.1 and 0.2m day^{-1}, much higher than those in bore 321.

7.6 Seasonal Streamflow and Shallow Groundwater

7.6.1 Discharge and Bore Water Level

The weir at S18 monitors streamflow from the top 40% (33ha) of Salmon catchment from which there is no evidence of a contribution to streamflow from the deeper groundwaters. Stream water and salt are generated by seepage from seasonal, perched groundwaters and by direct rainfall-runoff from the wetted channel and periodically from an expanded source area. The water level in the shallow bore 321 provides a record of the occurrence of the expanding source area and also of the relative state of the shallow groundwaters in relation to streamflow at the weir at S18.

7.6.2 Seasonal Response

Daily total flow (1000's m^3) at S18 and water level in bore 321 are shown in Figure 7.6 for June to November 1984. The S18 flow, as a proportion (%) of that at the main gauging station at SGS, is shown in Figure 7.7.

Flow at S18 was generally less than 20% of that at SGS until about the middle of July when it increased to more than 40% as the shallow groundwaters developed in the headwater area above S18. Flow had occurred at S18 before this time but was about half of the proportion at SGS once the groundwaters developed. The flow proportion varied from 25% to 40% of SGS through towards the end of August. The proportion decreased during periods of rainfall as significantly more runoff is generated from the steeper slopes and shallow soils downstream of S18. During recessions the flow proportion increases towards 40% which is also the ratio of the catchment areas (S18/SGS). The water level in 321 fluctuated about a 'base-level' of 255.5m AHD with rises and falls of about 0.1m.

Heavy rainfall during the first half of September caused a significant increase in flow at S18 and the water level in bore 321. The response of flow and water level for this period are shown in more detail in Figure 7.8. There is a close relationship between the response of the water level and of streamflow. The steady recession of flow following the peak discharge of each event is closely matched by the bore water level recession. The water level peaked at 0.5m above the 'base-level' of 255.5m on September 4.

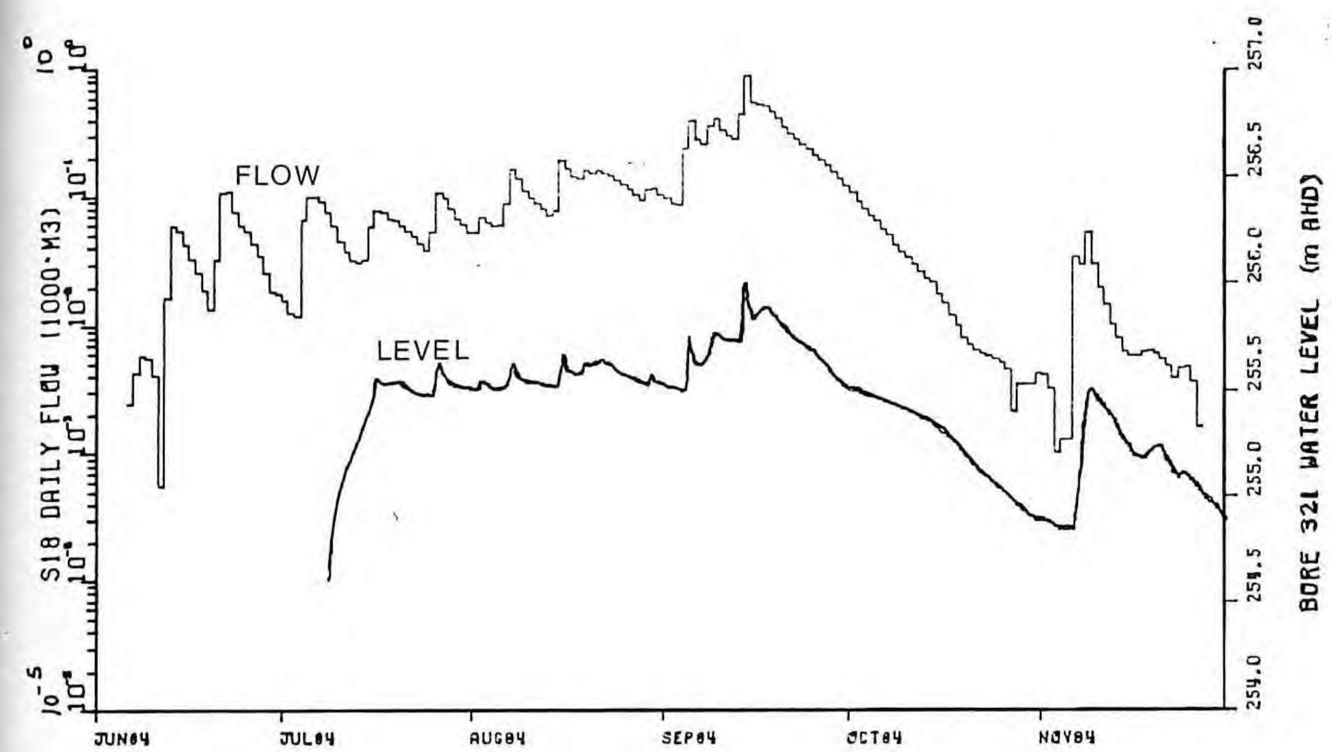


Figure 7.6 DAILY DISCHARGE AT S18 AND BORE 321 LEVEL

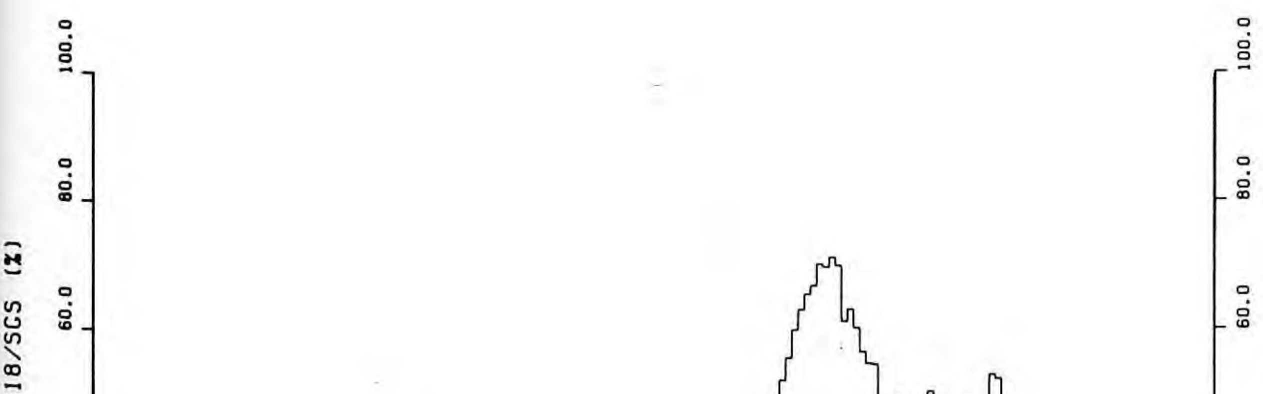


Figure 7.7 DAILY PROPORTION S18/SGS

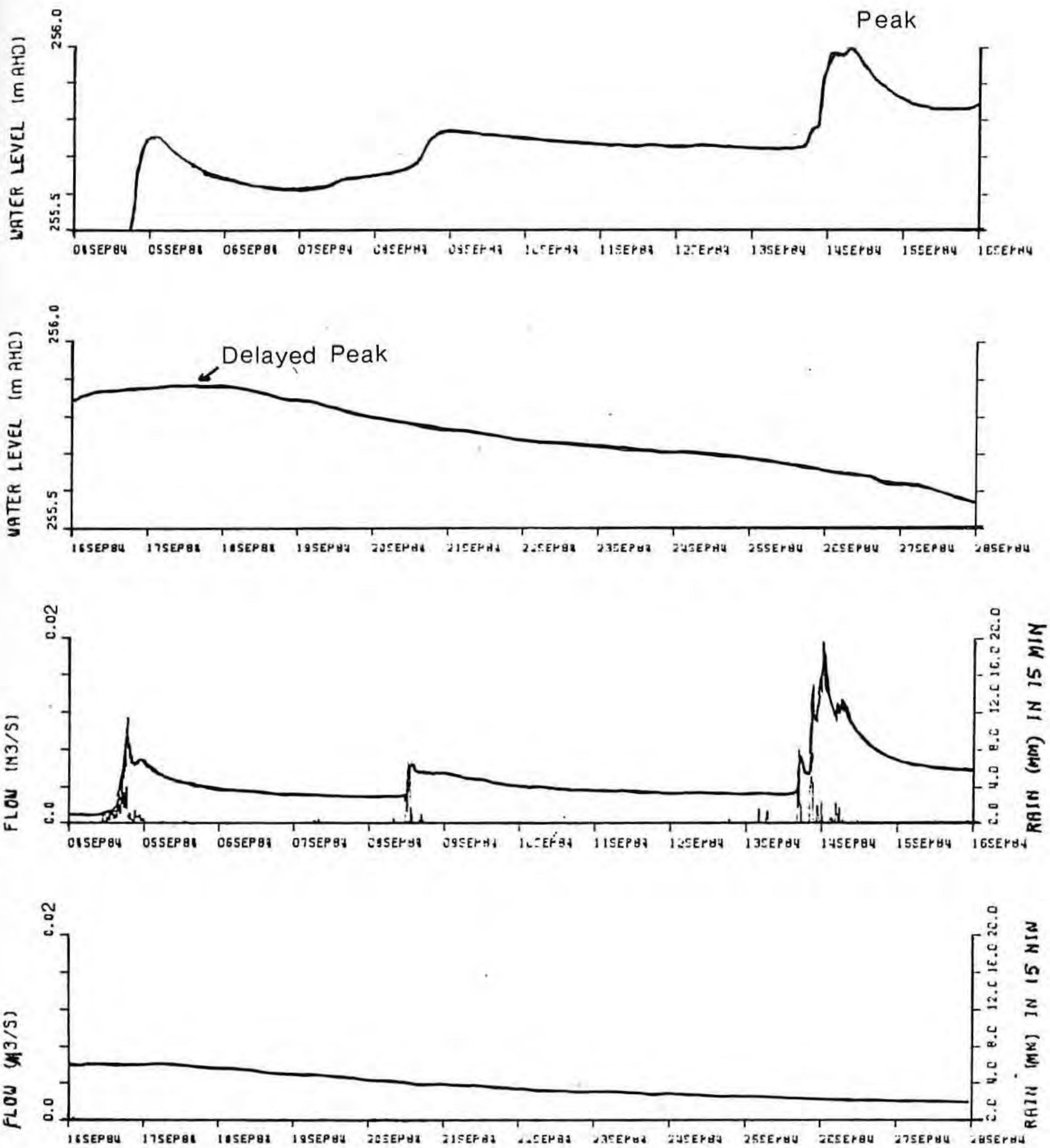


Figure 7.8 S18 DISCHARGE AND BORE 321 LEVEL

A notable feature of both the flow and water level recession is the increase in water level between September 15 and 18 and the relatively constant flow at S18 over this period. Little rain occurred during this period and therefore this may be a delayed response of shallow groundwater moving downslope into the headwater area. The delay between the major rainfall and second bore water level rise was about three days. A smaller, second increase in bore water level occurred during September 7, also about three days after the rainfall. Both of these examples of a delayed shallow groundwater response occurred after small increases in water level in the upslope bore 342 (not shown). These increases in bore 342 were the only such increases of water level above the perching layer in 1984 and is perhaps evidence for an increase in shallow groundwaters from upslope.

Flow and water level decreased from mid-September to late October. The flow proportion at S18 had increased to a seasonal peak of 70% by late September after which it decreased to about 45% and then to 0% by the end of October. The decrease from 70% to 45% coincided with the water level decline below 255.5m after which the level declined less rapidly. This water level corresponds with a change in soil to a yellow earth/clay. Shallow groundwaters were therefore likely to contribute less to streamflow from the less permeable material.

The seasonal peak water level in bore 321 was 0.25m below ground level and therefore areas of surface soil saturation were probably far less extensive in 1984 than in 1983 when water levels were closer to ground level for several weeks.

7.7 Stream Discharge - Bore Water Level Relationships

A plot of the relationship between discharge at S18 and water level in bore 321 is shown in Figure 7.9 for the period September 15 to October 15 1984. A relationship was 'fitted-by-eye' for water levels above 255.5m:-

$$q = 1.38 \times 10^{-3} \exp (3.66.h) \qquad 7.1$$

where q is discharge ($m^3 s^{-1}$) and h is water level above 255.5m.

From 255.5m to 255.4m a relationship:-

$$q = 4.31 \times 10^{-4} \exp (12.47h) \qquad 7.2$$

There is a significant change in slope of the discharge recession above and below 255.5m.

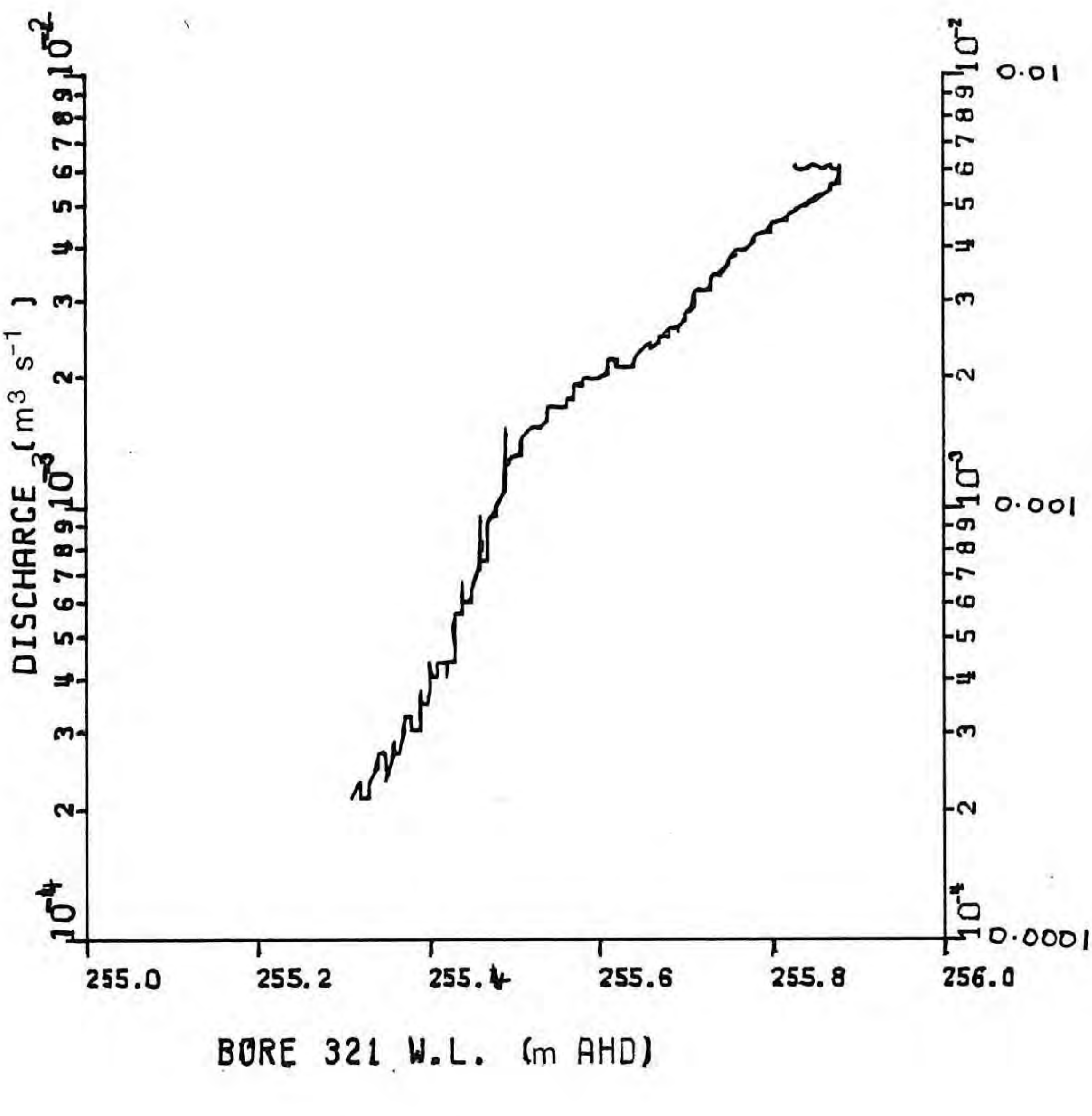
The general form of the relationship between discharge and water level is exponential. Bear (1979) developed equations for discharge for a spring of the general form:-

$$q = q_0 \exp (-B (t-to)) \qquad 7.3$$

where $B = A . T/S$ 7.4

where A is a constant of aquifer geometry (dimensions L^{-2}) and T and S are average transmissivity and aquifer storativity respectively. As water level is approximately linear with time during recession (see Figure 7.6) the discharge is exponentially proportional to the water level (h):-

$$q \propto e^{ah} \qquad 7.5$$



S18 DISCHARGE AND BORE 321 RECESSION RELATIONSHIP

Figure 7.9

7.8 Storm-Period Response

Stream discharge at S18 and water level responses in bore 321 are shown in Figures 7.10 for two storm events in September 1984 when bore water levels were higher than 255.5m. During the September 4 event water levels rose 0.25m in 6 hours and peaked about 6 hours after the peak discharge. In contrast, water levels during the September 13-14 event responded quickly to rainfall and peaked at or just after peak discharge.

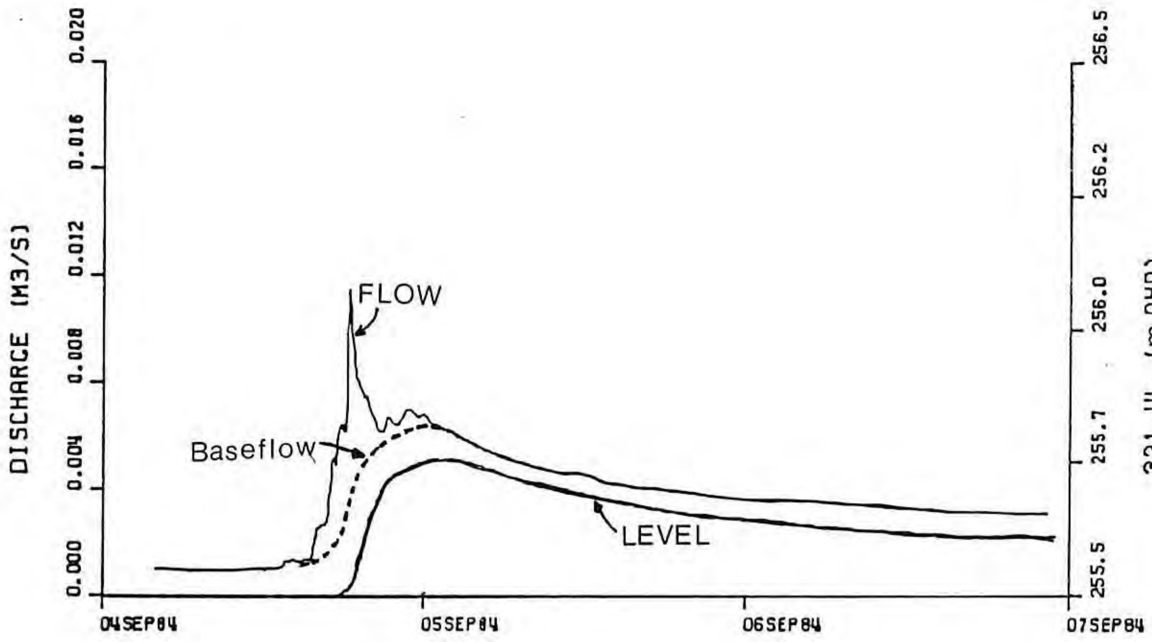
The bore water level response was used to construct a 'baseflow' hydrograph for the two events and then to estimate the proportions of direct rainfall-runoff (Q_r) and shallow groundwater contribution (Q_u) to the total stream discharge (Q_T) in response to the rainfall (R). The baseflow was defined (Figure 7.10) from the start of the hydrograph rise to just after the peak bore water level.

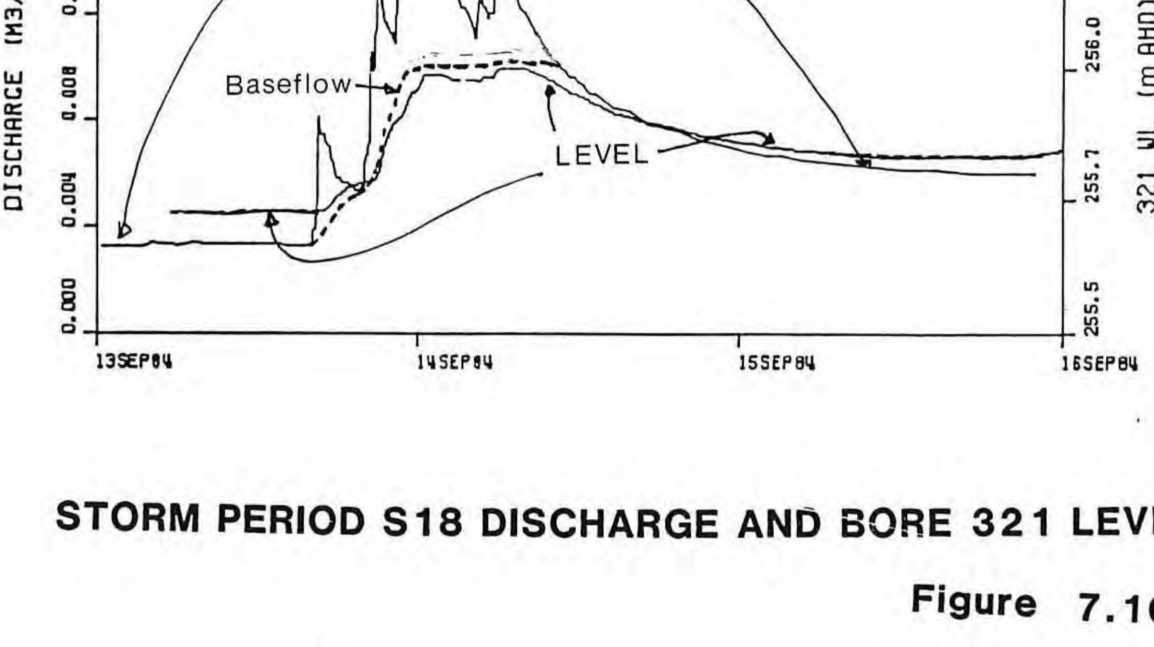
A pre-storm baseflow was subtracted from the total (Q_T) hydrograph to produce the net discharge (Q_T' and Q_u') in response to the total rainfall in each event. The results are listed in Table 7.2 and show that direct runoff was 39% and 27% of net total discharge for the two events or 0.57% and 0.93% of total rainfall. Therefore the shallow groundwater (Q_u') accounted for most of the streamflow with 61% and 73% of the discharge at S18.

The minimum size of the source area (A_c) for the direct rainfall-runoff (Q_r) can be estimated by assuming 100% runoff of rainfall (R):-

$$A_c = (Q_r/R) \,.\, A_o \qquad 7.6$$

where A_o is the total area at S18 (33ha).





STORM PERIOD S18 DISCHARGE AND BORE 321 LEVEL

Figure 7.10

This produces areas of $1881m^2$ and $3069m^2$ for the two events. The two small stream channels to S18 are each about 100m in length and therefore the average width of a source area based on the stream channels is:-

$$W = \frac{Ac}{2 \times 100} \qquad 7.7$$

which produces 9.4m and 15.4m. These are a little larger than observed and may be a result of an underestimate of the length of the wetted channel area which has been observed to 'grow' upslope past the seepage sites at S1 and S2 (Figure 7.3).

TABLE 7.2 STORM RUNOFF SOURCE PROPORTIONS

	R (mm)	Q_T (m^3)	Q_T' (m^3)	Q_u (m^3)	Q_u' (m^3)	Q_r (m^3)	Q_r/Q_T' (%)
4/9/84	50	284	239	191	146	93	39
Q/R(%)	-	-	1.45	-	0.88	0.57	-
13-14th Sept. 84	62	1002	705	812	515	190	27
Q/R(%)	-	-	3.45	-	2.52	0.93	-

Notes
(i) R : rainfall
(ii) Q : flow volume
(iii) T : total hydrograph; u : shallow groundwater; r : direct runoff
(iv) ' : net flow (less pre-storm discharge)

7.9 Summary

A 150m long transect of shallow bores at 30m intervals was installed to study the development of shallow groundwaters and relationships with saturated (source area) soil conditions and streamflow. The soil profile varied from red earth above mottled clays at the bottom of the transect to yellow earths and gravels over more pallid clays at the top of the transect.

Saturated soil conditions were observed to develop as shallow groundwaters perched (primarily) at the transition to yellow earth or mottled/pallid clay. The extent of a large area of saturated surface soil was observed to correspond closely with the water levels in bores on the transect.

Records of continuous bore water level showed relatively steep rises and falls in the near surface soils and successively 'flatter' recessions as water levels declined within different soil types. Rates of water level decline ranged from 0.02m day^{-1} in mottled soil to 0.2m day^{-1} in the upslope gravels.

Groundwater chloride concentrations typically ranged from 20 - 40 mg l^{-1} with lowest concentrations in the upslope gravels and highest concentrations in mid-transect where soils were more clayey.

Stream discharge at S18 increased significantly (as a proportion of that at SGS) when the water level in bore 321 rose well into the yellow-red earth material. Discharge response was observed to be synchronous with bore water level response, particularly during a wetter period in September 1984.

A second, more dampened increase in bore water level occurred about 3 days after the first rise and rainfall. Stream discharge was maintained during this secondary water level rise. During the subsequent recession an exponential relationship was obtained between discharge at S18 and water level in bore 321. This relationship has been reported in the literature for spring discharge from groundwaters.

At a finer time scale bore water levels were found to increase by more than 0.25m within six hours during two storm periods. The form of the bore water level response was used to construct a 'groundwater' component in each of the two stream hydrographs. This analysis indicated that about 61-73% of storm period streamflow volume may have been composed of discharge from shallow groundwater.

Shallow groundwater therefore have an important role in the generation of streamflow.

8. STREAM WATER AND CHLORIDE

8.1 Introduction

Part of the aims of this study was to determine the nature and relative significance of sources of stream water and chloride. A conceptual hillslope model was presented in Chapter 2 as the basis for sources of stream water and chloride. These were conceptualised as being direct rainfall-runoff, a contribution from a shallow, seasonal groundwater and a contribution from a deeper, more saline, permanent groundwater.

In this chapter the significance of direct runoff is first determined. Then the typical seasonal water and chloride responses are outlined as an introduction to the yearly water and chloride yields.

The yearly water and chloride data are then interpreted with two simple models of chloride and water sources. The within season implications of the conceptual source model are then presented.

8.2 Significance of Direct Rainfall Runoff

8.2.1 Observations of Surface Runoff

Observations during moderate rainfall on 15 July 1980 and 29 August, 1980 indicated that direct surface runoff from rainfall to streamflow was effectively limited to the channel area and near environs. Surface runoff occurred from rock outcrops and from areas of saturated surface soils (source areas). These areas probably constitute at least 6% of the total catchment (see Figure 7.3) after several weeks of heavy rainfall but are much smaller than this for most of the time.

Therefore the proportion of direct rainfall-runoff in streamflow is probably small.

The volume of direct runoff was estimated to have been about 33% of total volume for two events at the temporary weir at S18 in 1984 (Chapter 7). These estimates were based on the difference between total discharge and a shallow groundwater discharge inferred from water level response in a bore. Estimates of the minimum source area for the production of this runoff appeared plausible.

The contribution of direct runoff is required at the main gauging station for the interpretation of the relative source proportions for the whole catchment.

8.2.2 Streamflow Travel Time

The separation of the hydrograph into source components during storm runoff requires information on the duration (storage) time of water in the channel system. In particular the time for most surface (overland flow) runoff to pass the gauging station from the time of the start of hydrograph rise is required. In general terms:-

$$t = L/V \qquad 8.1$$

where t is travel time, L channel length and V is average velocity.

For Salmon catchment an effective travel distance of about 1100m was estimated for the length of stream channel from the SGS gauging station to the headwater seepage areas. Observations during storms indicated that relatively little additional surface runoff sources occur outside of this zone. Therefore travel times for storm flow can be considered as principally that of channel flow.

Beven et al. (1979) presented velocity-discharge information from dilution gauging of small, rough upland streams with pools and cobbled riffles in Great Britain. These data were fitted to:-

$$V = a.Q^{n} \qquad 8.2$$

with the result of eight channel reaches yielding a mean n of 0.32 with a range of 0.13 to 0.505.

For Salmon catchment, velocity and discharge were obtained by current metering at S18, S17, S9 and S14 (Figure 5.1). These data are plotted in Figure 8.1 and fitted to equation 8.2 to yield:-

$$V = 1.67 \; Q^{0.424} \qquad 8.3$$

This result is reasonably consistent with the result obtained by Beven et al. (1979) for similar conditions.

For a typical storm period discharge range of 0.01 - 0.1 $m^3 \; s^{-1}$, average travel time for the 1100m reach (equations 8.1 and 8.3) range between 1.3hr to 0.5hr. These imply average velocities of 0.2 - 0.6m s^{-1} which are consistent with those of Dunne and Black (1970b).

In this study a conservative travel time of 3hr is adopted which may produce an overestimate of the duration of direct surface runoff and thereby underestimate the volume of subsurface contributions.

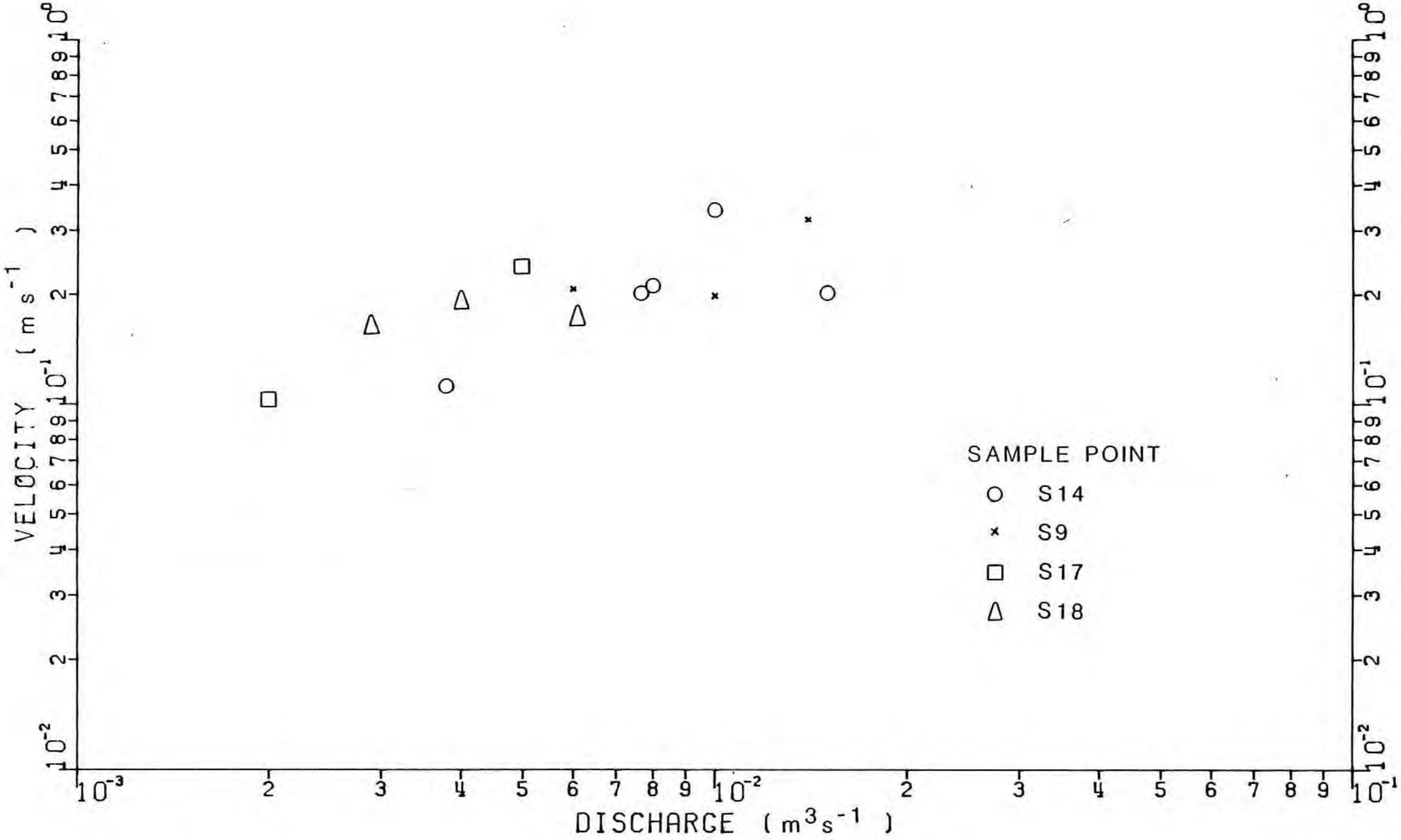


Figure 8.1
STREAM VELOCITY -DISCHARGE

That is, the adoption of 3 hours may result in an overestimation of direct runoff.

8.2.3 Storm Runoff Examples

Two rainfall and storm hydrograph events during 1983 are shown in Figures 8.2 and 8.3. In the first example during July 29-31 a five hour rain event produced 26mm with a peak intensity of 5.4mm in 15 minutes in the middle of the event. As a result discharge increased sharply from a pre-event baseflow of 0.01m^3 s^{-1} to a peak of 0.039m^3 s^{-1} which occurred within 3 hours of the hydrograph rise. Stream chloride decreased by 14 mg l^{-1} to a minimum of 60 mg l^{-1} which occurred just after the peak discharge. The stream recession from the peak was initially steep but then rapidly decreased to a more steady, sustained discharge which coincided with a steady increase in stream chloride concentration.

Although the bore 321 is more than 1100m upstream of the gauging station at SGS, it is of value to compare the storm period bore water level response with the discharge at SGS as shown in Figure 8.2. The bore water level rose 0.25m in 6 hours and peaked about 3 hours after the peak discharge.

The second example runoff event occurred one month later on August 31 - September 1 when 164mm of rain had fallen in the preceeding ten days. During the 2.5 hours to the end of August 31, 22mm of rainfall, with a peak intensity of 5.5mm in 15 minutes, produced a six-fold discharge increase to a peak of 0.17 m^3 s^{-1} within 2 hours. Stream chloride decreased sharply from 50 mg l^{-1} to a minimum of 32 mg l^{-1} just after the peak discharge. As with the July 29-31

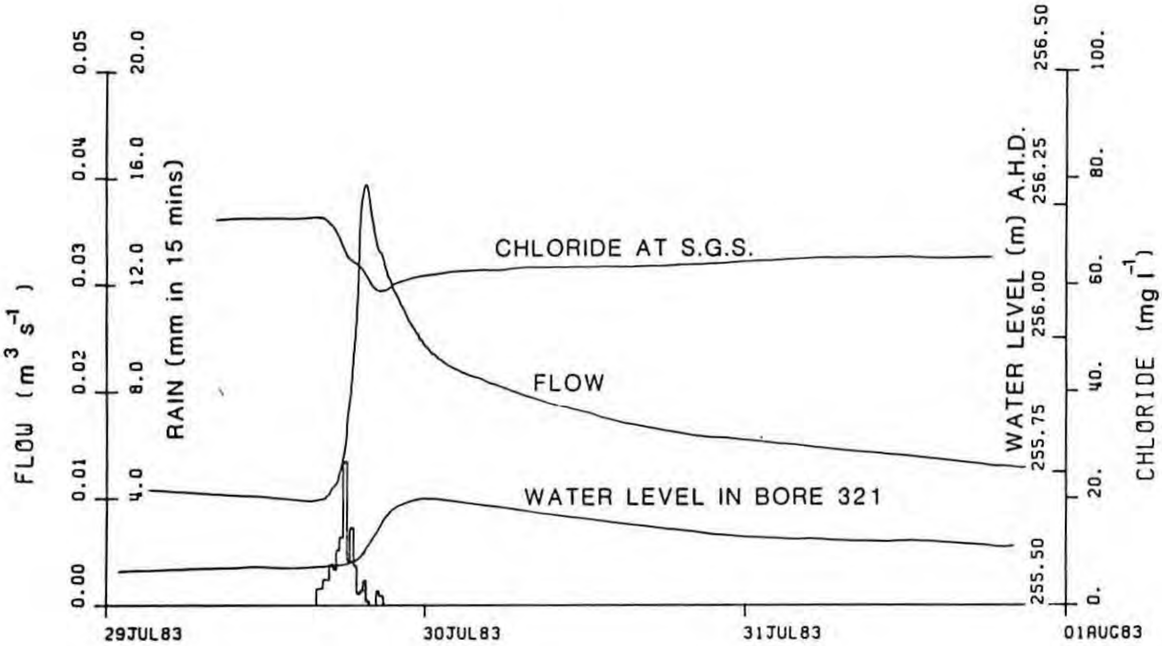


Figure 8.2 STORM RUNOFF EVENT : 29/7/83

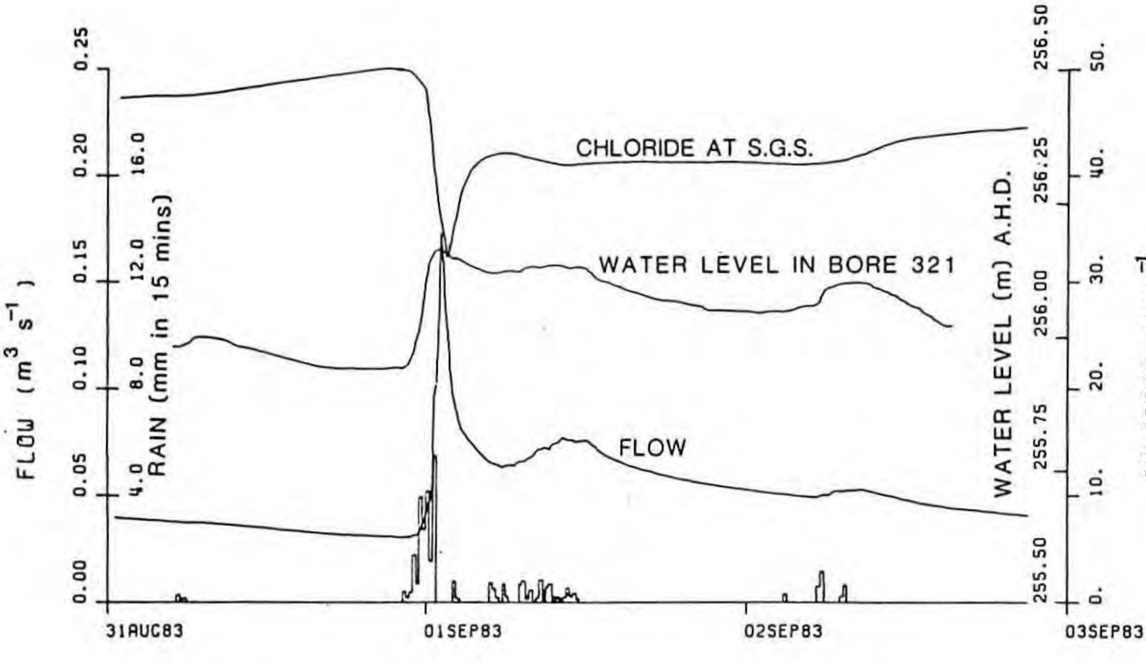


Figure 8.3 STORM RUNOFF EVENT : 1/9/83

event the discharge decreased steeply at first after the peak but then much less steeply to produce a sustained 'baseflow'. The stream chloride increases steeply from the minimum after the peak and then 'levels-off' at about 8 mg l^{-1} less than the pre-storm chloride concentration.

The water level in bore 321 (Figure 8.3) increased 0.45m in 3 hours and peaked at about the same time as the discharge at SGS. Both the stream discharge and bore water level recessions are of the same general shape, indicating that shallow subsurface flows may contribute significantly to streamflow.

8.2.4 Example Separations of Direct Runoff

The amount of direct rainfall-runoff as a proportion of total storm period volume and as a proportion of peak discharge will be estimated by a simple, 'geometric' method and a 'chemical' separation method.

Geometric Method

In the geometric model the groundwater contribution is assumed to vary linearly between the defined times of start and end of direct surface runoff. The start time is defined as the point on the hydrograph at which discharge increases from the pre-storm recession. The end of direct surface runoff is defined as 3 hours after significant rainfall. The simple assumption is shown in Figure 2.5.

Thus direct runoff (Q_r) is:-

$$Q_r = Q_T - Q_o \qquad 8.4$$

where Q_T is total discharge and Q_o is the linearly varying groundwater discharge.

Chemical Method

In the chemical model the concentration (C_o) of the groundwater (Q_o) is assumed to vary linearly between the defined start and end times. Thus from equations 2.11 and 2.12:-

$$Q_r = Q_T - Q_o \qquad 8.5$$

$$\text{and } Q_o = Q_T \; (C_T - C_r) \; / \; (C_o - C_r) \qquad 8.6$$

where C_T is concentration of total discharge and C_r is concentration of direct runoff.

Results

The direct runoff flow volume and peak discharge proportions for the events of 29/7/83 and 1/9/83 (Figures 8.2 and 8.3) are listed in Table 8.1. The direct runoff concentration (C_r) was assumed to be that of rainfall under canopy and a representative value of 10 mg l^{-1} was adopted (Table 4.2).

The chemical hydrograph separation method produces much less direct runoff than the geometric method with less than 20% by volume and less than 15% of the peak discharge. The geometric method produces about 2/3 direct runoff by volume and 1/3 by peak.

The chemical method probably underestimates the quantity of direct runoff because of the assumption of a linear response in the concentration. The concentration of the direct runoff is also likely to be higher than that of rainfall due to contact with solutes in the surface soils. This is evidenced by the fact that the minimum stream salinity occurs after the peak discharge, which it is reasonable to assume was the result of direct runoff.

Therefore the chemical method is not a useful means of separating storm period hydrographs because the simple assumptions (equations 2.11, 2.12) do not appear to apply. More information about variations in direct runoff salinity, shallow groundwater salinity and channel storage effects which are ignored in this model would be required.

For these reasons the geometric separation model will be used to estimate the yearly amount of direct rainfall runoff.

TABLE 8.1 EXAMPLE DIRECT RUNOFF SEPARATIONS

	Proportion of Volume			Proportion of Peak		
Event	Total Flow	Method		Peak Discharge	Method	
	m^3	Geom. %	Chem. %	$m^3\ S^{-1}$ %	Geom. %	Chem.
1700 29/7 220029/7/83	476	50	7	0.039	25	6
2330 31/8 0230 1/9/83	990	65	18	0.173	36	14

8.2.5 Yearly Contribution of Direct Rainfall-Runoff

The order - of yearly contribution of direct rainfall-runoff (Q_r) was calculated by applying the 'straight-line' (geometric) separation procedure (see Figure 2.5) to continuous discharge records. The quantity of Q_r in mm and as percentages of total discharge (Q_T) and rainfall are listed in Table 8.2

TABLE 8.2 YEARLY CONTRIBUTION OF DIRECT RUNOFF

Water Year	Rain (mm)	Q_T (mm)	Q_r (mm)	Q_r/R (%)	Q_r/Q_T (%)
1974	1493	367	15.3	1.0	4.2
1975	1001	84	1.6	0.2	1.9
1976	845	20	0.6	0.1	3.0
1977	972	74	5.5	0.6	7.4
1978	986	77	1.7	0.2	2.2
1979	831	17	0.6	0.1	3.5
1980	1284	138	3.0	0.2	2.2
1981	1473	173	5.4	0.4	3.1
1982	900	66	0.9	0.1	1.4
1983	1258	230	7.1	0.6	3.1
Mean	1105	125	4.2	0.4	3.2

Direct runoff varied from 0.1% of rainfall (1976, 1979 and 1982) to 1.0% (1974) with an average of 0.4% (4mm). As a proportion of total runoff, surface runoff ranged from 1.4% (1982) to 7.4% (1977) with an average of 3.2%.

The relatively high 7.4% in 1977 is possibly due to the more intense rainfalls, particularly in June and August, which produced more direct runoff even though total rainfall was the fourth lowest on record.

It is emphasised that these estimates of direct runoff are likely to be too high because of the method which assumes a linear groundwater discharge response between the defined times of direct runoff start and end. Most of the evidence indicates that the shallow groundwater responds rapidly to rainfall and contributes most of the discharge during the storm runoff period.

To illustrate this, the chemical separation procedure discussed above was applied to the storm events in 1983. This method produced just 2.8mm of direct runoff, significantly less than the 7.1mm by the geometric, 'straight-line', baseflow approach. The chemical method probably overestimates the contribution of groundwaters because the chemistry of direct runoff will be altered by contact with surface soils.

Therefore the 'true' quantity of direct rainfall runoff probably lies somewhere between the two estimates. However by either method the amount is small relative to total rainfall and to total yearly flow. Therefore direct runoff can be considered as a minor source component in yearly streamflow volume. Storm period peak stream discharge may be generated by direct runoff.

8.3 Seasonal Stream Water and Chloride

8.3.1 Flow Variability and Duration

The daily distributions of runoff (mm depth) for each of the ten years between 1974 and 1983 are shown in

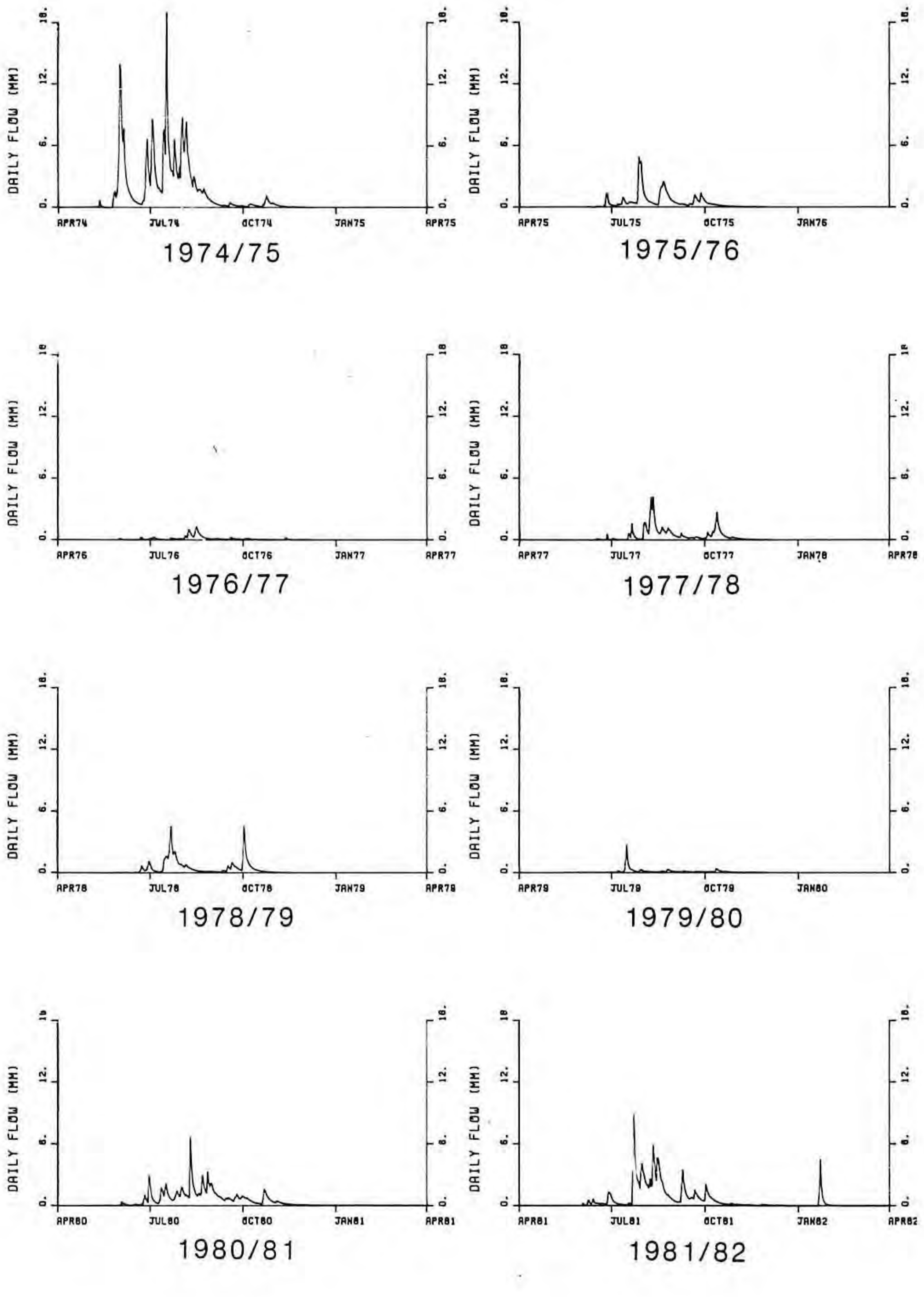


Figure 8.4
SEASONAL STREAMFLOW

Figure 8.4. Runoff in 1974 was much greater than any year except 1983.

Streamflow occurred for an average of 179 days (approximately six months) over the ten years with a range from 151 days in the driest year 1979 to 218 days in 1981/82. Flow occurred for 17 days in January-February 1982 producing 10mm runoff following 210mm rainfall over two days. Streamflow has always commenced in mid to late May or June following good rainfall and except for the 1982 summer has always ceased in late November or early December.

The absence of flow before heavy rains in May-June or after December indicates that streamflow is generated by a seasonal excess of rainfall over evaporation. Discharge from the deeper, permanent groundwater (Chapter 6), if it occurs at all, is insufficient to sustain a baseflow throughout the drier six months of the year.

8.3.2 Seasonal Stream Chloride

The stream water chloride concentrations shown in Figure 8.5 vary from more than 600 mg l^{-1} at the start of streamflow (1975) to minima of less than 50 mg l^{-1} during mid-winter flows. Larger discharge such as in 1974, 1983 and 1981 produce lower seasonal minima. Concentrations increase towards cease-to-flow although the final values are usually less than those of the initial flows.

Daily average chloride concentrations are plotted against daily accumulated flow volume in Figure 8.6. These data indicate an initial dilution or leaching of chloride as the concentration decreases with accumulating flow. The concentrations generally remain

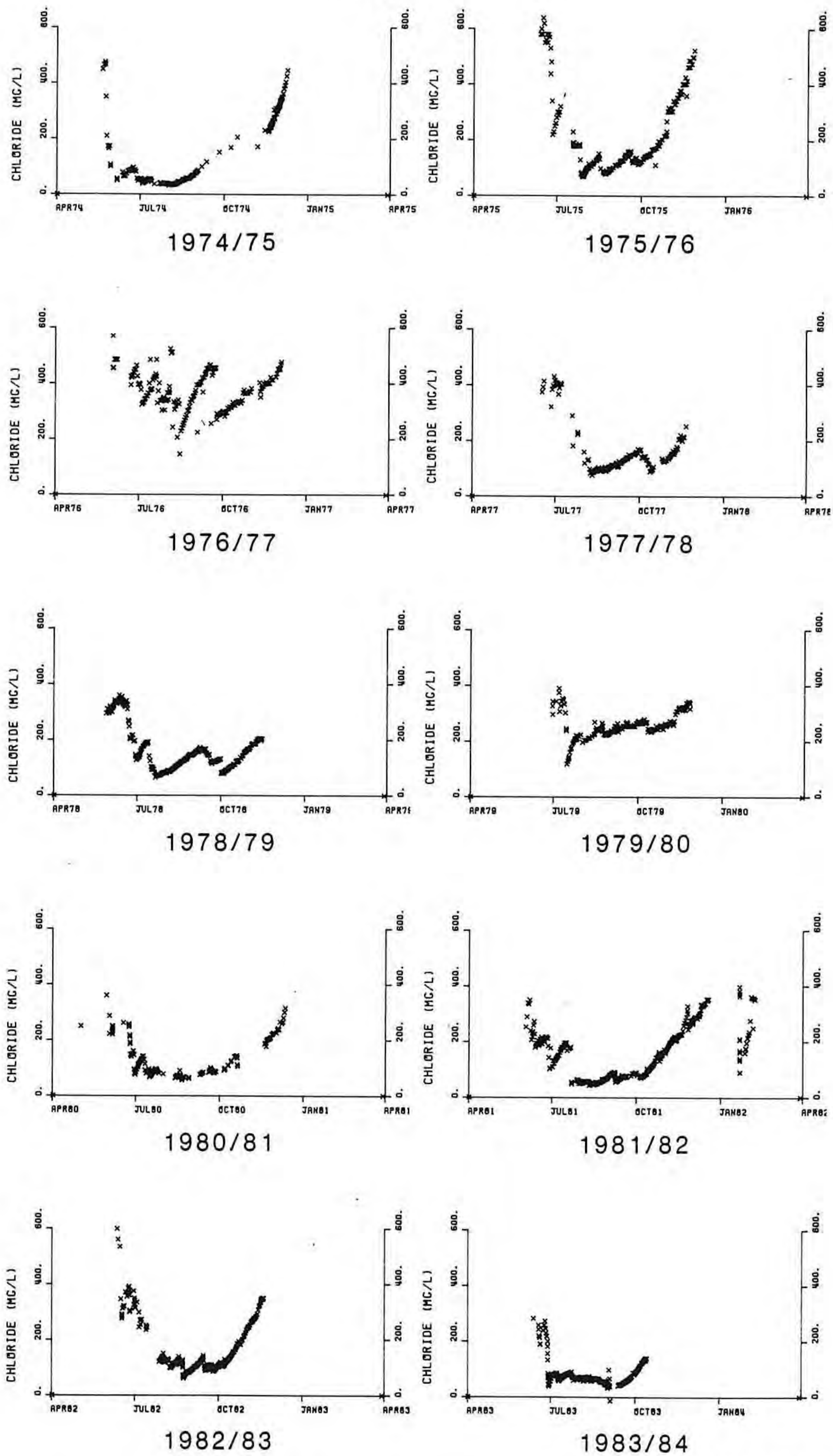
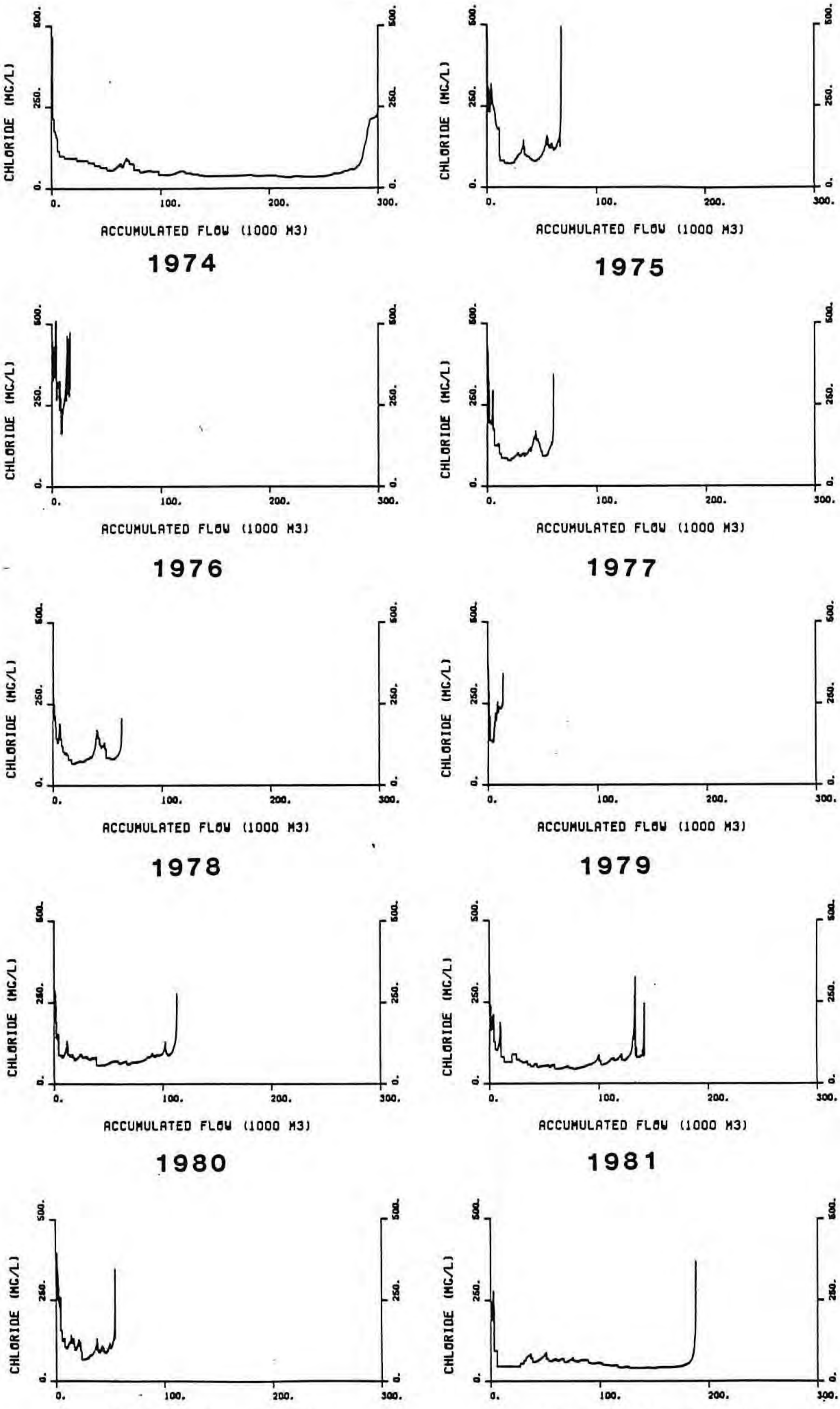


Figure 8.5
SEASONAL CHLORIDE CONCENTRATION



CHLORIDE CONCENTRATION VERSUS ACCUMULATED FLOW

Figure 8.6

fairly constant through the period during which most of the streamflow occurs, particularly in the wetter years such as 1974, 1983, 1981 and 1980. In the drier years the mid season concentrations increase and decrease more with varying flows.

Concentrations increase rapidly towards cease-to-flow (Figure 8.6). However this occurs with very little additional flow and does not result in much additional chloride load as can be seen in Figure 8.7 where the daily accumulated chloride is plotted against daily accumulated water. This figure also indicates the process of an initially higher rate of chloride leaching which decreases with increased water yield. This seasonal behaviour occurs every year although it is less pronounced in the driest years of 1976 and 1979.

In these years there was apparently insufficient streamflow (less than 20mm) to significantly deplete the initial store of chloride. Therefore stream chloride concentrations remained relatively high throughout the season (Figures 8.5 and 8.6).

8.4 Yearly Stream Water and Chloride

The seasonal accumulated daily chloride and streamflow data (Figure 8.7) indicated a similar response from year to year with evidence of a chloride leaching mechanism through the season. The yearly chloride and water yield data in Figure 8.8 show the non-linear relationship from year to year.

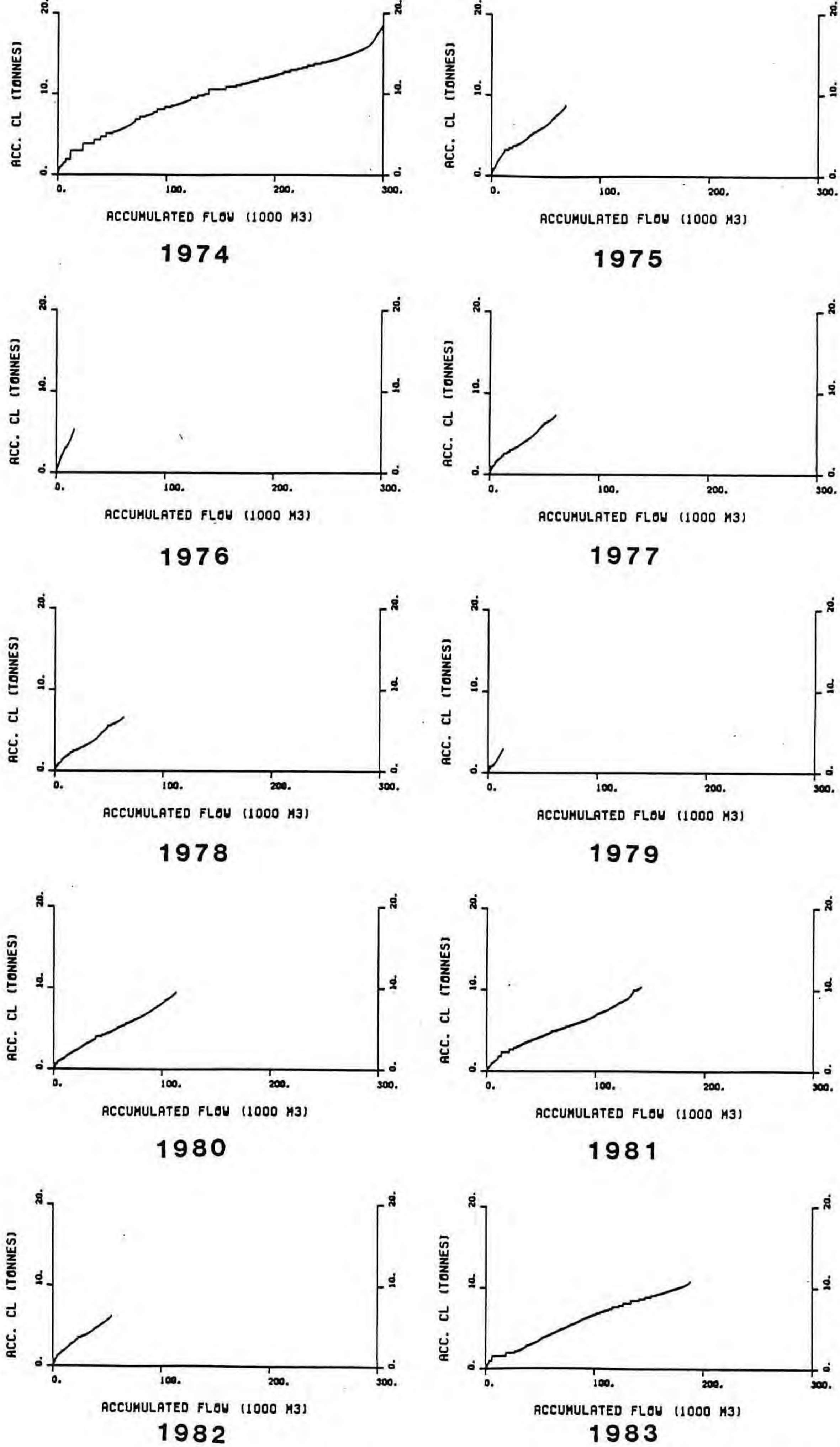


Figure 8.7 CHLORIDE LOAD VERSUS FLOW

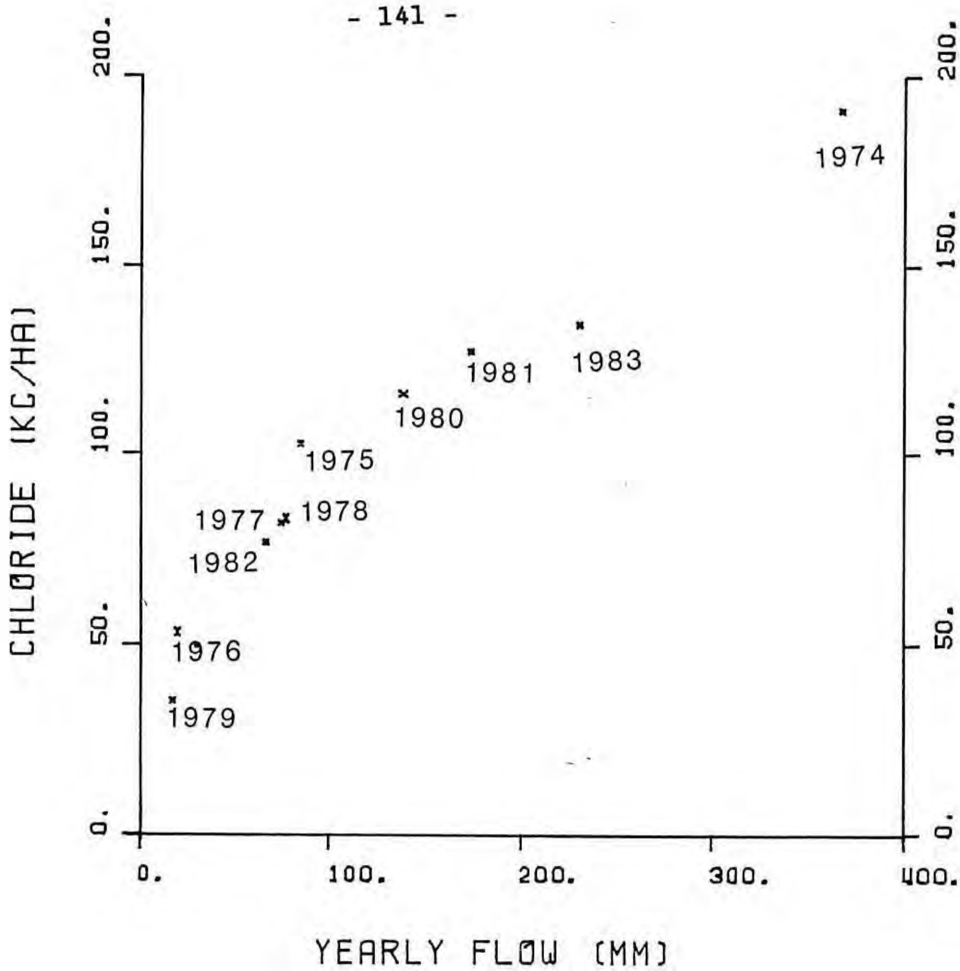


Figure 8.8 YEARLY CHLORIDE LOAD AND FLOW

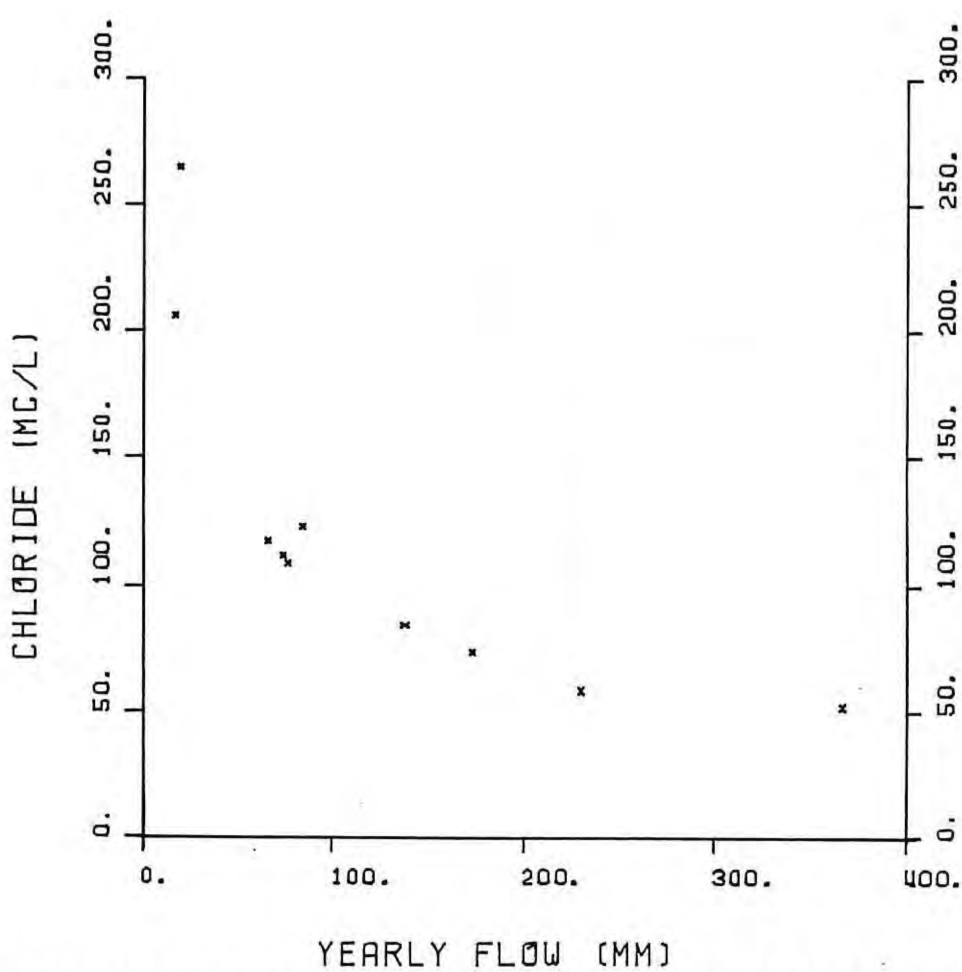


Figure 8.9 YEARLY CHLORIDE CONCENTRATION AND FLOW

A general relationship was fitted by regression:-

$$L = 10.38\ Q^{0.49} \quad \text{with } r^2 = 0.96 \qquad 8.7$$

where L is load (kg ha^{-1}) and Q is runoff (mm) (the yearly data are listed in Table 4.4).

It follows that the yearly flow-weighted concentration (C, mg l^{-1}) is:-

$$C = 1038.\ Q^{-0.51} \qquad 8.8$$

and is show in Figure 8.9.

Yearly chloride load is approximately proportional (equation 8.7) to the square root of accumulated flow. As a result, for flow above about 300mm yr^{-1} the yearly concentration asymptotes to 50 mg l^{-1}. Therefore additional load is produced at approximately a constant concentration which is not the case at lower flow.

Another important feature of the data is the significant differences in the chloride load between the first few years and the last years. In particular, there is a large difference in load between 1976 (53 kg ha^{-1}) and 1979 (35 kg ha^{-1}) although the water yields were similar at 20mm and 17mm respectively (Table 4.4). The load in 1975 was also larger than those in comparable years such as 1977 and 1978.

These differences were discussed in Chapter 4 on the water and chloride balances. There was more chloride output than input with evidence for a decrease in the chloride output/input ratio over the ten years.

Observations of the spatial sources of water and chloride (Chapter 5) and of the possibility of contribution by deeper groundwaters (Chapter 6) to streamflow were made. The deeper groundwater was much more saline than the seasonal, shallow groundwater and the deeper groundwater levels varied from a high in 1974 to a low in 1980. Therefore a contribution of water and particularly chloride to streamflow from the deeper groundwater could produce a net export of chloride from the catchment. A decrease in groundwater levels may also have resulted in a decreased contribution of water and chloride to streamflow.

Aspects of the sources of water and chloride are explored in the next section.

8.5 Sources of Water and Chloride

8.5.1 Conceptual Model

The conceptual model of the catchment as a hillslope section (see Figure 2.2) envisaged water and salt to be generated from three sources:-

(i) direct rainfall-runoff (Q_r)
(ii) discharge from a shallow, seasonal aquifer (Q_u)
(iii) discharge from a deeper, permanent and more saline aquifer (Q_g)

Therefore total stream water and chloride is composed of the sums of the (separate) component sources (equations 2.11 and 2.12):-

$$Q_T = Q_r + Q_u + Q_g \qquad 8.9$$

$$C_T Q_T = C_r Q_r + C_u Q_u + C_g Q_g \qquad 8.10$$

Where direct rainfall-runoff (Q_r) is small relative to total streamflow and particularly as the concentration (C_r) is much less than the groundwaters this component can be ignored. Therefore:-

$$Q_T = Q_u + Q_g \qquad 8.11$$

and

$$C_T Q_T = C_u Q_u + C_g Q_g \qquad 8.12$$

Solving:-

$$C_T = C_u + (C_g - C_u) \; Q_g/Q_T \qquad 8.13$$

Equation 8.13 is shown plotted in Figure 8.10b where the parameters were estimated by fitting equation 8.12 to the data listed in Table 4.4 in two groups:-

1976, 1979	$L = 2.4 \,.\, Q$	8.14
Others	$L = 62 + 0.35 \,.\, Q$	8.15

(See Figure 8.10a)

If the 1976 and 1979 are deeper groundwater contribution then $C_g = 240$ mg l^{-1} and from equation 8.15, $C_u = 35$ mg l^{-1}. The Q_g term is found by solving equations 8.14 and 8.15 for Q; producing $Q_g = 30$mm. Therefore equation 8.13 becomes:

$$C_T = 35 + (240 - 35) \times 30/Q_T \qquad 8.16$$

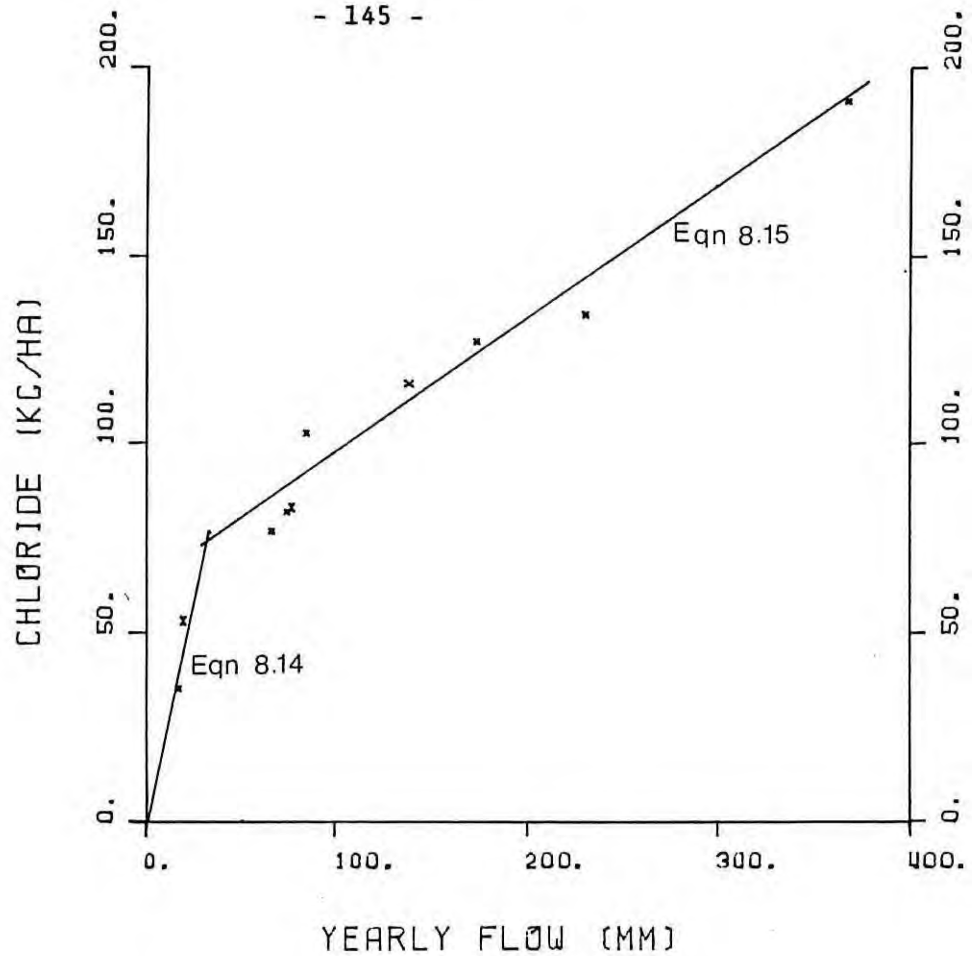


Figure 8.10a LINEAR SOURCE MODEL: YEARLY LOAD

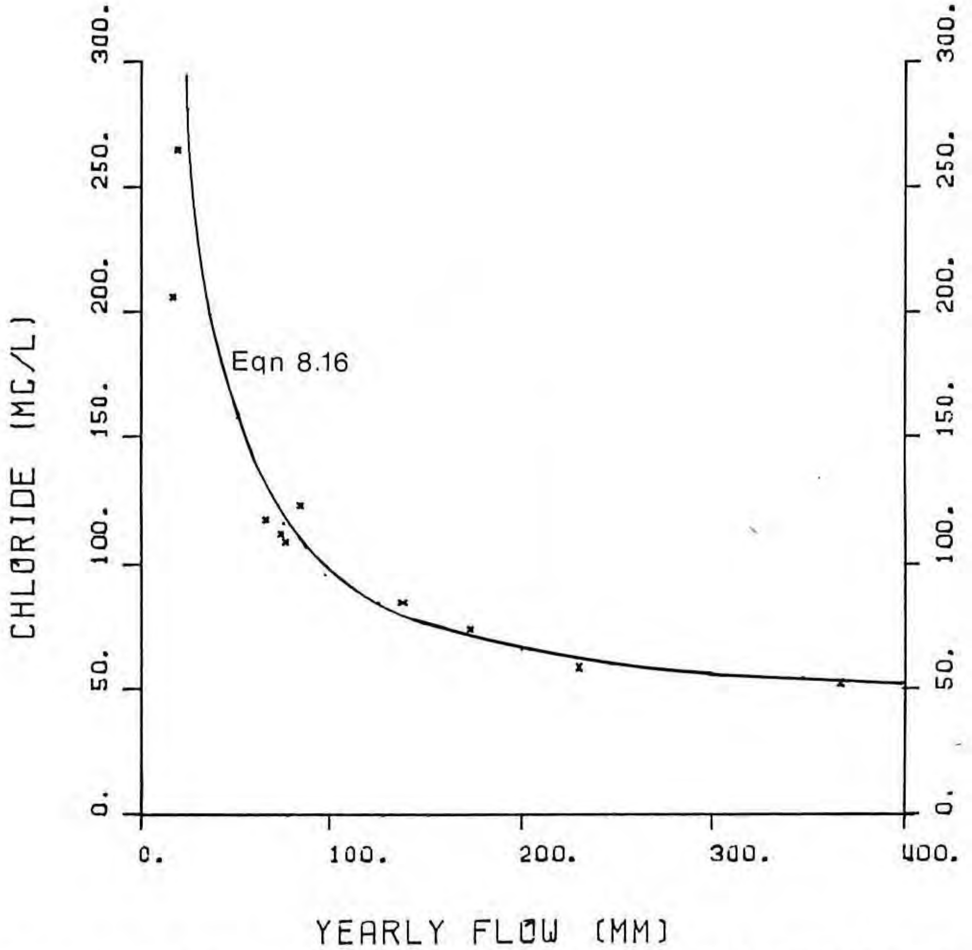


Figure 8.10b LINEAR MODEL: YEARLY CONCENTRATION

Both the shallow groundwater concentration (C_u) of 35 mg l^{-1} and the deeper groundwater (C_g) of 240 mg l^{-1} appear reasonable based on observations of water in shallow and deeper boreholes (Chapter 6 and 7).

The groundwater contribution of 30mm is of the right order, but is perhaps a little large because average annual recharge was estimated by Johnston (1983) to be only 25mm.

This model of sources of water and chloride basically assumes that each component occurs at a fixed concentration. Clearly this is not appropriate as less than 20mm occurred in total in 1976 and 1979 which is less than the nominal deeper groundwater contribution of 30mm alone. Some shallow groundwater Q_u and surface runoff, (Q_r) would have contributed in these years.

Therefore this conceptual, linear source component model, with the assumption of constant component concentrations, does not produce reasonable proportions of sources of stream water.

8.5.2 Exponential Source Model

An alternative approach is one based on a leaching model as discussed in Chapter 2.

The mechanism envisaged is of the accumulation of chloride in near surface soils from atmospheric saltfall and evaporation of deeper groundwater and subsequent leaching to the stream by the seasonal groundwater. The shallow groundwater functions as a 'carrier' medium for catchment solutes into streamflow. The model is (after Peck 1973, 1976; Duffy, 1984):-

$$L = C\,Q + M\,(1 - \exp(-K.Q)) \qquad 8.17$$

where Q is total flow (mm yr^{-1}), C is shallow groundwater concentration (mg l^{-1}), M is mass of yearly chloride accumulation (kg ha^{-1}) and K is a parameter (mm^{-1}) of the system storage (Peck, 1973).

The parameters of the model, C, M and K were estimated by first determining C, the 'carrier' concentration. A value of 35 mg l^{-1} was tried initially, based on the results of the previous model. However this produced individual year estimates of M for 1975 and 1974 which were not reasonable in that the store (M) for 1974 was significantly smaller than that for 1975. A value for C of 30 mg l^{-1} produced a more reasonable distribution of individual M's (Figure 8.11a) and is also consistent with the concentrations of shallow groundwaters in Salmon catchment where no deeper groundwaters occur (Chapter 7). Shallow groundwater concentrations in areas of no contributing deeper groundwaters (in the Darling Range) are also often around 30 mg l^{-1} (Peck et al., 1982a).

With a C of 30 mg l^{-1}, K and M were calculated separately for the 1974-1976 and the 1977-1983 periods and adjusted to produce:-

$$\text{1974-1976:} \quad L = 0.3\,.\,Q + 80\,(1 - \exp(-0.04.Q)) \qquad 8.18$$

$$\text{1977-1983:} \quad L = 0.3\,.\,Q + 65\,(1 - \exp(-0.04.Q)) \qquad 8.19$$

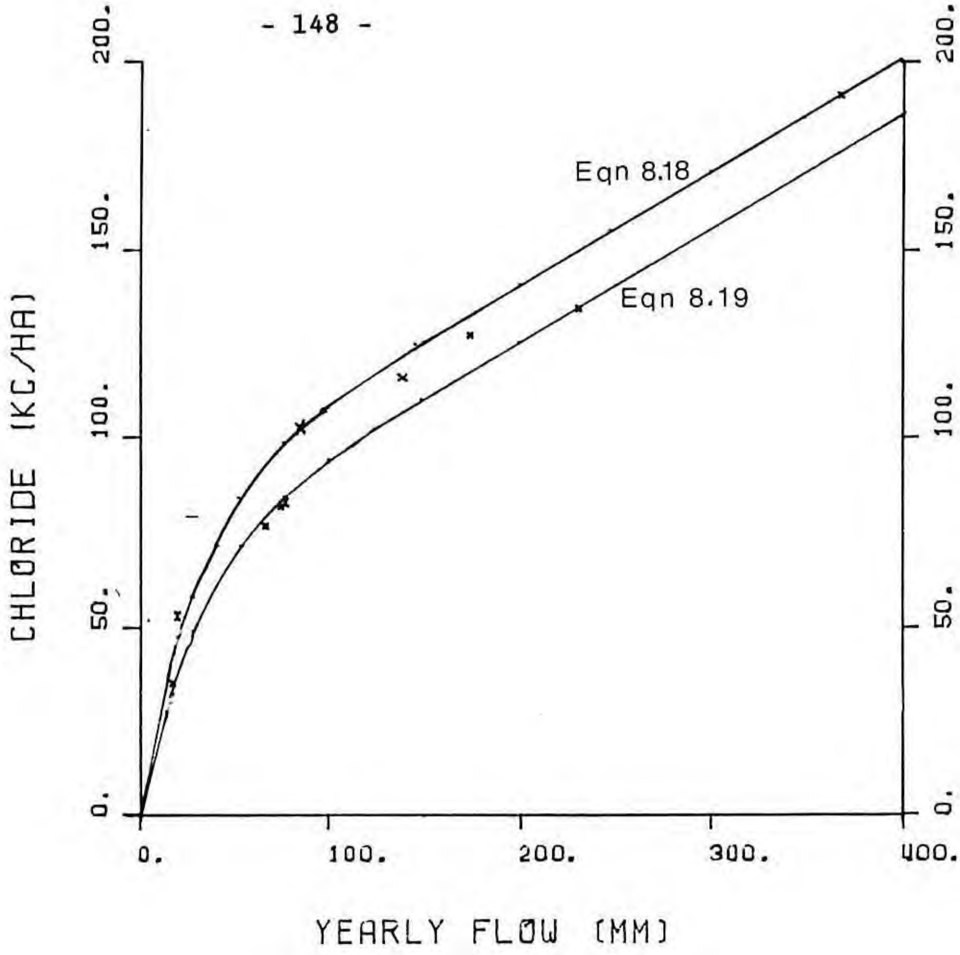


Figure 8.11a EXPONENTIAL MODEL: YEARLY LOAD

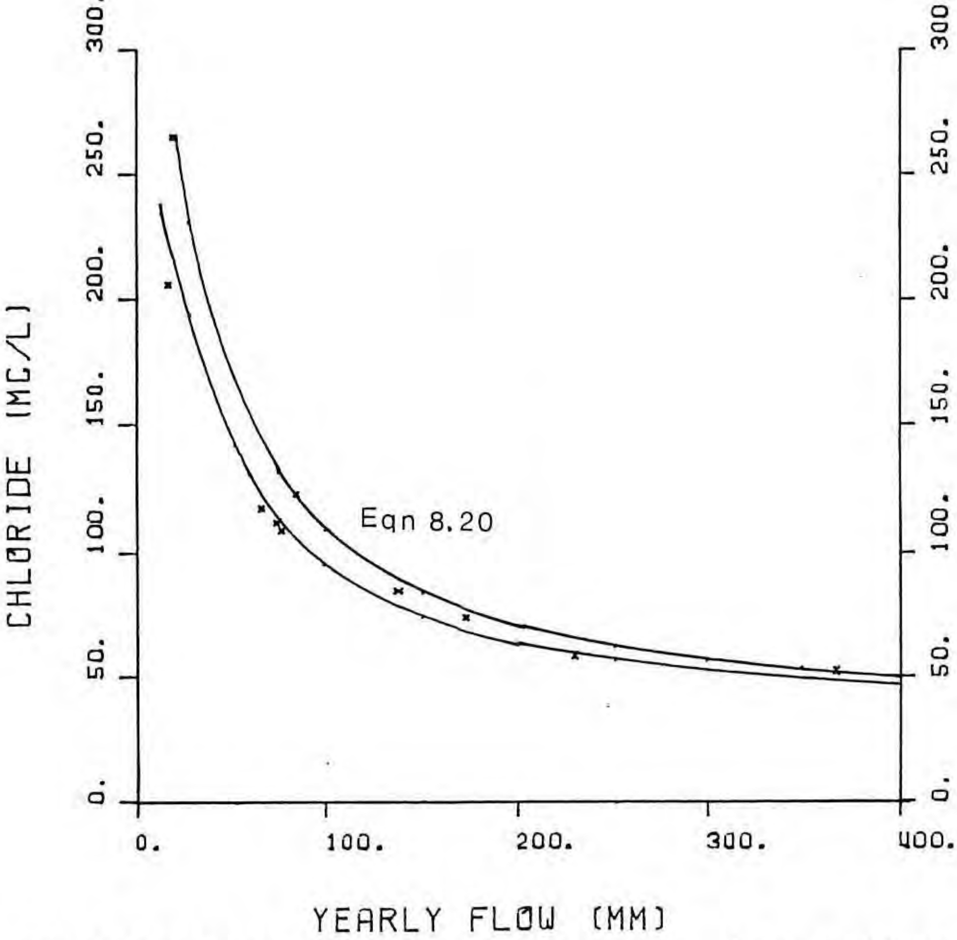


Figure 8.11b EXPONENTIAL MODEL: YEARLY CONCENTRATION

The mass of chloride M are significantly different between 1974-1976 and 1977-1983. This is possibly a result of the decline in deeper bore water levels between 1974 and 1980 (Chapter 6). A reduced flux of deeper groundwater could result in a smaller accumulation of chloride for subsequent leaching.

A value of K of 0.04 mm^{-1} was adopted for both periods. The inverse of K, termed W by Peck (1973, 1976), is 25mm. The W was considered by Peck to represent the "moisture storage" for a simple, well-mixed, one-dimensional leaching model. This is essentially the volume (containing M) to be leached.

In equation 8.17, W may represent the flux of deeper groundwater which is the source of accumulated chloride M. The concentrations of the groundwater are then (C_g = M/W) 320 mg l^{-1} and 260 mg l^{-1} chloride for the 1974-1976 and 1977-1983 periods.

The flow-weighted yearly concentration from 8.17 is:-

$$C = 30 + M\,(1 - \exp(-0.04.Q))/Q \qquad 8.20$$

and are shown in Figure 8.11b for 1974-1976 and 1977-1983.

8.5.3 Seasonal Analysis

In the preceeding analyses, the separation of groundwater components ignored the seasonal variation of discharge and stream chloride. In this section the effects of various assumptions about concentrations of sources and discharge rates and durations are investigated using the 1983 stream record.

The plot of logarithmic discharge and stream chloride at SGS in 1983 is shown in Figure 8.12. Discharge was greater than 10^{-3} m^3 s^{-1} for approximately 110 days between the middle of June and early October. For most of this time chloride was less than 100 mg l^{-1}, the exception being two weeks in late June.

By the time the shallow bores were effectively dry in early October (Chapter 7) stream discharge was 10^{-3} m^3 s^{-1} and chloride 100 mg l^{-1}. Chloride had increased to around 200 mg l^{-1} as discharge decreased to 10^{-4} m^3 s^{-1} by late October when flow ceased from upstream of the mid-catchment gauging site at S9. During the next month chloride increased to around 350mg l^{-1} as discharge fluctuated between 5×10^{-4} m^3 s^{-1} and 10^{-3} m^3 s^{-1}.

This increase in stream chloride concentration towards cease-to-flow for all years was noted in Figure 8.5. This is attributed to a combination of a decrease in the contribution of the shallow groundwater (Q_u) relative to the deeper, more saline groundwater (Q_g) and to evaporation.

The effects of the accumulation of chloride and subsequent leaching and of evaporation towards cease-to-flow are illustrated by using the linear source model (equations 8.11 and 8.12) to calculate the variation of deeper groundwater (Q_g) and the shallow groundwater concentration (C_u) through 1983. That is:-

$$Q_g = Q_T (C_T - C_u) / (C_g - C_u) \qquad 8.21$$

and

$$C_u = (C_T Q_T - C_g Q_g) / (Q_T - Q_g) \qquad 8.22$$

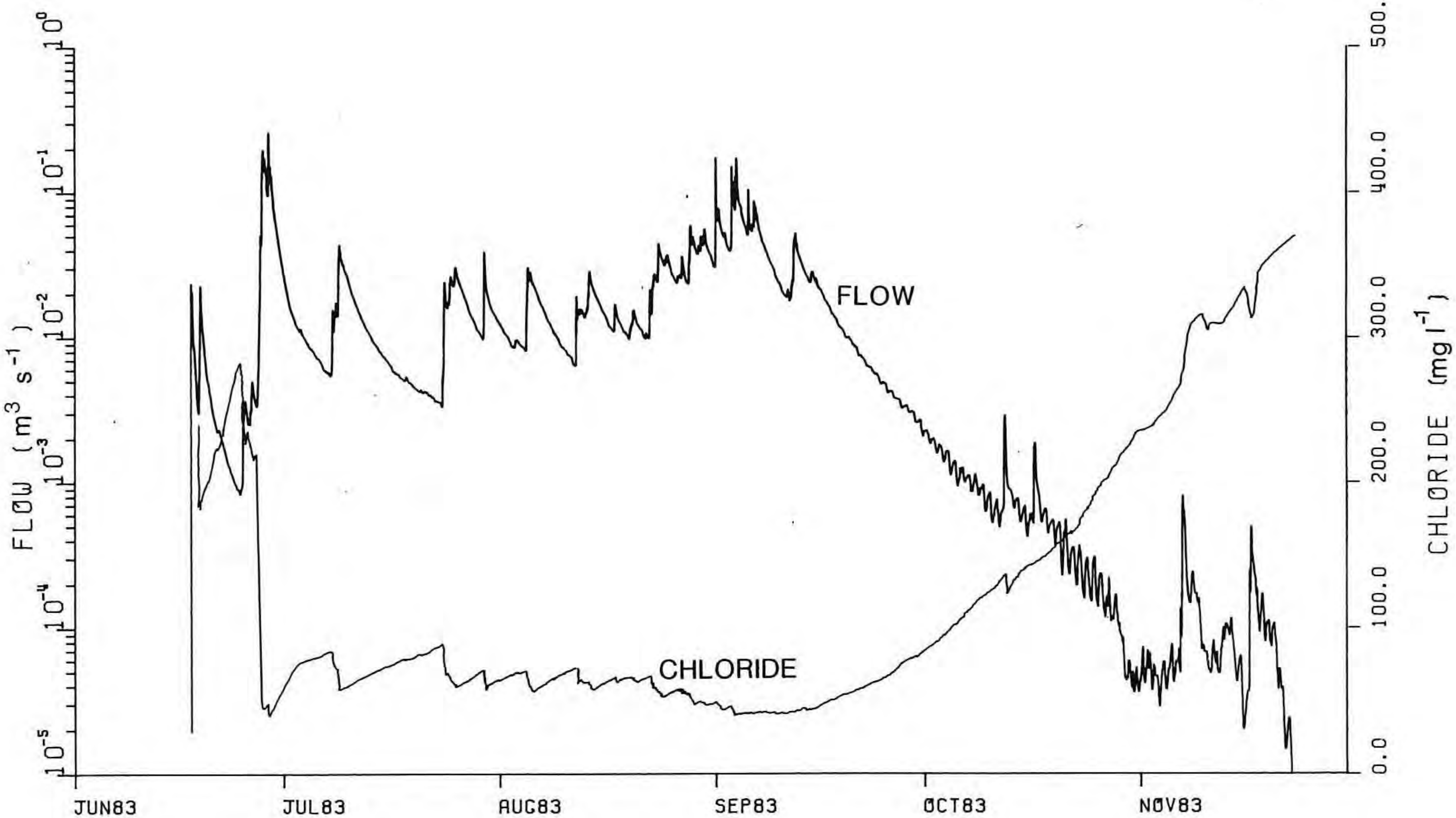


Figure 8.12
STREAM DISCHARGE AND CHLORIDE: 1983

For equation 8.21, C_g and C_u were set at (nominal) values of 200 mg l^{-1} and 40 mg l^{-1} and the resulting Q_g discharge through 1983 is shown in Figure 8.13. The important feature is the decrease from the start-to-flow in mid June to cease-to-flow in November. This is inconsistent with the variation of water levels in the deeper bores where minimum levels occurred in June and the peaks in September-October. Therefore groundwater discharge should <u>increase</u> throughout the period of flow, at least until late September, early October.

The general decrease of Q_g as shown is a result of the separation which 'forces' Q_g to be higher in the early period because stream chloride concentrations are higher. As flow continues and chloride is leached, concentrations decrease and thus Q_g decreases.

Therefore this result is not reasonable and constant concentrations cannot be assumed through a season.

If the contribution and concentration of deeper groundwater (Q_g and C_g) are assumed to be constant through the season then the shallow groundwater concentration can be calculated from equation 8.22. The yearly analysis indicated a possible groundwater contribution of 25 - 30 mm over a year. On this basis an instantaneous Q_g of 6×10^{-4} m^3 s^{-1} was selected along with a C_g of 200 mg l^{-1} for use in equation 8.22. The resulting C_u is shown in Figure 8.14.

The C_u varies considerably, although consistently through the flow period. Values in June are high, indicating significant additional solute load. From July until late September C_u decreases steadily from

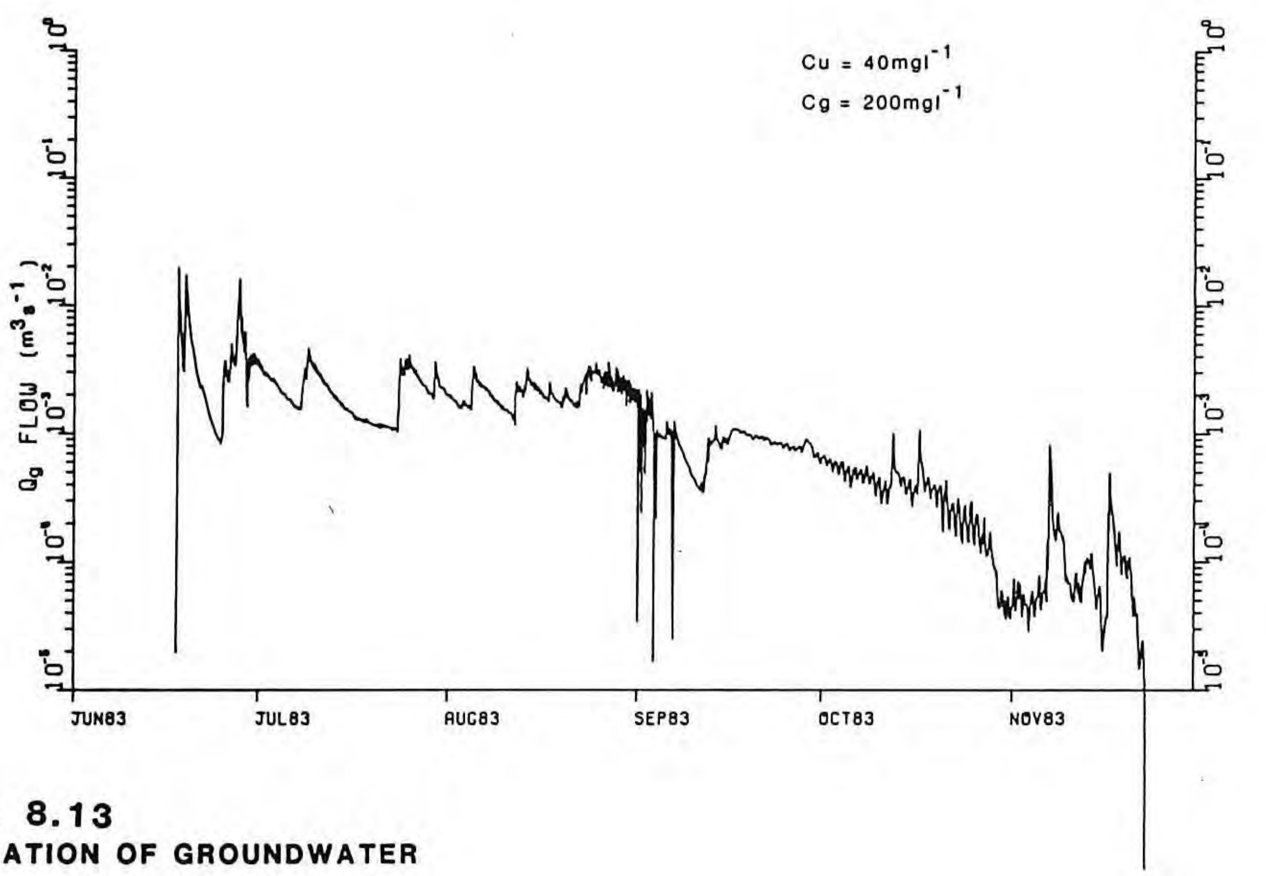


Figure 8.13
SEPARATION OF GROUNDWATER

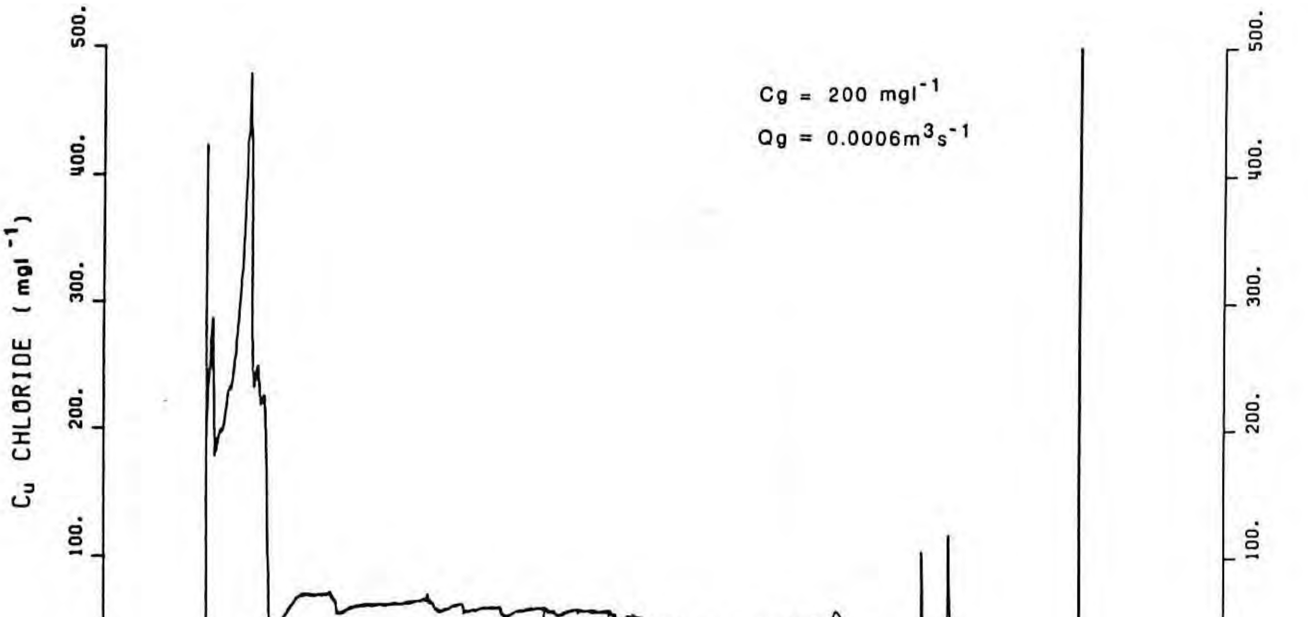


Figure 8.14
CALCULATED SHALLOW GROUNDWATER CHLORIDE

around 75mg l^{-1} to 40mg l^{-1}. Each decrease occurs after significant rainfall. Between such events the concentration is almost constant.

From late September C_u decreases to zero by early October. The decrease is a result of decreasing streamflow and separation using a fixed Q_g and C_g. The decrease can be interpreted as a decrease in Q_u rather than C_u as the shallow groundwaters dry-up (Chapter 7).

The isolated, sharp C_u peaks in October and November are the result of small runoff events. The higher C_u values are produced by a combination of leaching of accumulated solute and the separation of the fixed Q_g from an only slightly larger total discharge (Q_T). This produces a small Q_u term and thus a larger C_u.

It is clear from this analysis that the nominal shallow and deeper groundwater concentrations, as used in the separation procedure are not constant throughout the period of flow. The shallow groundwater concentration (C_u) is most likely to vary as was observed in the variation of chloride at seepage sites over the catchment (Chapter 5).

Therefore it is not possible to use the three component source model of streamflow (equations 2.11 and 2.12) with constant concentrations to estimate the proportions of shallow and deeper groundwaters through a season or with yearly data.

8.5.4 Source Contributions

The analysis of the probable contribution of direct rainfall runoff indicated that this component was relatively insignificant with an average of 4mm over ten years using a method which probably gave an overestimate. The contribution of chloride (directly) to streamflow by this component is of even less significance because the average concentration of throughfall is only about 10mg l^{-1} chloride.

The yearly flux of deeper groundwater was estimated to be about 25mm. However not all of this contributes to streamflow as the stream flows for only six months during winter. A constant deeper groundwater flux of 25mm yr^{-1} means a daily flux of 0.068 mm day^{-1} or about 6.5×10^{-4} m^3 s^{-1} which is very small in comparison with daily pan evaporations of 8-10 mm during summer. Therefore it is not surprising that there is no streamflow over summer. At most, the deeper groundwater could not contribute more than 12.5mm to the stream over the six months of flow and is likely to be less than this because of evaporation during this period.

Whereas there is likely to be little deeper groundwater water in streamflow the evidence is that a significant amount of chloride load is from the deeper groundwater. The exponential leaching model gave estimates of chloride stores of between 65 and 80kg ha^{-1} (5.3 to 6.5 tonnes). These stores could be developed in the shallow soils by a deeper groundwater flux of 26mm yr^{-1} at an average concentration of 260-320 mg l^{-1}. These concentrations are well within the range observed in the deeper bores.

Approximate source proportions of stream water and chloride are listed in Table 8.3. The Q_r components are from Table 8.2 and the L_r is taken as zero. The deeper groundwater component (Q_g) is assumed to be constant at 12mm which is probably too high. By difference, the shallow groundwater (Q_u) have contributed between 4mm (1979) to 350mm (1974).

TABLE 8.3 SOURCES OF STREAM WATER AND CHLORIDE

Water Year	Water (mm)				Chloride (kg ha^{-1})			
	Q_T	Q_r	Q_u	Q_g(i)	L_T	L_r(ii)	L_u(iii)	L_g
1974	367	15	350	12	191	-	105	86
1975	84	2	70	12	103	-	21	82
1976	20	1	7	12	53	-	2	51
1977	74	6	56	12	82	-	17	65
1978	77	2	63	12	83	-	19	64
1979	17	1	4	12	35	-	1	34
1980	138	3	123	12	116	-	37	79
1981	173	5	156	12	127	-	47	80
1982	66	1	53	12	77	-	16	61
1983	230	7	211	12	134	-	63	71

Notes (i) Q_g constant 12mm yr^{-1}

(ii) L_r : insignificant direct runoff chloride load

(iii) L_u : $L_u = 30$ (mg l^{-1}) x Q_u from equation 8.17

(iv) subscripts T, r, u, g represent total, direct, shallow groundwater and deeper groundwater respectively.

The proportions of chloride were calculated by using a constant shallow groundwater concentration of 30mg l^{-1} (equation 8.17) and then calculating the deeper component by difference. Most chloride is contributed by the deeper, more saline groundwaters and probably transported in seasonal groundwater discharge.

8.6 Summary

Direct rainfall-runoff is a relatively small component of total streamflow at about 3.2% averaged over ten years. This is possibly an overestimate because of the hydrograph separation procedure used. Separation using a chemical approach produced very high subsurface contributions and little direct runoff during storm runoff. This is considered to be a consequence of using a model which does not include storage effects and does not account for variations in direct runoff and subsurface chemistry.

Seasonally, streamflow occurs for only about six months from June to November in response to the seasonal rainfall (about 80% of yearly total). Stream chloride load and concentration indicate a significant leaching in the first flows of the season and an increase in concentration towards cease-to-flow.

Two models of stream chloride load were tested on the yearly water and chloride load data. A conceptual, three source model and an exponential leaching model were able to reproduce the general characteristics of yearly chloride load and water yield. The exponential model was better because of the non-linearity of the data and because the three source (linear) model produced unrealistically high deeper groundwater contributions. The inadequacy of the linear model was further demonstrated by analysis of the modelled responses through the 1983 flow period.

The shallow groundwaters are the dominant mechanism in the generation of stream water on Salmon catchment. Relatively little (if any) water is contributed by the deeper groundwater.

In most years the dominant source of stream chloride appears to be from the deeper groundwaters. The data indicate that this chloride is probably transported from the catchment in the seasonal groundwater discharge.

9. SUMMARY AND CONCLUSION

9.1 Summary

The aim of this study was to identify and quantify stream water and chloride generation in Salmon catchment. Catchment and sub-catchment water and chloride budgets were calculated, the spatial and temporal variability of streamflow sources were surveyed, characteristics of the shallow and deeper groundwater studied and two models of stream water and chloride sources were investigated.

The catchment water balance is dominated by evapotranspiration which accounts for about 90% of the average rainfall. Net interception was calculated to be about 16% of rainfall.

Chloride fall under forest canopy was significantly higher than that measured in gauges in well exposed sites in forest clearings. The difference was sufficient to change the catchment chloride budget (output/input) from significantly greater than 100% (exposed gauges) to approximately one of balance (canopy gauges). The chloride budget of about unity or greater is much greater than the catchment water output which has not exceeded 30% of rainfall.

The non-linear variation of chloride output/input relative to the water output/input, and the net output of chloride, indicated the contribution of deeper groundwater to streamflow.

A survey of sources of streamflow within the catchment identified discrete sources located within or close to the stream. These were often 'pipes' located in the stream bed or bank and several contributed significant quantities of water to streamflow. Considerable variations in the salinity of these sources were observed during winter and between the sources. These variations and subcatchment water and chloride budgets were used to classify the catchment into three hydrological zones.

In the topographically convergent headwater zone relatively fresh streamflow is generated from shallow, seasonal groundwaters. A more saline inflow, possibly from the deeper groundwater contributes in the more incised mid-catchment zone. The third zone is based on steeper, more shallow soil slopes and stream water and chloride generation is more efficient with evidence of a more saline inflow. These zones of observed hydrologic response correspond with those mapped by Bettenay et al. (1980) on the basis of topography, soils and vegetation.

The excess of catchment chloride output over input and the variations in stream source salinity within the catchment indicated that a more saline, permanent groundwater contributes to streamflow. A deeper groundwater was found to be areally extensive in the upper catchment but there was insufficient information for the lower catchment. Deeper bore water levels have decreased significantly over the ten years and it is therefore possible that the contribution of water and chloride to streamflow has also decreased. A conceptual deeper groundwater flow system was developed to account for observed salinity variations. Estimates of the catchment groundwater flux were reviewed and a value of 25-30mm yr^{-1} with a chloride concentration of between 200 and 500 mgl^{-1} was identified.

The development of seasonal shallow groundwater was observed on a transect of shallow bores in the headwater area. Perched groundwater developed in winter in the more permeable gravels and earths above clays and yellow earth-clays. Saturation of the profile on the lower half of the transect was observed to correspond to the development of an enlarged area of saturated surface soils forming a source area for direct rainfall runoff.

The duration of shallow groundwaters corresponded with the duration of stream discharge and the quality of the shallow groundwater was similar to that of the discharge at the small weir in the headwater area. At higher water levels, the recession discharge at the headwater weir was found to be proportional to the exponential, relative, shallow bore level. This finding is consistent with the theory of spring discharge from a groundwater system and implies that discharge is proportional to the exponential variation of the ratio of transmissivity to storage constant. The gradation of the profile from coarse surface material to increasing loam and clay may account for this result.

The water level response in a shallow bore was compared with the stream discharge hydrograph at the headwater weir over a period of a few weeks and during two storm events. A second increase in bore water level was observed to occur about three days after the rainfall. This was attributed to a wave of shallow groundwater moving laterally down the hillslope which had the effect of maintaining stream discharge for a short time.

Shallow bore water levels rose quickly in response to rainfall and peaked within a few hours of the peak stream discharge. This bore water level response was used to construct a groundwater component in the stream hydrograph. The separation indicated that shallow groundwater could account for a significant proportion of storm period stream discharge.

An analysis of the storm period contribution of direct rainfall-runoff was made by separating storm period hydrographs by a geometric method and a chemical method. These gave significantly different results with the chemical method producing very little direct runoff. This was considered to be a consequence of using 'constant' concentrations for the source components and to the absence of a storage term in the model. The simple, geometric hydrograph separation probably overestimated the direct runoff component on the evidence of shallow bore water level response. However the yearly proportion of direct runoff by this method was not significant at about 3% of total water yield and much less in terms of chloride yield. Therefore for the purpose of subsequent analyses this direct runoff component was included as part of the shallow groundwater contribution.

The yearly and seasonal stream chloride and water yield responses indicated an initial leaching of chloride in all years. After a period of leaching the rate of chloride production (and therefore concentration) decreased and continued at an approximately constant rate. Increases in stream chloride concentration towards the end of the six months of flow in early summer were not important in terms of chloride load.

Two simple, conceptual, streamflow source component models were tested on the yearly water and chloride data. In the first model streamflow is a <u>linear</u> combination of shallow and deeper groundwater contributions. The second model assumed that all stream water was from the shallow groundwater and that a large part of the chloride load was generated by the deeper groundwater which leached in an <u>exponential</u> manner.

Both models gave acceptable fit to the yearly data. However the exponential model is logically superior in terms of observed catchment processes. The linear model produces unreasonable estimates of the contribution of deeper groundwater, particularly in dry years, because of the assumption of constant source concentrations. This was demonstrated for streamflow during 1983. Therefore, although the hillslope model of the sources of stream water and chloride is a reasonable conceptual representation of the streamflow, the development of the set of equations based on a linear combination of sources is not applicable.

The formulation and particularly the parameters of the exponential model developed in this study are different from those of Peck (1976). For this forested catchment most of the streamflow is generated from the shallow groundwaters which leach salts from the deeper groundwater. The model parameters were estimated from the stream water and chloride which are integrated catchment responses. In Peck's model the parameters were estimated from catchment averages of experimental, mostly soil core, information. The water flux or recharge rate (probably too large) was estimated from a salt and water budget analysis for another catchment by Peck and Hurle (1973). Peck (1983) has estimated a value for recharge which was less than 14% of the 1976 value.

This study indicates that the exponential model can reproduce the yearly responses under essentially steady-state conditions. The applicability of developments from this approach, such as those of Peck (1976) and Peck et al. (1977) should be investigated. The development of such catchment scale models, in conjunction with more detailed, hillslope process models, could be a profitable research area, particuarly for prediction of the hydrologic effects of land use change.

9.2 Conclusions

The hillslope model of sources of streamflow was found to be a useful conceptualisation of catchment scale stream water and chloride generation processes. The analysis of catchment (integrated) inputs and outputs of water and chloride and the observations of the generation of water and chloride generally support the model concepts.

The catchment water balance is dominated by evapotranspiration with about 90% of average annual rainfall. Interception was calculated to be 16% of rainfall by difference between the fully exposed and under forest canopy gauges. Chloride input under canopy was greater than that measured in the exposed gauges.

Stream water yield averaged about 10% of rainfall and more than 95% of this was estimated to be generated from subsurface (groundwater) sources. Direct rainfall-runoff was not an important process in the generation of yearly stream water and chloride yields although storm period peak discharges were generated by this mechanism.

The hydrology of Salmon catchment is therefore dominated by subsurface processes and in particular by the seasonal, shallow groundwaters. This shallow groundwater is also the medium by which salt from the deeper, permanent groundwater system is leached from the catchment. As a result there is a net export of chloride from Salmon catchment. The export of chloride has decreased, relative to input, with a longer term decline in deeper groundwater levels.

Shallow groundwaters were also observed to correspond with the development of saturated surface soils (source areas) in the headwater area and to respond sufficiently during rainfall to contribute significantly to storm period streamflow.

The sub-catchment water and chloride yields and the location of streamflow sources were found to correspond with the hydrologic zones identified on the basis of topography, soils and vegetation mapping.

Two simple models of stream chloride and water sources were investigated. An exponential leaching model was found to reproduce the observed yearly catchment water and chloride responses better than the model of linear sum of source components.

9.3 Recommendations

9.3.1 The extent of and the dynamics of the ephemeral, shallow groundwater should be monitored on a larger scale than was attempted in this study. Continuous water level recording, in smaller diameter bores, is required to define the response of the soil water to rainfall.

9.3.2 The hydraulic characteristics of the shallow soils should be measured for interpretation of the groundwater response. Interception and/or tracer measurement of shallow, horizontal groundwater flows would be useful.

9.3.3 Continuous monitoring of discharge and water quality from the large soil pipes would assist in the interpretation of the significance of these during rainfall.

9.3.4 Isotopes, in conjunction with major ion chemistry, should be investigated for separating hydrographs.

9.3.5 The processes of saltfall under forest canopy should be investigated to determine the source of the apparent additional salt input.

9.3.6 Additional deep bores should be installed in the lower half of Salmon catchment to define the deeper groundwaters.

9.3.7 The source component model of streamflow be further developed to include a salt storage and leaching algorithm.

9.3.8 A numerical, hillslope model be developed to simulate the transient response of direct runoff, shallow groundwater and deeper groundwater contributions to streamflow for Salmon catchment. The Smith and Hebbert (1983) hillslope model should be tested with data from this study.

9.3.9 Catchment leaching models, such as the exponential model used in this study, should be developed further and applied to catchments undergoing a land use change (such as Wights catchment adjacent to the study site).

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