Impact of forest disturbance on jarrah (*Eucalyptus marginata)* forest hydrology

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DECLARATION

I declare that this thesis is my own account of my research and contains as its main content work which has not previously been submitted for a degree at any tertiary education institution.

J.K. Ruprecht

Date 6th May, 2018

Abstract

Globally, forests cover 31% of the Earth's land mass and are critical areas for water supply. In Australia, forested catchments provide 77% of urban water supplies to capital cities. However, recent studies have reported worldwide examples of forest damage resulting from drought or heat related events. The jarrah (*Eucalyptus marginata*) forests of south-west Western Australia (SWWA) have experienced both sudden and unprecedented forest collapse and profound reductions in streamflows. Projected further declines with climate change, reinforce the need to understand the hydrologic impact of forest disturbance and what management responses are needed to enhance forest resilience and productive capacity.

The aim of this thesis was to understand the impact of disturbance and climate on the hydrology of the forests of SWWA, with objectives to:

1. Examine the characteristics of forest hydrology;
2. Evaluate the hydrologic response to forest disturbance and climate variability; and

3. Evaluate forest water management options in the context of forest disturbance and climate change.

This thesis thus develops an understanding of the impact of disturbance and climate on the hydrology of the forests of SWWA. Using hillslope and paired catchment studies (Chapter 3), it develops an understanding of the process of infiltration and soil water dynamics and examines the hydrologic impact of forest disturbance. The studies demonstrate the important roles of infiltration, soil water dynamics, and groundwater on the forest water balance, and identify the major factors that impact forest disturbance and forest hydrology. These studies have improved understanding of factors contributing to catchment water balance, and streamflow generation processes.

Four catchments underwent land use change and the impact on catchment hydrology was studied by comparing with a control catchment (Chapter 4). These paired catchment studies evaluated the impact of converting forest to agriculture and of timber harvesting. They explored the streamflow generation mechanisms for forested and cleared catchments, the streamflow generation and salinity export changes due to clearing for agriculture, and the hydrologic impact of intense timber harvesting for increased water production.

The impacts of deforestation, forest thinning, bauxite mining, bushfires, dieback disease, and reforestation were evaluated using several paired catchment studies across SWWA (Chapter 5). The long-term implications for management of water yield, the impact of a range of disturbances at a catchment scale, and the impact of forest disturbances on stream salinity are also examined.

The relationship between the drying climate observed in SWWA over the last 40 years and observed changes in rainfall, groundwater levels, streamflow volumes and flow duration were studied in Chapter 6. The changing relationship between rainfall and streamflow and the likely implications of recent climate change scenarios are also studied. The major forest water issues that have been identified in this thesis are the declining water values in forested areas, such as less water volumes, shorter flow periods, and declining groundwater levels.

The adaptive strategies for forest ecosystems are identified to include resistance (protect highly valued areas), resilience (improve capacity to return to pre-disturbance conditions) and response (assist transition to new condition) are discussed in Chapter 7. Drivers identified (Chapter 7) by this thesis include (a) a drying climate with direct and indirect impacts on both the forest itself and on the overall water balance, (b) responses to historical forest management including forest harvest, deforestation and reforestation, (c) long-term impacts of bauxite mining and subsequent rehabilitation, and (d) the interaction of these forest disturbances at a catchment scale.

The major findings from this study include:

- The high saturated hydraulic conductivity of the sandy gravel topsoil overlies lateritic durcirust with a much lower saturated hydraulic conductivity;
- The presence of large infilled “holes” within the lateritic duricrust;
- Saturation above the lateritic duricrust was observed confirming subsurface flow concepts;
- Presence of vertical preferential flow observed confirming soil water concepts;
- The critical importance of the groundwater discharge area in streamflow generation;
- Increase in stream salinity directly linked to groundwater levels approaching the surface;
- The time to leaching of the salt from the catchment estimated at 200 years;
- Forest disturbances such as clearing, timber harvesting and forest thinning led to increased streamflow but with significant delays related to the presence or lack of a groundwater discharge area; and
- The extensive reduction in streamflow across the south west has ranged from 36 to 52% (1975 to 2000 compared to 2001 to 2012) seen as a delayed response to rainfall reductions from 1930 to 2000.

The challenge for the future is for forest hydrology research to influence current and future forest management to improve environmental and water supply outcomes for the forests of not only SWWA, but globally. Understanding the impact of land-use change on hydrology, water quality and on water resources, and separating this from climate variability and change, is a recurring problem globally. Further understanding is thus needed of the causes of changing forest hydrology and of management options to ultimately improve forest outcomes.

Acknowledgements

I am sincerely grateful to those people that have been involved in this research and the monitoring of the research catchments which were critical to this thesis.

I am grateful to Richard Harper (principal supervisor) and Bernard Dell (supervisor) for their assistance, invaluable advice and encouragement.

Chapters 3, 4 and 5 were based on a series of papers published in a range of scientific journals. I was principal author on each of these papers and was joined by a number of co-authors. In particular I would like to thank Nick Schofield who was an important supervisor, mentor and coach during the research that led to the published papers. The individual approaches, data processing and interpretations have been my own scientific work.

The comments of anonymous reviewers and referees of papers comprising Chapters 3 to 5 are gratefully acknowledged. I would also like to thank my daughter Liz for her constructive comments on chapters in progress.

Finally I thank my wife Sandy for her patience and perseverance whilst undertaking this research.

Publications used in this thesis[1]

Ruprecht, J.K. and Schofield, N.J., 1993. Infiltration characteristics of a complex lateritic hillslope. *Hydrological Processes*, **7**: 87-97.

Ruprecht, J.K. and Schofield, N.J., 1990. Seasonal soil water dynamics in the jarrah forest, Western Australia. I: Results from a hillslope transect with coarse-textured soil profiles. *Hydrological Processes*, **4**: 241-258

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Ruprecht, J.K. and Schofield, N.J., 1991a. Effects of partial deforestation on hydrology and salinity in high salt storage landscapes. I. Extensive block clearing. *Journal of Hydrology*, **129**: 19-38.

Ruprecht, J.K. and Schofield, N.J., 1991b. Effects of partial deforestation on hydrology and salinity in high salt storage landscapes. II. Strip, soils and parkland clearing. *Journal of Hydrology*, **129**: 39-55

[1] See Table 1.2 on page 32 for the summary of my involvement in each of these papers

Ruprecht, J.K., Schofield, N.J., Crombie, D.S., Vertessy, R.A. and Stoneman, G.L., 1991. Early hydrological response to intense forest thinning in southwestern Australia. *Journal of Hydrology*, **129**: 261-277.

Acronyms and terms

AFOLU	Agriculture, Forestry and Other Land Use
ARI	Average Recurrence Interval
BHC	Bauxite Hydrology Committee
Ea	Actual evapotranspiration
Eo	Potential evapotranspiration
EC	Electrical conductivity
FMP	Forest Management Plan
GAWS	Goldfields and Agricultural Water Supply
GCM	Global Circulation Model
GL	Gigalitre
gpg	gravity potential gradient
HRZ	High rainfall zone (mean annual rainfall > 1100 mm)
IBRA	Interim Biogeographic Regionalisation for Australia
IOCI	Indian Ocean Climate Initiative
IRZ	Intermediate rainfall zone (mean annual rainfall between 900 mm and 1100 mm)

IWSS	Integrated Water Supply Scheme
Ks	Saturated hydraulic conductivity
LAI	Leaf area index
LRZ	Low rainfall zone (mean annual rainfall < 900 mm)
MAR	Mean annual rainfall
ML	Megalitre
MTF	Mediterranean – type forests, which exist in a climate characterised by warm to hot, dry summers and mild to cool, wet winters.
P	Precipitation
PET	Potential evapotranspiration + see Eo
SWWA	South West Western Australia
TDS	Total Dissolved Salts, which involves the sum of the specific major ions (Na^+, K^+, Ca_2^+, Mg_2^+, Cl^-, HCO_3^-, CO_3^{2-} and SO_4^{2-}) in stream water. Another common method of measuring stream salinity is to measure the Total Dissolved Solids (TDSolids). This involves filtering a water sample to remove sediment and then weighing the residue after evaporation. TDS and TDSolids usually differ only

slightly because of minor ions and there is unlikely to be significant differences for stream salinity in SWWA.

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Chapter 1 Background and Thesis Aims

1.1 Introduction

Globally forests cover about 31% of the Earth's land mass and are critical areas for water supplies (FAO 2010). In Australia 77% of urban water supplies are from primarily forested catchments (NWC 2014). These forested catchments provide better water quality than other land uses such as agriculture, industry or urban areas with the forests also providing biodiversity and social values (FAO 2010).

However, issues such as urbanisation, agricultural expansion, increasing demand for water, and climate change are all placing greater pressure on forest areas (Geist and Lambin 2002). To meet both water supply demand and provide environmental services, sustainable forest management and protection are critical. However, what constitutes sustainable forest management for water is not clear. Changes in rainfall patterns and increasing temperatures associated with climate change are likely to cause widespread forest decline, not only globally (Allen *et al.* 2010) but regionally such as south-west Western Australia (SWWA) (Choat *et al.* 2012). In this region, droughts are predicted to increase in duration and severity (Choat *et al.* 2012; IOCI 2012). This changing baseline and the increasing demands on forests for water means a greater understanding of forest hydrology is essential to ensure better policies and decision-making.

The forests of SWWA, in the context of this thesis, are those defined as Jarrah (*Eucalyptus marginata*) Forest in the Interim Biogeographic Regionalisation for Australia (IBRA) when referring to the northern and southern jarrah forests and Warren when referring to the karri (*E. diversicolor*) or southern forests

(Environment Australia 2000). The forest types, as defined by the Forest Products Commission of Western Australia, are outlined in Fig. 1.1 and the extent of the south-west forests are shown in Fig. 1.2

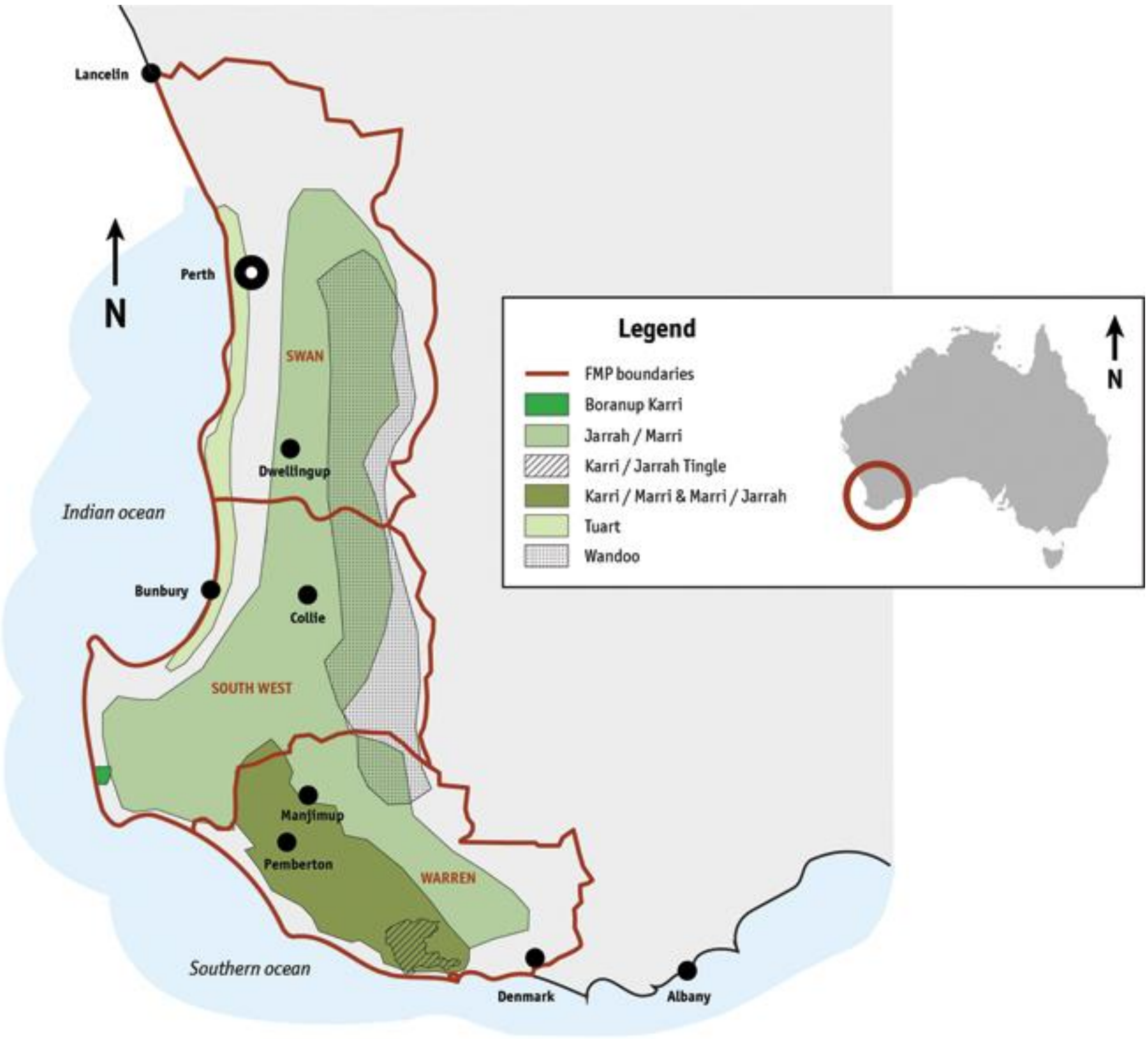


Figure 1.1 ***Forest types of south-west Western Australia (Forest Products Commission, 2014)***

Figure 1.2 Google Earth image showing south-west forest (forest is the darker green in the image)

Mediterranean-type forests (MTFs) in the Southern Hemisphere, such as in SWWA are very sensitive to a changing climate and the impacts on water supply and forest hydrology is considerable (Silberstein *et al.* 2012). In Mediterranean forests with an effective rainfall (P/E_o) lower than the potential evapotranspiration (E_o), the actual evapotranspiration (E_a) is only a fraction of potential evapotranspiration (Gracia *et al.* 2011). This means that MTFs grow under water limited conditions and is in contrast to temperate forests where actual evapotranspiration is close to the potential evapotranspiration (Gracia *et al.* 2011).

Given the water limiting conditions of the Mediterranean-type forests of SWWA, there is a more complex relationship with the impact of changing leaf area index (LAI) through forest management on evapotranspiration and streamflow than has been described in non-water limited forests (Schofield *et al.* 1989b).

The hydrology of the jarrah forests of SWWA have experienced major changes from a drying climate, forest disturbance, and forest management. The more recent drying climate since 2000, has resulted in an acceleration in the decline in streamflow, groundwater levels, and flow duration (Petrone *et al.* 2010). These observations of forest disturbance in combination with a changing climate provides an impetus to understand the complex interactions, and developing this understanding may assist global response strategies to climate change in other regions.

In a global sense, the hydrology of the jarrah forest is unique in producing little streamflow from moderate rainfall (Schofield *et al.* 1989b). This is attributed to the large soil water storage capacity of the forest soils which jarrah is well adapted to exploit during summer. Soil profiles are deeply weathered and formed over predominantly igneous substrates (Anand and Paine 2002). Consequently, the soils are both relatively infertile and often 20-50 m deep along interfluves (McArthur, 1991). The sequence of high winter rainfall accompanied by large soil water storage, followed by high summer evapotranspiration, explains a major part of the water balance (Gentilli 1989).

Superimposed on this water cycle is a distinctive climatic gradient across the forest (Gentilli 1989). Moving inland from the Darling Escarpment in the north and the coast in the south, mean annual rainfall decreases markedly from a maximum of 1278 mm to 635 mm at the eastern edge (Charles, et al. 2010). Along this rainfall gradient mean annual evapotranspiration increases and streamflow decreases. Annual pan evaporation increases in a general fashion from around 1200 mm on the south coast, to 1800 mm in the northern jarrah forest (Gentilli 1989; Luke *et al.* 1988). Thus, in the lower rainfall areas of the forest virtually all the rainfall is evaporated, resulting in the accumulation of salt in the soil (Hingston and Gailitis 1976). Salt accumulation does not occur to the same extent in the higher rainfall areas. This salt has its origin from the ocean and, as will be seen, this accumulation of salt in the regolith has profound influences on ground and surface water resources in the region.

The understanding of forest and water interactions is critical to SWWA as the forests have been utilised for water supply for Perth (capital city of Western Australia) and for irrigation, and have also been utilised for timber production and surface mining (for bauxite). Unlike the areas further inland where forests have been removed for agriculture, much of the jarrah forest has been retained for multiple use, with these being originally to protect water supply catchments, and also for timber production (Batini *et al.* 1980). The forests of SWWA have high conservation value, comprising one of the top 25 biodiversity hotspots in the world (Myers *et al.* 2000) and, being close to Perth, the forest is also a major recreation area.

The jarrah forest of SWWA has provided critical water supplies for both drinking water and for irrigated agriculture since the early 1890s. Up to the 1970s the water supply for Perth was at least 90% sourced from forested catchments, but this has declined to 18% in 2014. The modelled and observed streamflow at Water

Corporation dams (Fig. 1.3) shows a mean annual streamflow of less than 100 GL since 2001, compared to an average of 335GL from 1911-1974.

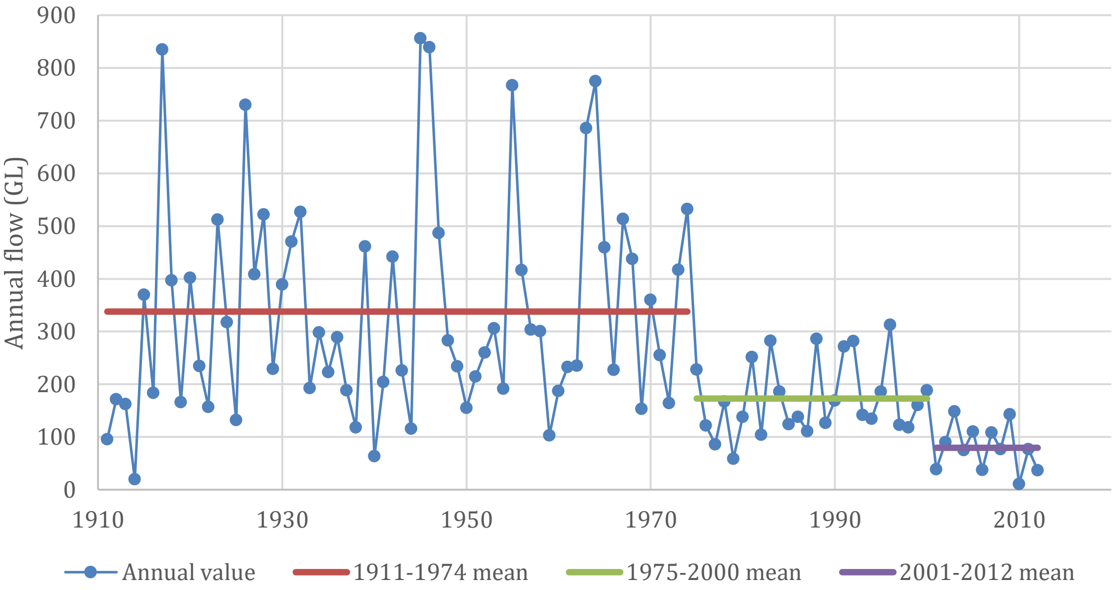


Figure 1.3 Annual streamflow at Water Corporation dam sites from 1911 to 2012 (modelled and observed).

Since the European settlement of Western Australia in 1829, land use practices in the jarrah forest have resulted in changes in water yield and serious water quality problems. The most significant forest disturbances which have affected water yield and quality are clearing for agriculture, forest harvesting and regeneration, and bauxite mining. Forest thinning for increased water yield also has the potential to affect water yield and water quality. The main impact on water quality is the release of salts from the regolith and consequent impairment of water supplies.

1.2 Broad biophysical context

1.2.1 Physiography

The main physiographic feature of south-western Australia is the Great Plateau (Mulcahy and Bettenay 1972). Near Perth, the Plateau is separated from the sedimentary Perth Basin (and its surface expression the Swan Coastal Plain) by the Darling Fault, an almost linear north-south feature which near Perth is expressed as the Darling Scarp up to 300 m high. The adjacent margin of the Plateau forms part of a stable Archaean shield composed largely of granite which has invaded linear belts of metamorphosed sedimentary and volcanic rocks, some isolated occurrences of which remain (Bettenay and Mulcahy 1972). Thin sheet-like dolerite intrusions occur abundantly in the basement rock, and are known to affect groundwater movement (Engel *et al.* 1987). Close to the Darling Scarp, deep V-shaped valleys occur, with shallow soils and frequent rock outcrops. Further inland, these valleys transform to broad U-shaped valleys and then to very broad valleys with long (1-2 km) sideslopes of low inclination (less than 10%) (Bettenay and Mulcahy 1972).

Deep weathering of the landscape

The inland areas are characterised by soil profiles that have been deeply weathered *in situ* (typically 20-50 m, but up to 70 m deep) to form what are locally termed laterite profiles (see Fig. 1.4). These profiles form over all types of geology (Churchward and Gunn 1983). Their occurrence and their consequent ability to store water in a Mediterranean climate are the reason that the jarrah forest occurs in this region (Churchward and McArthur 1978; Gilkes *et al.* 1973;

McArthur 1991). The profiles have a characteristic suite of horizons – with a sandy horizon near the surface that may be 0.3-0.5 m thick, a layer of ferricrete or bauxite gravel of around a metre in depth, a mottled zone that has an accumulation of iron and aluminium, and then a pallid zone that may be many metres deep and where most of the iron and aluminium has been removed (Anand and Paine 2002; Gilkes *et al.* 1973). Beneath the pallid zone is saprolite, which is partly weathered rock and then the basement rock.

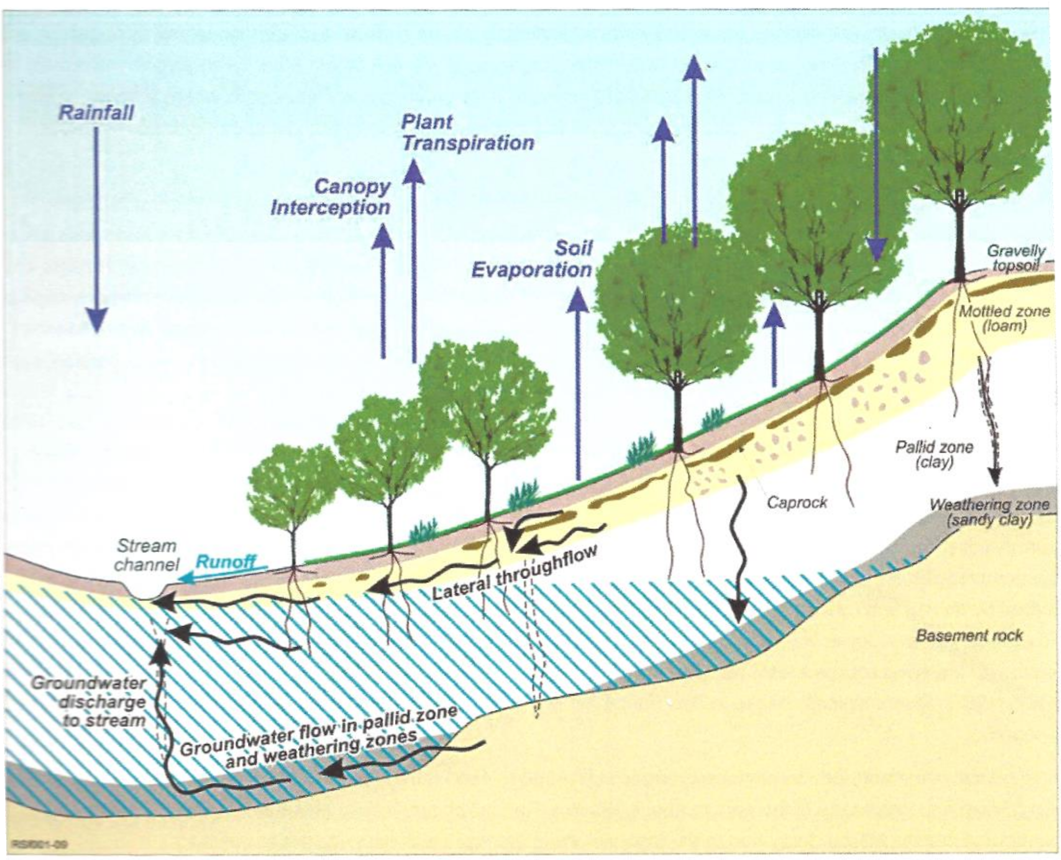


Figure 1.4 Diagram of runoff generation and soil profile in the jarrah forests of south-west Western Australia (CSIRO 2009)

In inland areas the drainage is ephemeral and sluggish, whereas near the Darling Scarp streams are often perennial and efficient. The greater incision near the Scarp is a result of the westerly drainage through this tectonically uplifted area and as a consequence there has been erosion or stripping of the deeply weathered profile (Churchward and Gunn 1983) and exposure of the underlying horizons of the laterite profile (Mulcahy 1973). Where there is deep landscape incision in valleys the bedrock may be exposed, with shallow soils and subsequent poor soil water storage (Harper *et al.* 2009b). In contrast, in interfluves and further inland there is often little erosion in the landscape and weathering and erosional products are retained in the landscape (Mulcahy 1973), including within the valleys. Soil water storage is consequently greater and groundwater systems may form within the valleys (Sharma *et al.* 1987b). Overall, the landscape is relatively flat with gentle to moderate gradients, except for the major river valleys such as those associated with the Helena and Murray rivers (Mulcahy and Bettenay 1972; Mulcahy *et al.* 1972).

Salt accumulation in the deep weathering profile

A large amount of salt occurs in the deep weathering profiles, with amounts of between 0.4 and 64.7 kg m^{-2} TDS (Johnston *et al.* 1980). For this region, the accepted theory for salt accumulation is that salt is transported from the ocean though the atmosphere to be deposited in rain, dry powder or dew. The process of entrainment of material from the sea surface, its injection into the atmosphere, and its transport and deposition to the land surface has been investigated in considerable detail and reviewed by Cryer (1986).

The geographic variation in salt contents reported by Hingston and Gailitis (1976) for SWWA found both decreasing chloride (Cl^-) concentration in rainfall and decreasing precipitation with distance from the coast. It is thus assumed that salt has accumulated in the landscape over long periods of time with accessions of salt at the soil surface conveyed into the soil via infiltrating water. Some of the salt has been exported via throughflow and groundwater discharge. However, when salt input exceeds salt output there has been salt accumulation in the soil (Johnston 1983; Johnston 1987a). The distribution of salt in the soil profile is influenced by a number of factors such as magnitude of salt input, average annual rainfall, depth of soil, presence of saturated zones, and soil properties (Hingston and Gailitis 1976; Peck *et al.* 1981).

Soil fertility patterns in the deep weathering profile

A major feature of the jarrah forest soils is that the soils formed on the deep weathering profiles are infertile (Hopper 2009) with primary minerals in the original rocks converted to clay minerals (predominantly kaolin) and iron and aluminium oxides (Gilkes *et al.* 1973). The subsequent pattern of soil fertility depends on the degree of landscape stripping (Churchward and Gunn 1983) with infertile soils in areas where the deep weathering profile is retained and fertile soils where it has been removed and the underlying unweathered bedrock exposed (Mulcahy 1973).

1.2.2 Climate

The climate of the jarrah forest is conventionally described as Mediterranean type with about 80% of rainfall falling between April and October and a seasonal summer drought that may last 4 to 7 months (Gentilli 1972). Further inland from the Darling Scarp in the north and from the coast in the south, the mean annual rainfall decreases markedly. Along this gradient mean annual evaporation increases and streamflow decreases. In the lower rainfall areas of the forest virtually all the rainfall is evaporated, resulting in the accumulation of salt in the soil (Schofield *et al.* 1989b). Salt accumulation does not occur to the same extent in the higher rainfall areas. Hydrological characteristics, such as average water yields and annual peak flows, and levels of soil salt storage, all vary significantly with annual rainfall.

1.2.3 The jarrah forest and current land use

Extent

The jarrah forest forms part of the broad-leaved evergreen forests of Australia (Ovington and Pryor 1983). On the basis of canopy density and height, the jarrah forest is classified as an open forest in the north and a tall forest in the south. As rainfall decreases towards the north and east the trees decrease in stature, thus forming a woodland or low forest. The trees and shrubs are considered to be sclerophyllous and the overall forest is comprised of several species dominated by jarrah (*Eucalyptus marginata*) itself, marri (*Corymbia calophylla*) and other tree species including yarri (*E. patens*) and bullich (*E. megacarpa*). There is an understorey of small trees (4 to 7 m), mainly of *Allocasuarina fraseriana*, *Banksia grandis*, *Persoonia longifolia* and *P. elliptica*, and a groundcover of woody shrubs including grass trees (*Xanthorrhoea preissii*) and zamia palms (*Macrozamia reidlei*) and an array of smaller species (Havel 1975).

Jarrah (and other associated tree species) generally occurs throughout the region (Fig. 1.1), however in some southern areas jarrah is replaced by pure stands of karri.

History of land use

Agricultural development has led to the deforestation or clearing of extensive areas of jarrah forest, particularly along the eastern edges, with a reduction in the area of jarrah forest from 53 000 km^2 (Lane-Poole 1921) to the current situation where 30,000 km^2 still has natural vegetation (Conservation Commission of Western Australia 2013). Small farms and orchards have been established on many of the more fertile valley soils on the western edge of the jarrah forest.

Northern jarrah forest silviculture

The northern jarrah forest has been subject to logging activity for about 100 years, and during this time it has nearly all been cut over at least once (Stoneman *et al.* 1989). Logging practices and intensity have varied greatly according to location, forest quality and the silvicultural practice of the day. Current forestry practice involves selective logging rather than clearfelling, with 135 to 185 km^2 harvested annually (Conservation Commission of Western Australia 2013).

Before 1918, timber cutting was not controlled. High-quality areas were sought and subjected to heavy selective cutting and clearfelling (Koch and Hobbs 2007; Stoneman *et al.* 1989). This gave rise to dense regrowth stands. In the period from 1920 to 1940, a group selection silvicultural system was used. This involved virtual clearfelling in a patchwork pattern, the objective being to create good regeneration conditions on the cut-over patches and to retain uncut patches with better growth potential (Stoneman *et al.* 1989).

Since 1940, the group selection cut areas and all other uncut or lightly cut forest have been subjected to single tree selection cutting. This has given rise to dense mixed-age stands (Stoneman *et al.* 1989). As a result of the varying logging practices, the forest now consists of a mosaic of various ages and structures of mixed-age and even-aged regeneration.

Southern forest silviculture

Logging in the southern forests of Western Australia commenced before 1900 and began with clearfelling of karri and intense selection cutting of jarrah. By the 1940s there was a change to selection cutting of karri. This converted back to clearfelling of karri and selection cutting of jarrah in the 1960s (Stoneman *et al.* 1989). In recent years the silviculture practice for karri and jarrah is more complex with retention of shelter trees and marri retention requirements (Burrows et al., 2011).

Bauxite mining

Bauxite mining is a major land use within the northern jarrah forest of SWWA. The upper layers of the deeply weathered profiles have economic accumulations of aluminium and this is removed as a strip mining operation. Approximately 10 $km^2 yr^{-1}$ of jarrah forest is annually cleared for bauxite, gold and coal mining (Conservation Commission of Western Australia 2013), with an estimate of over 180 km^2 cleared to date. The principal bauxite area covers 50 to 60% of the northern jarrah forest, including most of the developed metropolitan water supply catchments for Perth (Mauger *et al.* 1998). The higher grades of bauxite occur as discrete pods on hillslopes, ranging from 2 to 80 ha in size. Mining typically results in the clearing of all vegetation, topsoil removal, extraction of the top 4 to 6 m, and rehabilitation. Rehabilitation consists of ripping, topsoil replacement and establishment of understorey and eucalyptus overstorey species (Koch and Hobbs 2007; Nichols *et al.* 1985). The proportion of land cleared in a given catchment ranges from 20 to 60%, with an average of about 35% in the higher rainfall areas (Mauger *et al.* 1998).

When bauxite mining was initially approved in 1963, the impact of mining on water resources was considered to be minor (Steering Committee for Research on Land Use and Water Supply 1985). However, the escalation of the scale of mining gave rise to increasing concern, not only for the water resources, but also for the jarrah forest environment (Steering Committee for Research on Land Use and Water Supply 1985). In response to these concerns, a hydrology research programme was instigated. The initial concerns were, with respect to water resources, related to water quality, primarily reservoir turbidity and salinisation of the water supply. However, more recently the concern has been related to declining water yield from catchments rehabilitated after bauxite mining (Bari and Ruprecht 2003).

Dieback

Jarrah dieback is a plant disease caused by the soil-borne fungus *Phytophthora cinnamomi* (Podger 1968) that results in the death of not only jarrah but also a range of understorey species (Davison and Shearer 1989). The fungus is an introduced species which has been spread through the forest largely in soil carried on earthmoving machinery and transport vehicles. Thus, any land use practice that involves movement of contaminated soil has the potential to have an impact on the forest vegetation. The most severe dieback is located on the western margins of the northern jarrah forest, with severity decreasing progressively inland (Podger 1968). The most susceptible sites within this area are usually associated with water courses or poorly drained hill sites, but dieback of the vegetation has been found in nearly all soil types, topography and forest stands (Dell and Malajczuk 1989). Of the State Forest area, an estimated 20% is infected with *P. cinnamomi*.

Reforestation

In response to concerns over both stream and land salinisation, targeted reforestation of cleared areas was initiated in water resource recovery catchments (Schofield *et al.* 1988). The first areas targeted were water resource recovery catchments such as the Collie and Denmark river catchments with government programs. However, over time these government programs were replaced with private sector investment. Over the last 25 years there has been 300,000 ha of reforestation, primarily bluegums (*E. globulus*) established on cleared land (Harper *et al.* 2001; Harper *et al.* 2009a).

1.3 Jarrah forest hydrology

Studies into forest hydrology in SWWA have primarily focused on the drier jarrah forest and the moister karri forest. The forested areas of SWWA have been classified hydrologically into three categories (Loh *et al.* 1984):

1) High Rainfall Zone (mean annual rainfall >1100 mm);
2) Intermediate Rainfall Zone (mean annual rainfall between 900 mm and 1100 mm); and
3) Low Rainfall Zone (mean annual rainfall <900 mm).

The criteria of mean annual rainfall is based on isohyets developed in the late 1970s (Hayes and Garnaut 1979). Since then there has been a westward shift in many of the isohyets, but the isohyets from Hayes and Garnaut (1979) remain as the criteria for the rainfall zones as they correlate with salinity risk data such as soil salt storage and depth to groundwater.

Loh *et al.* (1984) described the average annual streamflow from the High Rainfall Zone ranging from 10 to 25% of mean annual rainfall, while streamflow volumes reduce rapidly as rainfall decreases eastwards across the intermediate rainfall zone. In the Low Rainfall Zone mean annual runoff is typically less than 5% of mean annual rainfall. Annual runoff in the Low Rainfall Zone is considered highly variable with over 30% of years with zero runoff (Loh *et al.* 1984). More recently no runoff has been recorded, in extreme rainfall years, even in some High Rainfall Zone catchments, given the decadal below average rainfall (Petrone *et al.* 2010).

With respect to stream salinity, Loh *et al.* (1984) described the High Rainfall Zone as having low stream salinity given groundwater discharged to the surface. In the Intermediate Rainfall Zone stream salinity was described as highly variable as groundwater may or may not discharge to the surface (Loh *et al.* 1984). Stream salinity was considered typically very low in the Low Rainfall Zone as groundwater levels are well below the ground surface.

1.3.1 Rainfall and evapotranspiration

Mean annual rainfall ranges from more than 1400 mm on the western boundary to 600 mm on the eastern boundary of the jarrah forest (isohyets across south-west shown in Fig. 2.4). Rainfall intensity is typically low and rarely exceeds the infiltration capacity of the undisturbed forested top soil (Sharma *et al.* 1987a).

There is moderate seasonal variation in annual rainfall, with typical coefficients of variation ranging from 10 to 20% of mean annual rainfall. Rainfall intensities are typically low to moderate. Williamson *et al.* (1987) reported 1 and 100 year recurrence interval intensities of 56 and 115 mm hr^{-1} (based on 6 min rainfall duration). At the sites monitored, 50% of total rainfall occurred at intensities less than 5.5 mm h^{-1} (Williamson *et al.* 1987).

The components of the evaporation water cycle are that some rainfall is intercepted by the canopy of the tree. Rainfall that is not intercepted will fall as throughfall or stemflow to the forest floor. From there the water may evaporate from the forest floor, or infiltrate and some of this water will then be transpired by the forest vegetation.

Measurements of evapotranspiration (from soil, interception and transpiration) in the jarrah forest have included water balance (simple and complex) (Schofield 1988; Williamson *et al.* 1987), ventilated chamber measurements (Greenwood *et al.* 1985), and heat pulse and ratio methods (Burgess *et al.* 2001). Simple water balance approaches have estimated evaporation, calculated as rainfall minus streamflow for eight relatively undisturbed research catchments, to typically range from 90 to 100% of annual rainfall. Evaporation, through this calculation was found to approximate rainfall up to about 1050 mm. A similar calculation for 29 catchments across the jarrah forest obtained a mean value of 90% of annual rainfall, and a range of 68 to 100% (Ian Loh personal communication, 1983).

Transpiration is a significant component of the water balance of the jarrah forest. The mature jarrah forest extracts water for transpiration from a considerable depth of the soil profile, and sometimes from the deeper permanent groundwater (Carbon *et al.* 1980). There have been reports of roots under jarrah trees to a depth of 43 m (Dell *et al.* 1983). Marshall and Chester (1992) measured annual tree water use of jarrah stands at the Del Park and Hansen catchments using the heat pulse method and found average annual tree water uptake was 485 mm, approximately 34% of annual rainfall. Macfarlane *et al.* (2010) using the Heat Ratio Method (an improved heat pulse method) measured annual water use of overstorey jarrah forest of 230 mm yr^{-1} in an old growth stand and 500 mm yr^{-1} in a regrowth stand in a site with annual rainfall of 1135 and 1235 mm in the two years of the study.

Greenwood *et al.* (1985) used the ventilated chamber technique (Greenwood and Beresford 1980) to measure evapotranspiration from the middlestorey and ground layer at Del Park catchment, and found the middlestorey trees evaporated 16% and ground layer 37% of rainfall respectively, to give a total of 51% of rainfall.

Annual average interception by mature jarrah forest has been calculated to be about 15% of annual rainfall (Stokes 1985; Williamson *et al.* 1987). A review of the interception data collected at the Del Park catchment showed that 13% of annual rainfall is lost though interception alone (Croton and Norton 2001). Schofield *et al.* (1989b) reported throughfall in a dense regrowth stand to be 79% of rainfall.

1.3.2 Streamflow generation

The types of streamflow generation commonly referred to in the literature include overland flow (Horton 1933), the partial area contribution concept (Betson 1964), the variable source area subsurface flow concept (Hewlett and Hibbert 1967) and the variable source area-overland flow concept (Dunne and Black 1970). More recently, the terms infiltration-excess overland flow (sometimes referred to as Horton overland flow) and saturation-excess overland flow (related to the variable source area concept) have been more widely used (Beven 2000).
The jarrah forest soils generally comprise highly permeable sandy to loamy surface horizons overlying clays of low permeability at 30 to 100 cm depth (McArthur 1991). During winter, rain infiltrates the soil, perches on the clay horizon and flows downslope to discharge to streams as shallow throughflow. Infiltration capacities of jarrah forest soils are high, and are rarely exceeded by rainfall intensities.

Average surface soil saturated hydraulic conductivities of approximately 10 m day^{-1} were measured with a ring infiltrometer and well permeameter at Del Park catchment (Ruprecht and Schofield 1993). In a study at Wights catchment, Sharma *et al.* (1987a) measured surface saturated hydraulic conductivities averaging 20.8 m day^{-1} (range 5.7 to 39.5 m day^{-1}). These values of surface saturated hydraulic conductivities are substantially greater than the observed rainfall intensities. Given these values of soil permeability, infiltration-excess overland flow is considered to be highly limited in extent, with saturation-excess overland flow considered a more common mechanism of streamflow generation, particularly for quick response streamflow ((Ruprecht and Schofield 1989a).

Shallow throughflow in the upper gravelly sand horizon is considered to be the major source of streamflow in the jarrah forest (see Table 1.1). Stokes and Loh (1982) calculated that about 90% of streamflow for Salmon catchment originated from throughflow. A perched water table was observed to form during winter in the sandy gravel or loamy surface soil above a relatively impermeable clay horizon (Stokes 1985).

Overland flow and groundwater flow are considered to make minor (4-6 %) contributions to annual flow volumes (Loh *et al.* 1984; Turner *et al.* 1987b; Williamson *et al.* 1987). Although annual flow is dominated by throughflow, overland flow and groundwater flow are considered influential in the delivery of water and salt to streams (Williamson *et al.* 1987).

Saturation excess overland flow is also not considered highly significant with respect to water yield in the jarrah forest, and is usually restricted to low landscape positions. According to Stokes and Loh (1982), these variable source areas contribute only about 5% of the annual flows, but dominate the generation of peak flood flows.

Table 1.1 Comparison of water yield components (%) for Wights and Salmon catchments (after Williamson *et al*., 1987)

Component	Wights (cleared)	Salmon (forested)
Direct runoff	16	4
Throughflow	77	90
Groundwater discharge	7	6

Groundwater flow is not considered to contribute significantly to water yield. In areas with high rainfall, extensive permanent groundwater systems are typically found in the freshly weathered (saprolitic) parent material above bedrock and in the lower section of the pallid clay horizon. Low in the landscape, bedrock frequently outcrops and soils are often shallow (Bettenay *et al.* 1980). This characteristic forces groundwater to discharge at the surface within first- or second-order stream catchments. Streamflow is rarely perennial, owing to evapotranspiration by vegetation flanking the streamlines usually being larger than groundwater discharge (Stokes 1985). However, where local geological features such as dolerite dykes occur, groundwater may converge to prolong baseflows in streams through summer.

In areas with low rainfall, groundwater does not normally discharge to streams, and often lies 15 m or more below the stream invert at the catchment outlet (Schofield *et al.* 1989b). The absence of groundwater discharge results in streams ceasing to flow at least 6 weeks before similar streams which have a groundwater discharge area, such as exist in higher rainfall areas. In areas of intermediate rainfall, conditions are transitional between those of the high and low rainfall areas, and depending on local conditions, groundwater may or may not discharge to streams. As will be seen in Section 4.2, these patterns have implications for the release of salts from the regolith and discharge into streams.

On an average annual basis, only 10 to 25% of rainfall in the high rainfall areas becomes streamflow and, in the low rainfall areas, this value can be as low as 1% (Schofield *et al.* 1989b). Only a very small portion of rainfall recharges groundwater with most lost by evapotranspiration. Nonetheless, a significant proportion of winter rainfall can be stored in the soil and transpired or evaporated during spring and summer. The capacity of the deep weathering profiles to store water explains the occurrence of a forest in a Mediterranean climate.

The annual rainfall to annual runoff relationship for a high rainfall forested catchment varies from about 3% of rainfall to about 15 to 20% of annual rainfall on a year to year basis (Fig. 1.5). In contrast, a low rainfall catchment has zero flow in 40% of years, ranging up to 7% runoff in wet years (Williamson *et al.* 1987). This is consistent with the mean annual runoff, for the period 1974 to 2000, for eight forested research catchments, from across the jarrah forest, which vary from 3 to 18% of mean annual rainfall (Fig. 1.6) (Bari and Ruprecht 2003), and also consistent with mean annual runoff from six larger gauged forested catchments (Fig. 1.6).

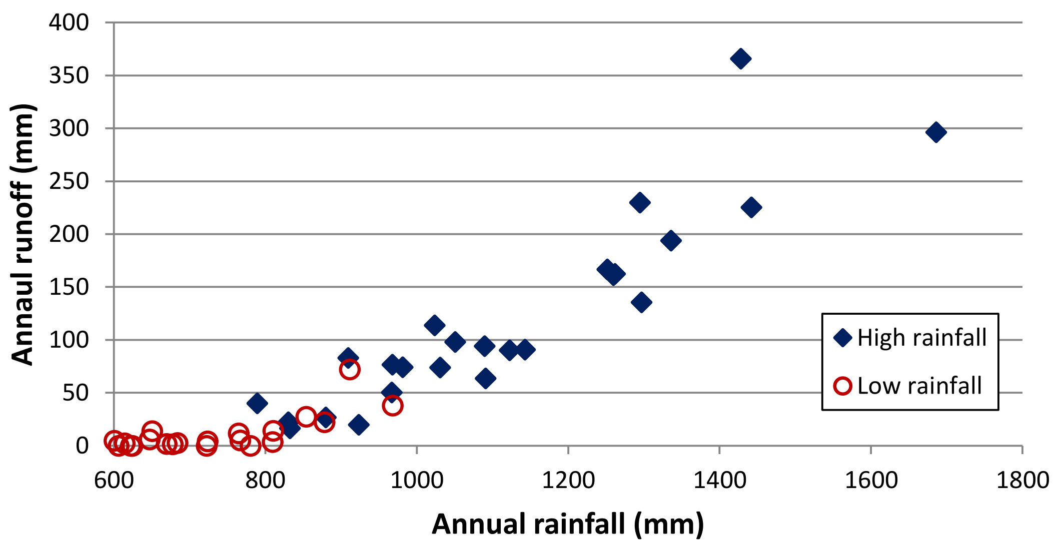


Figure 1.5 Annual runoff compared to annual rainfall for a high and a low rainfall forested research catchment (Salmon and Ernies, respectively) for the period 1974 to 1998 (Department of Water – Water Information Reporting)

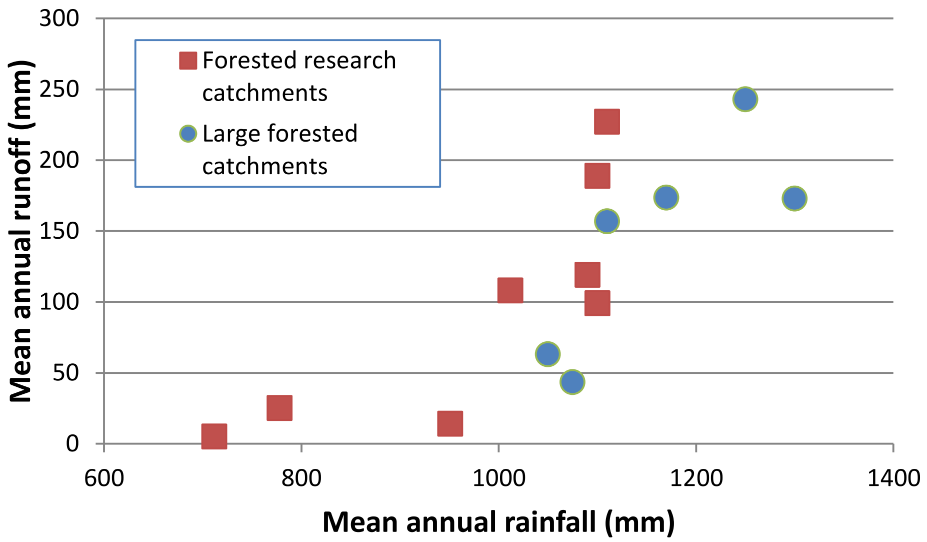


Figure 1.6 Mean annual runoff compared to mean annual rainfall, for the period 1974 to 1998, for forested catchments. The catchments are small and large sizes. (Department of Water – Water Information Reporting)

1.3.3 Soil water storage and movement

Unsaturated soil water dynamics play important roles in the hydrology of the jarrah forest. First, the deeply weathered lateritic profiles (up to 70 m in depth) provide a large soil water store (Sharma *et al.* 1987b) which leads to low water yields (Schofield *et al.* 1989b) and sluggish streamflow hydrographs (Loh 1974) compared to areas with shallow regolith. Second, the movement of water through the unsaturated zone to groundwater strongly affects groundwater levels.

Preferred pathways have been observed as pathways of water movement recharging groundwater aquifers in high rainfall catchments of the jarrah forest (Johnston 1983, Johnston 1987b). A critical condition for the process of vertical preferred pathways is the establishment of a perched aquifer in the upper soil horizon. Perched aquifers have been observed in the jarrah forest (Schofield *et al.* 1989b).

1.3.4 Groundwater and salt discharge

Groundwater in the jarrah forest is described by two systems — the perched, or unconfined, aquifer on the surface of the clay subsoil and the deeper (or permanent) aquifer which overlies the bedrock (Johnston 1983; Peck *et al.* 1981). The perched aquifer is ephemeral and, where it does not intersect the permanent groundwater, is fresh. Where saline permanent groundwater has contributed to the perched aquifer, shallow throughflow may carry substantial amounts of salt to the stream.

The permanent groundwater system in the high rainfall zone usually discharges to streams and keeps these areas leached of salt. As such, the high rainfall areas are generally of low salinity hazard. In the low rainfall areas, the deep groundwater tables in the forest are far below the stream bed and do not contribute to streamflow. As a consequence, salt has accumulated in the unsaturated zone and these areas pose a high potential salinity hazard (Stokes and Loh 1982) with changes in land use and consequent changes in landscape hydrology.

The deep unsaturated zone means that only persistent forest changes which increase recharge over long time periods (tens of years) are likely to cause long-term stream salinity problems. These conditions have occurred in the case of agricultural development (Hatton *et al.* 2003). Partial or transient land use changes in the intermediate rainfall areas (annual rainfall between 900 and 1100 mm) was considered to pose a major problem because the permanent groundwater is moderately saline and the watertable was not far below the streams (Mauger *et al.* 1998). However, in recent years, given the drying climate the groundwater levels in many of these intermediate rainfall areas has declined to be well below the valley floor in areas where deep weathering profiles occur in these locations (Croton *et al.* 2013).

Groundwater recharge in the jarrah forest generally occurs by two routes (Johnston 1987b). The first route is by matrix flow, and the second is by preferred pathways, which are generally old root channels, inserting water at depth into the soil matrix. Based on groundwater salinity data, groundwater recharge was estimated at 9% of annual rainfall at the Del Park catchment (Ruprecht *et al.* 1990). On the same basis, Johnston (1987a) estimated groundwater recharge at Salmon catchment (annual rainfall 1200 mm) at 2.2% of annual rainfall.

1.3.5 Stream salinity characteristics

The first reports of stream salinisation in Western Australia were produced around 1900. By the 1920s it was recognized that ringbarking trees for water supply (Bleazby 1917; Reynoldson 1909) and clearing native vegetation for agriculture (Wood 1924) resulted in rapidly increasing stream salinity in all areas of SWWA except those with high rainfall (>1000 mm yr^{-1}). Since that time the salinities of the major rivers have deteriorated dramatically (Mayer *et al.* 2005; Schofield and Ruprecht 1989). Research into stream salinisation commenced in earnest in the late 1960s.

Many theories have been put forward to explain the increasing salinity observed after clearing. The link between rising groundwater and increasing salinity was documented by Burvill (1947) and Bettenay and Mulcahy (1972). But there had not been definitive hydrological demonstration to confirm this link. This is particularly important in SWWA where aquifers have often developed in deeply weathered profiles that may be up to 70 m deep (McArthur 1991). In contrast to northern European and North American situations, where much of the literature has been developed, with sedimentary systems or shallow soils overlying bedrock (Kirkby 1978).

Stream salinity is governed by the relative volumes and salinities of overland flow, shallow throughflow and groundwater flow. As described above, permanent groundwater is the major source of salts, and throughflow and overland flow tend to dilute this source.

Annual mean stream salinities of forested catchments range from 80 to 400 mg L^{-1} Total Dissolved Salts (TDS) (Mayer *et al.* 2005). The lowest stream salinities (~ 100 mg L^{-1} TDS) occur in the low rainfall areas where groundwater does not discharge to streams (Schofield and Ruprecht 1989). In the high rainfall areas, where discharging groundwater salinities are low, stream salinities are again low (~ 150 mg L^{-1} TDS). Streams in the Intermediate Rainfall Zone have the highest salinities in all forest areas. Annual average values are commonly 250 mg L^{-1} TDS, but can approach 400 mg L^{-1} TDS. These high salinities result from a limited discharge of groundwater of moderate salinity, combined with throughflow of low salinity (Loh *et al.* 1984).

Stream salinity can also show strong seasonal and annual variations. Seasonally, the variations are usually associated with the proportion of throughflow, which is high in mid-winter and low in spring and autumn. Annual variability may be attributed to annual variations in throughflow due to rainfall variation and to longer-term changes in groundwater level resulting in greater or smaller contributions of groundwater to streamflow. This is because the source of the saltload is from groundwater which can vary with groundwater levels in the catchment and be diluted by streamflow from throughflow or overland flow (Turner *et al.* 1987b).

Forested catchments are known to export little salt and often accumulate salt in the soil profile (Mayer *et al.* 2005). In a forested catchment in a high rainfall area (>1100 mm of mean annual rainfall) the salt output/input ratio would be expected to be close to one since the catchment is relatively undisturbed and the high rainfall enables the throughflow of salt. In low rainfall areas, where groundwater is not discharging to the stream, the salt output/input ratio is likely to be much lower than one. For example, in very small forested catchments with mean annual rainfall less than 700 mm, less than 25% or even as low as 1% of salt accession in rainfall is exported from the catchment in streamflow (Williamson *et al.* 1987).

1.3.6 Stream sediment concentrations

Stream sediment concentrations in SWWA forests are generally below the level of detection (5 mg L^{-1} suspended solids less than 63 microns in size) (Western Australian Steering Committee for Research on Land Use and Water Supply 1987). The very permeable soils and associated lack of surface runoff, the generally low relief, the high amounts of litter and understorey vegetation all contribute to low sediment loads. Organic material from the natural decay of vegetation and disturbance due to logs falling naturally into stream channels have been the major sources of suspended material measured in undisturbed catchments. However, roads constructed for moving machinery and transport vehicles can generate increased sediment concentrations and loads, particularly at a local scale (Western Australian Steering Committee for Research on Land Use and Water Supply 1987).

1.4 Water context in the forests of SWWA

The initial focus in the forests of SWWA was utilization, initially for timber and then downstream water for domestic and irrigation purposes. However, the resource management issues, such as salinity, mining and dieback disease in the 1970s and 1980s meant a changing focus to understanding forest hydrology and management responses. However, there were earlier precedents to the resource management issues given the historical experience with forest removal to increase stream yield (Reynoldson 1909; Wood 1924).

The first major dam in the region was Mundaring Weir which was completed in 1902 to supply water to the goldfields of Western Australia. A sequence of below average rainfall years occurred during and following the completion of the weir and trees on a forested area of the catchment were ring-barked to improve flow. The Chief Engineer of the Goldfields Water Supply Administration (NC Reynoldson) in 1909 realised that the salinity of the reservoir had increased markedly. He identified the ring-barking as a major cause and recommended all alienated land be resumed and that all cleared land be re-forested (Power 1963).

The construction of Mundaring Weir was followed by further dam development along the Darling Scarp capturing downstream flows from predominantly forested catchments on short, westerly flowing streams. Churchman Brook, Canning, Serpentine, South Dandalup, and Wungong dams were constructed from 1924 to 1982 for water supply for Perth. Further south, irrigation dams (Harvey, Drakesbrook, Wellington, Stirling, Logue Brook and Waroona) were constructed along the Darling Scarp (Kelsall 2005; Morgan 2015).

In the 1950s and 1960s the Public Works Department of Western Australia (PWD) recognized a need to cease land release in three large river catchments that were under threat from salinity. In the mid-1960s the PWD proposed clearing control measures in the Wellington Dam catchment given rising stream salinity. These proposals were rejected by the Western Australian Parliament due to scientific uncertainties. In response, further research and monitoring were established in the Wellington Dam catchment in 1969 (Peck and Williamson 1987a) to investigate the impact of clearing on runoff and stream salinity. This was followed by the amendment of the *Country Areas Water Supply Act 1947* (*Country Areas Water Supply Act Amendment Act* 1976; 1978) to limit deforestation in water supply catchments and thus prevent stream salinization. This resulted in clearing controls in the Wellington Dam catchment in 1976 and the Mundaring Weir, Warren, Kent and Denmark river catchments in 1978 (Schofield *et al.* 1988), or in effect major current and future water supply catchments.

A foundation of water resource management in Western Australia in the 1970s and 1980s was the need to protect the existing forest cover on public land (Schofield and Ruprecht 1989; Schofield *et al.* 1988). Where there was any concern that timber harvesting or mining operations in forested areas may lead to salinity problems, extensive research programs were undertaken (Steering Committee for Research on Land Use and Water Supply 1985; 1987). Results from these programs have greatly improved the knowledge and understanding of forest hydrology.

The clearing of land across the SWWA has resulted in one million hectares of agricultural land affected by salinity, 450 plant species at risk of extinction, and 30 000 km of road and rail networks and 30 major rural towns at risk of increased maintenance costs due to the combined impact of salinity and waterlogging (Department of Environment 2003; Sparks *et al.* 2006).

In addition to the salinity concerns, the drying climate experienced since the mid-1970s (Bates *et al.* 2008b) and the consistent climate change predictions of increase aridity for the south-west have complicated both water management and forest management in SWWA. The climate change projections (IOCI 2012; Sadler *et al.* 1988) indicate a further drying of the climate and more frequent and extended dry periods. More recently, an additional issue has been the impact of a drying climate on not only water yield (Petrone *et al.* 2010), but also in stream ecology, and forest health. Following a drought in 2010, for example, there were widespread deaths in the jarrah forest (Matusick *et al.* 2013).

As mentioned, in the 1970s the forested catchments provided 90% of Perth's water supply. By the 1980s this had reduced to 60% and by the 2010s to 30% as initially groundwater and more recently seawater desalination have replaced surface water. The seawater desalination capacity is now 145 GL, with the monopoly water provider in the region, the Water Corporation, projecting that the contribution from forested catchments will reduce to zero before 2030 (Water Corporation 2011). However, the forested catchments continue to supply over 140 GL for irrigation on the Swan Coastal Plain.

Given these concerns of long-term supply of water quantity and quality it has been critical to understand the hydrology of SWWA and how this responds to forest disturbance and climate change. This has resulted in hillslope process studies, paired catchment experiments, catchment modelling, and intervention evaluation which form the core of this thesis.

1.5 Thesis aims and structure

Given the declining streamflows and salinity issues the research questions that need to be answered include confirming theories on how soil water moves through soil profile, what influences streamflow generation, what is the quantified impact of land use change, how distributed is the reduction in streamflows across the SWWA, what are the management implications and actions, and what further research needs to occur..

The aim of this thesis is to understand the impact of disturbance and climate on the hydrology of the forests of SWWA, and the objectives are:

1. To examine the characteristics of forest hydrology in SWWA;
2. To evaluate the hydrologic response to forest disturbance and climate variability of SWWA forests; and
3. To evaluate forest water management in the context of forest disturbance and climate change.

The thesis includes seven chapters with the overall structure outlined in Fig. 1.7. Chapter 2 reviews the relevant literature, Chapters 3 and 4 outline key research into hillslope hydrology and paired catchment studies respectively in the jarrah forests of SWWA. By understanding streamflow generation and the forest water

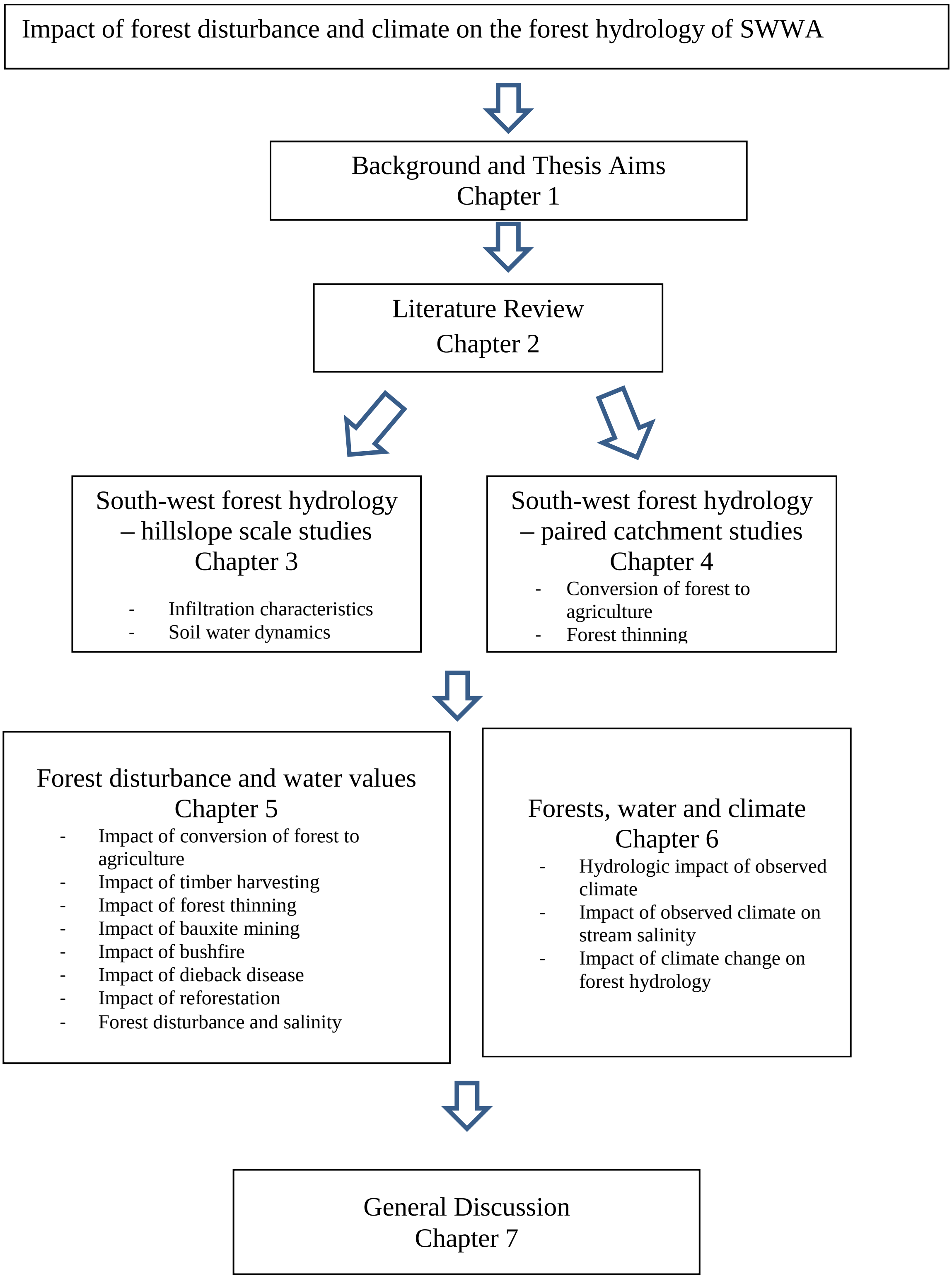


Figure 1.7 Schematic of the thesis structure

balance, the hillslope hydrology studies (Chapter 3) contribute to understanding how a changing climate or forest disturbance can impact on forest values. The paired catchment studies (Chapter 4) provide case studies on the impacts of various forest disturbances on water values. These case studies (paired catchment studies) are then put into context of broader studies and analysis to evaluate forest management (Chapter 5).

Chapter 6 examines the changes to forest hydrology from the observed climate and from climate change projections. Chapter 7 discusses management implications and forest values.

With respect to my role in the published papers that make up Chapters 3 and 4, a summary is provided in Table 1.2.

Table 1.2 Summary of my involvement in published papers

Chapter & Section	Paper	My role
3.1	**Ruprecht**, J.K. and Schofield, N.J., 1993. Infiltration characteristics of a complex lateritic hillslope. *Hydrological Processes*, 7: 87-97.	Developed research plan, undertook well-permeameter measurements, supervised construction of large infiltration sites, analysed data, and prepared and wrote journal paper. This work was supervised by Dr Schofield who reviewed the work plan and journal paper.
3.2	**Ruprecht**, J.K. and Schofield, N.J., 1990. Seasonal soil water dynamics in the jarrah forest, Western Australia. I: Results from a hillslope transect with coarse-textured soil profiles. *Hydrological Processes*, 4: 241-258.	For both papers I developed research plan, undertook some of the measurements, supervised measurement staff, analysed data, and prepared and wrote journal paper. Both were supervised by Dr Schofield who reviewed the work plans and journal papers.
3.3	**Ruprecht**, J.K. and Schofield, N.J., 1990. Seasonal soil water dynamics in the jarrah forest, Western Australia. II: Results from a site with fine-textured soil profiles. *Hydrological Processes*, 4:259 267.	

4.1	**Ruprecht**, J.K. and Schofield, N.J., 1989. Analysis of streamflow generation following deforestation in southwest Western Australia. *Journal of Hydrology*, 105: 1-17.	Developed research plan, analysed data, and prepared and wrote journal paper. This work was supervised by Dr Schofield who reviewed the work plan and journal paper.
4.2 4.3	**Ruprecht**, J.K. and Schofield, N.J., 1991a. Effects of partial deforestation on hydrology and salinity in high salt storage landscapes. I. Extensive block clearing. *Journal of Hydrology*, 129: 19-38. **Ruprecht**, J.K. and Schofield, N.J., 1991b. Effects of partial deforestation on hydrology and salinity in high salt storage landscapes. II. Strip, soils and parkland clearing. *Journal of Hydrology*, 129: 39-55.	For both papers I developed research plan, analysed the data, and prepared and wrote journal paper. Both were supervised by Dr Schofield who reviewed the work plans and journal papers.
4.4	**Ruprecht**, J.K., Schofield, N.J., Crombie, D.S., Vertessy, R.A. and Stoneman, G.L., 1991. Early hydrological response to intense forest thinning in southwestern Australia. *Journal of Hydrology*, 129: 261-277.	Developed research plan, analysed the data, and prepared and wrote journal paper. Drs Crombie and Stoneman provided silvicultural information for the paired catchments, and Dr Vertessy reviewed the manuscript. This work was supervised by Dr Schofield who reviewed the work plan and journal paper.
5	**Ruprecht**, J.K. and Stoneman, G. L., 1993. Impact of forest management activities on water values in native forests, *Journal of Hydrology, 150:369-391.*	Developed research plan, analysed the data, and prepared and wrote journal paper. The paper was revised and reviewed with Dr Stoneman.

Chapter 2 Literature review

2.1 Introduction

The role of forests in relation to the sustainable management of water resources remains a significant and contentious problem in many parts of the world (Calder 2007). This is despite extensive research into forest hydrology and significant advances into how forests interact with the climate and disturbance, both worldwide and Australia (Calder 2007).

One of the major characteristic traits of Mediterranean– type vegetation is that typically it is limited by water availability. At a regional scale, the ratio of rainfall to potential evapotranspiration has low values in Mediterranean ecosystems, often below one, which in effect means that actual transpiration in Mediterranean ecosystems never reach the potential evapotranspiration. In addition, a Mediterranean-type climate is characterized by the driest period of the year being during the summer season. These represent profound differences from temperate ecosystems where many forest hydrological studies have been undertaken and concepts developed.

In a worldwide sense, the hydrology of the jarrah forest of SWWA is unique in producing low levels of streamflow from moderate rainfall. This is attributed to the large soil water holding capacity of the deep weathering profile which jarrah (*Eucalyptus marginata*) is well adapted to exploit (Schofield *et al.* 1989b). Despite this low streamflow the jarrah forest was historically the dominant water supply to the state capital of Western Australia – Perth (WAWRC 1988).

Forests cover a large proportion of the earth's surface (31%) and are important for environmental, social, cultural and economic values (FAO 2010). Water and forests are two interdependent natural resources, consequently, the study of the interface between these two resources, forest hydrology, is an important field of research.

The National Research Council (National Research Council 2008) explained the general principles of managing forests for water (Fig. 2.1) under three main categories: modifiers of forest hydrology, and general and specific hydrologic responses. These principles show the general magnitude and direction of water movement through direct hydrologic responses to changes in forests over short time scales and in small areas (National Research Council 2008). However, forest hydrology needs to predict hydrologic responses in forest landscapes that are changing over large areas and/or over long time scales. In SWWA the forest landscapes are not only changing in response to past management actions, but also in response to climate change. These two factors, past management and climate change, have already had profound influences on forest water but their dynamic nature has complicated the ability and confidence in predictions of future forest hydrologic response.

Eagleson (1978) considered the water balance of a catchment as controlled by the interplay of climate, soil and vegetation. A proposed refined model separates soil and vegetation characteristics from management or disturbance (Fig. 2.2). Consistent with the general principles of hydrologic response, the three primary factors which influence the hydrologic response are:

1) Climate;

2) Land / Vegetation characteristics; and

3) Forest / Landuse management.

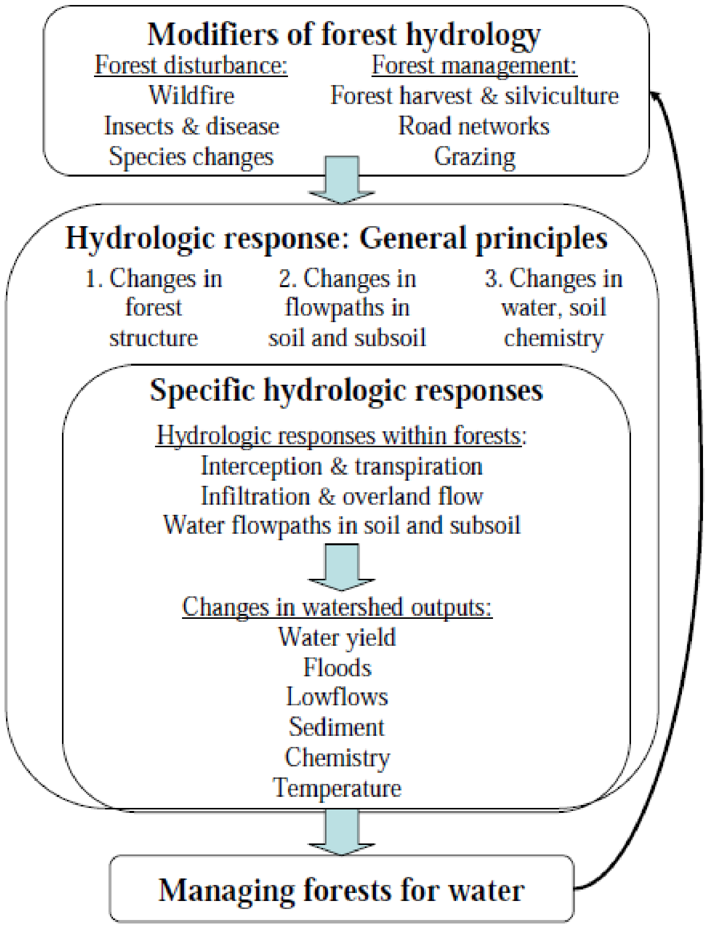


Figure 2.1 General principles of forest hydrologic response (National Research Council 2008)

Furthermore, the National Research Council study (National Research Council 2008) defined forest hydrology as the study of water in forests including:

- Its distribution, storage, movement and quality;
- Hydrologic processes within forested areas; and

- The delivery of water from forested areas.

Forest hydrology research uses field measurements; experiments and modelling to characterise and predict hydrologic processes and their responses to disturbance and management of forests. Paired catchment studies have been a major empirical approach in forest hydrology (Bren 2015; Neary 2016). In paired catchment studies, one catchment is left as a control and the other is treated (subject to disturbance or management such as timber harvesting or forest thinning). The two catchments (control and treated) are selected based on similarity of catchment characteristics such as size, slope, vegetation type, and landuse history. The measured changes in the relationship of streamflow and water quality between the treated and the control catchment are then used to quantify the impact of the forest disturbance or management.

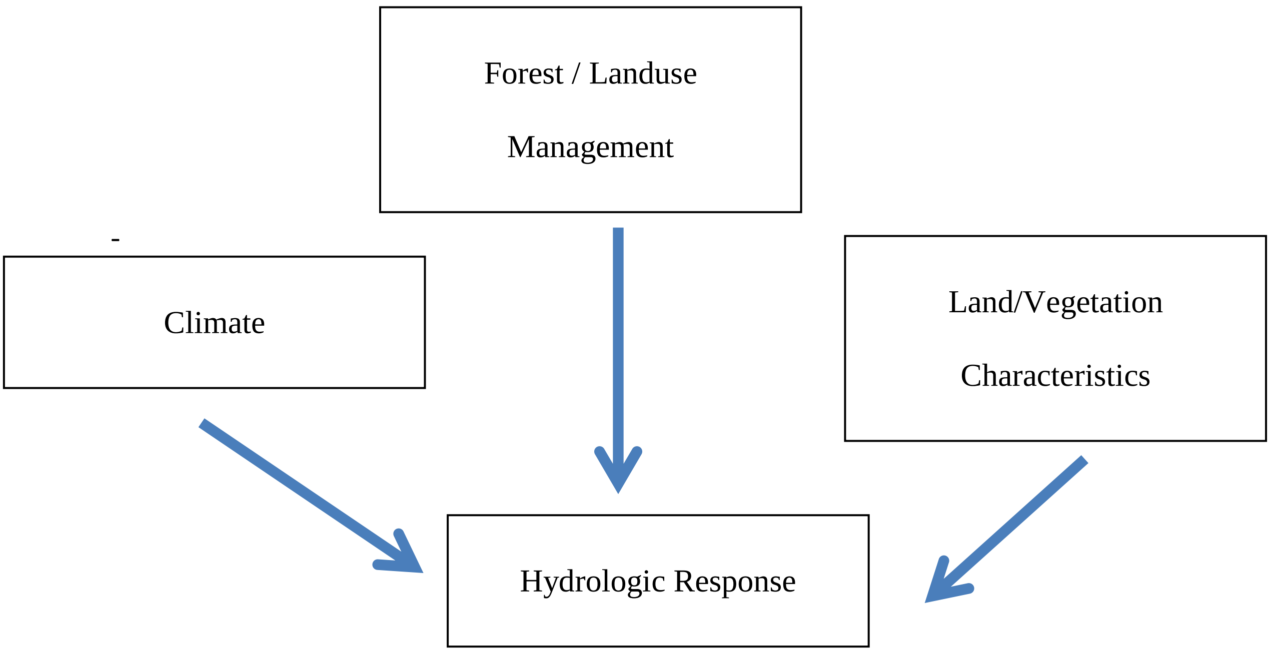


***Figure 2.2* Primary factors that influence the hydrologic response (adapted from Eagleson (1978)**

In addition to paired catchment studies, process measurement and plot-scale studies are used to understand the hydrology of catchments. Modelling at a point, plot, hillslope, experimental catchment and at a water resource catchment scale is used to further understand the hydrology and understand alternative management or climate scenarios.

Thus, while paired catchment studies appear to be a useful tool for catchment studies that examine the influence of land-use changes, the problem of resolving climate influences is harder because traditional pairs will be in a similar climate regime. Much hydrological experimentation in catchments was undertaken before the effects of climate change on hydrology were considered to be of importance (Bates *et al*. 2008b). Thus, this review will integrate the key factors impacting on the hydrologic response and provide strategic direction for further research.

2.2 Paired catchment approach

Defining the impact of land-use change on hydrology, water quality and on water resources is a recurring problem globally. Identifying the influence of land-use change has been difficult to separate from climate variability and in particular decadal variability and longer-term climate change.

Thus, recognizing and calculating the hydrological impacts of land-use change are not inconsequential exercises. DeFries and Eshleman (2004) have summarised the complexities as the:

(1) Relatively short period of hydrological records;

(2) Relatively high natural variability of most hydrological systems;

(3) Difficulties in 'controlling' land-use changes in real catchments within which changes are occurring;

(4) Relatively small number of controlled small-scale experimental studies that have been performed; and

(5) Challenges involved in extrapolating or generalizing results from such studies to other systems.

The high natural variability of hydrological systems and the difficulties in controlling land-use changes in real catchments resulted in the widespread use of "paired-catchment" studies (Best *et al.* 2003; Bosch and Hewlett 1982; Hewlett and Hibbert 1967; Neary 2016). The concept of a paired catchment study is that two catchments that are similar in a range of catchment characteristics are monitored simultaneously. One of the pair is subjected to a land-use change following an initial calibration period and the other remains in its original condition as a control over the subsequent measurement period (see Fig. 2.3) (Bren and Lane 2014). These approaches have been applied for more than a century, and a considerable literature has been generated.

Bren and Lane (2014) assessed the length of calibration required and based on the Nash-Sutcliffe coefficient of efficiency determined that reasonable calibration was achieved after 60 days, while more complex models achieved good calibration after 300 days. Studies in SWWA have used three years or more to achieve a level of confidence in pre-treatment calibration (Croton and Reed 2007; Williamson *et al.* 1987).

Best *et al.* (2003) raised some uncertainties with paired catchment studies, including:

- Whether the generalisations from paired catchments can be applied to larger catchments; and

- Whether the impacts of climate variability can be separated from the effects of land use change.

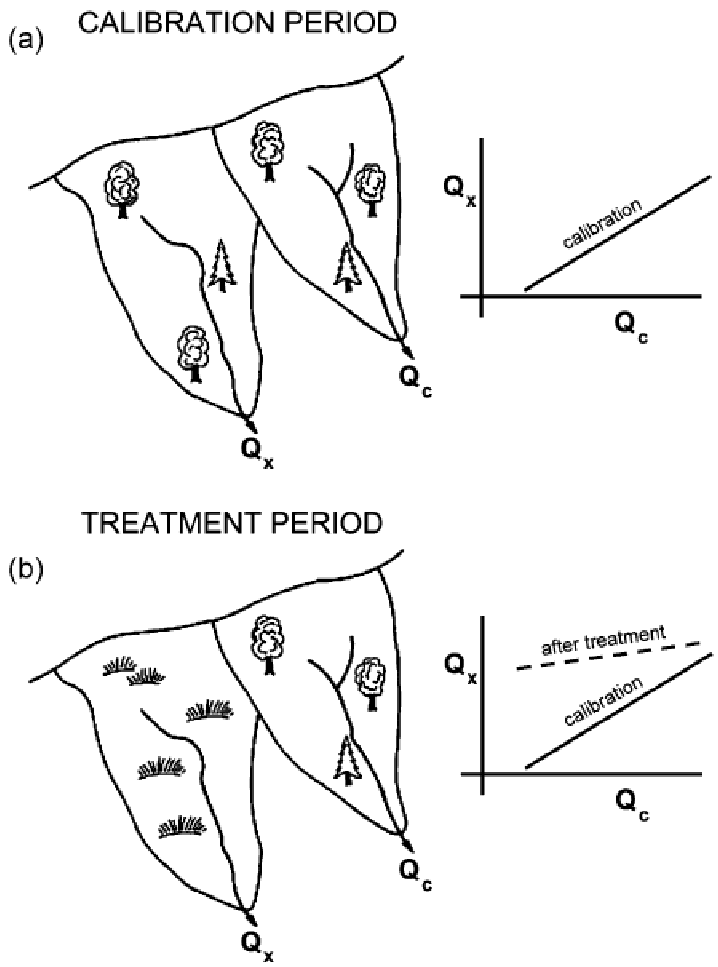


Figure 2.3 Sketch of a paired catchment study (from Hewlett 1982)

2.3 Paired catchments studies into the effects of forest disturbance on streamflow

The paired catchments studies into forest disturbance can be considered in four broad categories:

- Timber harvesting and regeneration;

- Conversion of eucalypt forest to pine plantation;
- Afforestation experiments; and
- Deforestation experiments.

Many of the timber harvesting and regeneration paired catchment water balance studies undertaken to date have focused on determining the immediate effects of timber harvesting on streamflow. However the longer term impacts on water yield from timber harvesting and regeneration have become more topical and the focus of research and investigation (Hawthorne *et al.* 2013; Lane *et al.* 2010; Scott and Prinsloo 2008; Webb *et al.* 2012).

Afforestation experiments relate to converting pasture or crops to pine or eucalypt plantations or native forest. Deforestation experiments study the conversion of forest to pasture or cropping.

In the following two sections paired catchments studies across the world and in Australia are summarised.

2.3.1 Northern hemisphere studies

Hibbert (1967) reviewed 39, mainly deciduous hardwood and conifer, forested catchment experiments across the world and summarised the results as:

- A reduction of forest cover increased water yield;
- Afforestation decreased water yield; and
- The response to treatment was variable and unpredictable.

Bosch and Hewlett (1982) summarised the results from 94 experimental catchments in USA, Canada, Japan, South Africa, New Zealand and Australia undertaken to determine the effect of vegetation change on water yield. Since 1982, many additional paired catchment studies have been reported in the literature and have been summarised (Brown *et al.* 2013; Brown *et al.* 2005; Hornbeck *et al.* 1993; Sahin and Hall 1996; Stednick 1996).

The conclusions from Bosch and Hewlett (1982) were consistent with those of Hibbert (1967) and included:

1) Reducing forest cover causes as increase in water yield;
2) Increasing forest cover causes a decrease in water yield;
3) Coniferous and eucalypt cover types cause ~40 mm change in annual water yield per 10% change in forest cover;
4) Deciduous hardwoods are associated with ~25 mm change in annual water yield per 10% change in cover;
5) Reductions in forest of less than 20% apparently cannot be detected by measuring streamflow; and
6) Streamflow response to deforestation depends on both the mean annual precipitation of the catchment and on the precipitation for the year under treatment.

Stednick (1996) summarised that the streamflow responses to vegetation conversion depend both on the region's annual precipitation and on the precipitation for the year under treatment. Stednick (1996) also emphasised that changes in water yield from harvesting less than 20% of a catchment couldn't be determined by streamflow measurements.

The annual yield change resulting from treatment in high rainfall areas appears to be independent of the variation of rainfall from year to year (Bosch and Hewlett

1982; Stednick 1996) and more a function of forest regrowth or leaf area index (Stednick 1996). Part of the variability in reported annual water yield increases and streamflow response following timber harvesting, including partial harvest, may be due to the physical location of the harvest units with respect to the source area of streamflow (Bosch and Hewlett 1982; Stednick 1996).

Stednick (1996) discussed the process of hydrologic recovery needing to focus on not just water yield, but also peak flows and lowflows. The other uncertainty related to the baseline and the aim for recovery. This is particularly relevant to the SWWA given the current drying climate. Ebel and Mirus (2014) discussed a disturbance-response-recovery concept. The disturbance-response-recovery trajectory is considered hysteretic; a similar conclusion was reached by Hatton (2001) for SWWA in relation to deforestation, salinization, and widespread reforestation.

Sahin and Hall (1996) applied fuzzy linear regression analysis to data from 145 experiments, which included the 94 catchments from Bosch and Hewlett (1982) and additional catchments from Australia (Ruprecht and Schofield 1989a), tropical rainforests (Bruijnzeel 1990), and north east USA (Hornbeck *et al.* 1993). Collectively they found that for a 10% reduction in cover, the yield from conifer-type forest increased by some 20 to 25 mm, whereas that for eucalyptus type forest increased by only 6 mm. Both values were somewhat lower than those previously published, as was the 5 mm decrease in yield associated with a 10% afforestation of scrubby vegetation. A 10% reduction in the cover of deciduous hardwood forests gave a 17 to 19 mm increase in yield, broadly in line with earlier estimates (Sahin and Hall 1996). Wang *et al.* (2011) reviewed the relationship between forest cover and catchment water yield data from 118 catchments in northern China using published data from a variety of sources and found that forest cover was not correlated with annual precipitation at micro (<50 km^2) and meso scales (50–1000 km^2), while they were positively correlated at the macro (>1000 km^2) scale. The multiple stepwise regression analysis by Wang *et al.* (2011) indicated that runoff was influenced by altitude, annual precipitation, forest cover, and potential evapotranspiration (PET) in Northeast China.

Sahin and Hall (1990) conclusion that a 10% reduction in cover results in the yield from a conifer-type forest increasing by some 20 to 25 mm is based on a small number of data points and extrapolation. In contrast Stednick (1996) is considered more conservative in his conclusions given the small number of data points for less than 20% of a reduction in forest cover (see Brown et al. 2005).

2.3.2 Tropical and sub-tropical studies

Bruijnzeel (1990) reviewed 20 paired catchment studies in humid tropical and subtropical sites. Bruijnzeel (1990) concluded that the direction and magnitude of water yield changes related to forest clearance in tropical forests were similar to those reported by (Bosch and Hewlett 1982; Hibbert 1967; Stednick 1996). Bruijnzeel (2004) in summarising the paired catchment studies, concluded that the removal of more than 33% of forest cover resulted in significant increases in annual streamflow in the first three years.

2.3.3 New Zealand studies

Dons (1986) found an average annual runoff decline of 83 mm in a large basin vegetated with scrub that had been 28% afforested with *Pinus radiata*. Fahey and

Jackson (1997) in studies on the South Island, reported streamflow increases, in the first year after clearing, ranging between 22 and 67 mm per 10% of forest area cleared. Rowe and Pearce (1994) presented results generated by harvesting native forest catchments and establishing radiata pine plantations in New Zealand. The catchments were small (1.6 to 4.6 ha) and in a high-rainfall (2000 to 2500 mm) zone. They found that "in the year after treatment (e.g. deforestation) streamflow generally increased by 200 to 250 mm". This was followed by rapid colonisation of the newly planted catchments with bracken and other species which led to a rapid decline in streamflow which returned to pre-treatment levels after an average of about 5 years. Streamflow yields then continued to decline for another 2 to 3 years before stabilising at a level about 250 mm yr^{-1} lower than the pre-treatment level.

2.3.4 South African studies

Van Wyk (1987) reported mean annual streamflow declines for three grassland catchments afforested with pines in the south-western Cape Province of 32, 35 and 47 mm per 10% of catchment afforested. Scott and Prinsloo (2008) studied the longer-term effects of afforestation with *Pinus radiata* and *Eucalyptus grandis* on streamflows from two paired-catchment experiments in South Africa. They found that the conversion to pine caused peak reductions in yield over a 5-year period of 44 mm yr^{-1} or 7.7% yr^{-1} for each 10% of catchment planted when the trees were between 10 and 20 years old. The conversion to a eucalypt plantation caused peak reductions over a 3-year period of 48 mm yr^{-1} and 10% yr^{-1} for each 10% of catchment planted. However, Scott and Prinsloo (2008) concluded that as the plantations matured (over 30 years of age in the case of pines and over 15 years of age in the case of eucalypts) the flow reduction trend was reversed, and streamflow effects appeared to be tending toward pre-afforestation levels.

2.3.5 Australian studies

Australian paired studies can be categorised into:

1) Timber harvesting;
2) Conversion of forest to pine plantation;
3) Conversion of pasture to pine plantation; and
4) Conversion of forest to agriculture.

Timber harvesting

The major published studies of the impacts of timber harvesting in eastern Australia are Karuah, Tantawangalo, Coranderrk, North Maroondah, Reefton, Yambulla and East Keiwa are summarised below (see Table 2.1).

Karuah, New South Wales

The Karuah hydrology project is the longest running native forest paired catchment study in the state of New South Wales (NSW), where different eucalypt species occur. Following logging in 1983, Cornish (1993) reported that water yields increased significantly by about 150 to 250 mm yr^{-1} in five of the six treated catchments and the magnitude was related to the proportion of the basin harvested.

Webb *et al.* (2012) reported on the longer term impacts of logging and regeneration at Kuruah, and found that the initial increase lasted for greater than 5 years in the logged and unburnt Bollygum catchment, but had returned to pre-treatment levels within 2.5 years in the logged and burnt Corkwood and Jackwood catchments, and within 2 years in the Kokata and Coachwood plantation catchments. A significant reduction in streamflow was reported in three catchments – Corkwood (113.5 mm yr^{-1}), Bollygum (72.7 mm yr^{-1}) and Kokata (68.9 mm yr^{-1}). The Corkwood catchment had returned to the pretreatment level within 7 years post-harvest. A continuing suppression of streamflow after 27 years was described in two of the catchments, Bollygum and Kokata. Kokata experiencing a further decline from 2005 onwards to a mean annual reduction of 172.4 mm. By contrast a significant increase in streamflow relative to the pre-treatment level was recorded in the Jackwood catchment after 23 years (157.6 mm yr^{-1}).

Table 2.1 Summary of timber harvesting paired catchment studies in eastern Australia

Study	Pre-disturbance	Disturbance	Response
Karuah, NSW 2 controls with 6 treated catchments	Forest	Timber harvesting, with no regeneration burn, with regeneration burn, and with hand planted	Five catchments experienced a significant increase in streamflow ranging from 120 mm to 319.6 mm (varied in proportion to proportion (%) of catchment logged). This initial increase lasted for greater than 5 years in the logged and unburnt catchment, but had returned to pre-treatment levels within 2.5 years in the logged and burnt catchments, and within 2 years in the plantation catchments. After 27 years two catchments showed a continuing streamflow reduction, three catchments showed no change and one catchment showed an increase in streamflow
Tantawangalo, NSW 1 control with 2 treated catchments	Forest	Timber harvesting, with one catchment patch-cut and another thinned	Streamflow increased by 10% in the first 3 years after logging at patch-cut catchment, and by 31% for the first 4 years at the thinned catchment. Streamflow then returned to pre-treatment levels for 1 year and subsequently declined by 20% below pre-treatment levels over the next 4

Study	Pre-disturbance	Disturbance	Response
			years at patch-cut catchment. The thinned catchment returned to pre-treatment levels over the same period.
Coranderrk, Vic Control with 3 treated catchments	Forest	Timber harvesting, with subsequent burning and hand-reseeding	The annual flow increased for the next 3 years (relative to the control), reaching a maximum of almost 300 mm. 8 years after the logging the annual flow had returned to pre-treatment levels. Annual flow then continued to decline with a maximum reduction of around 200 mm yr^{-1}. Thirty-four years after treatment streamflow was still below the pre-treatment flow.
North Maroondah, Vic 4 controls with 10 treated catchments	Forest	Timber harvesting	Myrtle Creek catchment showed initial increase of 320 mm post treatment, followed by a reduction in flow of 80 to 50 mm, 7-11 years post treatment Monda catchment experienced initial increase, but reduction not observed complicated by insect infestation.
Reefton, Vic Controls with 2 treated catchments	Forest	Timber harvesting	Increased flow of 70 mm or 32% of pre-treatment.
Yambulla, NSW Control with 5 treated catchments	Forest	Integrated logging with regeneration burn in 3 of the treated catchments	In all five treated catchments an increase in streamflow was detected following the 1979 wildfire and/or integrated logging activities that occurred at various intervals. A subsequent reduction of streamflow to below that of a mature stand was not detected in three of the catchments but was detected in the two that had been subjected to integrated logging and wildfire.
East Kiewa, Vic Control with 1 treated catchment	Forest	Timber harvesting, followed by regeneration burn	Small increases in total streamflow following treatment

Tantawangalo, New South Wales

Lane and Mackay (2001) reported on an 11-year paired-catchment study implemented to assess the effect of differing logging practices on streamflow in mixed-species eucalypt forest at Tantawangalo Creek, NSW, Australia. Following a 3-year calibration period, Ceb catchment was retained as a control, Wicksend catchment was patch-cut to remove 22% of basal area, and Willbob catchment was thinned to remove 12% of basal area. The treatments resulted in an initial increase in monthly total streamflow and baseflow at both treated catchments. In annual terms, streamflow increased by 10% in the first 3 years after logging at the patch-cut catchment, and by 31% for the first 4 years at the thinned catchment. Streamflow then returned to pre-treatment levels for 1 year and subsequently declined by 20% below pre-treatment levels over the next 4 years at the patch-cut catchment (Lane and Mackay 2001), whilst the thinned catchment returned to pre-treatment levels over the same period (Lane and Mackay 2001).

Coranderrk, Victoria timber harvesting

The Coranderrk Project, located 70 km east of Melbourne, is considered Australia's longest-running paired catchment study (Bren *et al.* 2010). Three small catchments (Slip Creek, Blue Jacket Creek, and Picaninny Creek) originally carrying old growth mountain ash (*Eucalyptus regnans*) and mixed forest species were gauged from 1958. In 1971/72 Picaninny Creek catchment was substantially clear-felled and regenerated, predominantly with mountain ash. The annual flow increased for the next 3 years (relative to the control), reaching a peak of almost 300 mm increase relative to the control catchment. Flow then declined and by 8 years after the logging the annual flow had returned to pre-logging levels. Annual flow then continued to decline with a maximum reduction of around 200 mm yr^{-1}. Thirty-four years after logging the flow was still below the pre-treatment flow and showing no sign of recovery, although there were year to year variations associated with rainfall and drought (Bren *et al.* 2010).

North Maroondah, Victoria timber harvesting

The Maroondah catchments comprised 10 treated experimental catchments (3-Black Spur, 3-Ettercon, 3-Monda and 1-Myrtle) and four control catchments. The area has an extensive research and publication history, summarised by Watson *et al.* (1999). The results for the Myrtle group showed a clear treatment effect with an increase in water yield for 2-3 years following clear-felled harvesting treatment, falling to a decrease in water yield 10 years after treatment (Watson *et al.* 1999). The results for the Monda group were initially similar to the Myrtle group with a short term increase in water yield. However, at the age when a decreased water yield would be expected an insect infestation occurred causing forest dieback and regeneration (Watson *et al.* 1999).

Reefton, Victoria timber harvesting

The aim of the Reefton study in Victoria was to assess the effects on both water yield and water quality of some forest practices commonly used in the management of mixed species eucalypt forest in Victoria. There were six gauged catchments ranging in size from 70.4 to 521.2 ha. The pre-treatment conditions were summarised in Wu *et al.* (1984). The initial results to 1989 were assessed by Mein and Nandakumar (1993) who reported an increased flow of 70 mm or 32% of pre-treatment following 20% of the catchment cleared.

Yambulla, New South Wales

The Yambulla project was established in 1977 within five small catchments in Yambulla State forest, located 50 km southwest of the township of Eden in south-eastern NSW (Webb and Jarrett 2013). A sixth catchment was added to the project in 1979 following an extensive wildfire that burnt four of the original catchments (MacKay and Cornish 1982). In all five treated catchments an increase in total streamflow, baseflow and stormflow was detected following the 1979 wildfire and/or integrated logging activities that occurred at various intervals. A subsequent reduction of streamflow to below that of a mature stand was not detected in three of the catchments but was detected in the two that had been subjected to integrated logging followed by a wildfire, and a wildfire followed by salvage logging, respectively. The streamflow reduction was negligible and short-lived in each case meaning that overall there was a cumulative increase in streamflow in the post-disturbance period. These results contribute to a growing body of evidence indicating that catchment-scale hydrological responses to disturbance of mixed species eucalypt forests do not follow the response often reported for wet Mountain Ash forests (Webb and Jarrett 2013).

East Kiewa, Victoria

A paired-catchment experiment was initiated in 1978 to study the hydrologic impacts of logging in the East Kiewa valley mountain area of north east Victoria, Australia. The study sites were mostly vegetated with *Eucalyptus delegatensis* (alpine ash). One catchment, Slippery Rock Creek (136 ha) was retained as the control, and Springs Creek (244 ha) underwent logging of 30% of the area after 4 years of calibration (Lane *et al.* 2006). The study found there were small increases in total streamflow following treatment, and small but statistically significant increases ($p < 0.005$) in mean event flows (Lane *et al.* 2006).

Conversion of native forests to pine plantations

The major published paired catchment studies of the water impacts of converting native forest to pine plantation are Lidsale, Croppers Creek, Toolara, and Stewarts Creek which are summarised below (see also Table 2.2).

Lidsdale, New South Wales

The experimental catchment at Lidsdale in the Central Tablelands of NSW, Australia was established in 1967 (Putuhena and Cordery 2000). The water balances of the two forest catchments were observed for a 27-year period; 11 years before and 16 years after the forest conversion. In January 1978, L-6 was covered by eucalypt forest. During February 1978, the forest was cleared and windrowed, and then burned in April 1978. During winter of the same year the catchment was planted with *P. radiata* (Putuhena and Cordery 2000). Data on precipitation, canopy interception, forest floor interception and water yield, combined with analysis of the rate of vegetative growth and development of the forest floor litter, enabled investigation of the effects of forest conversion on the water balance components for the whole period (1967–1993). The age of a *P. radiata* plantation during the first 16 years of its growth greatly affected the streamflow and other water balance components. For the first 4 years after forest conversion, the rates of evapotranspiration and streamflow changed completely. Transpiration and the evaporation of intercepted rainfall ceased after the forest was cleared. The changes in the first 4 years were followed by a further transformation of the whole evapotranspiration process as the pine plantation developed. A trend of increasing evapotranspiration and canopy and forest floor interception losses as the plantation grew, with decreases in runoff, was followed by an equilibrium situation in which streamflow, and the evapotranspiration from soil water storage were smaller than for the native forest (Putuhena and Cordery 2000).

Croppers Creek, Victoria

In 1975 the Croppers Creek paired catchment project in north-east Victoria was initiated to determine the hydrologic impact of intensively-managed radiata pine on water flows and water quality compared to the native (eucalypt) forest (Bren and Hopmans 2007; Bren and Papworth 1991). One catchment (Clem Creek) was converted from native forest to plantation in 1980 after a five year ‘‘calibration’’ period. Adjoining ‘‘control’’ catchments (Ella and Betsy Creek) consist of native, mixed species eucalypt forest.

The conversion to radiata pine led to an increase in water yield of up to 300 mm yr^{-1} in the Clem Creek catchment immediately after clearing. This response has declined but in general the runoff from the pine catchment is still higher than that of the eucalypt catchment except for ‘‘drought’’ years (Bren and Hopmans 2007).

Table 2.2 Summary of conversion of forest to pine paired catchment studies in eastern Australia

Study	Pre-disturbance	Disturbance	Response
Lidsdale, NSW Control with 1 treated catchment	Forest	100% cleared then pine plantation	Initial increase of 20% with clearing followed by a 2-4% reduction in streamflow 14 to 16 years post treatment

Study	Pre-disturbance	Disturbance	Response
Croppers Creek, Vic 2 controls with 1 treated catchment	Forest	100% cleared then pine plantation	Initial increase of 300 mm with clearing, but streamflow still above control 25 years post treatment, except in dry years
Toolara, Qld Control with 1 treated catchment	Forest	90% cleared then pine plantation	3-5 year increase in streamflow, followed by decline to pre-treatment conditions
Stewarts Creek, Vic Control with 1 treated catchment	Forest	100% cleared then pine plantation	Initial increase of approximately 380 mm Returned to pre-treatment in 18 years, but subsequently streamflow was above pre-treatment at 24-27 years post treatment

Toolara, Queensland

A paired catchment study was conducted over a 10-year period on the hydrology of an exotic *Pinus* plantation in the coastal lowlands of south-east Queensland, Australia. Each catchment was instrumented with a stream monitoring station, tipping bucket rain gauge, and a network of piezometers to monitor the shallow perched water table (Bubb and Croton 2002). After a 6-year calibration period, a harvest treatment was imposed on one catchment with clearfelling of approximately 90% of the catchment area, which contained a mature (44-year-old) *Pinus elliottii* plantation. This subsequently was re-established with a second rotation plantation of a hybrid of *P.elliottii* × *P.caribaea* var. *hondurensis*. The control catchment (Crayfish) contained a *P. elliottii* plantation similar to the clearfelled site. The interim results showed a temporary, 3 to 5-year increase in stream flows, followed by a decline to preharvest flows. Definition of the decline was based on only 1 year of data and is considered very preliminary (Bubb and Croton 2002).

Stewarts Creek, Victoria

Stewarts Creek (Victoria), represents a set of decommissioned research catchments with 9 years of pre-treatment data and 25 years of post-treatment. The treatment for one catchment was a conversion from native eucalypt forest to pine. After 8 years, the treated catchment had similar hydrologic variables compared to the native forest control. However, Mein *et al.* (1988) concluded that interception loss was likely to be greater and total runoff consequently smaller in a mature pine forest compared to a native eucalypt forest.

Conversion of pasture to pine

The major published paired catchment studies of the water impacts of converting pasture to pine plantation are Pine Creek and Red Hill and these are summarised below (see also Table 2.3).

Pine Creek, Victoria

The 3 km^2 Pine Creek catchment in Victoria has been used extensively in studies assessing impact of pine afforestation in the Murray Darling Basin (Lane *et al.* 2005; van Dijk and Keenan 2007; Zhang *et al.* 2007). In 1986 and 1987 the whole of the catchment was converted from open grassland to *Pinus radiata* plantation, with preliminary analysis showing that annual stream salinity and salt load have decreased significantly (Zhang *et al.* 2007). However, there is no published adjacent control catchment to isolate the climate influence.

Red Hill, NSW

The Red Hill hydrology project consisted of two paired catchments nested within a larger catchment established in 1989 near Tumut. The treatment consisted of afforestation with *P. radiata.* A small (195 ha) afforested catchment (Red Hill) was compared with a small (135 ha) Kileys Run catchment which remained under pasture, and was also compared with the much larger (832 ha) planted Saw Mill Creek catchment. This design was chosen so that effects of both treatment and catchment size could be evaluated. Preliminary results indicated that additional annual pine water use was in excess of 200 to 250 mm by age nine years (Major *et al* 1998).

Table 2.3 Summary of conversion of pasture to pine paired catchment studies in eastern Australia

Study	Pre-disturbance	Disturbance	Response
Pine Creek, Victoria Treated catchment, no control	Pasture	Pine plantation	95 mm yr^{-1} reduction in water yield estimated
Red Hill, NSW Control with 1 treated catchment	Pasture	Pine plantation	Additional annual pine water use was in excess of 200 to 250 mm by age nine years. However there was no calibration period prior to treatment.

Conversion of forest to agriculture

The major published paired catchment studies of the water impacts of converting native forest to agriculture (cropping or pasture) are Brigalow and Babinda are summarised below (see also Table 2.4).

Brigalow, Queensland

The Brigalow Catchment Study (BCS) commenced in 1965 with a pre-clearing calibration phase of 17 years to define the hydrology of 3 adjoining catchments

(12–17 ha). After two catchments were cleared in 1982, three land uses (Brigalow forest *Acacia harpophylla*, cropping, and grazed pasture) were monitored for water balance, resource condition and productivity, providing information for scientific understanding and resource management of the major land uses of the brigalow bioregion (Cowie *et al.* 2007).

Hydrologic characteristics were then compared for the following 21 years. In their pre-disturbance state, the catchments behaved similarly, with average annual runoff being 5% of annual rainfall (Thornton *et al.* 2007). Once cleared, total runoff from the cropping catchment increased to 11% of annual rainfall and total runoff from the pasture catchment increased to 9% of annual rainfall. However, timing of the individual runoff events varied between land uses (Thornton *et al.* 2007).

Babinda, Queensland

The Wyvuri paired catchment experiment which is located near Babinda on the eastern slopes of the Graham Range was established in 1969. Two catchments were maintained in an undisturbed condition until June 1971, when the northern catchment was logged on an unconstrained salvage basis prior to clearing in July 1973 for the establishment of tropical pastures (Cassells *et al.* 1985). The initial response was an increase in streamflow of 293 mm yr^{-1} (10.2%) in the first 2 years after clearing.

Table 2.4 Summary of conversion of forest to agriculture paired catchment studies in eastern Australia

Study	Pre-disturbance	Disturbance	Response
Brigalow Control + 2 treated catchments	Forest	Agriculture – cropping and pasture	Cropping - Increase in streamflow of 6% of annual rainfall Pasture – Increase in streamflow of 4% of annual rainfall
Babinda Control + 1 treated catchment	Rainforest	Agriculture – tropical pasture	Streamflow increase of 293 mm yr^{-1} (10.2%) in the first 2 years after clearing

2.4 Paired catchment studies across SWWA forests

An extensive range of small paired catchment experiments has been conducted in the jarrah forest. The treatments cover nearly all the major land uses in the jarrah forest, including bauxite mining, forest harvesting and regeneration, forest thinning and clearing for agricultural development. In each case, the experimental method involved paired catchment studies, which compared water yield from a treated catchment with that from an untreated, control catchment. More recently, the use of complex distributed catchment models has been used to complement the paired catchment approach (Brunner and Simmons 2012; Croton *et al.* 2005).

Paired catchment studies have been undertaken in the south-western forests of Australia since the 1970s. These paired catchment studies (Fig. 2.4) have included:

1) Collie Research Catchments – research into forest clearing for agriculture;
2) Southern Forests research catchments – research into logging and regeneration;

3) Forest thinning research – research into forest thinning to increase water yield; and
4) Bauxite Mining – research into managing water yield and water quality issues with mining and rehabilitation.

In the following sections the research catchments and their treatments are explained, but the results from key catchment experiments are described in detail in Chapter 4, and an overall review of the studies described in Chapters 5 and 6.

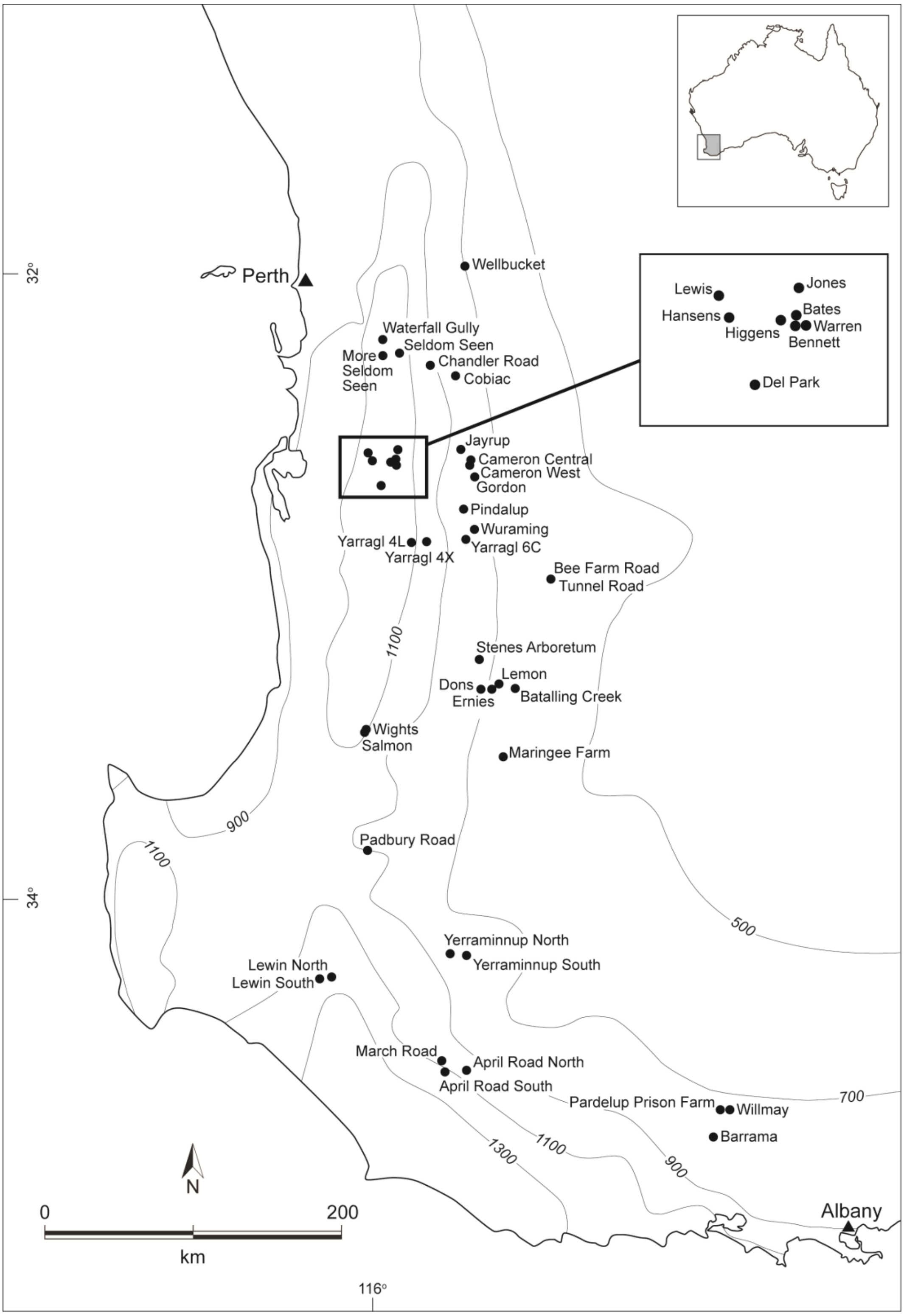


Figure 2.4 Location of research catchments across south-west forests of Western Australia

2.4.1 Paired catchment studies into deforestation

In the late 1960s, concern developed about the impact of continued agricultural development on the salinity of streams and rivers in SWWA. A particular area was the Collie River Basin which has been developed to provide water supplies for irrigation and for country towns up to 300 km away. At that time there was a significant body of opinion that the basic cause of salinity was not agricultural development but climate trends; in particular, sequences of years with exceptionally high or low rainfall (Schofield *et al.* 1988). In order to resolve this question, the Australian research organisation CSIRO began a field study in collaboration with the local water authority (the Public Works Department of Western Australia, now part of the Department of Water in Western Australia) and with the aid of funding from the then Australian Water Resources Council (Peck and Williamson 1987b).

The experimental catchments (Table 2.5; Fig. 2.5) were established in two groups: a pair in a relatively high (about 1200 mm yr^{-1}) rainfall zone, and a set of three further inland (about 800 mm yr^{-1}) (Peck and Williamson 1987b). Although salinity is not considered to be a problem in the higher rainfall area, it was expected that effects of agricultural development would be detectable and that they would be evident much earlier than any changes further inland.

An internet search (Google scholar – 26th November 2016) identified over 80 publications that have used the data from the Wights catchment in their analysis. This supports the critical importance of long-term monitoring to understand the hydrologic impact of clearing forested areas.

Table 2.5 Research catchments into impacts of agricultural clearing in SWWA

Catchment	Catchment area (km^2)	Mean annual rainfall (mm)	Treatment
Wights	0.94	1120	100% cleared
Salmon	0.82	1120	Control – Open jarrah forest
Ernies	2.70	820	Control – Open jarrah forest
Lemon	3.44	820	Lower 53% cleared, remaining 47% open jarrah forest
Dons	3.50	800	Parkland clearing 4%, strip clearing 20%, soil unit clearing 14%, remaining area open jarrah forest

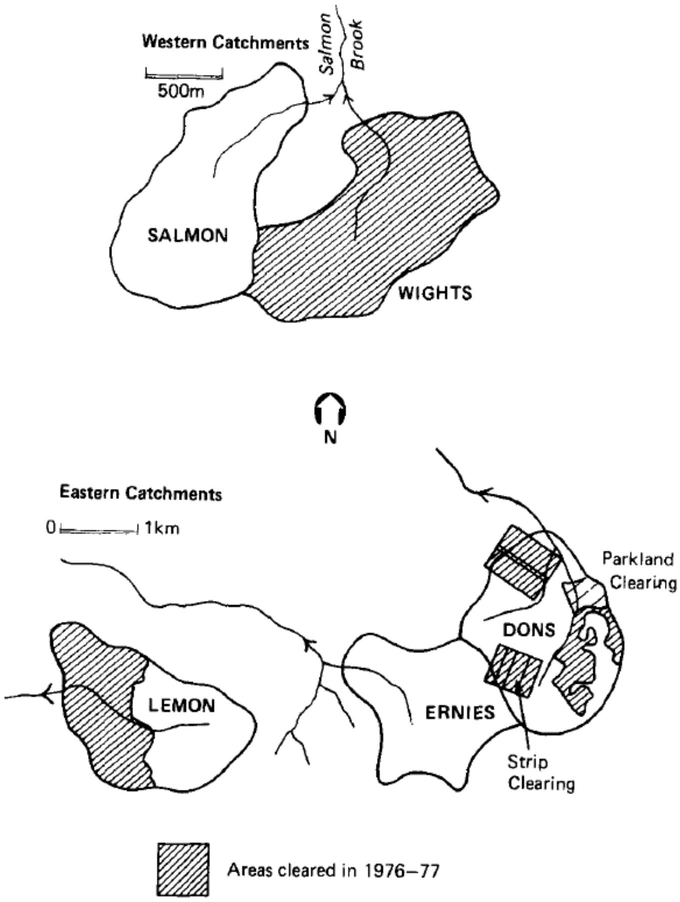


Figure 2.5 Pattern of clearing for Collie research catchments (Williamson et al. 1987)

2.4.2 Paired catchment studies into the impact of timber harvesting and regeneration

In response to concerns regarding the impact of logging to water quality, in particular stream salinity, another research programme was establish to quantify any water quality impacts (Western Australian Steering Committee for Research on Land Use and Water Supply 1987). The main objectives of the research programme were to:

1) Determine the magnitude and duration of any increase in stream salinity and sediment concentration resulting from the proposed logging operations;
2) Consider the long term effects of logging and regeneration on water yield; and
3) Propose, if necessary, improved logging practices to preserve water quality.

In the northern jarrah forest, an initial catchment experiment into the impacts of timber harvesting was conducted at the Wellbucket catchment (Table 2.6). In the southern forests, four research catchments were treated with a range of logging operations, with three catchments remaining as untreated controls (Table 2.6).

2.4.3 Paired catchment studies into forest management to increase runoff

Studies into the option of increasing runoff into existing reservoirs for the Perth water supply have been undertaken since the early 1980s (Steering Committee for Research on Land Use and Water Supply 1987). These early studies were followed by a more intense investigation by the Water Corporation titled the Wungong Catchment Trial (Reed *et al.* 2012).

Eight catchments (Table 2.7) have been treated with a range of forest management or thinning treatments to measure the water yield increases.

Table 2.6 Research catchments investigating impact of timber harvesting

Catchment	Catchment area (km^2)	Rainfall zone	Treatment
Lewin South	0.9	High	Heavy selection cut of jarrah/marri, karri gully clearfelled – no stream buffer
Lewin North	1.13	High	Control – jarrah/marri, and karri forest
April Road South	1.79	Intermediate	Control - Jarrah, marri and karri forest
April Road North	2.48	Intermediate	Jarrah, marri and karri forest clearfelled and then replanted with karri – stream buffer retained
March Road	2.61	Intermediate	Clearfelled and then replanted – no stream buffer retained
Yerraminnup South	1.83	Low	Heavy selection cut jarrah forest – stream buffer retained
Yerraminnup North	2.53	Low	Control – jarrah forest
Wellbucket	4.65	Low	Heavy selection cut of jarrah forest – stream buffer retained

Table 2.7 Research catchments investigating impact of forest thinning on hydrology

Catchment	Catchment area (km^2)	Rainfall zone	Treatment
Hansen	0.78	High	Uniform thinning, reducing basal area from 35 to 7 m^2 ha^{-1}
Higgens	0.60	High	Uniform thinning, reducing basal area from 37 to 14 m^2 ha^{-1}
Jones	0.69	High	Operational thinning, reducing basal area from 43 to 17 m^2 ha^{-1}
Gordon	2.09	Intermediate	Control - Forest
Bates	2.70	High	Control
Yarragil 4L	1.28	Intermediate	Operational thinning
Yarragil 6C	4.58	Intermediate	Intensive treatment
Yarragil 4X	2.73	Intermediate	Operational thinning
Wuraming	4.4	Intermediate	Control - Forest
Lewis	2.01	High	Control, later mined for bauxite
Cobiac	3.64	Intermediate	66% of catchment thinned, reducing basal area from 26.4 to 15.7 m^2 ha^{-1}
Chandler Road	17.50	Intermediate	25% of catchment previously mined and rehabilitated 55% of catchment thinned (predominantly bauxite mining rehabilitation) resulting in reducing basal area from 26 to 16 m^2 ha^{-1}

2.4.4 Paired catchment studies into impacts of bauxite mining on forest hydrology

Bauxite mining is a major land use within the Jarrah Forest of SWWA, and typically involves clearing of all vegetation, removal of topsoil, blasting of caprock, excavation of the top 1-3 m of caprock, topsoil spreading, and revegetation (Koch and Hobbs 2007). Bauxite mining typically occurs over 20 to 30% of a catchment at a first-order stream level. Approximately 1000 ha of forest is cleared annually for mining. The principal bauxite mining area covers 50–60%

of the Northern Jarrah Forest and most of the water supply catchments for the Perth metropolitan area and surface water irrigation catchments (Steering Committee for Research on Land Use and Water Supply 1985).

The Seldom Seen and More Seldom Seen catchments were mined from 1969 to 1994 (Table 2.8) with Waterfall Gully as the control catchment. Del Park catchment had hillslope and hydrologic process studies undertaken at native forest and mined sites. The West Cameron, Central Cameron, Gordon, and Jayrup catchments were part of the Joint Intermediate Rainfall Zone Research Programme (JIRZRP) (Mauger *et al.* 1998). The intermediate rainfall zone was considered the most at risk zone for salinity increases from bauxite mining. Tunnel Road and Bee Farm Road were within the low rainfall zone of the jarrah forest (Table 2.8).

Table 2.8 Research catchments investigating water impacts of mining

Catchment	Catchment area (km^2)	Rainfall zone	Treatment
Waterfall Gully	8.74	High	Control - Forest
Seldom Seen	7.53	High	Treated - Mined from 1969 to 1994
More Seldom Seen	3.2	High	Treated - Mined from 1969 to 1994
Del Park	1.31	High	Treated
Warren	0.87	High	Treated
Bennetts	0.88	High	Treated
Lewis	2.01	High	Treated
West Cameron	1.87	Intermediate	Treated – 33% cleared for mining
Central Cameron	4.73	Intermediate	Treated – 27% cleared for mining
Gordon	2.13	Intermediate	Control - Forest
Jayrup	45.8	Intermediate	Treated – 13% cleared for mining
Tunnel Road	2.07	Low	Treated
Bee Farm Road	1.81	Low	Control - Forest

2.4.5 Paired catchment studies into impacts of reforestation on hydrology

A paired catchment study was initiated in the Upper Hay River to assess the impact of reforestation on streamflow and stream salinity (Table 2.9). The Willmay catchment was selected as the 'control' catchment on which the land use did not change during the experiment. In 1993, the other two catchments were planted with *Eucalyptus globulus*. Trees were planted on 45 ha (40%) of the cleared areas of the Pardelup subcatchment and on 15 ha (17%) of the Barrama catchment.

Table 2.9 Research catchments investigating water impacts of reforestation

Catchment	Catchment area (km^2)	Rainfall zone	Treatment
Willmay	1.53	Low	Control - Pasture
Barrama	0.94	Low	17% planted with *E. globulus*
Pardellup	1.49	Low	40% planted with *E. globulus*

In addition to the three catchments identified in Table 2.9, catchment studies into reforestation have been undertaken at Batalling Creek and Maringee Farm gauged catchments in the Collie River Basin and at Balingup Brook tributary at Padbury Road in the Blackwood River Basin (Schofield *et al.* 1989a).

2.5 Integrative approaches

The paired catchment studies in Western Australia have a relatively unique aspect which is their long term nature. For example the Seldom Seen and More Seldom Seen mined catchments have been operated for nearly 50 years.

These catchment studies have been complemented with steady state modelling (Mauger 1996; Schofield 1988), whilst in recent years complex dynamic models have been developed (Bari and Croton 2000; Croton and Barry 2001).

A comparison of the paired catchment studies across SWWA with the Australian studies is examined in Chapter 5, whilst Chapter 6 examines the interrelationship of forests, water and climate.

Chapter 3 SWWA forest hydrology – hillslope studies

The understanding of infiltration characteristics of jarrah forest soils and soil water dynamics is important for understanding streamflow generation given the unique nature of the jarrah forest soil profile. As described in Chapter 1, the jarrah forest has a very deep soil profile and consequently a very large soil water holding capacity which influences streamflow generation. Quantifying the infiltration characteristics and effects on soil water storage will assist in evaluating the impact of forest disturbance and climate variability and change on forest water values.

This chapter consists of three studies and is each is given in whole as a separate section within the chapter. The topics addressed are:

3.1 Infiltration characteristics

3.2 Soil water dynamics coarse textured soils

3.3 Soil water dynamics fine textured soils

The overall aim was to understand the key components that influence streamflow generation in jarrah forest catchments. This would assist evaluating the impact of forest disturbance on water values.

All experiments were undertaken in the Del Park catchment which is described in each section.

3.1 Infiltration characteristics of a jarrah forest soil[2]

3.1.1 Abstract

Large infiltration ponds (10-15 m^2) were used, in conjunction with a ring infiltrometer and a well permeameter, to determine the infiltration characteristics of a complex lateritic soil profile in the jarrah (*Eucalyptus marginata*) forest of SWWA. Simultaneous measurements of soil water content and soil water potential allowed a description of the infiltration and redistribution in the soil profile. The infiltration ponds effectively measured the conductivity of a subsurface lateritic duricrust which was found to have a relatively high saturated hydraulic conductivity (K_s) of 2.7 m d^{-1}, despite its apparently massive and extensive nature. Removal of the topsoil identified large (~ 1 m^2) infilled holes penetrating the duricrust over about 6% by area. Measurements indicated that these large 'holes' had a high K_s value (~ 10 m d^{-1}), whereas the remaining duricrust had a lower K_s value (~ 2 m d^{-1}). These results have implications for probable maximum flood design calculations and assessing the hydrological impact of extensive open-cut bauxite mining.

3.1.2 Introduction

Interest in characterizing the infiltration behaviour of jarrah forest soils stems from two sources. Firstly, nearly all of the surface water supply catchments of SWWA are situated in the jarrah forest. Based on current probable maximum flood estimations, the dam spillways of these catchments need upgrading at a cost

[2] Published as: **Ruprecht**, J.K. and Schofield, N.J., 1993. Infiltration characteristics of a complex lateritic hillslope. *Hydrological Processes*, 7: 87-97.

of about $US 67 million. A substantial part of this money could potentially be saved if more realistic infiltration data were available. Significant parts of these catchments are covered by shallow lateritic duricrust sheets which have been assumed to have low permeability.

Secondly, extensive open-cut mining for bauxite in the jarrah forest involves excavation typically to a depth of 5 m and completely removes the lateritic duricrust (Steering Committee for Research on Land Use and Water Supply 1985). This process is likely to strongly affect the near surface soil water dynamics and streamflow generation mechanisms.

Only surface saturated hydraulic conductivities *(K_s)* of native jarrah forest soils have been reported previously (Sharma *et al.* 1987a). Their site was characterized by texture contrast or duplex soils in which sandy high conductivity soils overly low conductivity clay subsoils (Bettenay *et al.* 1980). At this site the K_s values of the surface soils were high (mean 21 m d^{-1}). No measurements of subsoil K_s values were taken, but ponding occurred at its surface for extensive periods during winter (Ruprecht and Schofield 1989a), indicating its low conductivity. Moore *et al.* (1986) measured K_s values with a mean of approximately 6.5 m d^{-1} for forest soils of eastern NSW, Australia, whereas (Talsma and Hallam 1980) in the ACT, Australia measured a mean K_s value of 13.8 m d^{-1}.

In this study the approach was to construct large infiltration ponds to measure the subsoil K_s value. Supplementary information on the topsoil saturated hydraulic conductivities was obtained with a ring infiltrometer and a constant head well permeameter.

3.1.3 Site description

The infiltration experiments were located within the Del Park research catchment, which is situated approximately 100 km south of Perth and 5 km NNW of Dwellingup (Fig. 3.1), within the high rainfall zone (>1100 mm yr^{-1}) of the northern jarrah forest. The climate of the region is characterized by high winter (June-August) rainfall and hot dry summers (December-February). The average catchment annual rainfall (1975-89) was 1170 mm, 76% of which fell from May to October. The long term (1938-86) annual rainfall from a rain gauge at Dwellingup was 1276 mm. The average annual pan evaporation (1969-86) at Dwellingup was 1505 mm.

The catchment is located within the south-western province of the Archaean Yilgarn Block. The catchment bedrock is generally granitic with a number of intruding dykes composed mainly of dolerite. *In situ* weathering of the basement rocks has led to the development of a deep (>20 m) lateritic profile on the hillslope.

The elevation of the study area varies from 277 to 307 m, with an average inclination of 10%. The vegetation of the study area has been described in detail by Ruprecht *et al.* (1987) and was defined as an open forest dominated by jarrah *(Eucalyptus marginata)* and marri *(Corymbia calophylla)* with a well defined middle storey of sheoak *(Allocasuarina fraseriana)* and banksia *(Banksia grandis).*

Soils of the middle and upper slopes of the study area consist of gravelly sands overlying the lateritic duricrust, known locally as caprock. The depth to caprock ranges from 0 to 1 m and averages 40 cm. Dark organic-rich soil is evident to about 10 cm depth. The caprock structure varies from massive to unconsolidated and is typically about 2 m thick. Underlying the caprock are substantial depths (>10 m) of sandy or silty material which is commonly called the pallid zone which in places increases in clay content with depth.

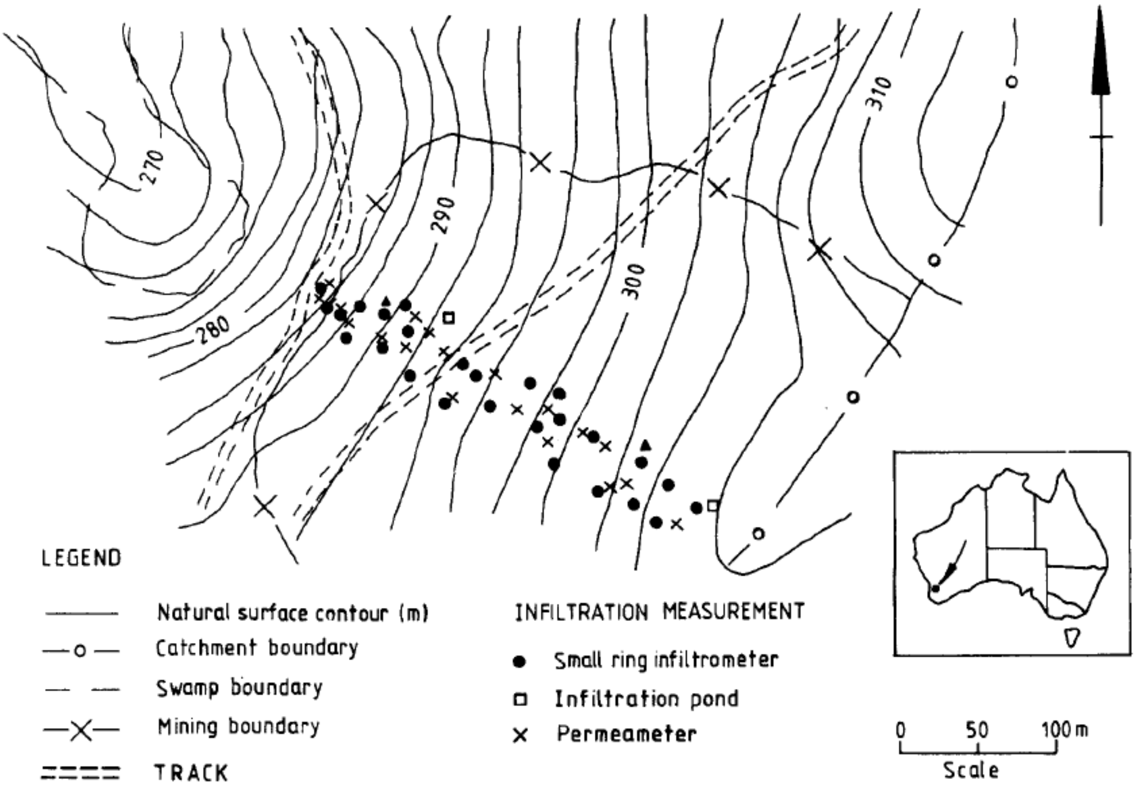


Figure 3.1. Location map of Del Park research catchment and infiltration experiments

3.1.4 Experimental procedure

The scale at which measurements are taken should be greater than the 'representative elementary volume' (Bear 1979; Williams and Bonell 1988; Youngs 1983) to overcome macroscale variability. Youngs (1983) has suggested that instead of measuring soil physical properties at a point, bulk properties of the whole system are more appropriate for many purposes. In this study it was suspected that the use of traditional infiltrometers and permeameters would be inappropriate for the soil profile with the duricrust layer. The approach taken was to construct large infiltration ponds (~ 10 m^2). Measurements were also taken of the topsoil with a ring infiltrometer and well permeameter for comparison.

Large infiltration ponds

Two large-scale infiltration experiments were carried out within the study area (Fig. 3.1). The area of the midslope pond was 11.3 m^2 and the upslope pond 14.6 m^2. The infiltration ponds were constructed by excavation of a trench with a narrow backhoe bucket (300 mm wide) down to the caprock. The perimeter depth to caprock was approximately 0.40 m. Owing to the variable ‘micro-relief’ of the caprock, final topsoil excavation took place by hand. The sides of the trench not supported by soil were formed with timber and concrete poured into the formwork.

A large tank (1.1 m^3) with a constant head device was set upslope of the infiltration pond. The constant head/water level device consists of an air inlet and water outlet tubes fitted to the base of the tank and supported on a frame within the infiltration pond (Fig. 3.2). A water tanker was at hand to periodically refill the storage tank.

The experimental sites incorporated neutron access tubes and mercury manometer tensiometers. The neutron moisture meter calibration and analysis of soil water data for these soils has been carried out previously (Ruprecht and Schofield 1990a). Methods of measuring and analysing the neutron moisture meter and tensiometer data are described by Ruprecht and Schofield (1990b). Before the infiltration pond experiment began, the soil water content and soil water potentials were measured. During the experiment the volume of water infiltrated was measured approximately every 10 minutes, whereas the soil water potential measurements were taken approximately every 30 minutes and the soil water content measured hourly. An infiltration pond experiment in progress is shown in Fig. 3.2.

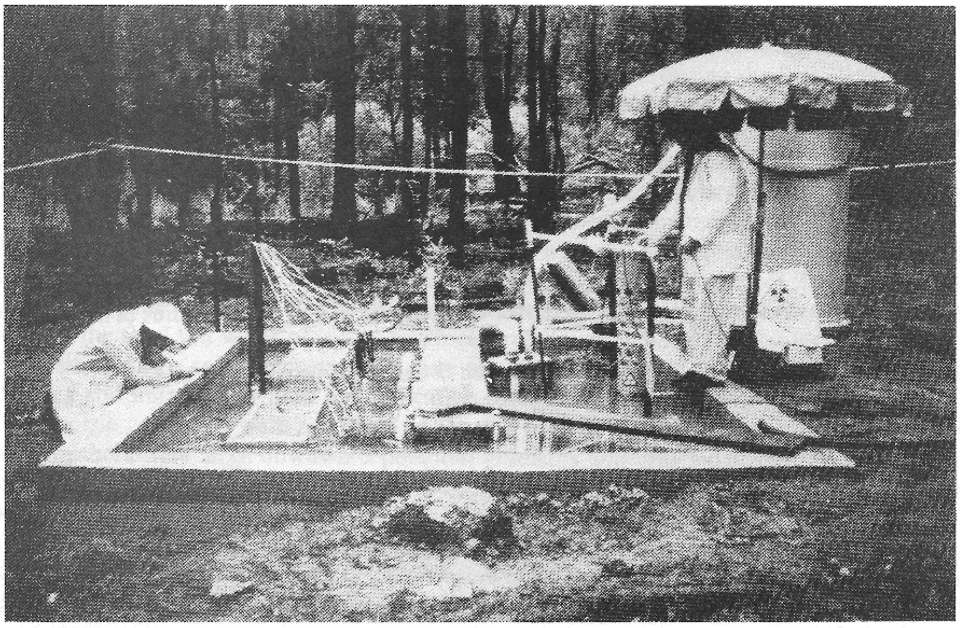


Figure 3.2. Infiltration pond experiment in progress

Infiltrometer

The ring infiltrometer consisted of a 300 mm diameter ring and a constant head device attached to a cylindrical storage tank similar to, but smaller than, that used in the infiltration pond experiments.

Permeameter

A constant head well permeameter specifically designed for permeable soils (Bell and Schofield 1990) was used to measure the saturated hydraulic conductivity of the surface soils overlying the caprock. The method involves forming a cylindrical hole with a soil driver, filling it with water to a prescribed depth, maintaining a constant head of water in the hole and measuring the inflow rate to the hole.

3.1.5 Theory

Infiltration ponds and ring infiltrometer

Steady-state infiltration into a uniform soil is described by (Philip 1969) as:

$$\nabla\,(D\,\nabla\,\theta) = \delta K/\delta z \qquad (1)$$

where θ is the volumetric water content ($m^3\ m^{-3}$), *D* is the soil water diffusivity ($m^2\ s^{-1}$) and K is the hydraulic conductivity ($m\ s^{-1}$).

Usually both *K* and D depend strongly on θ and the matric potential φ**.** By assuming (Philip 1969)

$$K = K_s \exp(\alpha\ \varphi) \qquad (2)$$

then

$$\phi = \int_{\Phi_i}^{\Phi} K(\Phi)\mathrm{d}\Phi \qquad (3)$$

where φ is the matric flux potential and accounts for capillary effects in the unsaturated zone, $\mathbf{\Phi}$ is the pressure head, $\mathbf{\Phi_i}$ is the initial pressure head and a is the sorptive number.

Solving Equation (1) for the boundary condition of a shallow circular pond, Wooding (1968) found that the complex series solution was almost identical to the simple expression (Scotter *et al.* 1982)

$$q = \alpha\phi + 4\phi/\pi r \tag{4}$$

where q is the flux density into the soil averaged over the source area and $\boldsymbol{r}$ is the radius of the shallow circular pond.

If the soil is initially at 'field capacity' or drier, then equation 4 can be approximated to

$$\phi = K/\alpha \tag{5}$$

Thus Equation (4) can be rewritten as:

$$q = K[1 + 4/(\pi\alpha r)] \tag{6}$$

For the above theory to be useful, q must approach the steady-state value within a reasonable time period (Scotter *et al.* 1982).

Measurement of a is typically performed by using a twin ring infiltrometer (Scotter *et al.* 1982) or by measuring the initial and final water contents and sorptivity (White and Scully 1987).

Field soils in eastern Australia examined by a range of authors (Scotter *et al.* 1982; Talsma 1987; White and Scully 1987) gave values of a ranging from 2 to

92 m^{-1} and did not indicate a strong dependence on soil texture. For the experiments reported here a value of *a* of 20 m^{-1} was assumed for the reasons described by Bell and Schofield (1990).

As the infiltration ponds were so large (10 to 15 m^2), the influence of capillarity on the estimate of saturated hydraulic conductivity *(K_s)* was assumed to be negligible. Consequently, the long term flow-rate was assumed to approximate the value of K_s.

Permeameter

Of the various theoretical analyses of the constant head permeameter, Bell and Schofield (1990) found that of Philip (1985) as reliable as any. Philip (1985) carried out an analytical saturated-unsaturated flow analysis and obtained the solution for K_s as (equation 7):

$$K_s/Q = (\gamma^2 - 1)^{-1/2}\left[\frac{4.117a^2\gamma(1-\gamma^{-2})}{\ln[\gamma + (\gamma^2-1)^{1/2}] - (1-\gamma^{-2})^{1/2}} + \frac{4.028a + 2.517a\gamma^{-1}}{0.5\alpha \ln[\gamma + (\gamma^2-1)^{1/2}]}\right]^{-1} \qquad (7)$$

where *Q* is the steady discharge rate from the borehole, a is the radius of the borehole, y = *H/a, a* is the sorptive number and *H* is the constant head of water above the base of the borehole. Details of parameter values and a suitable procedure for estimating K_s are given by Bell and Schofield (1990) and have been followed here.

3.1.6 Results

The infiltration experiments on the study area were carried out in July (mid-winter) 1987.

Large infiltration ponds

The first large infiltration pond experiment was located at a midslope site and was five hours long. There was a significant reduction in the infiltration rate from 12 to 2-4 m d^{-1} during the first 10 minutes (Fig. 3.3a). After this initial response the infiltration rate gradually decreased to 2.0 m d^{-1} after four hours. Variability in the infiltration rate may have been caused by fluctuations in the head of water in the pond during re-filling of the tank. The cumulative infiltration for the midslope site is shown in Fig. 3.3b. There is a relatively constant slope in the cumulative infiltration line until 200 minutes. At this point the infiltration rate appears to be declining slightly. The long term infiltration rate was estimated to be 2.1 m d^{-1}.

Before beginning infiltration there was a downward total soil water potential gradient through the soil profile, although from 1 to 2 m there was only a small downward gradient (Fig. 3.4). The osmotic potential was considered negligible due to the low solute concentrations in the shallow soils. An increase in the total soil water potential was measured down to 1.8 m, 18 minutes after beginning the experiment. Positive matric soil water potentials were measured 30 minutes after beginning infiltration from 1 to 2 m. During and immediately after the infiltration experiment, low downward potential gradients and large positive matric potentials were measured in the 1-2 m depth interval. At depths less than 1 m and greater than 2 m there were large downward potential gradients and only small positive matric potentials were evident. This process of a saturation zone occurring within the caprock layer is consistent with the natural soil water dynamics during winter rainfall (see Section 3.2).

The soil water content data (Fig 3.5) show most change occurring from 3 to 5 m. The maximum volumetric water contents reached during this infiltration experiment were generally less than the saturated water content based on porosity values for the soil (Ruprecht and Schofield 1990a).

The second large infiltration pond experiment was located at an upslope site (Fig. 3.1) and was three hours 30 minutes long. The infiltration rate fluctuated between 3 and 6 m d^{-1} initially and then settled to approximately 3.4 m d^{-1} (Fig. 3.3). The maximum soil water content profile (Fig. 3.5) highlights the small increases in soil water content through the duricrust layer and the large increases below. The long term infiltration rate for the upslope infiltration site was estimated to be 3.5 m d^{-1}.

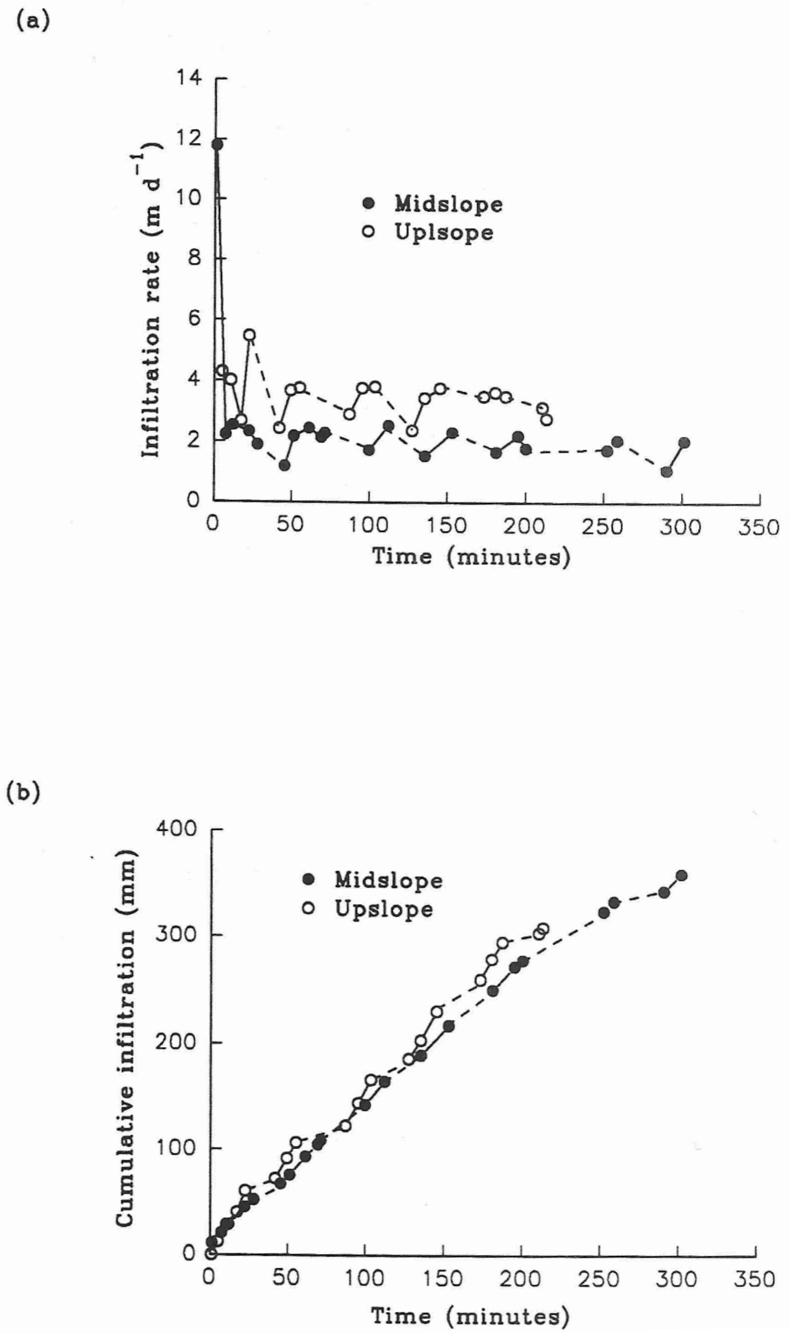


Figure 3.3. Infiltration pond experiments. (a) Infiltration rate; (b) cumulative infiltration

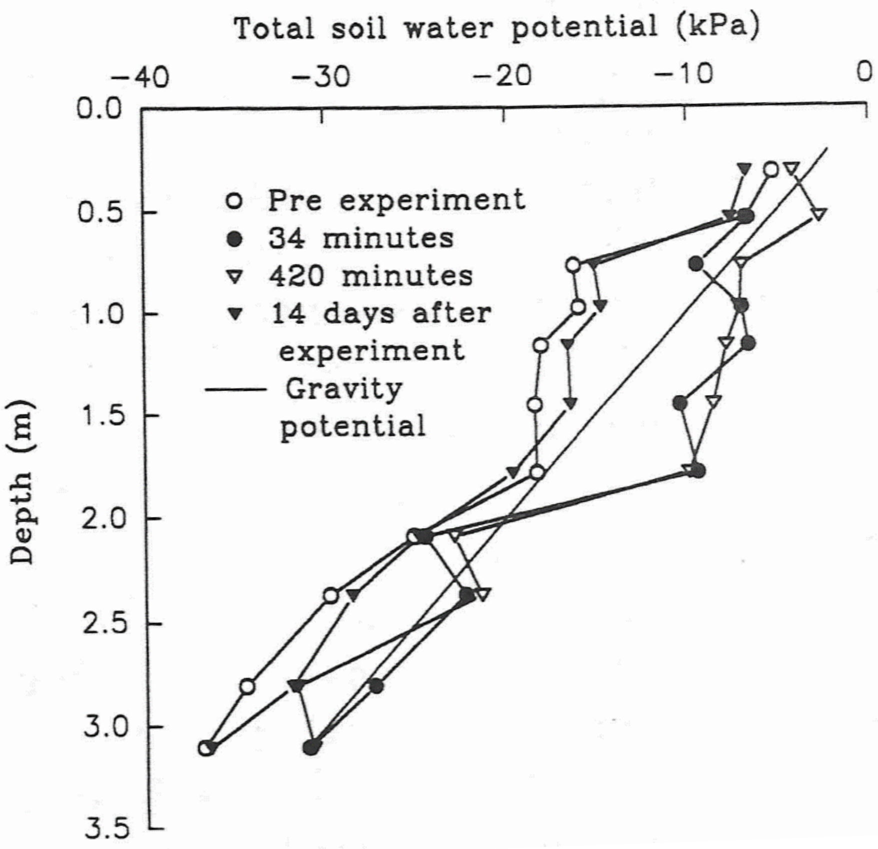
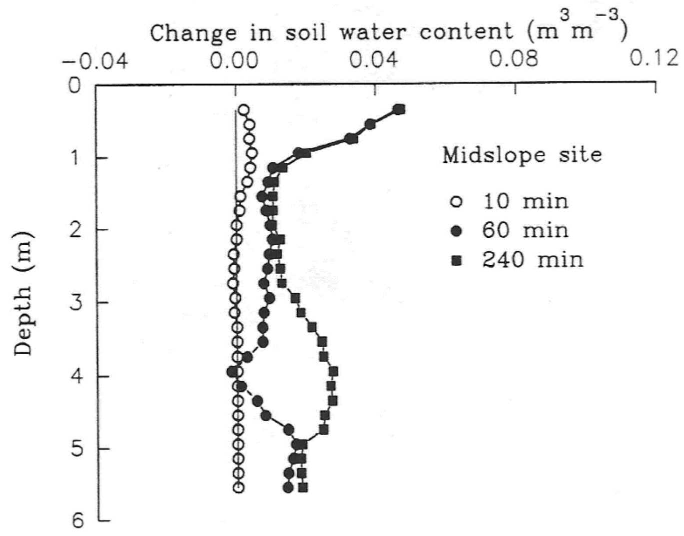


Figure 3.4. Soil water potential measurements during infiltration pond experiment at the midslope site

(a)



(b)

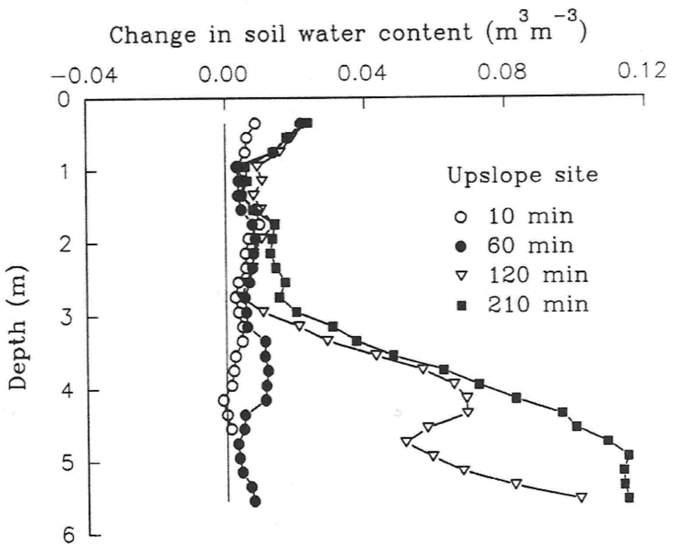


Figure 3.5. Maximum increases in soil water content for infiltration pond experiments

Infiltration experiments on large infilled holes in the caprock layer

At both infiltration ponds the surface soils above the caprock were removed by using a high pressure water hose, exposing the caprock layer (Fig. 3.6). The caprock was predominantly a cohesive layer, but at both sites there was a 0.8-

1.6 m diameter 'hole' penetrating through the caprock layer but infilled with coarse gravel and rocks. Infiltration experiments were conducted on these holes at both sites using the constant head device. The results from these measurements (Table 3.1), when compared with the infiltration pond results, show that the holes in the caprock provide 30-36% of the total infiltration flux but are only 5-15% of the area, respectively.

Based on these measurements the caprock was considered to consist of two basic elements. One was the massive caprock layer with a K_s value of 2 m d^{-1} and the second was the large infilled holes with a K_s value of 10 m d^{-1}. Together these two elements made up the caprock layer with a K_s value of 2.7 m d^{-1}. Further exploration of the representative nature of the subsurface soils observed within the infiltration ponds was undertaken by water-stripping the surface soils of an area cleared of vegetation in preparation for surface mining for bauxite. The presence of infilled holes within the caprock layer was found to be fairly common.

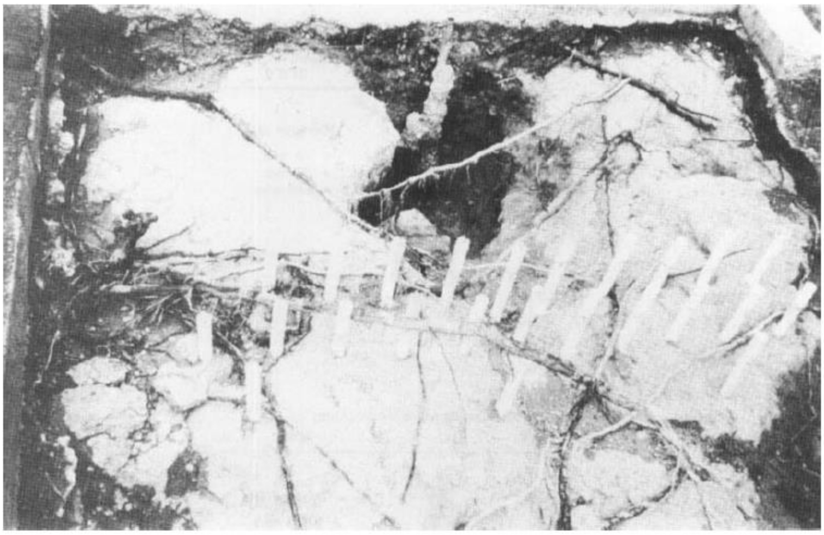


Figure 3.6. Top of caprock exposed by the removal of overburden exposed at the midslope infiltration pond site

Table 3.1 Results from constant head infiltration experiments in 'holes' in the caprock

Site	Infiltration rate		Area	
	K_s (m d^{-1})	(m^3 h^{-1})	(%)[(1)]	(%)[(2)]
Midslope	12	0.29	30	5
Upslope	8.6	0.77	36	15

(1) Percentage of infiltration rate from holes compared with infiltration from infiltration pond.
(2) Percentage of surface area of holes compared with infiltration pond.

However, as a result of the physical difficulties in stripping the surface soils from large areas, only a qualitative picture of the extent of large 'holes' in the caprock layer could be gained by this method.

The spatial extent of caprock on the hillslope was estimated by sampling the full length at the hillslope with a penetrometer. The caprock was found to occur on the ridge and 400 m downslope to within 50 m of the swamp at the base of the slope. The sampled depths to caprock had a mean value of 0.4 m. The maximum sampling depth allowed was 2 m. The presence of rocks and gravel within the holes found at the two infiltration sites indicates that caprock may not exist even if a 'strike' is recorded. The number of samples >1 m was approximately 6%. This estimate of the extent of holes in the caprock is underestimated due to there being no way to determine if caprock or a loose rock was sampled.

Infiltrometer and well permeameter

Twenty-eight ring infiltrometer and 20 well permeameter experiments were carried out on the hillslope, as shown in Fig. 3.1. Using Equation (6) for the ring infiltrometer and Equation (7) for the well permeameter, values of K_s were calculated and are summarized in Table 3.2; the distribution of values is shown in Fig. 3.7.

Although the mean and median values of K_s were similar for the two techniques, the distributions show some differences. The ring infiltrometer data show a typical log normal distribution, whereas the well permeameter data appear to show a bi-modal distribution. A mean topsoil K_s value was taken as 13 m d^{-1} from the results in Table 3.2.

Table 3.2 Comparison of estimates of K_s for the surface soils

	No of experiments	K_s (m d^{-1})		
		mean	median	sd[(1)]
Ring infiltrometer	28	13.1	10.9	9.2
Well infiltrometer	22	12.7	10.6	6.9

(1) Standard deviation

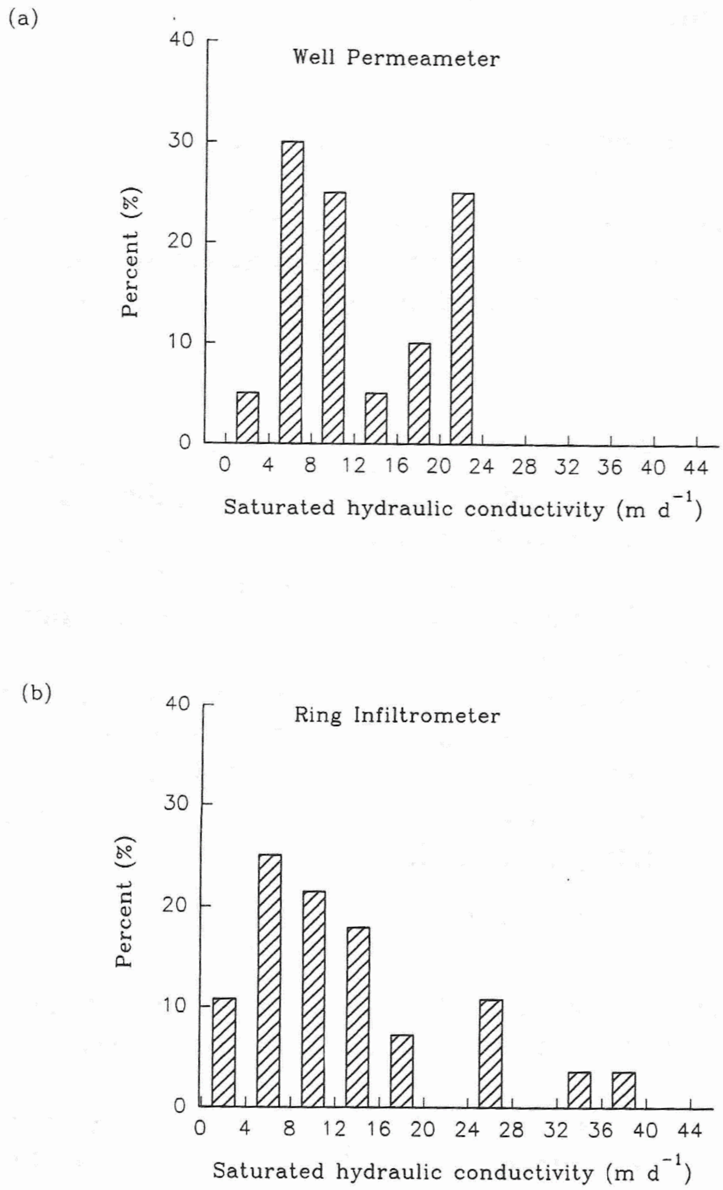


Figure 3.7. Distribution of measured values of *K*, for the surface soils. (a) well permeameter; (b) ring infiltrometer

3.1.7 Discussion

Infiltration experiment methodology

The infiltration pond experiments were able to provide information on the saturated hydraulic conductivity of the subsurface caprock layer within the study area. The values at the two sites tested were of the same order. However, it should be recognized that the nature of the caprock can vary significantly from site to site (Shearer and Tippett 1989) and consequently caution should be exercised in generalizing the results.

K_s value for caprock

The large infiltration ponds, with their walls bedded on the caprock, essentially measured the permeability of the caprock at a scale considered realistic for natural conditions. The two estimates of K_s were similar and unexpectedly high for the caprock layer. The relatively small increases in soil water content during the infiltration experiments were considered to be due to the low infiltration capacity of the caprock.

K_s value for topsoil

Estimates of the mean K_s value for the topsoil were similar for the infiltrometer ring and permeameter (13 m d^{-1}). These measurements were of the same order but a little lower than those obtained by Sharma *et al.* (1987a) for sandy surface soils in the Collie catchment, 100 km south of this site. The estimates of the mean K_s value of the topsoil were also of similar magnitude to that measured in eastern Australia (Moore *et al.* 1986; Talsma and Hallam 1980).

Soil water flow mechanisms

The measurement of infiltration characteristics at this site can lead to a better understanding of soil water flow processes. The topsoil has a high K_s value (~ 13 m d^{-1}), whereas the underlying caprock K_s value is an order of magnitude lower. Nevertheless, a K_s value of ~ 2.4 m d^{-1} for the caprock implies that the infiltration capacity of the soil profile would be very rarely exceeded by rainfall intensity in this region. This explains the observed general absence of saturated throughflow and overland flow at this site (Schofield *et al.* 1985).

Macroporosity of the caprock

The caprock appeared to be characterized by macroporosity at two scales. The large (~ 1 m^2) infilled holes were observed at both infiltration ponds and covered 10% of the combined areas. Stripping of other sites nearby also suggested a high frequency of such holes. However, sampling with a penetrometer located a smaller frequency of large holes. This is probably because the penetrometer hit sizeable loose rocks within some of the holes. The estimated K_s value for the holes was close to that of the topsoil and suggests that infilling by the topsoil may have occurred.

The second scale of macroporosity appears to be at the small root diameter (~ 1 cm) or less. Small jarrah roots were observed to penetrate the caprock at a number of locations. This small scale macroporosity accounts for the relatively high K_s value (2.0 m d^{-1}) of the visibly massive caprock.

The total macroporosity of the caprock available to water appears to be small from observations of change in water content (0.015 m^3 m^{-3}) during pond infiltration.

3.1.8 Conclusions

The upper soil profile of the jarrah forest consists of two distinct horizons of differing saturated hydraulic conductivity. The sandy gravel topsoil, with a K_s value of ~ 13 m d^{-1}, overlies a lateritic duricrust with a K_s value of ~ 2.7 m d^{-1}.

The high K_s value of the duricrust was attributed to two scales of macroporosity, one being large (~ 1 m^2) infilled holes (K_s ~ 10 m d^{-1}) at an estimated frequency of 6-10% by area, and the other being <1 cm^2, leading to a duricrust K_s value of 2.4 m d^{-1}. The high values for K_s explained the general absence of significant saturated throughflow or overland flow at this site.

The large infiltration ponds were considered to be a viable technique for estimating the duricrust K_s value. The results of this study can be applied to the estimation of probable maximum floods and assessing the hydrological impacts of bauxite mining.

This study provided an understanding of how water infiltrates into the soil profile. The next two sections in this chapter study the soil water dynamics after the water has infiltrated into the soil profile.

3.2 Seasonal soil water dynamics - coarse-textured soil[3]

3.2.1 Abstract

Seasonal soil water dynamics were measured on a hillslope transect in the jarrah forest of SWWA over the period 1984-86 using mercury manometer tensiometers,

[3] Published as: **Ruprecht, J.K**. and Schofield, N.J. 1990. Seasonal soil water dynamics in the jarrah forest, Western Australia. I: Results from a hillslope transect with coarse-textured soil profiles. *Hydrological Processes*, 4: 241-258.

gypsum blocks, and a neutron moisture meter. The soil water potential gradients indicated downward vertical drainage flux through winter and spring. There was generally a change to an upwards flux in early summer which was sustained through to autumn. A shallow ephemeral saturation zone was identified in and above a duricrust layer, lasting up to three days after heavy, late winter rainfall. The annual maximum to minimum unsaturated soil water storage on the hillslope was approximately 400 mm to 6 m depth and 480 mm to 15 m depth. This did not change significantly in years of substantially different winter rainfall. The magnitude of seasonal soil water storage was similar to other forested areas with deep soil profiles. The depth of observable infiltration was dependent on annual rainfall. This was consistent with the observation that groundwater levels responded to rainfall over the whole hillslope in wet years but only responded on the lower slopes in dry years. The average summer drying rate of the soil profile to 6 m depth of 3.5 mm day^{-1} was within the range of values reported for forests elsewhere. In late summer, following an extended drought period, the drying rate decreased downslope but increased midslope.

3.2.2 Introduction

This hillslope hydrology study involved the measurement of various components of the hydrological cycle before, during, and after bauxite mining with the aim of gaining a quantitative understanding of the hydrological impact over time. This study describes the seasonal soil water dynamics on the undisturbed (pre-mining) hillslope over the period 1984-86.

Unsaturated soil water dynamics plays an important role in the hydrology of the jarrah forest. Firstly the deeply weathered lateritic profiles (up to 70 m deep) provide large soil water stores (Sharma *et al.* 1987b) which lead to low streamflow volumes of 0 to 30% of annual rainfall (Schofield *et al.* 1989b) and exceptionally sluggish storm hydrographs (Loh 1974). Secondly the nature of the movement of water through the unsaturated zone to groundwater can strongly affect groundwater levels. Rising groundwater levels following forest clearing have been shown to lead to large increases in streamflow (Section 4.1) and stream salinity (Schofield *et al.* 1988; Williamson *et al.* 1987). Measurements of soil water dynamics beneath native forest have taken place at other locations in Western Australia (Sharma *et al.* 1987b), in eastern Australia (Nicolls *et al.* 1982; Talsma and Gardner 1986) and at other locations around the world (Hodnett and Bell 1986; Hudson 1988).

3.2.3 Site description

The hillslope transect under study is located within the Del Park catchment, approximately 100 km south of Perth and 5 km NNW of Dwellingup (Fig. 3.8). The climate of the region is characterized by high winter rainfall and hot, dry summers. The catchment annual rainfall over the period 1975 to 1986 was 1128 mm, 76% of which fell from May to October. The long-term annual rainfall (1938 to 86) at Dwellingup was 1276 mm. The average annual pan evaporation (1969 to 86) at Dwellingup was 1505 mm.

The hillslope transect has an elevation varying from 277 to 307 m above sea level (Fig. 3.9). Valley slopes are generally moderate with an average inclination of 10%. The hillslope transect had a mean slope of 9% and was 400 m in length. The bedrock is generally granitic with a number of intruding dolerite dykes.

In situ weathering of the basement rocks has led to the development of a deep lateritic soil profile ranging from 13 to 48 m depth on the hillslope (Chapter 2). The surface soils on the middle and upper slopes of the transect were gravelly sands of depth ranging from 0-1 m and averaging 40 cm. Dark organic-rich material was evident to about 10 cm depth. These surface soils overlied a duricrust layer known locally as caprock. The caprock was extensive and massive but was perforated by a moderate frequency of large (0.5-2 m^2) holes (Chapter 3, Section 2), infilled with coarse gravel and rocks. The caprock was typically about 2 m thick. Underlying the caprock were substantial depths (>10 m) of sandy or silty material which in places increased in clay content with depth. On the lower slopes, caprock was absent and the upper 3 m of the soil profile was typically colluvial or alluvial gravelly soils. Below 3 m, a finer-textured pallid zone occurred.

The vegetation on the hillslope transect has been described in detail by Ruprecht *et al.* (1987). The overstorey vegetation was predominantly jarrah and marri on the middle and upper sites and bullich *(E. megacarpa)* on the lower slope. Throughout the hillslope the middlestorey was dominated by bull banksia (*B. grandis*) and sheoak (*Allocasuarina fraseriana*). The forest density of the hillslope was calculated in terms of basal area (29 m^2 ha^{-1}), projected canopy cover (45%) and leaf area index (1.4). The rooting habit of the dominant jarrah is to proliferate radial roots above the caprock (about 90% of root volume) and to drop vertical sinker roots (Kimber 1974) through 'holes' in the caprock. Roots have been observed in this area to 40 m depth (Dell *et al.* 1983). The valley floor is occupied by a permanent swamp.

3.2.4 Experimental procedure

Soil water potential measurement

Mercury manometer tensiometers and gypsum blocks were installed at three locations (lower slope - T1, middle slope - T2, upper slope - T3) on the hillslope transect (Fig. 3.9). The tensiometers and gypsum blocks were installed in sets of twelve, from depths of approximately 0.25 to 3.0 m in 0.25 m intervals. In each set the tensiometers and gypsum blocks were placed in two parallel lines on the contour, spaced 25 cm apart. They were all installed in May 1984. Tensiometers accurately measured total soil water potential in the range 0 to ~ 75 kPa. At lower (more negative) soil water potentials, vapour bubbles nucleated within the instrument and disturbed the readings. Gypsum blocks were used to extend the range of tensiometers. Their accuracy was somewhat poorer than the tensiometers but they were useful in measuring very low matric potentials during summer.

The gypsum blocks were read by measuring electrical resistance with an AC bridge. The effect of temperature on the gypsum blocks was determined by installing adjacent thermistors. The change in electrical resistance was found to range from 3% per degree at 15°C to 1.6% per degree at 30°C (K. Baldock personal communication 1988).

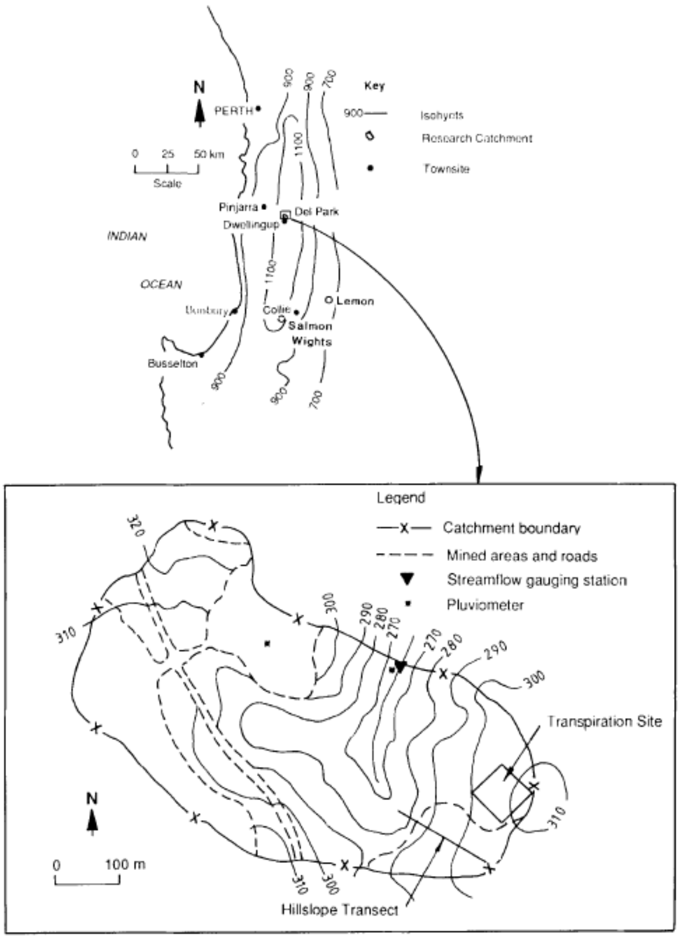


Figure 3.8 Location and descriptive map of Del Park catchment

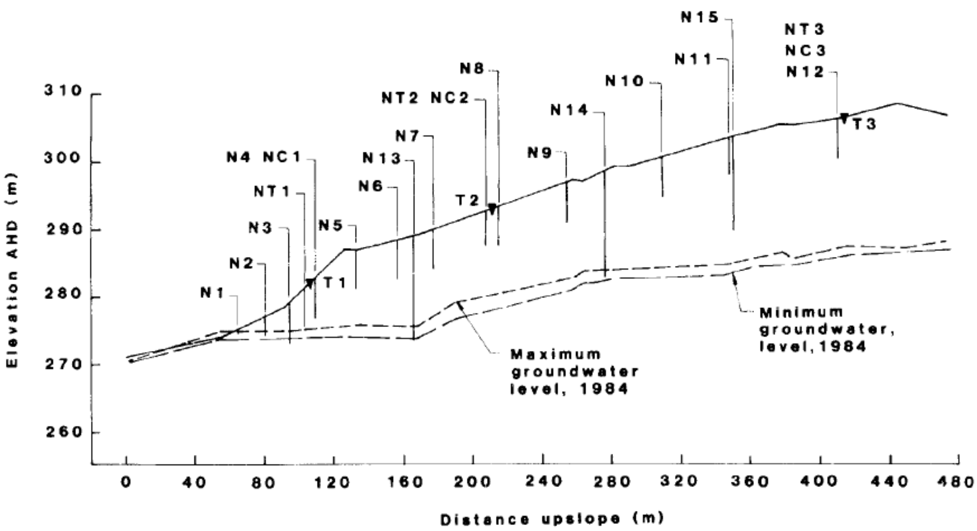


Figure 3.9 Cross-section of hillslope transect showing neutron tube (N) and tensiometer (T) sites and groundwater levels

Soil water content measurement

Soil water content was measured with a Didcot neutron moisture meter. A total of 21 access tubes were installed on the hillslope transect as shown in Fig. 3.9. The access tube locations ranged from adjacent to the valley floor to within 40 m of the catchment divide. Of the 21 access tubes, 18 were installed to 6 m depth in May 1984 and three were installed to approximately 14 m depth in March 1986.

Soil water monitoring commenced on 9 May 1984. The tensiometer and gypsum block data were collected on the same day as the neutron moisture meter data. The monitoring frequency was initially weekly but after one year was extended to two weekly. Depth sampling for the 6 m access tubes was at 0.20 m intervals, starting at 0.15 m below ground level. For the deeper access tubes, measurements were taken from an initial depth of 0.15 m at 0.30 m depth intervals. Calibration of the neutron moisture meter was based on measurements of changes in volumetric soil water content and neutron count ratio from two sampling times (Ruprecht and Schofield 1990a). Further details on the installation and operation of the soil water instrumentation are given by Hodnett and Schofield (1985) and Schofield *et al.* (1985).

3.2.5 Results

Rainfall and evaporation

Rainfall over the period of study (1984-1986) was on average 7% below the long-term (1938-1986) average of 1276 mm (see Table 3.3). The annual rainfall at Del Park was above average in 1984 and below the long-term average for 1985 and 1986. Pan evaporation, measured at Dwellingup, was slightly below average for the three years, particularly in 1986, which was also a low rainfall year. The seasonal rainfall pattern was typically characterized by an initial large rainfall event in March or April.

Table 3.3 Rainfall and evaporation data

Station	Period of record	Average over record	1984	1985	1986	Average over study
509 263(1) rainfall (mm)	1975-1986	1128	1338	1126	1051	1175
009 538(2) rainfall (mm)	1938-1986	1276	1324	1202	1052	1193
009 538(2) evaporation (mm)	1969-1986	1505	1411	1445	1295	1384

(1) Located adjacent to gauging station for the Del Park catchment
(2) Location at Dwellingup, *5* km SSE of Del Park catchment

At the beginning of June there was generally the onset of frontal winter rains. These rain bearing fronts usually persisted to August or September and were followed by scattered rain to November. Extended dry periods with occasional summer storms were experienced from December to March. The maximum periods with no rain ranged from 19 days in 1984/85 to 33 days in 1986/87. During those summer periods pan evaporation averaged 7 mm day^{-1}.

Soil water potential

The seasonal variation in total soil water potential at the lower slope site T1 is shown in Fig. 3.10a for the winter-spring period 6 June 1985-4 December 1985. From mid-July to October the profile was dominated by gravity drainage. By 5 November 1985 there was a significant reduction in total water potential with a zero gradient horizon from 0.6 to 2.6 m. By 4 December 1985 a significant upwards potential gradient had occurred over most of the profile.

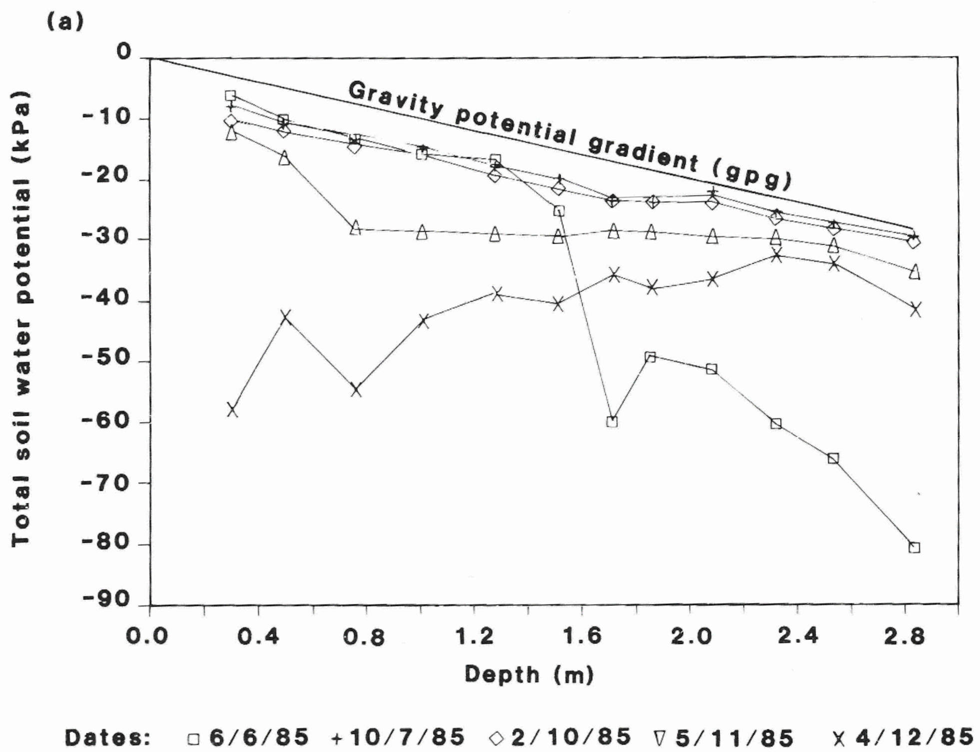
The gypsum block data for the summer of 1985/86 are shown in Fig. 3.10b. On 13 February 1986 there was an upwards potential gradient over most of the profile. However, by 19 March 1986 a downward potential gradient had formed to 1.3 m with an upwards potential gradient from 1.3 to 2.2 m. This reverse in potential gradient was due to 136.8 mm of rain falling between 13 February 1986 and 19 March 1986. The soil water potentials at site T2 did not decrease below the tensiometer range for nearly all the period of record.

The matric soil water potentials were particularly close to zero in 1984. Fig. 3.11a depicts the total water potential profiles from winter to the beginning of summer of 1986. The total water potential profile for 6 June 1985 indicates a sink at approximately 1.0 m with the top 1.0 m having a downward gradient, while from 1.0 to 1.8 m there was an upward gradient. This sink was faintly evident on 10 June 1985. By the 5 November 1985 a downward potential gradient had been restored over the soil profile. For the 4 December 1985 and 8 January 1986 there was an upward gradient down to 0.8 m.

Through the winter and spring of 1986 there was a downward gradient with matric potentials close to zero (Fig. 3.11b). There was evidence of a section with close to a zero gradient at approximately 1 m depth. There was also evidence of positive soil water potentials on 3 September 1986 (Fig. 3.11b). These potentials are considered to be due to heavy rainfall (92.6 mm) over the preceding 14 days. However there was little rain (0.8 mm) on the preceding three days. This implies that the positive matric soil water potentials were sustained for at least three days.

The soil water potential data at site T3 (not shown) were similar to site T2 with relatively small changes in soil water potential through a year (Ruprecht and

Schofield 1989b). The small changes in soil water potential at T2 and T3, compared to T1, are attributed to the dominance of caprock from 0.5-2.5 m at T2 and T3. This material has only a limited capacity for transmitting water and this is principally via large (~ 2 m^2) holes filled with gravels and medium-sized rocks and via smaller root channels and rock fractures (see Section 3.1.6 and Fig. 3.6).



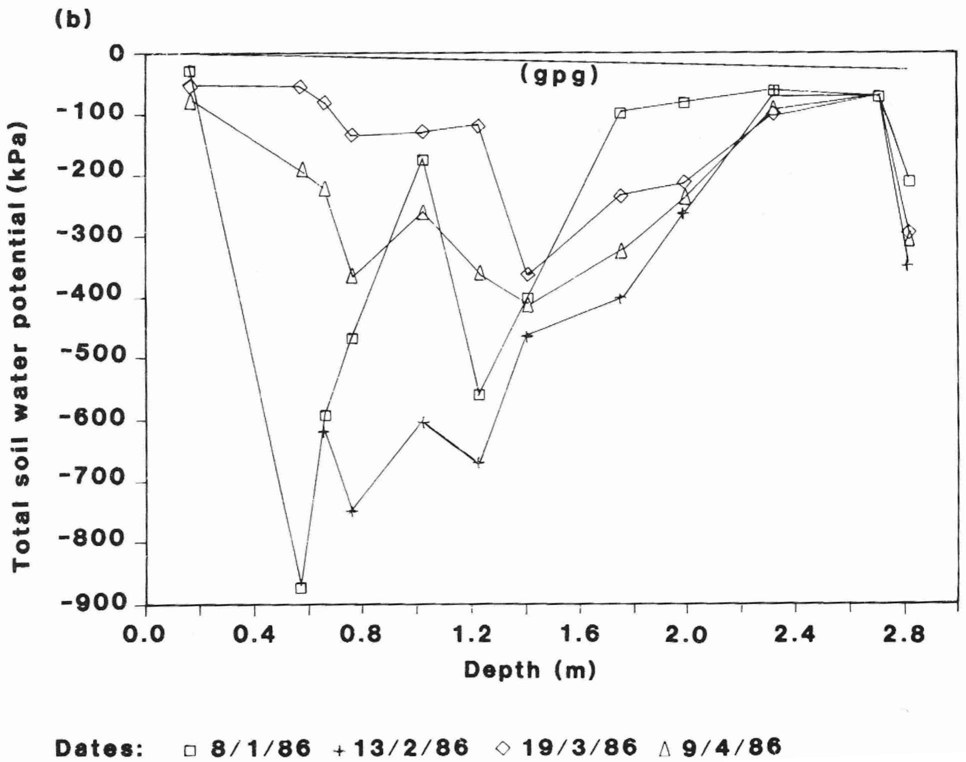


Figure 3.10 Soil water potential at TI: (a) 6/6/85 to 4/12/85; (b) 8/1/86 to 9/4/86

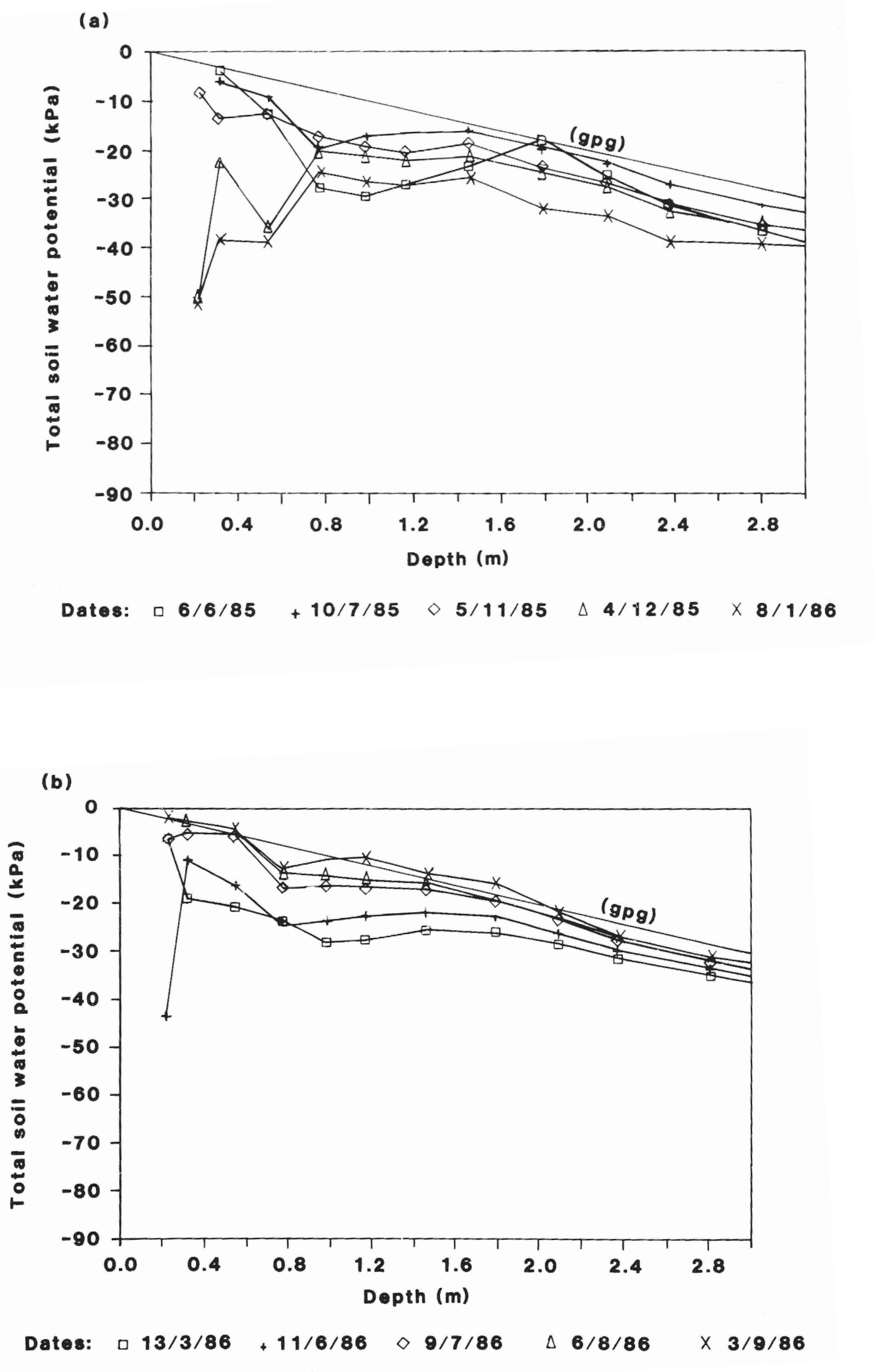


Figure 3.11. Soil water potential at T2: (a) 6/6/85 to 18/1/86; (b) 13/3/86 to 3/9/86

Variations through a storm period

The tensiometers were monitored through a storm event in order to gain some understanding of the short time-scale dynamic behaviour of the soil water potential profile. The results were reported by (Schofield *et al.* 1985). At site TI, the main impact of the rainfall was to reduce matric potentials and bring the total potential close to the gravitational potential line. Saturation in this profile, however, was not evident. At site T2 a significant part of the profile (0.3 to 0.4 m, 0.8 to 1.8 m and 2.2 to 2.8 m) became saturated. Saturation of the 0.3 to 0.4 m layer persisted through the day and probably represents ponding on the surface of the caprock.

The matric soil water potentials remained positive at 1.8 m depth for approximately three days. As with set T2, T3 showed three sections of the soil profile that were saturated. However, in this profile the saturation was short-lived. It is clear from these results that caprock may become ‘saturated’ during high intensity storm events. This most likely occurs because the relatively few water transmitting paths through the caprock become filled.

Soil water content response of individual access tubes.

The progression of a wetting front through winter and spring was apparent at NT1. The wetting front reached 2.8 m by 11 June 1986, 3.5 m by 9 July 1986, and 5.5 m by 3 September 1986. In the period from 1 May 1986 to 11 June 1986, 165 mm of rainfall was recorded, and the lower slope (which includes NT1) had an increase in soil water storage of 104 mm. This increase only occurred in the top 3 m. After approximately 350 mm of rain and three months of winter an increase in soil water content at depths greater than 5 m was observed. The reduction in soil water content at NT1 in the top 4.5 m commenced on 2 October 1985 and below 4.5 began on 4 December 1985.

The maximum and minimum soil water content profiles over this period of measurement for one of the deeper access tubes (N15) on an upslope location showed a significant change in soil water content down to 9 m (Fig. 3.12). The water table at this location was approximately at 18 m depth.

Seasonal variation in soil water storage

A clear seasonal pattern was apparent in soil water storage for the lower, middle, and upper locations (Fig. 3.14). Soil water storage was seen to peak in August to September and reach a minimum during the February to May period, depending on year and location. The three sites had maxima which coincided for 1985 at approximately mid-August. However, in 1986 the maximum for the lower slope was a month delayed (end of September) compared to the middle and upper slope locations (end of August). This is consistent with the observation of McCord and Stephens (1987) of a significant lateral component to unsaturated flow on a hillslope with no apparent sublayers (other than caprock) of lower permeability. The lower slope location showed a greater decrease in soil water storage than either the middle or upper sites during summer, particularly 1986/87. For all three sites, soil water storage decrease accelerated at the beginning of December.

This is considered to be due to increasing evaporative demand with the onset of summer. During late summer early autumn (April-May), the soil water storage remained approximately constant. This is probably due to the autumn rainfall balancing evaporation and drainage (Fig. 3.13).

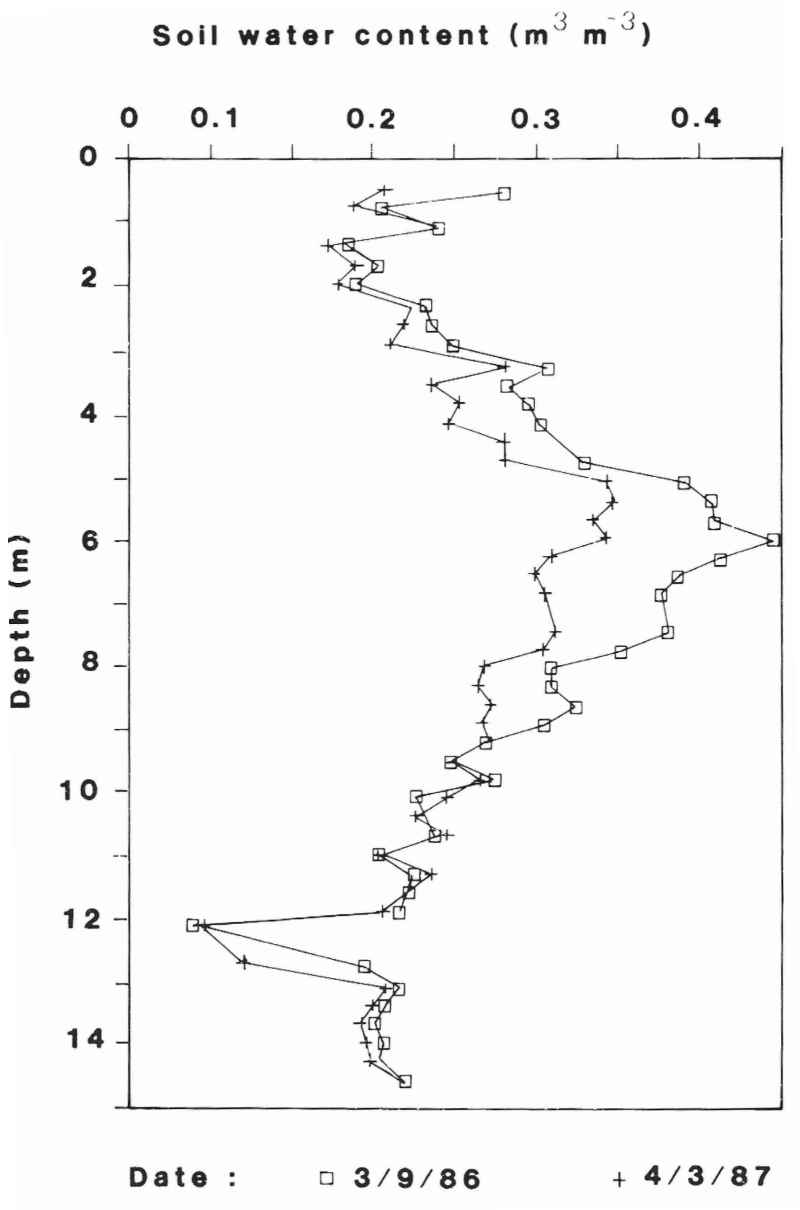


Figure 3.12. Maximum and minimum soil water content profiles for access tube N15

Comparison of the soil water storage maxima for the 0 to 3 and 3 to 6 m profiles for the lower site shows a delay of approximately 2 weeks for the 3 to 6 m interval to attain a maximum water storage compared to the 0 to 3 m interval. At the midslope site the 0 to 3 and 3 to 6 m intervals were similar in their seasonal behaviour although the 0 to 3 m interval had a significantly smaller amplitude of soil water storage change, indicating the lower water storage capacity of caprock (Fig. 3.15). Below 6 m depth there was an increasing delay in time before maximum soil water storage occurred. Also the amplitude of soil water storage was reduced. The rate of progression of the 'wetting front', based on observations of the maxima, was on average 0.2 m day^{-1}.

However, as the 'wetting front' progressed deeper into the profile the rate of progression slowed down, decreasing from 0.23 m day^{-1} from 0 to 3 to 3 to 6 m to 0.08 m day^{-1} from 6 to 9 to 9 to 12 m.

The upper slope shows similar soil water storage variation as the midslope. Again there does not appear to be any time lag between maxima of the 0 to 3 m and 3 to 6 m depth intervals. This is consistent with water being rapidly transmitted through but not significantly retained in the caprock layer.

In each year, the amplitudes of the middle and upper sites were similar but about 20% smaller than the lower slope (Table 3.4). The difference was attributed to the small water storage in the caprock on the middle and upper slopes, although there is the possibility of some lateral flow to the lower slopes. The soil water storage minima were generally in February or March and the maxima in September or October. A reduction in both the maxima and minima for the lower site, from 1985 to 1986, infers that there was no significant change in the amplitude. The maxima and minima for the middle and upper locations did not change significantly from 1985 to 1986.

The annual rainfall within the catchment for 1985 and 1986 was 1126 mm and 1051 mm, respectively. However, the rainfall from March to October was 1023 mm and 874 mm, respectively. Despite there being l50 mm less rainfall from March to October in 1986 there was only a slight change (increase) in the amplitude of the soil water storage. This implies that in below average rainfall years the reduction in rainfall is likely to mostly be reflected in a reduction in streamflow, groundwater recharge, or transpiration.

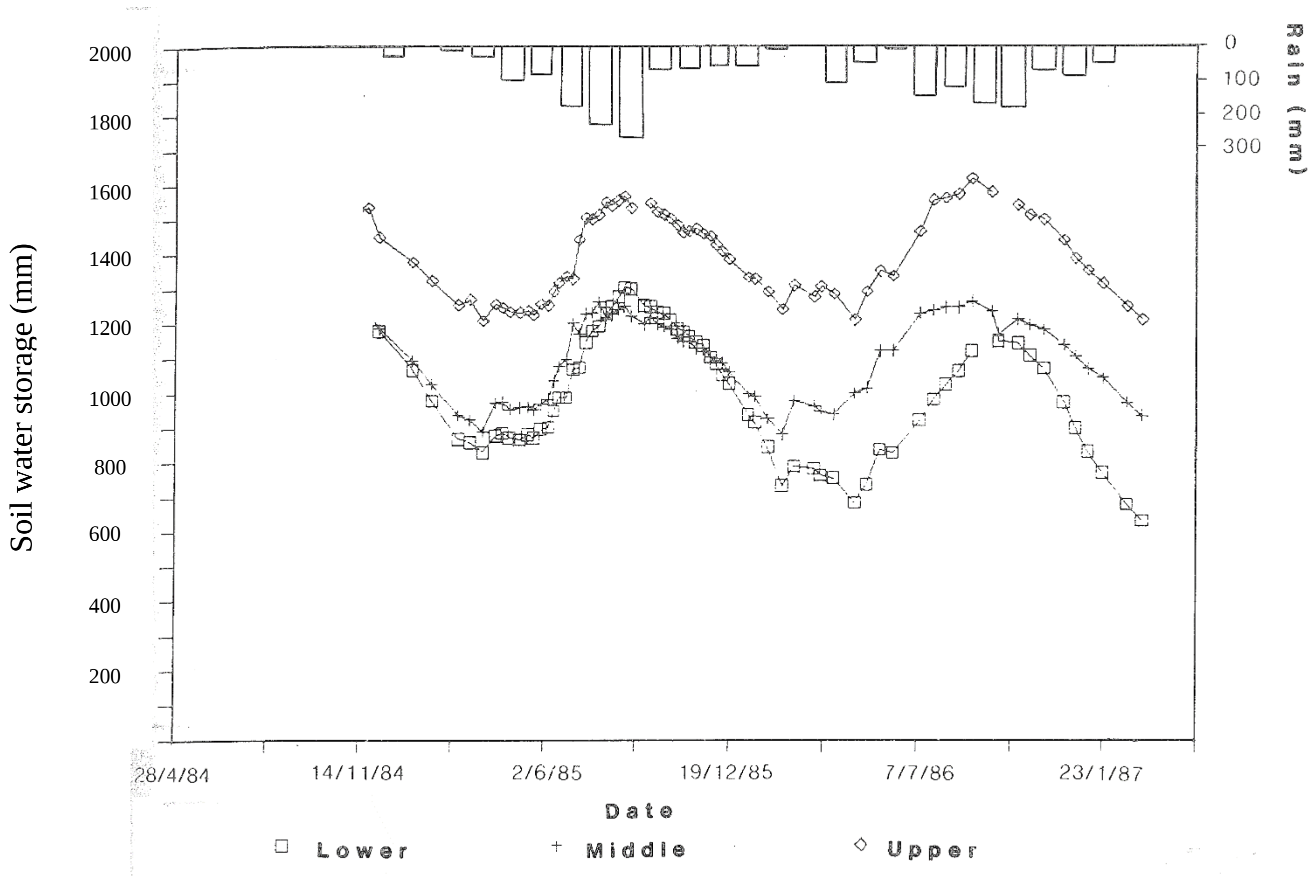


Figure 3.13 Time trends of soil water storage for the lower, middle, and upper sites

A reduction in streamflow from 205 mm in 1985 to 155 mm in 1986 was observed. It appears that the soil water storage is replenished by about the same amount in both low and average rainfall years.

At the midslope location two access tubes were installed to 14 m depth. The amplitudes of soil water storage for 3 m depth intervals at this location (Fig. 3.14), show that changes in soil water storage occur down to the 9 to 12 m zone, for the relatively low rainfall year of 1986. The total unsaturated soil water change for 0 to 6 m was 375 mm and for 0 to 15 m was 480 mm. The soil water storage change from 6 to 12 m accounted for approximately 22% of the total soil water storage change.

Table 3.4 Minima, maxima and amplitudes of soil water content to 6 m depth for downslope, midslope and upslope sites

	1985			1986			Ave
	Min (mm)	Max (mm)	Amplitude (mm)	Min (mm)	Max (mm)	Amplitude (mm)	Amplitude (mm)
Downslope	828	1298	470	685	1146	461	466
Midslope	886	1247	361	880	1262	382	372
Upslope	1205	1565	360	1239	1614	375	368
Hillslope average			397			407	402

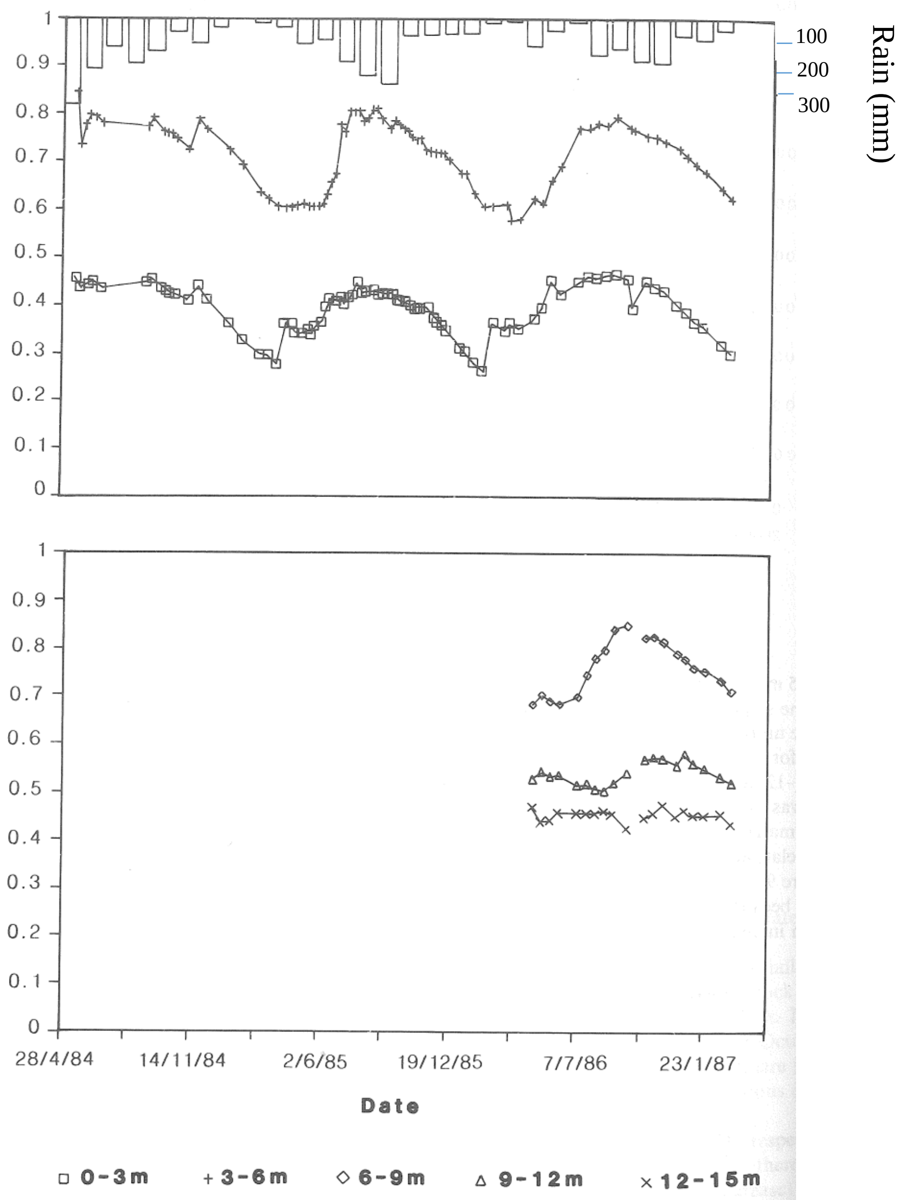


Figure 3.14 Time trend of soil water storage at the midslope site for 3 m depth intervals

The relationship between rainfall and the change in soil water storage for the three hillslope sites is shown in Fig. 3.15. The rates of change in soil water storage were plotted instead of absolute change in soil water storage because of the varying time interval from 17 to 42 days. The soil water storage of the midslope location increased during the initial winter rainfalls. By about mid-July, there was no further increase in soil water storage and any large additional rainfall events must have infiltrated rapidly beyond 6 m. This occurred earlier for the middle location than the upper location. At the upper location, the period of water accumulation during winter to 6 m depth continued typically to early September. At the lower slope, water storage increased to September or October. This reflects the differing natures of these profiles. In general, the soil water storage rate is approximately 4 mm day^{-1} lower than the rainfall rate for the lower slope. This translation is considered to be a measure of the losses to the 6 m profile by evapotranspiration and drainage past 6 m. The middle and upper sites show a less consistent translation of the soil water storage to rainfall of 2.5 mm day^{-1} on average.

The 3 to 6 m layer increased in soil water storage significantly in the period 11/6 to 9/7, the 6 to 9 m layer increased from 9/7 to 3/9, and the 9 to 12 m layer increased from 1/10 to 4/11. The 12 to 15 m depth interval soil water storage did not seem to change significantly from month to month apart from a random variation. It is therefore assumed that the drying through the profile from the 4 November 1986 to 5 March 1987 represents soil water evaporation and transpiration over the dry summer period.

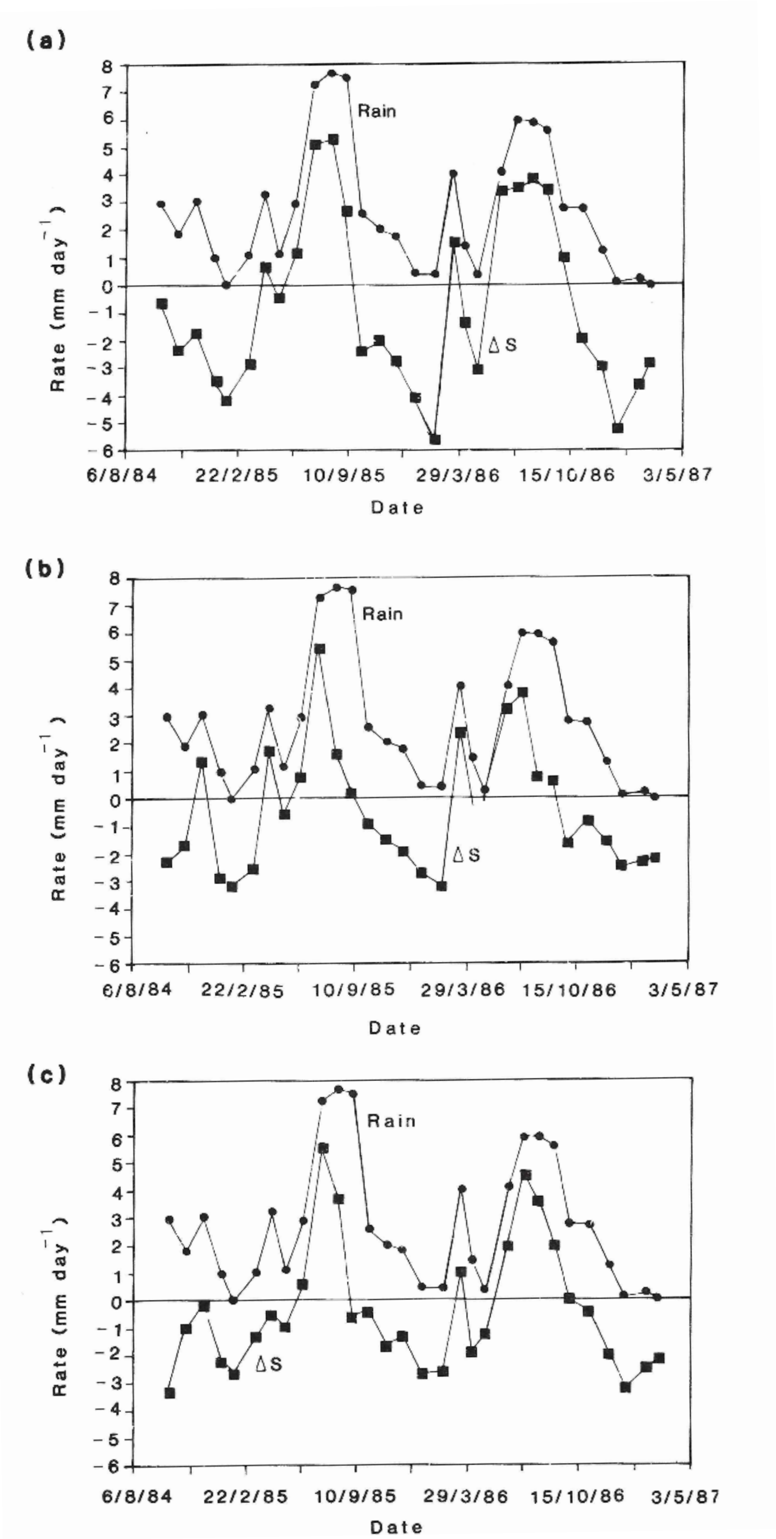


Figure 3.15 Rates of change in soil water storage compared to rainfall at (a) lower, (b) middle and (c) upper sites

The rates of change in soil water storage at the midslope location, which included the two 15 m access tubes, showed progressive wetting with depth (Fig. 3.16).

Soil water drying rates

Under conditions of negligible drainage, neutron moisture meter measurements can be used to calculate soil water drying rates (rate of decrease in soil water storage), which can be assumed to approximate the sum of soil water evaporation and transpiration. At the lower site, soil water drying rate from the top 6 m of the soil profile during the summer periods ranged from 2 to 5.7 mm day^{-1}, with an average of 3.7 mm day^{-1}. At the middle and upper sites the soil water drying rate averaged 2.5 mm day^{-1} and ranged from 1.7 to 3.2 mm day^{-1}. The 15 m soil profile monitored at the midslope location showed an average soil water drying rate during the 1986/87 summer of 3.5 mm day^{-1}, and a maximum of 4.3 mm day^{-1}.

The soil water drying rate at the lower location, as a percentage of pan evaporation, averaged 65% through the summer periods. However, at the end of the 1984/85 and 1986/87 summer periods there was a considerable reduction to about 45% of pan evaporation. The reduction in the soil water drying rate relative to pan evaporation indicates that soil water content and consequently soil water potential has reduced appreciably and reduced transpiration from this unsaturated zone.

At the midslope site, the drying rate gradually increased from 3.1 mm day^{-1} for 4 November to 10 December to 3.5 mm day^{-1} for 6/1 to 17/2 and to 4.4 mm day^{-1} for 17 February to 5 March. This increase in drying rate occurred at the end of a very long dry period which was preceded by a low rainfall winter. This indicates that transpiration at this site was not soil water limited. The pattern of drying through the profile indicates that transpiration had removed similar amounts of water per unit depth over the 0 to 9 m profile, and about half this amount from 9 to 12 m. There is no evidence at this site of water being transpired first from the upper horizon and then progressively down the profile. There was a reduction in soil water drying rate at the end of the 1986/87 summer in the top 6 m of the soil profile at the upper location. Whether this occurred over the entire unsaturated profile of 18 to 20 m is unknown.

3.2.6 Discussion

Annual soil water storage amplitudes

The annual soil water storage amplitude of 400 mm for a depth of 6 m on the Del Park hillslope is within the range of other results from forest in Australia: 215 mm in jarrah forest at Salmon catchment (Sharma 1984), 340 mm in jarrah forest at Wights catchment and 190 mm in jarrah-wandoo (E. *wandoo)* forest at Lemon catchment (Sharma *et al.* 1987b), and 480 to 680 mm in eucalypt forest at Bushrangers Creek, ACT (Talsma and Gardner 1986).

For the two years of measurement at Del Park, the annual amplitudes of soil water storage were similar (within 2.5%) despite significantly different (15%) winter rainfall. This result is contrary to that of Sharma (1984) who observed a range of 150 to 300 mm over a four year period and Talsma and Gardner (1986) whose measured amplitude varied from 480 mm in average rainfall years to 680 mm in a drought year. This difference between Del Park and the other sites cannot be readily explained.

Soil water drying rates

The average summer drying rate of 3.5 mm day^{-1} for 6 m unsaturated soil profile at Del Park is within the range of measurements for forest elsewhere: 2.2 mm day^{-1} at Salmon catchment (neutron method) (Sharma *et al.* 1987b), 2.4 mm day^{-1} in Victoria (Langford and O'Shaughnessy 1979), 4.5 mm day^{-1} in New South Wales (lysimeter) (Dunin and Aston 1984), 3.4 to 4.1 mm day^{-1} in Wisconsin (neutron method) (Sartz 1972), and 1.5 to 1.8 mm day^{-1} in the U.K. (neutron method) (Hudson 1988).

An increase in the soil water drying rate was identified on the hillslope transect during December. This coincides with increased evaporative demand as measured at Dwellingup by pan evaporation from December to February. In late summer, at the lower site on the hillslope transect, there was a significant decrease in the soil water drying rate. At this site, the depth to the water table was about 6 m. Thus at the time of large negative soil water potentials in the unsaturated zone, vegetation may have been extracting water directly from the groundwater. This argument may be supported by the observation that groundwater levels on the lower slopes decline after December. At the midslope site on the hillslope transect with an unsaturated zone of 18 m, there was no indication of transpiration being soil water limited.

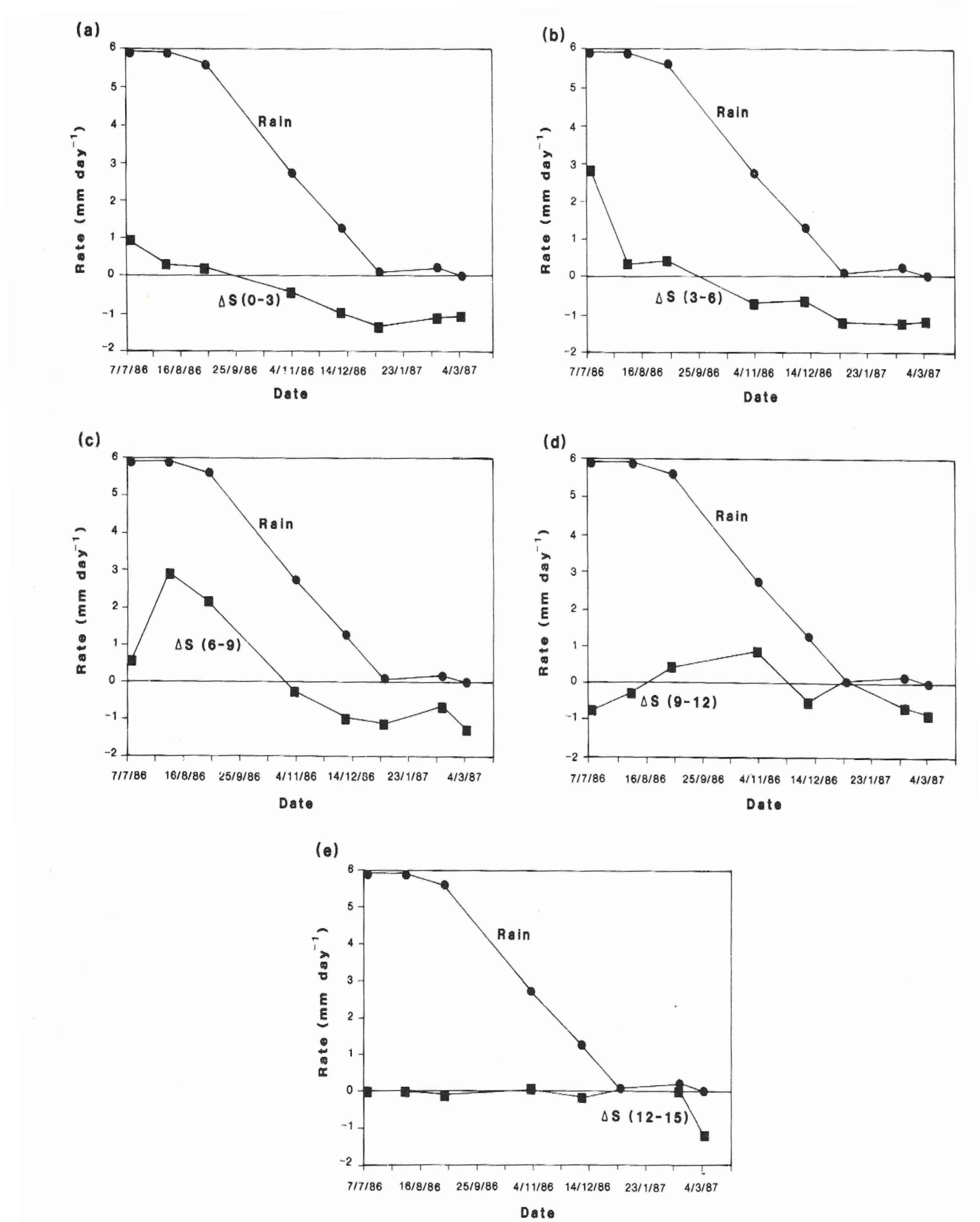


Figure 3.16 Rates of change in soil water storage compared to rainfall at the middle site for 3 m depth intervals

Pattern of soil water extraction

The pattern of summer drying through the profile indicated that transpiration removed similar amounts of water per unit depth per unit time over the 0 to 9 m profile and half this amount from 9 to 12m. This is in contrast to the observations of Nnyamah and Black (1977) for a Douglas fir forest. They found the zone of maximum root water uptake moved progressively from shallower to deeper depths as the soil dried. However, this was only for a soil profile of 0.6 m depth which is difficult to compare to a 12 m soil profile.

Areas of rapid soil water penetration

It was found on the midslope of the hillslope transect that, after a certain amount of rainfall in winter, there was no increase in soil water storage and any large additional rainfall events infiltrated rapidly beyond 6 m. Schofield (1986) found similar results when numerically simulating the water movement in a gravelly sand. The soil water content increased to a value below saturation, and remained close to this value during the input of additional rainfall. This phenomenon was attributed to the attainment of a sufficiently high unsaturated hydraulic conductivity to allow rapid transport of additional water deep into the profile. The field observations are also consistent with theoretical computations by Scotter (1978) on macropore flow under unsaturated conditions.

Distribution of groundwater recharge on the hillslope

Over the three years of study (1984 to 86) there was successively decreasing annual rainfall. Groundwater observations on the hillslope transect imply that recharge occurred over the whole transect in 1984 only. In 1985 and 1986 groundwater levels indicated that recharge was limited to the lower slopes. The two deep access tubes installed in March 1986 on the midslope and upslope did not show any soil water change at depths beyond 12 m. This may also indicate the limiting depth of infiltration although steady-state flow, rapid macropore flow or simply very low fluxes might not be detected by the neutron moisture meter method.

3.2.7 Conclusions

Soil water movement was dominated by drainage during winter and early spring over the hillslope. This changed to an upward flux in early summer, which was maintained through to autumn.

The amplitude of annual soil water storage change of 400 mm for a depth of 6 m and 480 mm for 15 m depth were in the range of values found elsewhere in Australia. Annual soil water storage amplitudes did not change significantly from year to year despite differences in rainfall.

In wetter than average years, groundwater level response to rainfall was observed over the whole hillslope transect, while in the drier than average years, the groundwater response was restricted to the lower slopes. This was consistent with observed changes in soil water content.

The soil water drying rate during summer (3.5 mm day^{-1}) was in the range of values for forest measured elsewhere in the world. The soil water drying rate at the lower site on the hillslope transect reduced significantly in late summer in the unsaturated zone, whereas the drying rate increased at the midslope location.

The caprock layer was found to become saturated and to perch water on its surface following heavy rain. The saturated conditions were generally short lived. Soil water was considered to pass through the caprock predominantly via preferred channels, some as large as 2 m^2.

The hydrology of the hillslope was characterized by the absence of surface runoff and negligible saturated throughflow. There was some evidence suggestive of unsaturated throughflow. The predominant unsaturated water movement was considered to be in the vertical plane.

3.3 Seasonal soil water dynamics - fine-textured soil profile[4]

3.3.1 Abstract

Seasonal soil water dynamics were measured at a fine-textured, upslope site within the jarrah forest of SWWA and compared to the results from a coarse-textured hillslope transect. Gravity drainage dominated during winter and early spring. This reversed in early summer and an upward potential gradient was observed to 7 m depth. A shallow ephemeral saturation zone was observed above a clay pan at 1.5 m depth. This saturation zone persisted through late winter and early spring, contrasting with the short-lived saturation in the duricrust on the hillslope transect.

The annual maximum to minimum unsaturated soil water storage was about 530 mm, 50 mm greater than the hillslope transect and higher than most values

[4] Published as: **Ruprecht, J.K**. and Schofield, N.J., 1990. Seasonal soil water dynamics in the jarrah forest, Western Australia. II: Results from a site with fine-textured soil profiles. *Hydrological Processes*, 4:259 267.

reported elsewhere in Australia. Significant soil water content changes following winter rain were generally restricted to 6 m but at one site occurred to 9 m. These depths were significantly less than the coarser-textured hillslope transect. Soil water drying rates averaged 5 mm day^{-1} during extended dry periods compared to 3.5 mm day^{-1} on the hillslope transect. The drying rate occurred uniformly through the profile until late summer when a significant decrease in the upper 3 m was observed.

3.3.2 Introduction

This site was established to support interpretation of transpiration measurements being carried out with a ventilated chamber (Schofield *et al.* 1985) and is hence called the 'transpiration site' (Fig. 3.18). Soil classification at this site revealed a fine-textured profile, possibly associated with a dolerite intrusion. Since fine-textured soil profiles are commonplace in the Darling Plateau of Western Australia (Bettenay *et al.* 1980; Churchward and Dimmock 1989), it was considered useful to contrast the soil water dynamics of this site with the coarse-textured hillslope transect (Section 3.2).

3.3.3 Site description

The general site characteristics are given in Section 3.2.3. The soil profile of the transpiration site is characterized by a gravelly topsoil (~ 50 cm deep) above an unconsolidated duricrust overlying a clay pan at 1.5 to 2.0 m depth. Below the clay pan, silty clays grade to clayey silts over the sampled depth of 18 m. The vegetation of the transpiration site is high density (basal area 47 m^2 ha^{-1}) jarrah forest, with greater proportions of jarrah and bull banksia than the hillslope transect.

3.3.4 Experimental procedure

Mercury manometer tensiometers and gypsum blocks were installed at site T4 (Fig. 3.17) between May and December 1984 in the same manner as on the hillslope transect. At site T5 a set of gypsum blocks and thermistors were installed to 12 m depth, at 0.5 m intervals over the depth range 0.5 to 1.5 m and then at approximately 1 m intervals. A total of 10 neutron access tubes were installed at the site ranging in depth from 6 to 18 m. Monitoring of the site instrumentation was the same as for the hillslope transect (Section 3.2.4).

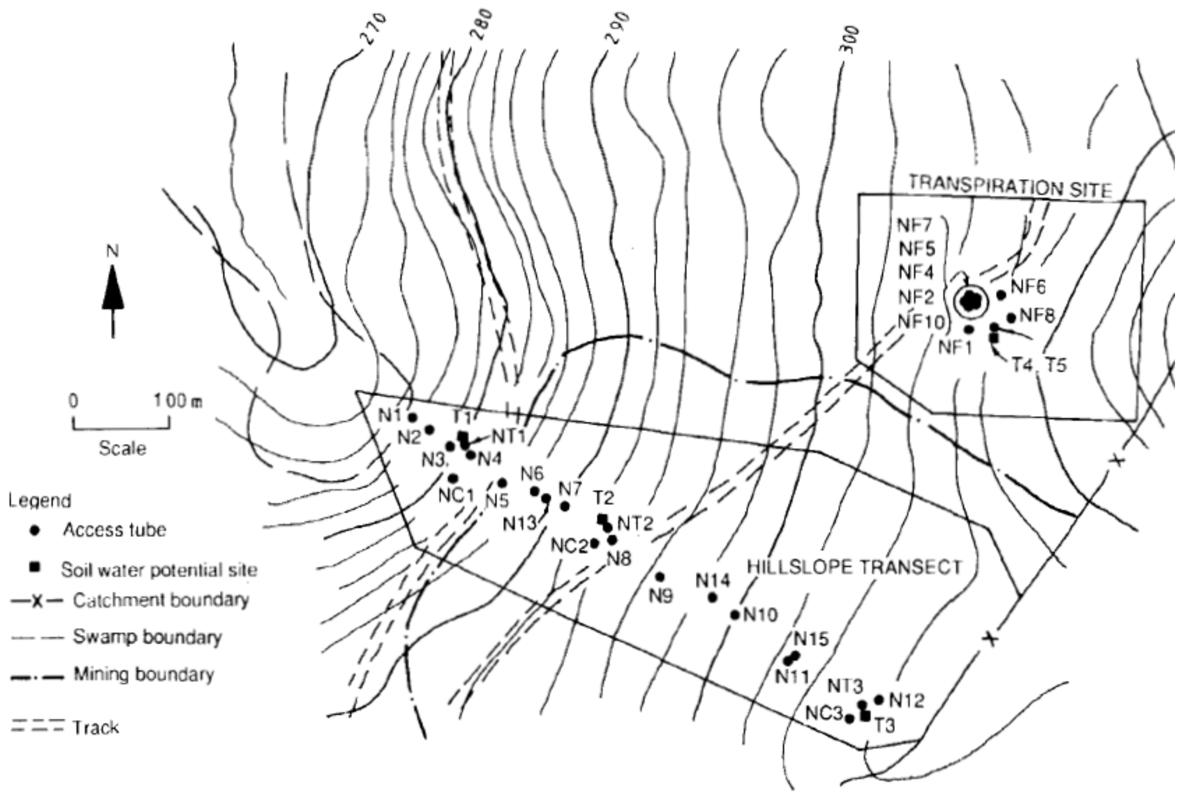


Figure 3.17 Map of instrumentation and the study sites within Del Park catchment

3.3.5 Results

Soil water potential

Through the winter there was a downward potential gradient in the soil profile at T4 (Fig. 3.18a). However the matric potentials were significantly less than zero below 2 m. From 31 July 1986 to 22 October 1986 there were positive soil water potentials from 0.7 to 2 m depth, indicating saturation. There was a substantial lowering in total soil water potential from 18 December 1985 to 31 January 1986, particularly in the top 1.5 m (Fig. 3.18b). There was also a change from a downward gradient to an upward gradient to at least 2.5 m during this period. This upward gradient was still evident at 1 May 1986. The only significant rainfall (73 mm) during the December to May summer period fell in late February and this is reflected in the soil water potentials of 26 February 1986.

At site T5 there was a downward potential gradient (Fig. 3.19) to mid-November (20 November 1985). By the end of December (30 December 1985) there was a significant upward gradient to about 6 m. From 6 m to 12 m there was a zero gradient horizon. On the 22 February 1986 there was 73 mm of rain, which significantly changed soil water potentials down to 5 m by 26 February 1986. During the summer of 1986/87 an upward potential gradient was recorded to 7.3 m depth.

Soil water content response of individual access tubes.

The progression of a 'wetting front' was evident in most access tubes as exemplified by NF2 in Fig. 3.21a. At 9 July 1986 the wetting front had reached 2.5 m, by 7 August 1986 the front had reached 4.0 m, by 3 September 1986 the front had reached 4.7 m and by 1 October 1986 had reached 5.6 m. The profile dried at a relatively constant rate down to 5 m depth (Fig. 3.20b).

A different behaviour during wetting was noted at access tube NF3, as shown in Fig. 3.21. The soil water content from 24 July 1986 to 3 September 1986 remained constant from 4 to 6 m, while from 2 to 4 m and from 6 to 9 m (limit of access tube) there was an increase of about 0.05 $m^3\ m^{-3}$. This coincides with a zone of positive soil water potential down to 2 m at T4. This access tube is the only location which had significant infiltration past 6 m.

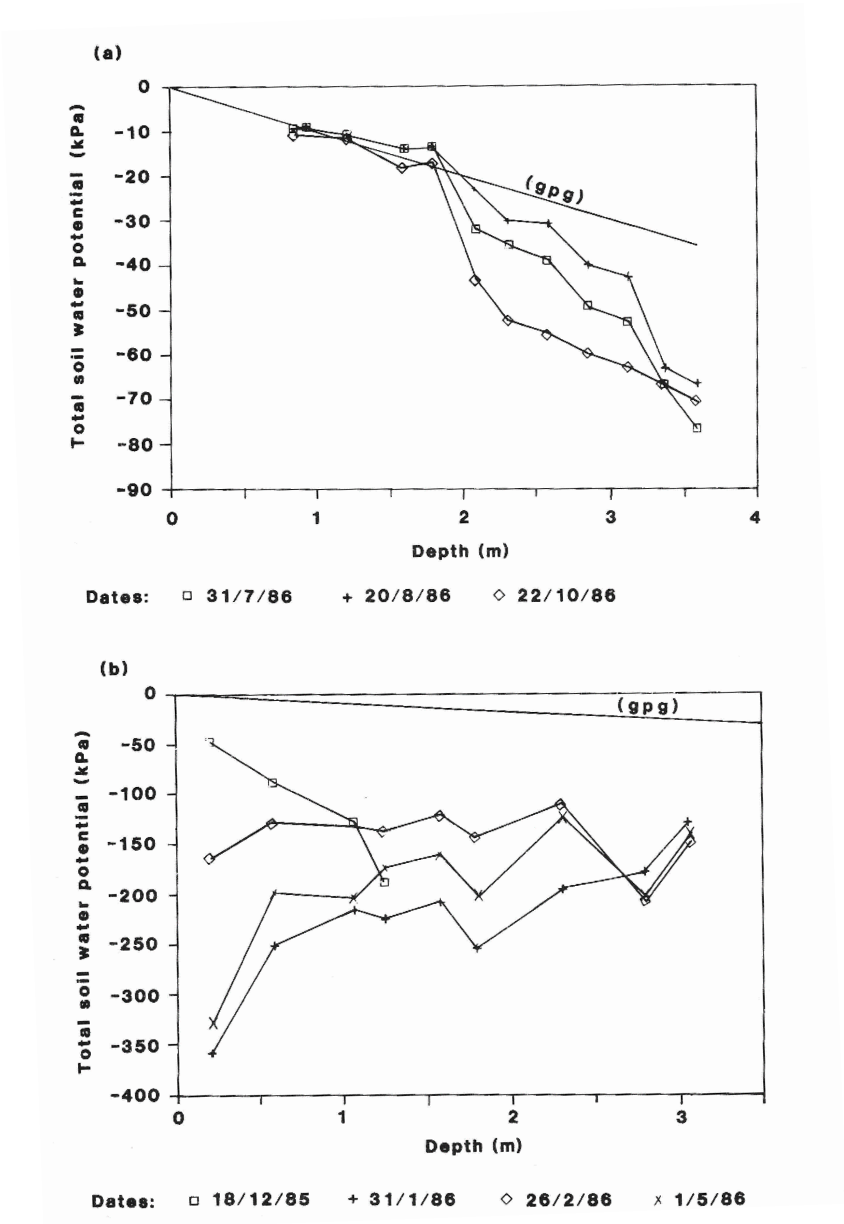


Figure 3.18 Soil water potential at Site T4: (a) 31/7/86 to 22/10/86; (b) 18/12/85 to 1/5/86

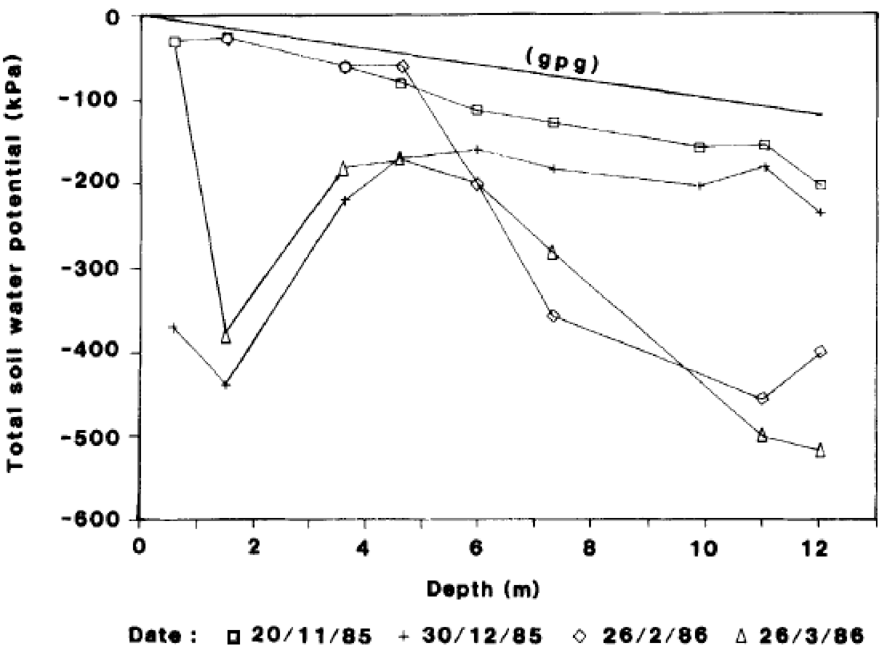


Figure 3.19 Soil water potential at Site T5: 20/11/85 to 26/3/86

Seasonal variations of soil water storage

The variation in soil water storage, averaged for all access tubes, is only discernible down to 6 m (Fig. 3.23). There is a small trend from 6 to 9 m but no pattern is detectable below 9 m. The seasonal pattern in soil water storage shows a lag of approximately 50 days between maxima of the 0 to 3 m and 3 to 6 m intervals in 1985/86 (Fig. 3.23). The 6 to 9 m depth interval shows a small response to rainfall followed by a gradual decay in soil water storage in 1986. This implies that infiltration occurred to 9 m but soil water depletion was slow. There was no lag between the peaks for the 0 to 3 m interval and the 6 to 9 m interval, which may indicate that preferential flow had occurred into this zone. There was, however, a lag in the commencement of the response of soil water storage to rainfall by the 6 to 9 m zone compared to the 0 to 3 m zone. This may be due to the time required for saturation to occur above the clay pan to allow preferred flow to take place.

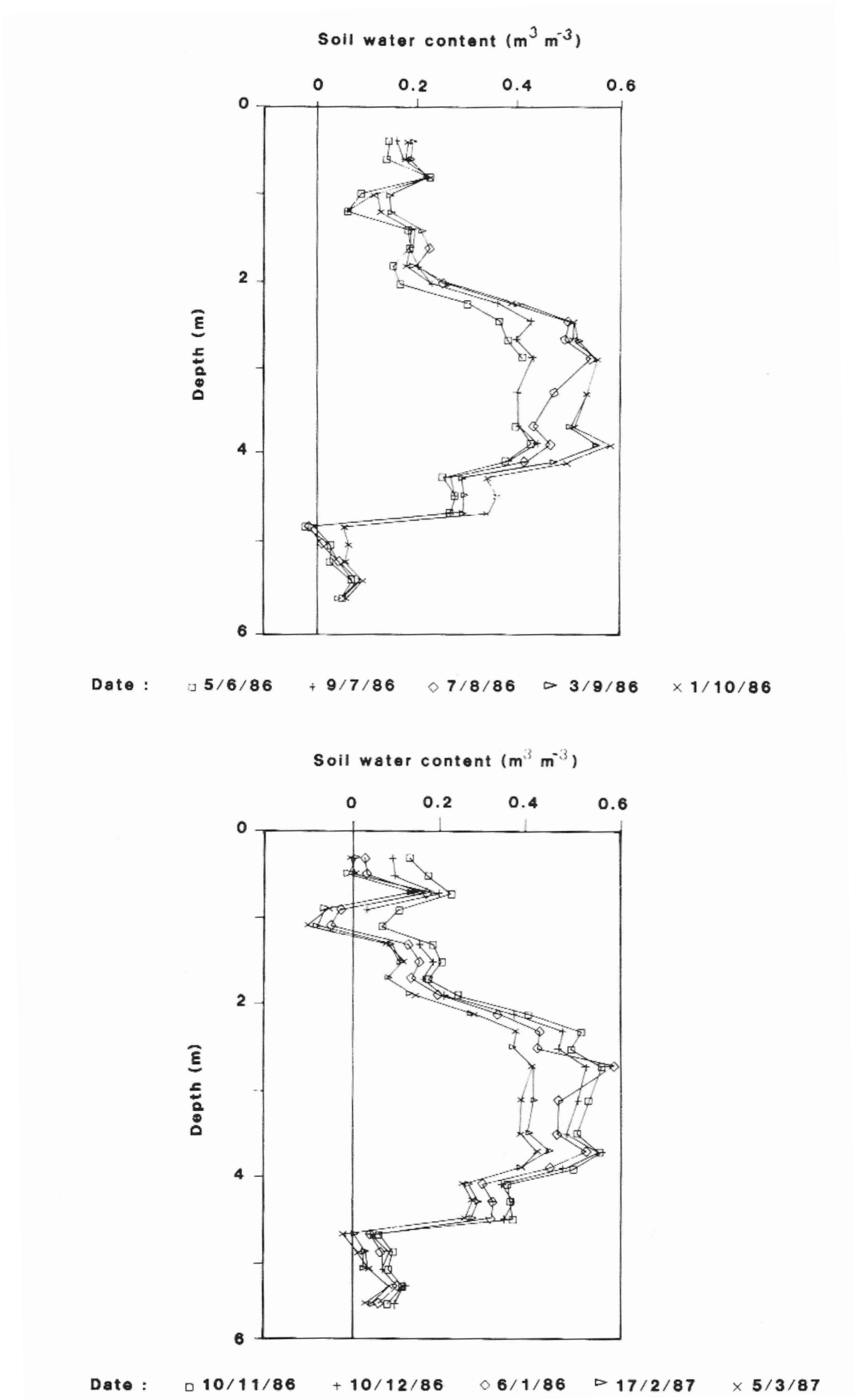


Figure 3.20 Soil water content profiles for access tube NF2: (a) 5/6/86 to 1/10/86; (b) 10/11/86 to 5/3/87

The 9 to 12, 12 to 15, and 15 to 18 m depth intervals show no discernible seasonal variation of soil water content with time. However there are statistically significant ($p -= 0.001$, $p <0.003$) trends of a small linear increase in soil water storage of (0.015 and 0.02 mm day^{-1} respectively), over the last two years of data, for the 12 to 15 m and 15 to 18 m depth intervals. These trends may indicate a very slow redistribution of soil water from the 6 to 9 m depth interval or from earlier, wetter years.

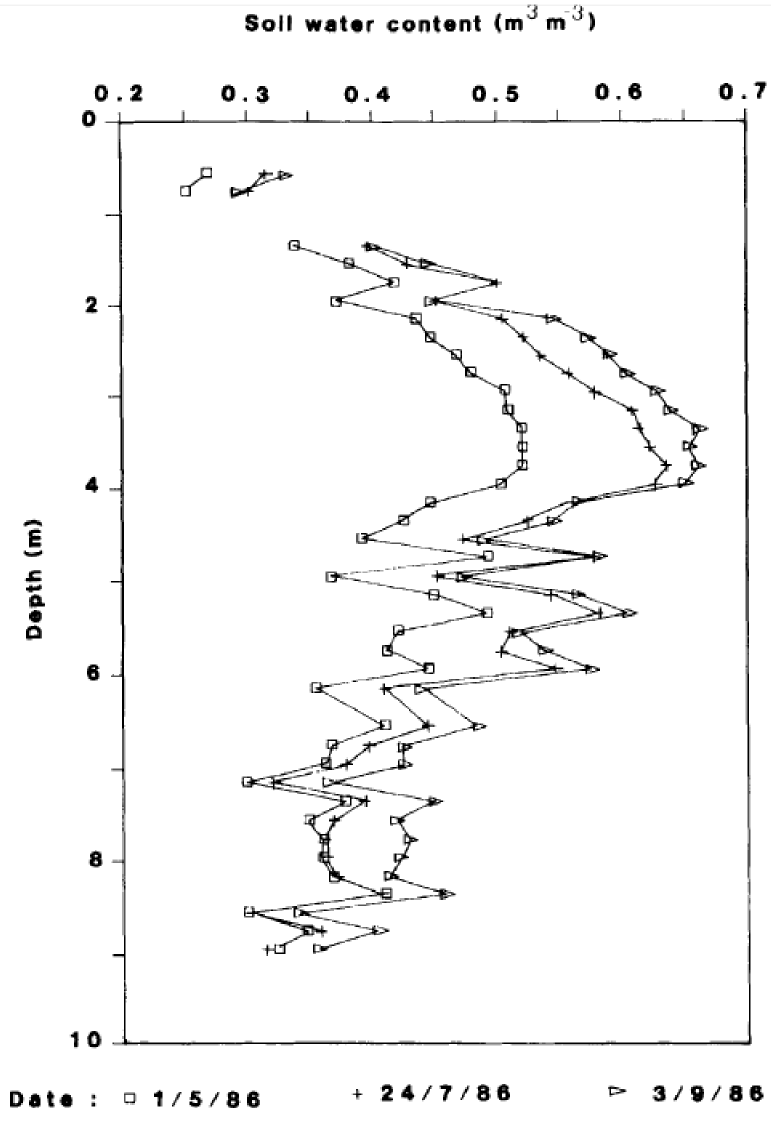


Figure 3.21 Soil water content profiles for access tube NF3: 1/5/86 to 3/9/86

The amplitudes for the 0 to 3 and 3 to 6 m depth intervals were 255 mm and 210 mm, respectively. There was only a slight seasonal trend for the 6 to 9 m interval resulting in a 90 mm amplitude. The total amplitude to 6 m was 430 mm and to 18 m was 530 mm.

The response of soil water storage to rainfall is rapid for the top 3 m, but becomes progressively slower for deeper intervals (Fig. 3.24 a-f). The response of the 3 to 6 m interval is typically delayed by a month from the onset of significant winter rains. Below 9 m there are random responses which are typically less than the accuracy of measurement (± 0.5 mm day^{-1}). During the summer (10 December 1986 to 5 March 1987) the decrease in soil water storage from the top two zones (0 to 3 and 3 to 6 m) was approximately equal. Below 6 m there was little change in soil water storage. This indicates that soil water extraction by evaporation and transpiration removed similar amounts of water per unit depth over the 0 to 6 m profile.

Drying rates of soil water storage

The soil water drying rate averaged *5* mm day^{-1} during the summer, reaching a maximum of 5.6 mm day^{-1}. At the end of the 1986/87 summer there was an indication of a reduction in the soil water drying rate from 5 to 3 mm day^{-1}. This reduction in soil water drying rate occurs predominantly in the 0 to 3 m zone (Fig. 3.24a) and indicates that transpiration was soil water limited at this time.

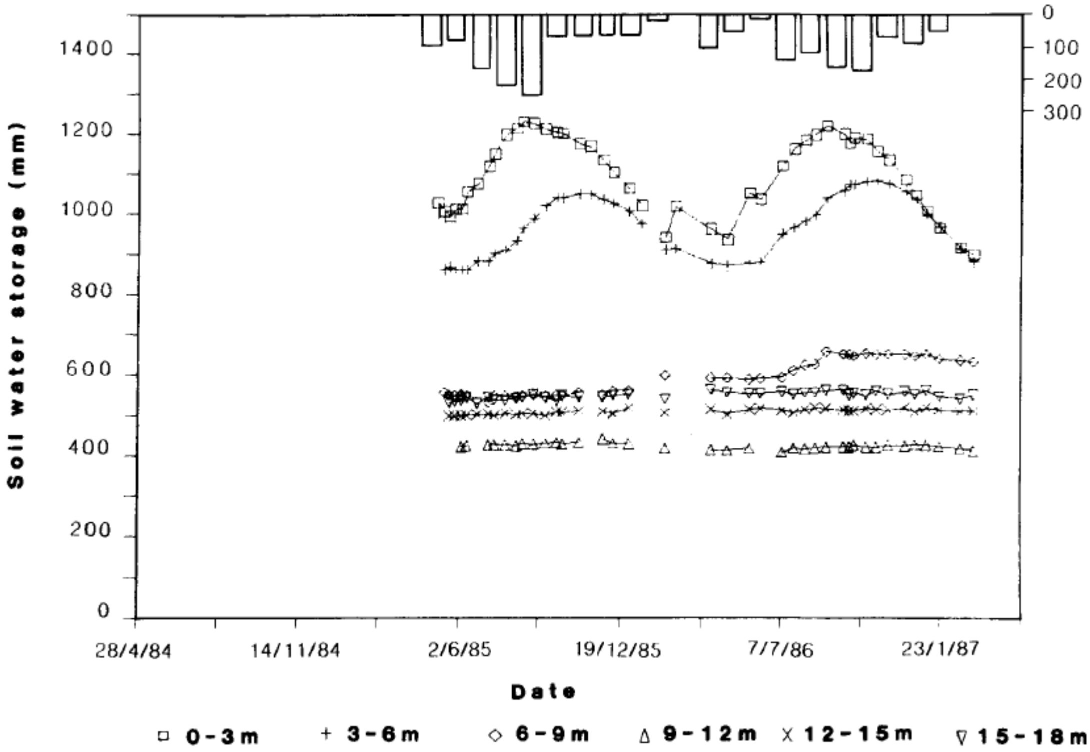


Figure 3.22 Time trends in soil water storage for 3 m depth intervals

3.3.6 Discussion

Perched aquifers

Saturation of the soil profile at site T4 occurred from 0.8 to 1.8 m and persisted through the winter and early spring. The saturation occurred above the low permeability clay pan. The long duration of saturation at site T4 contrasts with the short-lived and infrequent periods of saturation at T2 and T3 on the adjacent hillslope transect. This implies that areas with shallow (<2 m) clay subsoils may be zones of preferential throughflow generation and sources of streamflow as suggested by Stokes and Loh (1982). These sites may also be susceptible to the dispersal of *Phytophthora cinnamomi* spores in throughflow with subsequent downslope infections as indicated by Kinal (1986).

Annual soil water storage amplitude

The annual soil water storage amplitude was 530 mm for an 18 m soil profile but changes in soil water storage were generally restricted to the upper 6 m. This compares to an annual soil water storage amplitude of 480 mm and changes in soil water storage measured down to 12 m for the hillslope transect. The higher volume storage of water and smaller depth of infiltration at the transpiration site are probably related to the higher porosity and lower hydraulic conductivity of the clay soils (Carbon *et al.* 1980).

Soil water drying rates

The average drying rate of the soil profile at the transpiration site was 5.0 mm day^{-1} for the extended dry period of 1986/87. This is considerably higher than the drying rates measured on the hillslope transect. This may be due to the higher density of vegetation at the transpiration site.

Pattern of soil water extraction

Soil water extraction occurred down to 9 m. There was no significant trend beyond this depth. This compares to soil water extraction measured down to 12 m at the hillslope transect. The pattern of soil water change throughout most of summer was for equal rates of water extraction, for the 0 to 3 m and 3 to 6 m depth intervals. However, at the end of summer (5 March 1987) there was a reduction in the drying rate in the top 3 m.

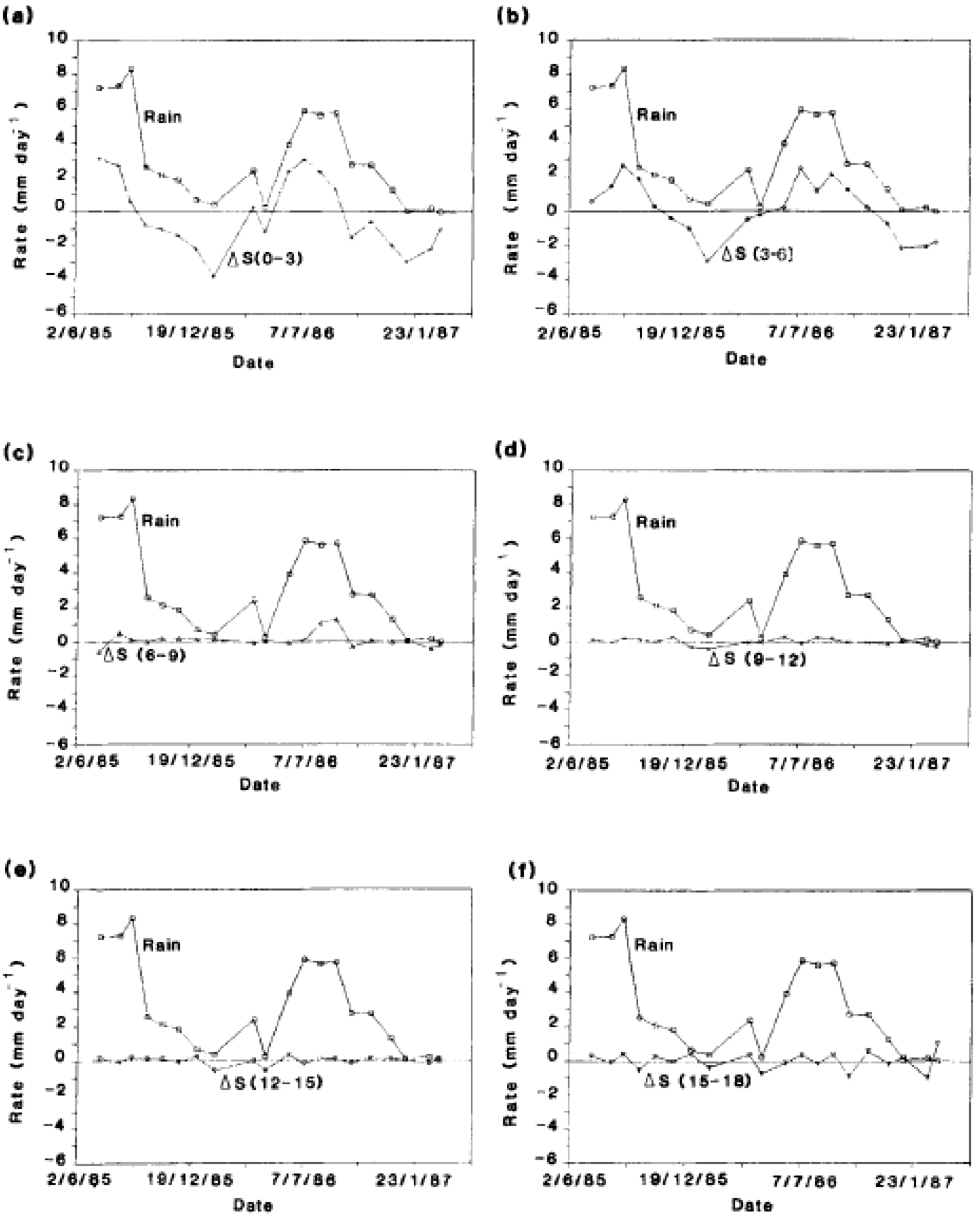


Figure 3.23 Rates of change in soil water storage: (a) 0 to 3 m, (b) 3 to 6 m, (c) 6 to 9 m, (d) 9 to 12 m, (e) 12 to 15 m, and (f) 15 to 18 m

Mechanism of soil water movement

There is some evidence at the transpiration site of 'preferential' flow of soil water from the 0 to 3 m horizon to the 6 to 9 m horizon following saturation occurring in the upper horizon. Other observations of rapid preferential flow in fine-textured subsoils have been described for the jarrah forest by Johnston (1987b).

3.3.7 Conclusions

- Infiltration was dominated by gravity drainage during winter and early spring;
- Soil water potential data indicated that upward potential gradients occur down to at least 7 m during summer;
- A shallow saturation zone was observed throughout middle and late winter above the clay pan;
- The amplitude of annual soil water storage change of 530 mm for the study site was generally above typical values found elsewhere in Australia;
- A small but statistically significant increase in soil water storage of about 10 mm yr^{-1} occurred over the 12 to 18 m depth range over the monitoring period;
- Significant seasonal soil water content changes were generally restricted to 6 m depths.
- There was evidence of preferential flow to 9 m following the development of saturation in the surface soils; and
- Soil water extraction occurred at uniform rates over the top 6 m of the soil profile except in late summer when the drying rate for the 0 to 3 m interval decreased substantially.

The next chapter concerns paired catchment studies investigating the hydrologic impact of forest disturbances. The knowledge and understanding established from the hillslope studies in this chapter are key to understanding and explaining the changing hydrology that occur with these forest disturbances.

Chapter 4 SWWA forest hydrology – paired catchment studies

This chapter examines the hydrologic impact of forest disturbance using four paired catchment studies. In particular, these studies evaluate the impact of converting forest to agriculture and of timber harvesting. Section 4.1 explores the streamflow generation mechanisms for forested and cleared catchments, whilst sections 4.2 and 4.3 explore both streamflow generation and salinity export. Section 4.4 evaluates the hydrologic impact of intense timber harvesting for increased water production.

This chapter consists of four catchment experiments:

4.1 Conversion of forest to agriculture – high rainfall zone

4.2 Conversion of forest to agriculture (extensive clearing) – low rainfall zone

4.3 Conversion of forest to agriculture (parkland clearing) – low rainfall zone

4.4 Forest thinning experiment – high rainfall zone

The conversion of forest to agriculture in the high rainfall zone (Section 4.1) is focused on the processes that generate streamflow. The conversion of forest to agriculture in the low rainfall zone (Sections 4.2 and 4.3) is focused on not only streamflow but also stream salinity. As described in Section 1.3.5 stream salinisation is a critical water resource management issue in Western Australia.

4.1 Change in streamflow generation from converting native forest to agriculture[5]

4.1.1 Abstract

The clearing of native vegetation and establishment of agricultural plants on a small catchment in SWWA resulted in a large streamflow increase (~ 30% rainfall yr^{-1}). This increase was brought about by a decrease in transpiration and interception loss. Streamflow increased markedly in the first year after clearing (~ 10% rainfall) and continued to increase at a slower rate for a further five years, when a new streamflow equilibrium was reached. Explanations of the time trend of streamflow increase were sought in terms of streamflow generation mechanisms. The initial increase in streamflow was attributed to the impact of the immediate decrease in interception loss (~ 13% of rainfall). The subsequent linear increase in streamflow was closely correlated with the expansion of the groundwater discharge area, and the cessation of streamflow increase was considered to result from the attainment of a new groundwater recharge discharge equilibrium. Evidence from other catchments which have undergone forest reductions show that the permanent groundwater system is instrumental in controlling the streamflow response following forest reduction.

4.1.2 Introduction

This section examines the mechanisms of streamflow generation following total or partial deforestation of a catchment. The analysis is based primarily on the

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[5] Published as: **Ruprecht, J.K**. and Schofield, N.J., 1989. Analysis of streamflow generation following deforestation in southwest Western Australia. *Journal of Hydrology*, 105: 1-17.

measured hydrological responses of Wights catchment (Fig. 4.1), following conversion from forest to agriculture. Additional evidence is also examined from numerous other experimental catchments in SWWA that have been subjected to some form of forest reduction.

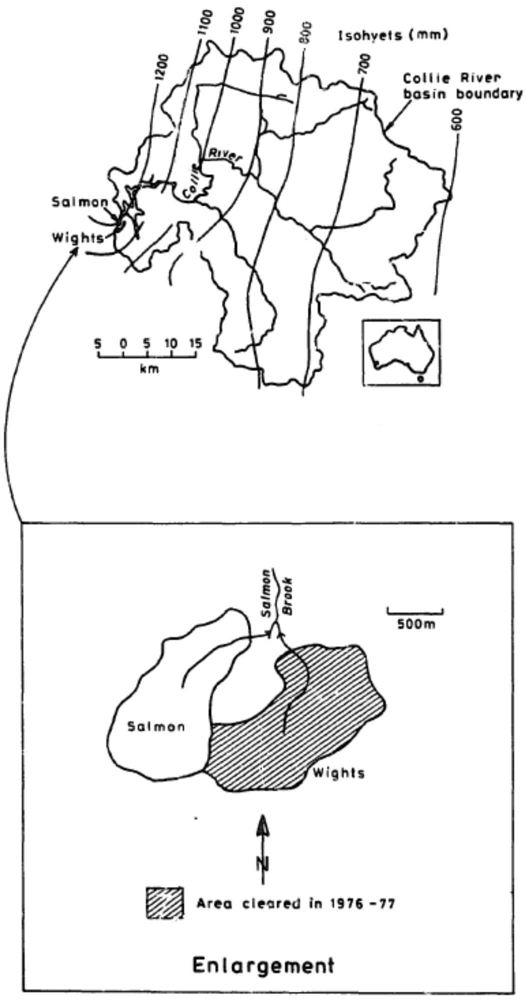

Figure 4.1 Location map for Wight's and Salmon experimental catchments

4.1.3 Methods

Wights catchment and the adjacent, paired Salmon catchment are located about 170 km southwest of Perth in the Collie River basin (Fig. 4.1). The area has a Mediterranean climate with cool, wet winters (June to August) and warm to hot, dry summers (December to February). Wights catchment has a long-term average annual rainfall of 1200 mm yr^{-1} rainfall. The mean annual Class A pan evaporation at Collie, 20 km northeast of Wights catchment, is 1630 mm yr^{-1}. Wights catchment area is 94 ha. The soils are mainly gravelly and sandy laterites, with some yellow podzolics, overlying a kaolinitic clay. The regolith averages 30 m in depth to a granitic basement. The vegetation at the beginning of the experiment was an open forest dominated by jarrah, subdominants of marri, and bull banksia. Swamp areas were characterised by yarri (*E. patens*) and marri. There was also a well-defined understorey.

Instrumentation on the experimental catchments included neutron moisture meter access tubes, piezometer networks, pluviometers, saltfall gauges, and stream gauging stations. Details of the instrumentation are given by Williamson *et al.* (1987). The hydrological components measured were rainfall, saltfall, interception, soil water content, permanent groundwater piezometric levels, ephemeral groundwater levels and streamflow.

In the summer of 1976/77 merchantable timber was logged from the whole of Wights catchment, the remaining timber heaped and burnt and the area aerially sown to clover and grass in May to June 1977. In 1978, the area was leased and grazed with sheep and horses. Further agricultural developments involving tree stump removal, pinwheel-raking, bulldozer-levelling and ploughing were carried out in the autumn of 1980. Salmon catchment remained in a forested condition ever the period of investigation.

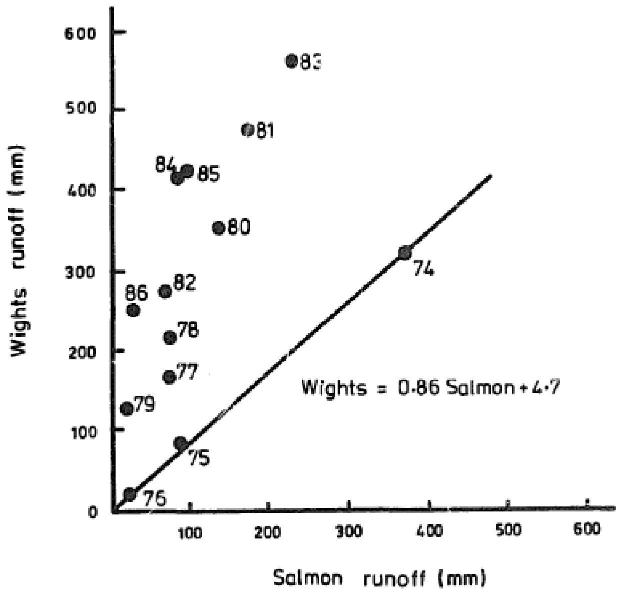

Figure 4.2. Annual streamflow relationship of Wights to Salmon

4.1.4 Observed streamflow response of Wights to clearing

Streamflow trend

To remove the effects of climatic variation, the paired catchment approach was used. A three year pre-treatment period (1974 to 76) was allowed for calibration. The streamflow relationship between Salmon and Wights catchments prior to clearing is shown in Fig. 4.2. This relationship has been used to predict streamflow increases for the period 1977 to 86 (Table 4.1).

Three main characteristics of the trend of annual streamflow increase are apparent and are denoted by sections A, B and C on Fig. 4.3. Section A represents an initial jump in streamflow in the first year after clearing. Section B shows a relatively linear streamflow increase for the next six years. Section C indicates a cessation of the streamflow increase approximately seven years after clearing. It is assumed that section C represents a new hydrologic "equilibrium" in which the catchment has adjusted to the change in vegetation. However, more years of data collection will be required to verify this assumption.

Rainfall-streamflow relationship

Annual rainfall versus annual streamflow is plotted in Fig. 4.4. Prior to clearing the rainfall-runoff relationship has been approximated by a linear relationship. It could be assumed that the rainfall-runoff relationship will approach a gradient of one at high annual rainfall, so that if more data were available a nonlinear relationship would be more appropriate (1974 to 76 data in Fig. 4.4). The 1974 to 76 regression indicates that zero runoff would occur for an annual rainfall of around 800 mm yr^{-1}. A second linear rainfall-runoff regression was performed for the years of assumed new hydrologic equilibrium (section C in Fig. 4.3). This regression (1983 to 86 data in Fig. 4.4) has a similar gradient to the 1974 to 76 data, but is substantially displaced towards higher runoff.

Extrapolation of the 1982 to 86 regression indicates that zero runoff would occur for an annual rainfall of 400 mm yr^{-1} which is well below the minimum annual rainfall recorded in the area (605 mm yr^{-1} at Collie in 1940). The displacement of the 1974 to 76 and 1983 to 86 regressions in the rainfall range 800 to 1200 mm is about 300 mm. Since in 90% of years the annual rainfall is in this range, an increase in streamflow of this magnitude can be expected to have a similar frequency. Moreover, in low rainfall years (e.g. <800 mm yr^{-1}) significant streamflows should occur, where little would occur in the forested condition.

Table 4.1 Streamflow increase on Wights catchment due to agricultural clearing

Year	Rainfall (mm)	Measured flow (mm)	Predicted flow (mm)	Measured-predicted flow	
				(mm)	(% rainfall)
1974	1326	320	321	-1	0
1975	1027	81	77	4	0
1976	822	19	22	-3	0
1977	877	164	68	96	11
1978	943	217	71	146	18
1979	781	128	19	109	14
1980	1165	351	124	227	19
1981	1347	481	154	327	24
1982	837	277	62	215	26
1983	1147	561	202	359	31
1984	1050	417	75	342	32
1985	1105	421	79	343	31
1986	770	249	25	224	29

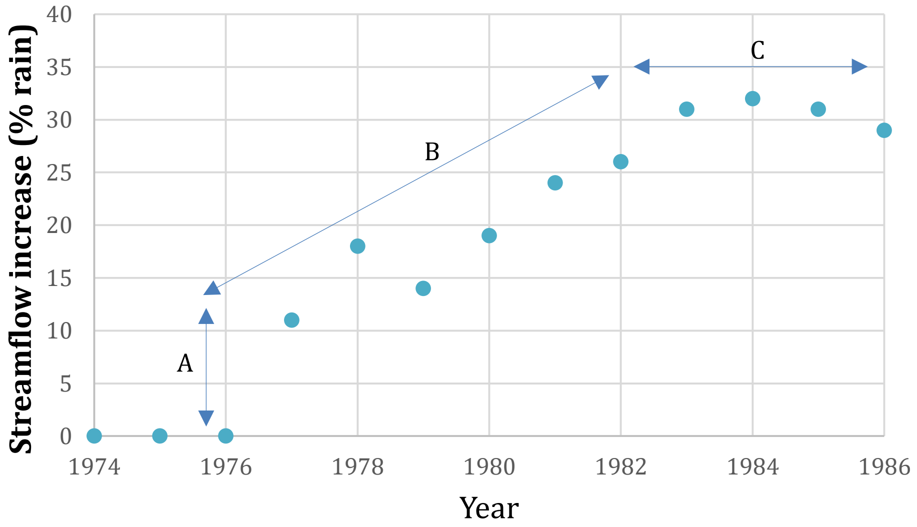

Figure 4.3. Time trend of streamflow increase for Wights catchment (A is initial response, B is transient and C is new equilibrium)

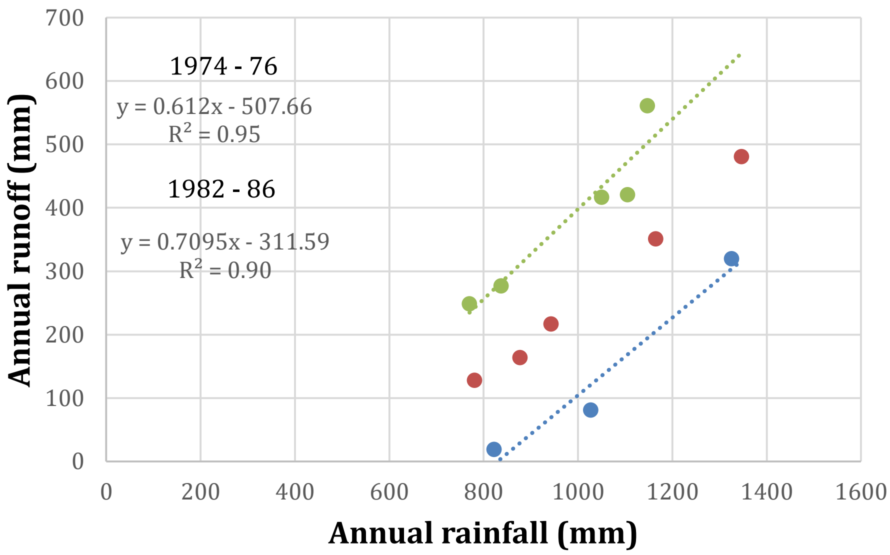

Figure 4.4. Annual rainfall to streamflow relationship for Wights catchment for the pre-clearing period (1974 to 76 – blue dots) and for the post clearing period (1977 to 1981 red dots and 1982 – 1986 green dots)

Streamflow characteristics

Streamflow rates and daily rainfall totals for Wights and Salmon catchments for 1986 are shown in Fig. 4.5. The streamflow characteristics of the two catchments are clearly different with Wights being perennial and having high peak flows in early winter, while Salmon is ephemeral and does not generate streamflow until mid-winter. From mid-July the streamflow hydrographs had a marked similarity in shape, indicating that the major streamflow generation mechanisms may have been retained following clearing for agriculture.

The streamflow response of Wights catchment varies with season. During February and March there was only minor streamflow for 20 to 40 mm of rain.

However, from June to August similar rainfall events produced consistently larger streamflow responses.

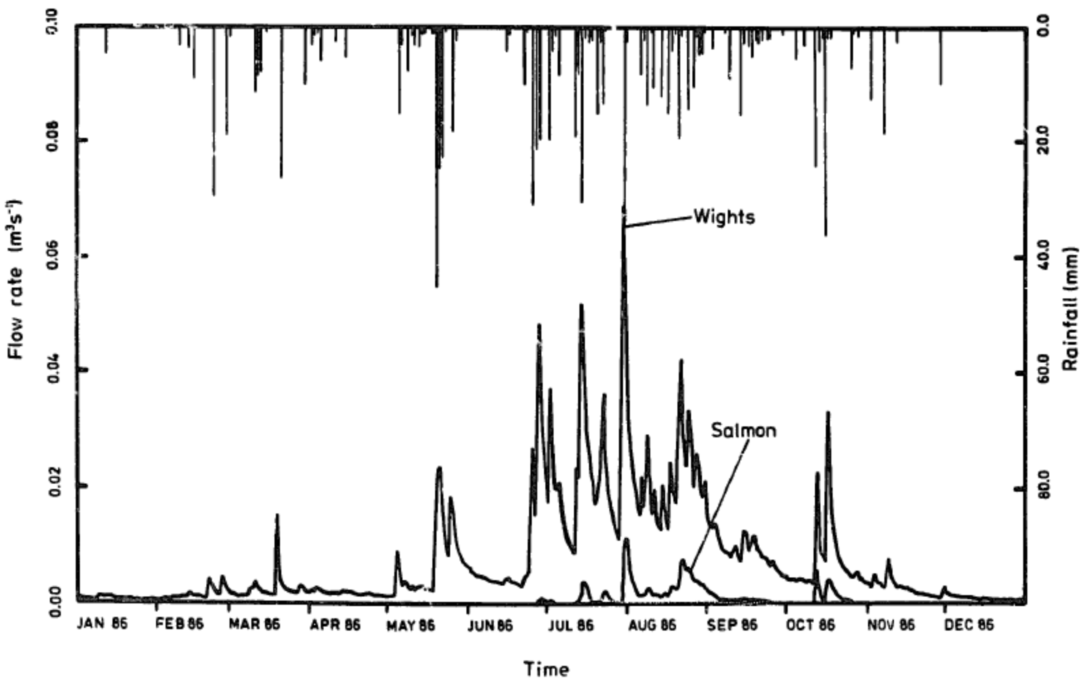

Figure 4.5. Streamflow and rainfall for Wights and Salmon catchments for 1986

4.1.5 Mechanisms of streamflow generation after clearing

The general hydrological changes which occurred following agricultural development of Wights catchment were an increase in streamflow (Fig. 4.3), an increase in unsaturated water content (Sharma *et al.* 1987b), a rise in the permanent groundwater system (Peck and Williamson 1987a), and a decrease in evapotranspiration (Williamson *et al.* 1987).

In order to explain Wights streamflow response to clearing in terms of streamflow generation mechanisms, three hypotheses have been identified and will be examined. These hypotheses are:

1. That the increase in streamflow is generated from increased infiltration excess overland flow (Hortonian flow);
2. That the increase in streamflow is generated from increased throughflow brought about by increased net precipitation; and
3. That the increase in streamflow is generated primarily from an expanded permanent groundwater discharge area which leads to increased saturation excess overland flow, throughflow, return flow and groundwater flow.

Hypothesis 1: Increase in streamflow is generated from increased infiltration excess overland flow

Infiltration excess overland flow is considered an important streamflow generation mechanism for fine-textured soils, non-wetting soils, when surface sealing occurs and as a result of surface compaction (Section 1.2.2). The surface soils of Wights catchment are coarse-textured, and texture induced Hortonian flow is considered to be negligible. Surface sealing and crusting have been found to be dominant factors reducing infiltration capacity for bare or unprotected soils (Morin and Benyamini 1977; Skaggs and Khaleel 1979), while planting grass reduced surface runoff from 50% to zero (Hino *et al.* 1967). The establishment of pasture on Wights catchment is likely to minimise surface sealing effects.

No observations of non-wetting soils have been made on Wights catchment. Nor have there been any measurements of the degree of compaction caused by clearing and agriculture.

(Sharma *et al.* 1987a) measured saturated hydraulic conductivities (K_S) and sorptivities (S) for Salmon catchment and Wights catchment after the establishment of agriculture. The mean saturated hydraulic conductivity of Wights catchment (131 mm hr^{-1}) was significantly less than the mean for Salmon catchment (866 mm hr^{-1}), indicating that agricultural development has had a significant impact.

The relationship between rainfall intensity and the time to ponding for three different soil conditions on Wights catchment was determined by Sharma *et al.* (1987a) and is shown in Fig. 4.6. These soils represent: a location where the infiltration capacity was the least in the catchment (curve 1); a soil group with the least infiltration capacity (curve, 2); a soil with average infiltration parameters (curve 3). Also shown in Fig. 4.6 are the 50 and 90 percentiles and maximum rainfall intensities as calculated from six minute time intervals by Williamson *et al.* (1987). Significantly higher rainfall intensities can occur at smaller time intervals as noted by (Marsh 1974), who found the peak rainfall intensity was five times greater than the peak 6 min intensity. It is clear from Fig. 4.6 that the time to ponding would be in excess of 30 min for the 90 percentile rainfall intensity even for the lowest infiltration capacity site.

For the maximum rainfall intensity recorded it would take some 30 min to ponding for the least infiltration capacity soil group. On this basis the probability of significant infiltration excess overland flow is very small.

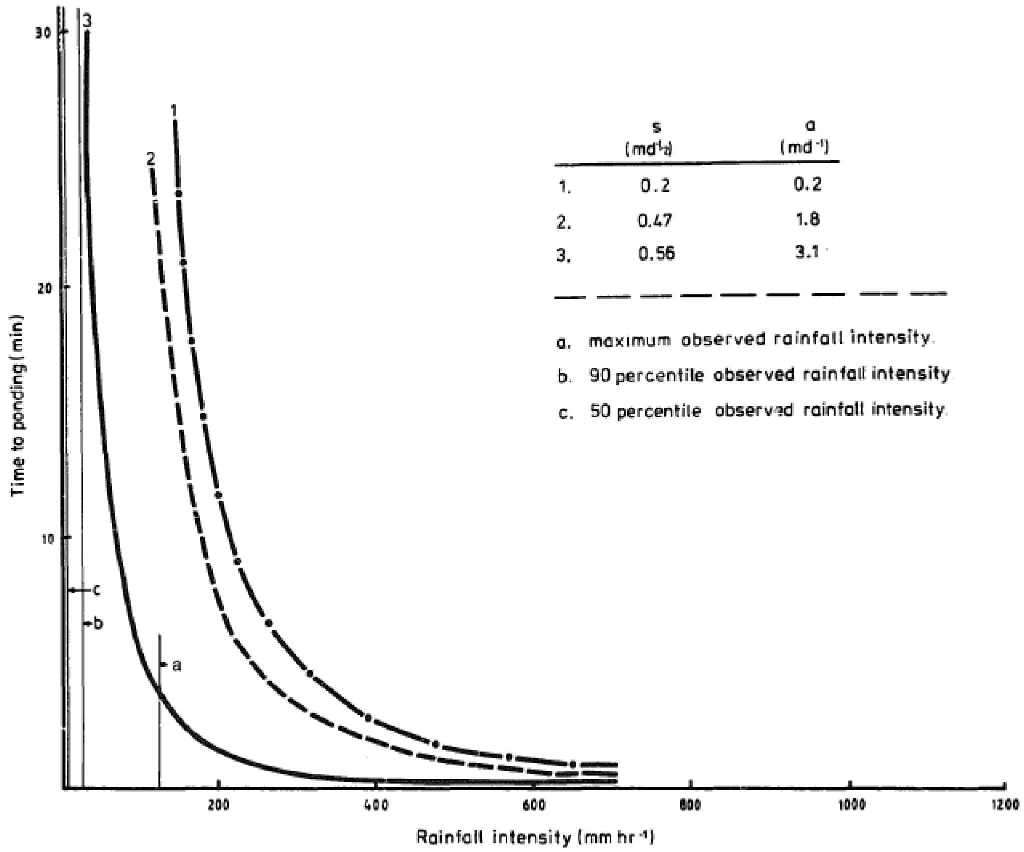

Figure 4.6 Time to ponding from infiltration excess

Hypothesis 2: Increase in streamflow generated by increased throughflow

The existence of an intermittent, shallow, perched groundwater system on Wights catchment was demonstrated by two shallow bores (RW11 and RW12) with their locations shown in Fig. 4.7 and the water levels over the period of monitoring, from September 1980 to March 1982, shown in Fig. 4.8.

The shallow groundwater system forms in the gravelly-sand A horizon, above a low permeability mottled and pallid clay zone. RW11 is situated adjacent to the stream channel. As shown in Fig. 4.8, the water level in this bore rises abruptly at the commencement of winter rains and remains elevated at a fairly consistent base level throughout winter. Superimposed on this base level are hydrograph peaks which occur in response to individual rainfall events. During the period monitored the water level did not reach the soil surface and saturation excess overland flow would not have occurred. Lateral water movement at this point would have been primarily saturated throughflow.

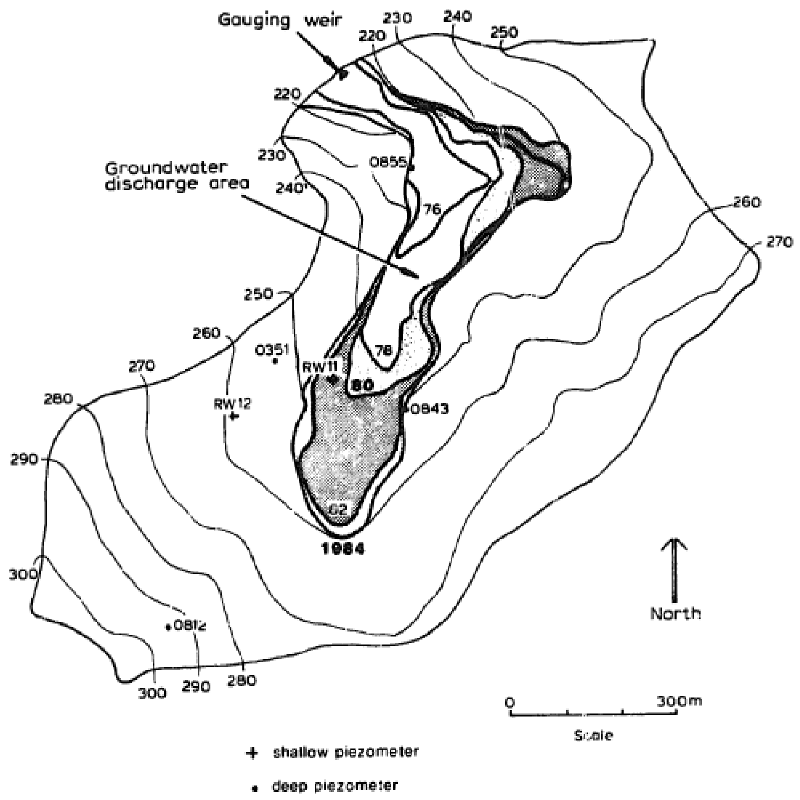

Figure 4.7 Growth of groundwater discharge area for Wights catchment

RW12 was located upslope of RW11 and at some 14 m greater elevation in a mid-slope position. The bore hydrograph showed that this site drained more rapidly than RW11. As with RW11, saturation to the soil surface did not occur but the water level was elevated for a significant portion of the winter.

In hypothesis 2, increased streamflow following clearing is attributed to increased discharge from the shallow groundwater system. The assumed sources for increasing throughflow are increased net precipitation through decreased interception loss, and decreased transpiration. Possible increases in soil water evaporation and subsoil infiltration are considered to be minor by comparison.

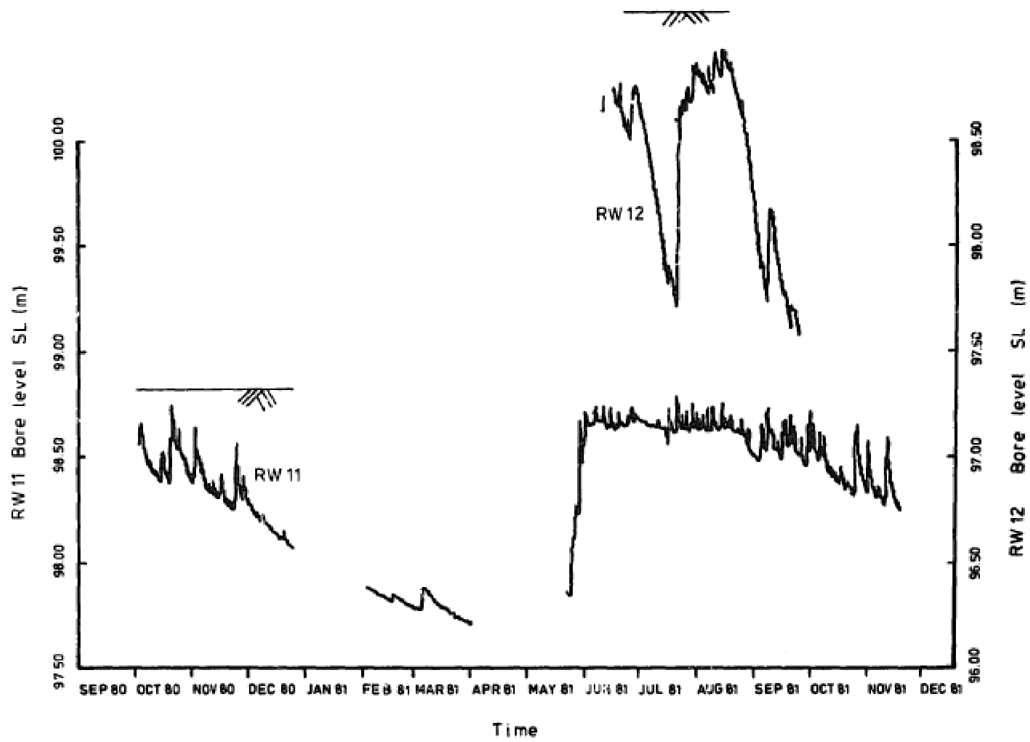

Figure 4.8 Shallow bore responses on Wights catchment

The decrease in interception loss on Wights catchment following clearing was calculated by Williamson *et al.* (1987) to be 13% of rainfall, assuming that interception by pasture was zero. This amount would be a direct and immediate increase in recharge to the shallow groundwater system.

Williamson *et al.* (1987) also estimated that evapotranspiration (excluding interception) from Wights catchment decreased by 15% of rainfall following agricultural development. Some of this decreased evapotranspiration may have been from the surface aquifer, leading to increased throughflow.

Evidence of elevated soil water content following clearing for agriculture was provided by Sharma *et al*. (1987b) from the analysis of neutron probe data. These authors found that there was a significant and rapid increase in soil water content to 6 m depth. They concluded that a new equilibrium water content profile would have been achieved within two years of clearing. Increased recharge to the shallow groundwater system would lead to higher water tables, greater saturation overland flow, and higher peak flows. The greater depth of water in the perched aquifer and the greater areal extent of saturation would contribute to the higher baseflows observed (Fig. 4.5).

Since clearing has an immediate effect on interception and transpiration, and a rapid effect on soil water content, any increases in throughflow associated with these hydrological changes would be rapid. It is proposed that the initial jump in streamflow (section A in Fig. 4.3), which accounts for about one-third of the total streamflow increase, is due to these mechanisms. However these changes would not bring about the continued linear increase in streamflow observed in the subsequent six years (section B in Fig. 4.3).

Hypothesis 3: Increase in streamflow due to increased groundwater-saturated source areas

In this hypothesis it is proposed that the increase in streamflow following clearing is generated from an expanded permanent groundwater system, which in turn leads to increased overland flow, throughflow and groundwater flow contributions to streamflow.

The behaviour of the deep groundwater system was monitored by a network of deep piezometers. The effects of agricultural clearing are seen by the responses of four piezometers (Fig. 4.9), located as shown in Fig. 4.7. At lower elevations (piezometers 0843 and 0855) the piezometric levels rose to the surface by 1980 and have remained above this level since. At higher elevations (piezometers 0351 and 0812) the rise in piezometric levels were delayed but continued to rise for several years.

The rise in permanent groundwater levels has been shown to be due to increased groundwater recharge following clearing of the native vegetation. Sharma *et al.* (1987b) demonstrated elevated soil water contents following clearing. The mechanisms of recharge have been investigated in detail by (Johnston 1987b). He showed that recharge is spatially discontinuous at a small-scale (1 to 100 m) and that the existence of a perched shallow groundwater may lead to rapid recharge at localised sites. It is also clear that the loss of native deep-rooted vegetation means that water infiltrating the subsoil can no longer be extracted directly by plants and will eventually recharge the deep groundwater irrespective of the mechanism involved.

One outcome of rising groundwater levels is an increased area of the discharge zone. The discharge area on the catchment has been defined from the intersection of the natural surface (1 m contours) with the piezometric surface (computer contoured from the network of 30 piezometers). The growth in discharge area following clearing is shown in Fig. 4.7. The discharge area has increased from a pre-clearing value of 3.5 ha (1976) to 17 ha in 1984 since when it has remained relatively stable. The time trend of the groundwater discharge area shows a near-linear increase for six years following clearing after which it has levelled-off (Fig. 4.10). This time trend has the same shape as the streamflow increase (sections B and C in Fig. 4.3). A regression of streamflow increase as a percentage of rainfall *(SI)* with groundwater discharge area *(GDA)* shows a good linear relationship between these variables (Fig. 4.11), taking the form:

$$SI = 1.8GDA - 10.2 \qquad (1)$$

$$R^2 = 0.90$$

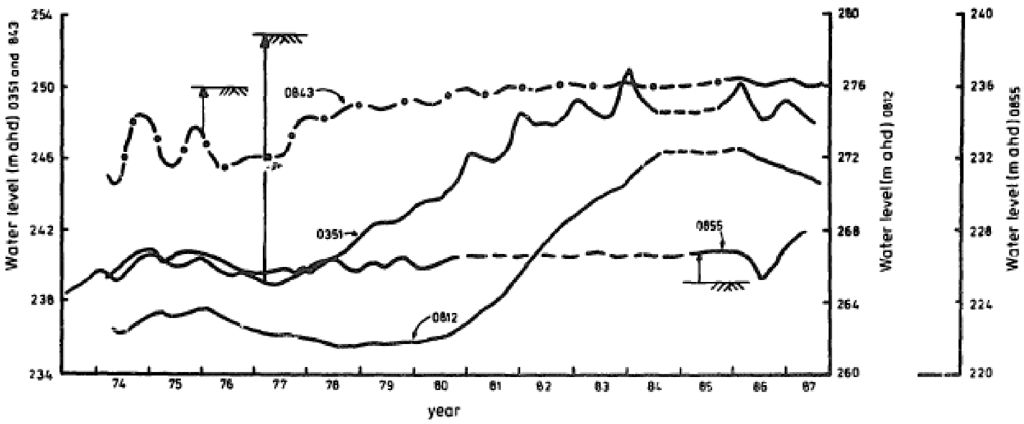

Figure 4.9 Hydrographs of piezometers on Wights catchment

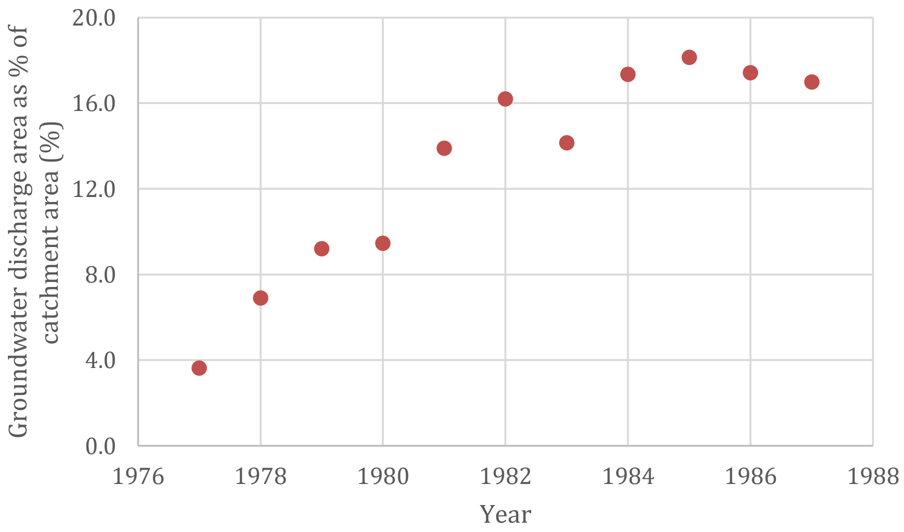

Figure 4.10 Time trend of groundwater discharge area on Wights catchment

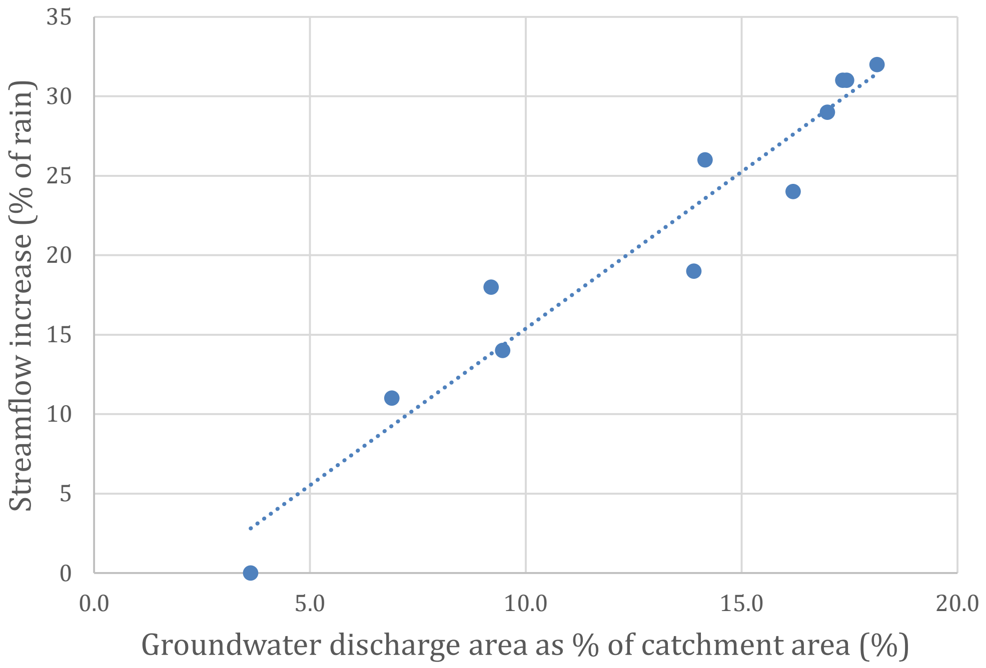

Figure 4.11 Relationship between streamflow increase and groundwater discharge area

From the above evidence, it is proposed that the increase in groundwater discharge area has played a major role in streamflow generation following clearing. The expanded groundwater system has led to increases in groundwater, throughflow and overland flow components.

Two estimates of the contribution of groundwater discharge to streamflow for Wights catchment in 1983 have been made. Hookey (1987) used the Prickett and Lonnquist (1971) groundwater model with a simple recharge component added. He modified saturated hydraulic conductivities and recharge rates to simulate the observed discharge area and potentiometric levels over time. The model indicated that by 1983 about 80% of the additional recharge was being discharged and that discharge areas would increase by a further 7 ha. The annual volume of additional groundwater discharge at 1983 was computed at 77 mm. Williamson *et al.* (1987)

estimated the groundwater discharge for 1983 to be 38 mm, using the mass balance approach of Stokes and Loh (1982).

Increased "direct runoff" following clearing of Wights was analysed by Williamson *et al.* (1987). They calculated the proportion of direct runoff by using a linear hydrograph separation technique between the start and end times of storm hydrographs. Pre and post-clearing comparisons for years 1974 and 1983 showed that direct runoff increased from 4 to 16% of the total runoff.

The major contribution to Wights 1983 streamflow was calculated by Williamson et al. (1987) to be throughflow (77%). This component was found to be an even higher proportion of streamflow (~90%) in the forested Salmon catchment (Stokes, 1985). Further evidence for the predominance of throughflow on Wights is the similar hydrograph shapes of Wights and Salmon in 1986 (Fig. 4.5), the sluggish regressions and large "baseflows" of Wights (Fig. 4.5), and the rainfall-runoff relationship following clearing (Fig. 4.4).

Since the groundwater discharge area occupied some 18% of the catchment area by 1983, it might be expected that the proportion of direct runoff would have increased more substantially. However, it is unlikely that the groundwater discharge area would be fully saturated to the soil surface throughout the year or even winter. This contention is supported by the shallow bore RW11, which was close to the discharge area perimeter in 1980 and 1981. The water level in this bore was elevated close to the surface during winter and declined over spring, summer and autumn (Fig. 4.8). Thus the discharge area would have a significant soil water deficit at the start of winter and may remain partially unsaturated during most of winter. Saturated throughflow in the high conductivity surface soils would be the primary mechanism of desaturation, as noted by Schofield (1986).

4.1.6 Observations from other catchments subjected to forest reduction in SWWA

Several other small research catchments have undergone forest reduction to varying degrees in SWWA. The effects of forest reduction on streamflow are summarised in Table 4.2. Four of the catchments, which had groundwater at the soil surface within the catchment, had an increase in annual streamflow ranging from 14.5 to 32.5% with a mean of 20%.

The catchments which did not have groundwater at the soil surface, had a range of increases in annual streamflow from 0.4 to 4.8%, with a mean of 3.5%. This supports the contention that the main increase in streamflow will not occur unless or until the groundwater system rises to the soil surface. It should also be noted that groundwater intersection of the soil surface in forested conditions is associated with higher rainfall areas. In lower rainfall areas, the rise in water tables to the ground surface will depend primarily on the level of forest reduction, and even in the case of total clearing, water table rises and consequent streamflow increases will be substantially delayed.

Table 4.2 Summary of streamflow increases of research catchments (at 1987) following forest reduction (from Steering Committee for Research on Land Use and Water Supply (1987))

Catchment	Long term rainfall (mm)	Treatment	Forest reduction	Post-treatment monitoring	Average annual streamflow increase			Max annual streamflow increase		Groundwater at surface
					mm	% rain	% flow	mm	% rain	
Wights	1200	Clearing	PCF 100-0	76-86	239	23.9	272	359	32.8	Yes
Lemon	800	Clearing	PCF 100-46	76-83	17	2.1	279	38	4.8	No
Dons	800	Clearing	PCF 100-62	76-83	11	1.4	286	38	4.8	No
March Rd	1070	Timber harvesting	CC 65-0	82-85	121	11.3	147	196	18.3	Yes
April Rd North	1070	Timber harvesting	CC 65-0 buffer 10% of area	82-85	104	9.7	167	155	14.5	Yes
Lewin South	1220	Timber harvesting	CC 70-11 BA 44-7	82-85	116	9.5	81	178	14.6	Yes
Yerraminnup S	800	Timber harvesting	CC 70-10 Buffer 12% of area	82-85	20	2.3	83	38	4.5	No
Wellbucket	700	Timber harvesting	CC 38-20 BA 16-11	77-81	2	0.3	128	3	0.4	No
Yarragil 4L	1120	Forest thinning	CC 55-22 BA 35-11 LAI 1.9-0.6	83-85	17	1.9	293	31	3.1	No

CC = crown cover (%), BA = basal area (m^2 ha^{-1}), PCF = percentage of catchment forested, LAI = leaf area index

4.1.7 Concluding discussion

Agricultural development of Wights catchment has brought about a large increase in streamflow, currently averaging 31% of rainfall. The source of additional streamflow following agricultural development is the reduction in evapotranspiration brought about by replacing native deep-rooted species with agricultural shallow-rooted species.

The increase in streamflow over time has a characteristic form which has been divided into three components: (1) an initial jump immediately following clearing; (2) a linear increase for a subsequent seven years; and (3) the establishment of a new equilibrium flow.

The initial jump in streamflow (~ 10% of rainfall) has been attributed to the immediate impact of decreased interception loss (~ 13% of rainfall) increasing recharge to the intermittent shallow groundwater system and consequently increasing throughflow.

The subsequent linear increase in streamflow, amounting to about 20% of rainfall, is attributed to the expansion of the discharge zone of the permanent groundwater system. Evidence for this was the close correlation between groundwater discharge area and streamflow increase. Insufficient data are available to quantitatively describe the way in which the groundwater system has affected the streamflow generation mechanisms. However, it is recognised that both permanent groundwater discharge and direct runoff have increased substantially following clearing and together could account for about 10% of rainfall. The remaining 10% of rainfall could be derived from increased throughflow off the groundwater saturated area, and from throughflow upslope of the groundwater discharge area that would previously have infiltrated in areas which are now saturated. Each of these streamflow components are dependent on the groundwater saturated area, and thus account for the progressive streamflow increase.

The cessation of increases in streamflow and groundwater discharge areas in 1983 is taken to mean that the permanent groundwater system has attained a new "quasi-equilibrium" in which groundwater discharge equals the enhanced recharge following clearing. This section emphasises the role of the permanent groundwater system in generating additional streamflow over time, and demonstrates why long delays in significant hydrologic response are encountered in lower rainfall areas, where the depth to groundwater is large.

The next section of this chapter will consider the impact of extensive clearing of forest for agriculture in a low rainfall zone on the hydrology and salinity.

4.2 Impact on stream salinity from converting native forest to agriculture - Extensive block clearing[6]

4.2.1 Abstract

A 344 ha experimental catchment in SWWA was partially cleared (western 53% of the catchment) in 1976 to study the effects of agricultural development on water quantity and quality. The impact on the groundwater system in the cleared area was dramatic. Initial rates of rise were only 0.11 m yr^{-1} but this increased after 10 years to an average of 2.3 m $year^{-1}$.

Groundwater rises of 15 m in the valley and 20 to 25 m on the lower sideslopes were observed over 13 years. A small seep (groundwater discharge area) appeared for the first time in 1988 and by 1989 it covered an area of 1 ha. Streamflow initially increased by 30 mm yr^{-1} (4.0% rainfall) compared with a native forest average streamflow of 8 mm yr^{-1} (1.0% rainfall). However, since the seep area developed, the increase in streamflow has been 50 mm yr^{-1} (6.6% rainfall). Stream salinity was low prior to clearing (30 mg L^{-1} Cl^-) and remained low for 9 years after clearing. However, since 1987, stream salinity increased dramatically as the ground water approached the ground surface, and by 1989 reached an annual average of 290 mg L^{-1} Cl^-. The daily maximum in 1989 was 2200 mg L^{-1} Cl^- compared with 92 mg L^{-1} Cl^- from 1976 to 1986. The catchment changed from net salt accumulation pre-clearing to net salt export after 1987. Thirteen years after

[6] Published as: **Ruprecht, J.K**. and Schofield, N.J., 1991. Effects of partial deforestation on hydrology and salinity in high salt storage landscapes. I. Extensive block clearing. *Journal of Hydrology*, 129: 19-38.

clearing, the groundwater level, stream yield, stream salt load and stream salinity had not reached equilibrium but were all still increasing.

4.2.2 Introduction

Stream and land salinization are major environmental and economic problems affecting several arid and semi-arid regions of the world. Dudal and Purnell (1986) estimated that 7% of the global land surface and 42% of Australasia is salt-affected. In the United States and Canada the total area of salt-affected land has been estimated at 2.2 x 10^6 ha (McKell *et al.* 1986).

Non-irrigated or 'dryland salinity' affects some 810,000 ha of the northern Great Plains of the United States and Canada (Brown *et al.* 1983). Stream salinities of SWWA have increased since European settlement in 1829 to the extent that only 48% of the divertible surface water resources remain fresh (less than 500 mg L^{-1} total soluble salts (TSS) (Western Australian Water Resources Council 1986). By 1989, some 443 000 ha or 2.8% of agricultural land was severely salt-affected (Western Australian Bureau of Statistics, personal communication, 1989) and was increasing at the rate of 18 000 ha $year^{-1}$ (Schofield 1989).

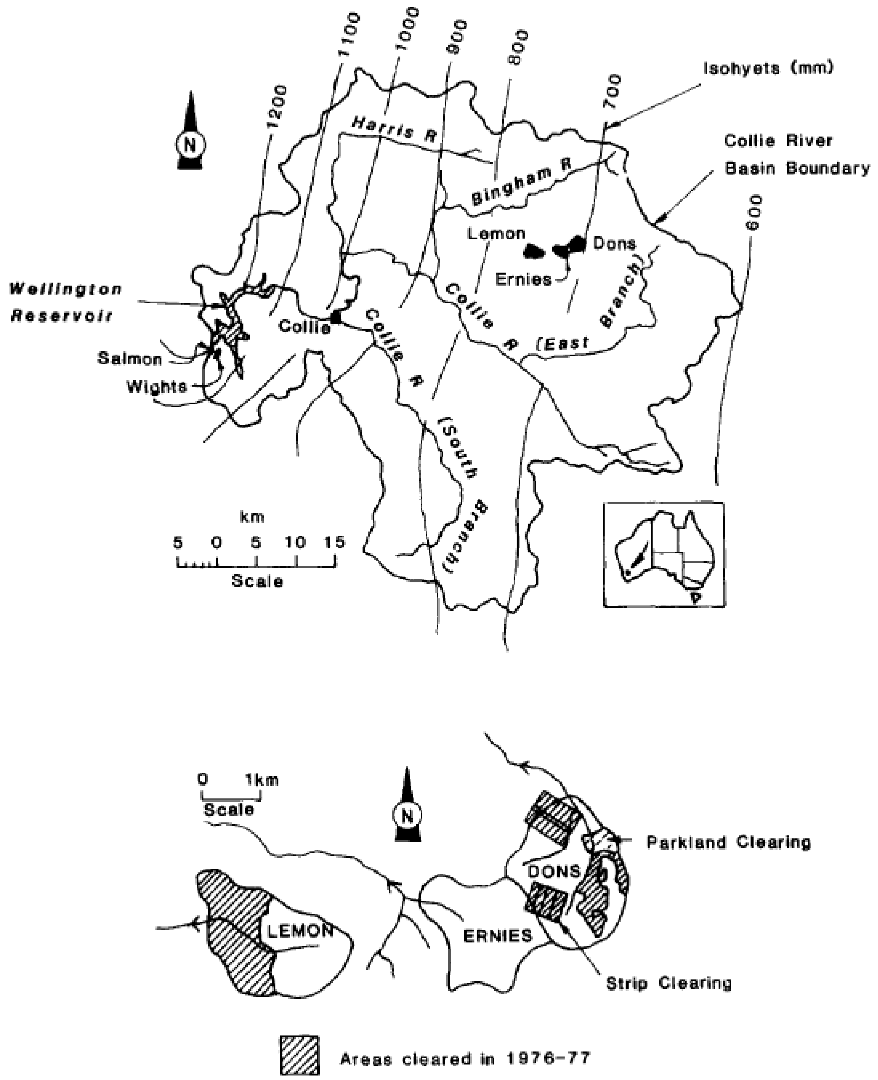

Figure 4.12 Location map for experimental catchments in the Collie River Basin (after Williamson *et al*. 1987)

The substantial increases in land and stream salinity in SWWA have resulted from the extensive clearing of native, perennial, deep-rooted vegetation (mainly eucalypt forest and woodland) for the development of annual, shallow-rooted agricultural pastures and crops (Peck 1983; Schofield *et al.* 1988; Wood 1924). This land use change has led to increasing groundwater recharge and rising water tables (Peck and Williamson 1987a). Salt stored in the unsaturated soil profile has been mobilized and transported to the soil surface and streams (Williamson and Bettenay 1979).

Increases have also occurred in streamflow (Williamson *et al.* 1987). In all areas with significant salt storage (mostly less than 1100 mm yr^{-1} rainfall areas), stream salinities have increased as a result of agricultural development (Schofield and Ruprecht 1989).

In the early 1970s, five experimental catchments were established in the Collie River basin (Fig. 4.12) to confirm the cause and quantify the process of salinization. The results from two of the catchments in a high rainfall (approximately 1200 mm yr^{-1}) area were described by Peck and Williamson (1987b) and Ruprecht and Schofield (1989a). The three remaining catchments in the eastern lower rainfall (approximately 750 mm yr^{-1}) part of the Collie River Basin form the basis of this and Section 4.3) (Ruprecht and Schofield, 1991). Lemon catchment (the subject of this paper) is the most representative experimental example of agriculturally induced salinity in Western Australia.

4.2.3 Site description

The partially cleared Lemon catchment, and the nearby forested control, Ernies catchment, lie approximately 250 km southwest of Perth (Fig. 4.12). Their areas are 344 ha and 270 ha, respectively. The climate is Mediterranean, with cool, wet winters (June to August) and warm to hot, dry summers (December to February). The long-term average annual rainfall (1926 to 1979) is 750 mm (Lemon) and 720 mm (Ernies). The Class A pan evaporation is approximately 1600 mm for both catchments (Luke *et al.* 1988). Rainfall usually exceeds pan evaporation for only 4 months of the year (May to August).

The catchments were selected as being representative of the eastern part of the Darling Range (Bettenay *et al.* 1980) and as having the potential to exhibit characteristics commonly associated with salinity problems which follow some years after clearing. The geology of the catchments is typically medium even-grained granites which grade locally into weakly foliated gneissic granites (Bettenay *et al.* 1980). Dolerite dykes are abundant and typically underlie most divides and spurs. The slopes of the catchments are typically around 1:12 and the valleys are broad and flat.

The soils are predominantly gravelly and sandy laterites with some yellow podzolic soils, overlying kaolinitic clay. The regolith averages 30 m in depth. The vegetation at the beginning of the experiment was an open forest dominated by jarrah with a light admixture of marri (Bettenay *et al.* 1980). There was a well-defined tree and shrub understorey. The upper storey had an average height of 20 to 35 m and average basal area of 24 m^2 ha^{-1}. On the lower valley sides wandoo (*E. wandoo*) was more dominant. In the low-lying drainage lines, tree species were mainly absent but where present consisted of flooded gum (*Eucalyptus rudis)* and paperbark (*Melaleuca pressiana).*

Within Lemon catchment, soil Cl^- storage averaged 20 kg m^2 while the average maximum Cl^- concentration of the ground water was approximately 6000 mg L^{-1} (Turner *et al.* 1987a).

4.2.4 Experimental method and instrumentation

Experimental method

The experimental approach was the paired catchment method, one being treated (Lemon) and the other remaining a control (Ernies). The treated catchment had the western half (53.5%) of its area cleared (Fig. 4.13). A pasture of grasses and clover was established for grazing sheep on the cleared areas. Ernies catchment remained as a forested control throughout the experiment. The pre-treatment calibration period was 1974 to 1976. The clearing within Lemon catchment took place between November 1976 and March 1977. From 1977 to 1989 the hydrological response of Lemon catchment to agricultural development has been monitored.

Measurements

Both catchments were instrumented to measure components of the water and salt balances. Rainfall was measured near the catchment outlet with a pluviometer (Fig. 4.13). The chloride content of rainfall was obtained by periodic sampling of storage rain gauges from 1974 to 1983 (Williamson *et al.* 1987).

Streamflow was measured at a sharp-crested V-notch weir. The electrical conductivity of stream water was continuously monitored using a toroidal cell (Brown and Hamon 1961). Water samples were also collected at regular intervals by an automatic pump sampler to determine the relationship between Cl^- and electrical conductivity. In stream waters of SWWA, the dominant anion of total soluble salts is Cl^-. Consequently, measurements of salinity, salt storage and saltfall are expressed in terms of Cl^- ion in this Section, unless otherwise stated.

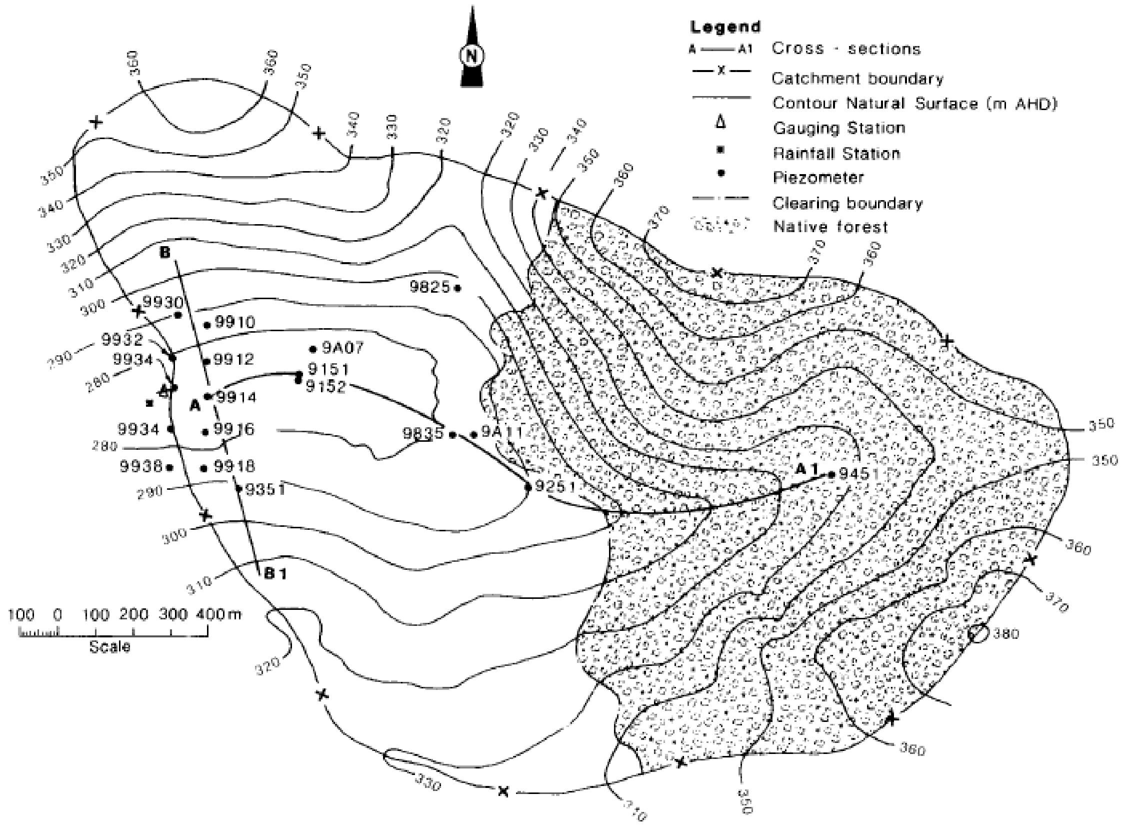

Figure 4.13 Lemon catchment map

Groundwater piezometers were installed in holes drilled by auger to about 40 m or to hard rock. Screen lengths of 1 to 3 m were located at the bottom of each piezometer. In both catchments at least five piezometers were distributed between lower valley and near-divide sites. Additional piezometers were placed across the catchment outlet to determine subsurface flows across the catchment boundary. Piezometers were also installed on a 400 m grid over the cleared area of Lemon catchment.

The depth of water in the piezometers was measured at monthly intervals for 3 years and then at 2 or 3-monthly intervals. Water samples were taken at the same time with a bailer to measure electrical conductivity, the chloride ion concentration, and, less frequently, the concentration of major ions.

4.2.5 Results

Rainfall and saltfall

The long-term rainfall record at Collie Post Office (30 km southwest of the study area) shows a trend of decreasing rainfall since the mid-1940s. This decline in rainfall in SWWA is well documented (Broadbridge 1988; Pittock 1983). During the study period of 16 years, the long-term average rainfall (1926 to 1979) was only exceeded three times, with an extended dry period from 1975 to 1982 (Table 4.3). The average rainfall for the study period was 703 mm for Lemon catchment compared with a long-term (1926 to 1979) average of 750 mm.

Saltfall averaged 3.4 g m^{-2} Cl^- with a coefficient of variation of 30% over the study period. This compares with an average soil salt storage of 20 kg m^{-2} Cl^- for Lemon catchment.

Groundwater

The subsurface hydrology of the Darling Range is characterized by a dual groundwater system, an upper ephemeral perched ground water above the clay B horizon and a permanent ground water overlying the bedrock (Schofield *et al.* 1988; Williamson *et al.* 1987). The response of the deep groundwater system to forest clearing and establishment of pasture was dramatic. Rises of 15 to 25 m in ground water were observed from within the cleared area during the study period (Figs. 4.14(a) and 4.15(a)).

Table 4.3 Lemon Catchment rainfall, streamflow and chloride output

Year(1)	Rain	Streamflow		Saltfall(2)	Chloride flow		
	mm	mm	% rain	$g\ m^{-2}\ Cl^{-}$	$g\ m^{-2}$	$mg\ L^{-1}$	O/I(3)
1974	979	48.5	5.0	4.5	1.38	28	0.31
1975	739	4.8	0.6	2.9	0.17	36	0.06
1976	594	0.8	0.1	1.9	0.02	22	0.01
1977	650	7.2	1.1	2.7	0.03	37	0.11
1978	727	25.4	3.5	3.0	0.86	34	0.29
1979	605	3.0	0.5	2.2	0.11	37	0.05
1980	731	12.3	1.7	3.6	0.43	35	0.12
1981	768	46.3	6.0	4.5	1.67	36	0.37
1982	532	10.1	1.9	2.8	0.43	42	0.15
1983	821	55.9	6.8	6.2	1.85	33	0.30
1984	676	24.3	3.6	2.9	0.85	35	0.29
1985	737	30.4	4.1	3.2	0.88	29	0.28
1986	580	19.8	3.4	2.5	0.73	37	0.29
1987	558	8.5	1.5	2.4	0.51	60	0.22
1988	924	67.4	7.3	4.0	6.61	98	1.65
1989	662	52.0	7.9	2.8	14.9	286	5.3

(1) Water year commencing 1 April

(2) From 1974 to 1983 the input in saltfall was taken from observations reported in Williamson et al (1987). From 1984 to 1989 saltfall was estimated by assuming a constant chloride concentration in saltfall of 4.3 $mg\ L^{-1}$

(3) Chloride output to input ratio

The groundwater level rises within the cleared areas contrast with the forested area of the catchment where groundwater level declined by 2 m (piezometer 9451, Fig. 4.14(a)). This decline was consistent with piezometers in the forested Ernies catchment where groundwater levels in valley piezometers declined by an average of 1.0 m and upslope piezometers by an average of 2.0 m. This probably reflects the generally below-average rainfall over the study period.

The annual minimum groundwater level+ expressed as depth to ground water, for the four bores from a transect along the valley (A-A1 in Fig. 4.13) is shown in Fig. 4.14(a) and the groundwater levels for the transect in Fig. 4.14(b).

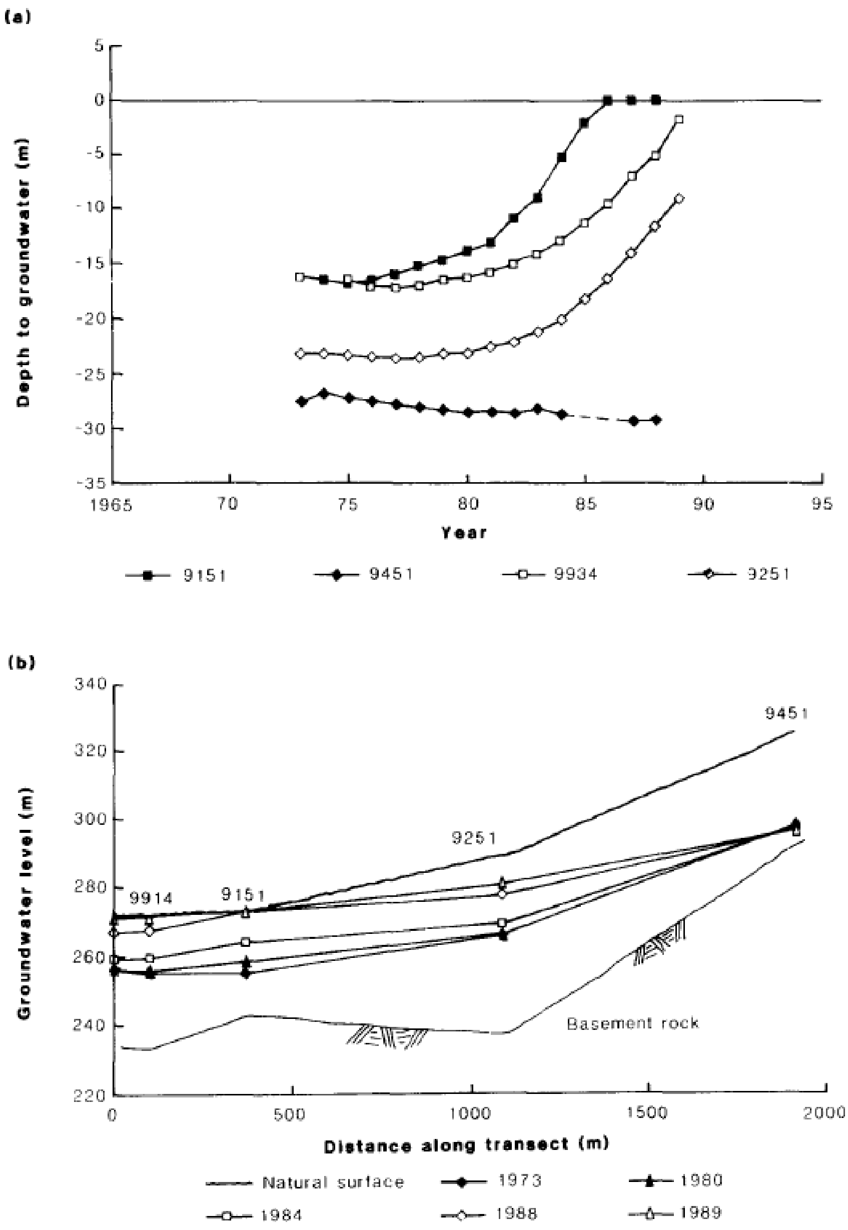

Figure 4.14 Groundwater response to clearing in Lemon catchment: (a) depth to ground water based on annual minima; (b) transect along valley (A-A1)

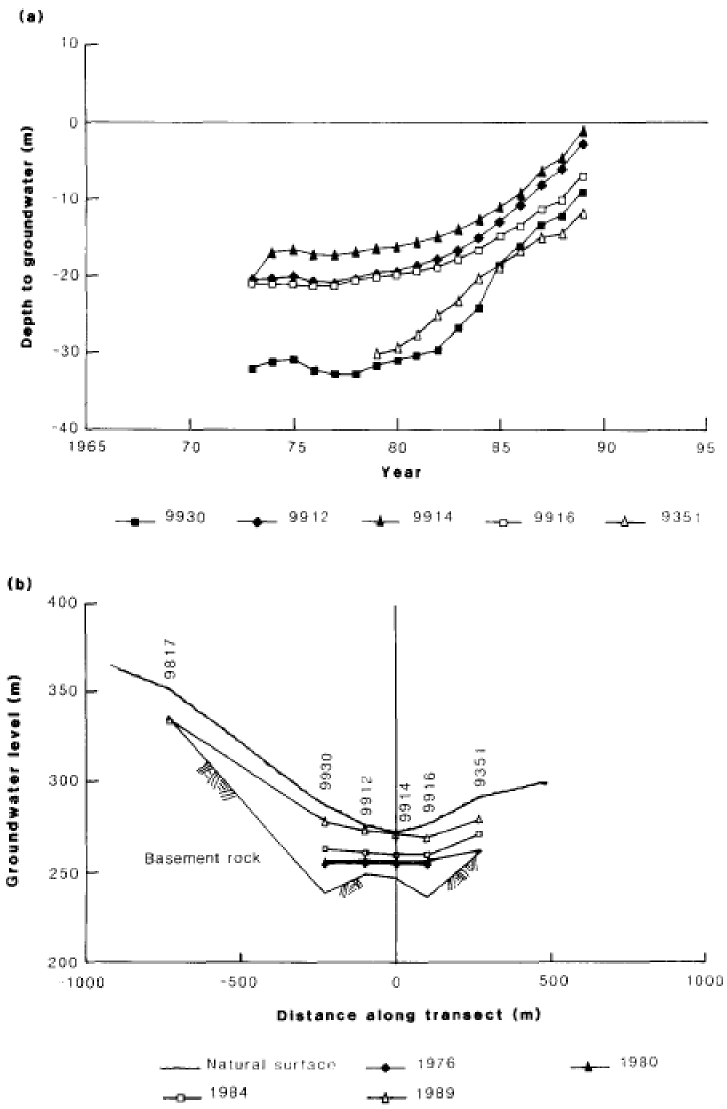

Figure 4.15 Groundwater response to clearing across the catchment outlet: (a) depth to ground water based on annual minima; (b) transect across catchment outlet (B B1)

The natural surface and basement rock information was interpolated from the drilling log information at the piezometer locations. The groundwater piezometric level rose to the ground surface for the first time in 1987 at bore 9151. This may have been the first location owing to the bedrock configuration (Fig. 4.14(b)).

The transect across the valley (B-B1 in Fig. 4.13) (Fig. 4.15) shows a relatively flat groundwater level prior to clearing. Since the groundwater level began rising, the lowest piezometric level has consistently been at piezometer 9916, which is offset from the lowest point in the valley surface. The rate of groundwater level rise in the cleared area was initially relatively slow (Figs. 4.14(a) and 4.15(a)), with average rates of rise of approximately 0.11 m yr^{-1} in the first four years (1977 to 1980). However, from 1981 to 1985 the average groundwater rise was 1.45 m yr^{-1} and for 1986 to 1989 the rise was 2.3 m yr^{-1}. A maximum rate of groundwater rise of 4.8 m yr^{-1} was observed. This high rate of rise continued until the piezometric level reached the ground surface.

Groundwater salinity

The salinities of ground water measured in piezometers within Lemon catchment averaged 770 mg L^{-1} Cl^- prior to clearing and 700 mg L^{-1} Cl^- in April 1989, though the difference was not statistically different (t-test, $P > 0.20$). These results apply to the base of the deep aquifer where the piezometer screens were located. Nested or multi-port piezometers would have been necessary to identify groundwater salinity changes at the water table and through the aquifer.

Streamflow

The annual streamflow to rainfall relationship has undergone some fundamental changes due to clearing (Fig. 4.16(a)). Prior to clearing, approximately 700 mm of annual rainfall was needed for streamflow to occur, equivalent to 400 mm in the first 4 months of the water year (commencing on April 1). By 5 years after clearing a 'quasi-equilibrium' in the annual streamflow to rainfall relationship was observed (Fig. 4.16(a)). For this 'quasi-equilibrium', maintained from 1982 to

1988, only approximately 100 mm was needed in the first 4 months of the water year for streamflow to commence. However, 1989 was an outlier to this relationship and coincides with the emergence of a groundwater discharge area (1 ha) surrounding the catchment outlet.

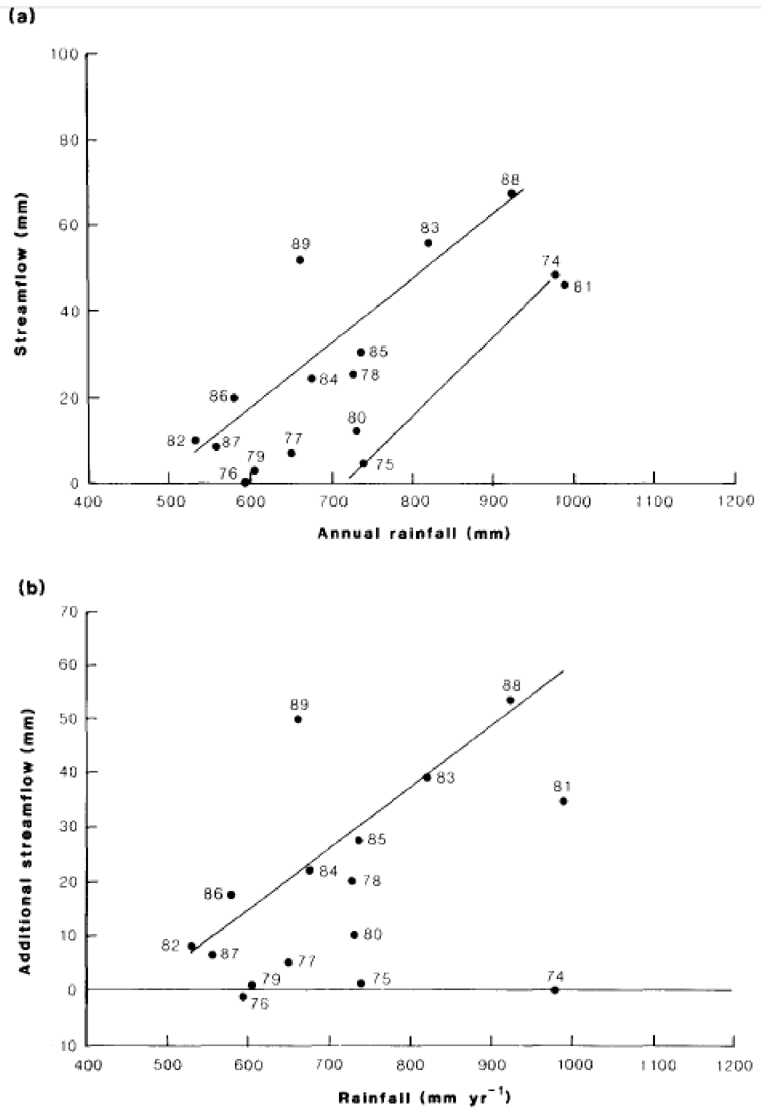

Figure 4.16 Streamflow response to clearing of Lemon catchment: (a) streamflow to rainfall relationship; (b) additional streamflow to rainfall relationship

The additional streamflow generated from Lemon catchment as a result of clearing is shown in Fig. 4.16(b). The additional streamflow was determined by using the 3 year pre-treatment record between Lemon and Ernies (control) catchments to derive a pre-treatment relationship. This relationship was based on a regression of Ernies streamflow on Lemon streamflow, derived by Williamson et al. (1987), when both catchments were forested. This regression equation was used to predict the annual streamflow from Lemon with no clearing. This analysis also suggests that a hydrological quasi-equilibrium had been reached, forming a linear streamflow to rainfall relationship and that 1989 was an outlier to this relationship.

The streamflow hydrograph comparison between Lemon catchment and Ernies catchment for a pre-treatment year (Fig. 4.17(a)) shows the similar hydrograph response. The comparison of streamflow for Lemon and Ernies (Fig. 4.17(b)) for 1988 shows that the major change due to clearing is the earlier commencement of streamflow. However, when Ernies catchment does commence flowing, the hydrographs are very similar. In 1989 (Fig. 4.17(c)), Lemon catchment streamflow is significantly greater than Ernies streamflow over the entire record. From Figs. 4.17(b) and 4.17(c) it is apparent that Lemon catchment became perennial over the summer of 1988/1989 in contrast to the ephemeral Ernies catchment for all years and Lemon catchment pre-1989. Table 4.3 lists the annual streamflow and rainfall over the study period. From Table 4.3, 1989 has the highest streamflow value as a percent of annual rainfall.

Stream salinity

The flow-weighted mean annual stream Cl^- concentrations (mg L^{-1}) shown in Fig. 4.19(a) highlight the substantial rises in stream salinity from 1987 onwards and in particular 1989. The average pre-treatment Cl^- concentration was 29 mg L^{-1} which was only slightly lower than the 1978 to 1986 average of 35.4 (± 3.5) mg L^{-1}. However, from 1987 to 1989 there was an order of magnitude increase to 286 mg L^{-1}. The annual stream Cl^- concentrations for Lemon catchment relative to streamflow are shown in Fig. 4.19(b). The significant increases for 1988 and 1989 are apparent.

The relationship between groundwater level and stream Cl^- is shown in Fig. 4.18(c). From 1978 to 1986 the influence of depth to groundwater in the valley of Lemon catchment is negligible. In 1987, one bore approximately 100 m upslope of the catchment outlet had a piezometric level above the ground surface and by 1989 the valley floor became an extensive groundwater discharge area (approximately 1 ha). The stream Cl^- chemograph (Fig. 4.19(a)) for 1986 is very low with no major changes with streamflow. In 1989 (Fig. 4.19(b)), there were major fluctuations in stream Cl^- relative to streamflow, ranging up to 2200 mg L^{-1}.

Catchment salt balance

The atmospheric input of salt (saltfall) and the Cl^- load in the streamflow for Lemon catchment are shown in Table 4.3. The mean saltfall was 3.1 (± 1.2) g m^{-2} Cl^-, with a mean concentration of 4.4 mgL^{-1} Cl^-. Prior to catchment clearing, the output to input ratio (O/I) of salt averaged 0.13 and ranged from 0.01 to 0.31. In the 10 years after clearing but prior to ground water reaching the soil surface (1977 to 1986), O/I ratios were higher, with an average of 0.36 and a range of

0.05 to 0.96. 1988 shows an O/I ratio greater than one, while in 1989 the O/I ratio increased to 5.3.

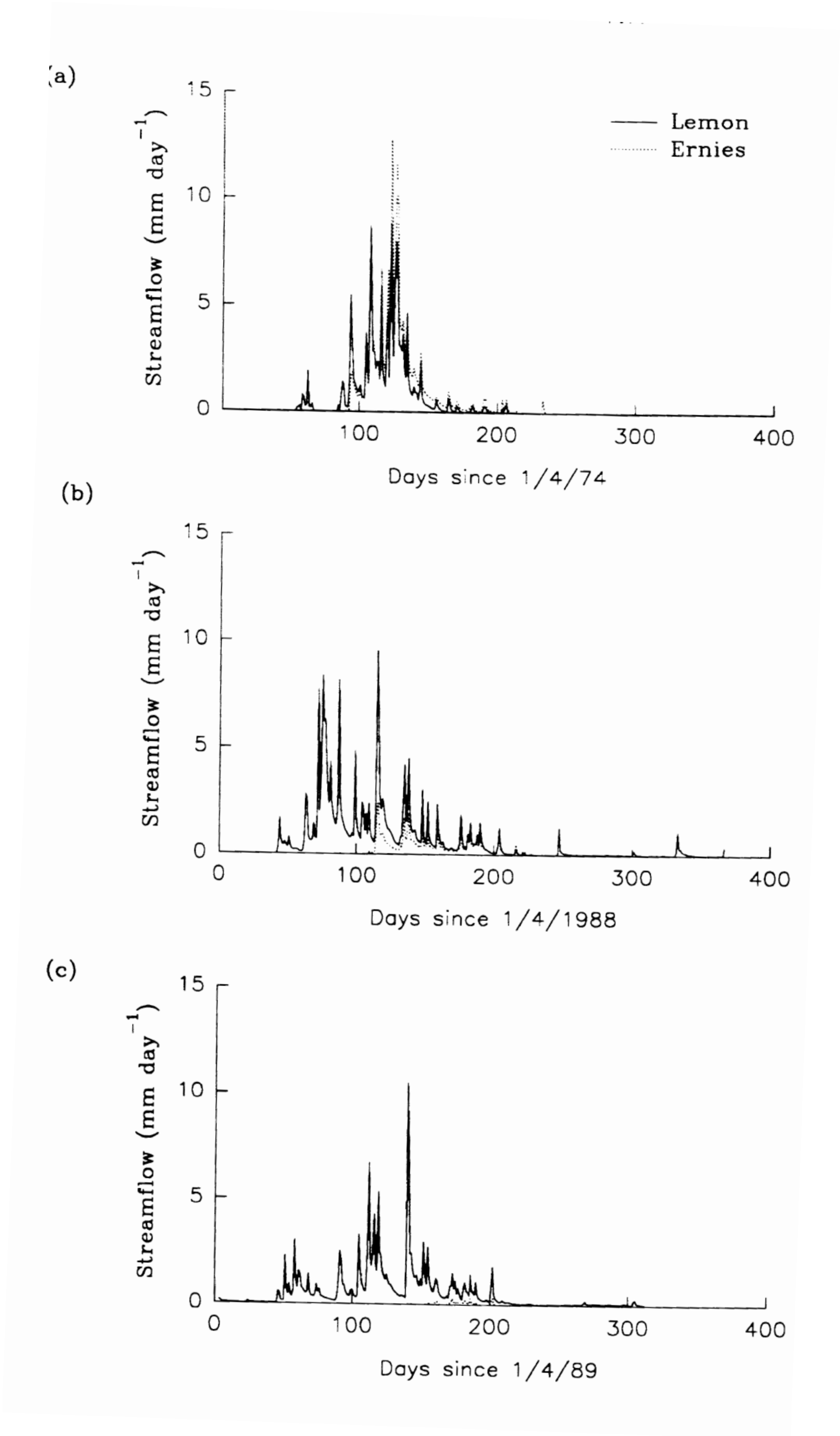

Figure 4.17 Streamflow hydrographs for Lemon and Ernies catchments: (a) 1974; (b) 1988; (c) 1989

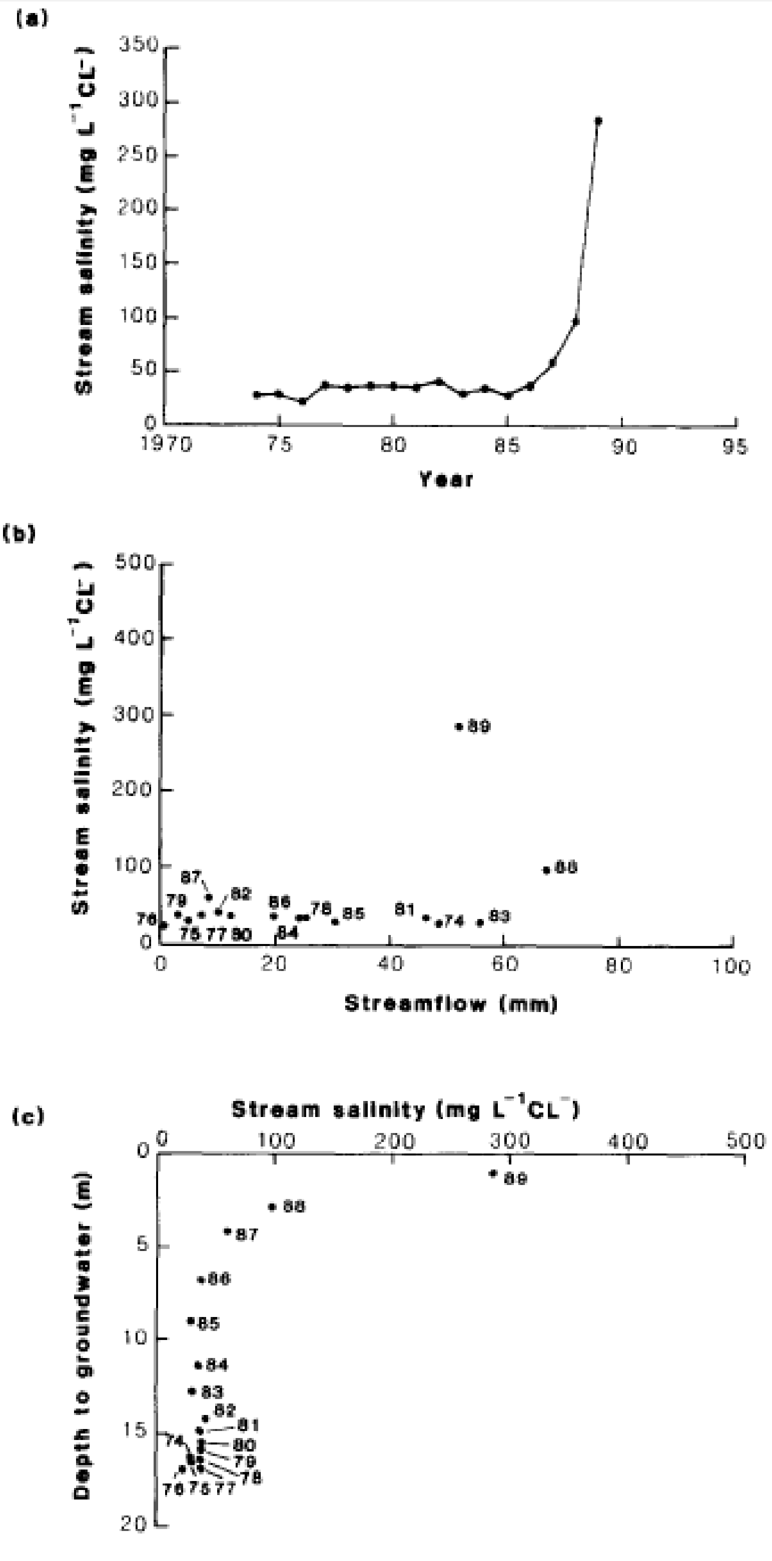

Figure 4.18 Stream salinity response to clearing of Lemon catchment: (a) annual stream salinity with time; (b) stream salinity to streamflow relationship; (c) stream salinity increase relative to depth to groundwater

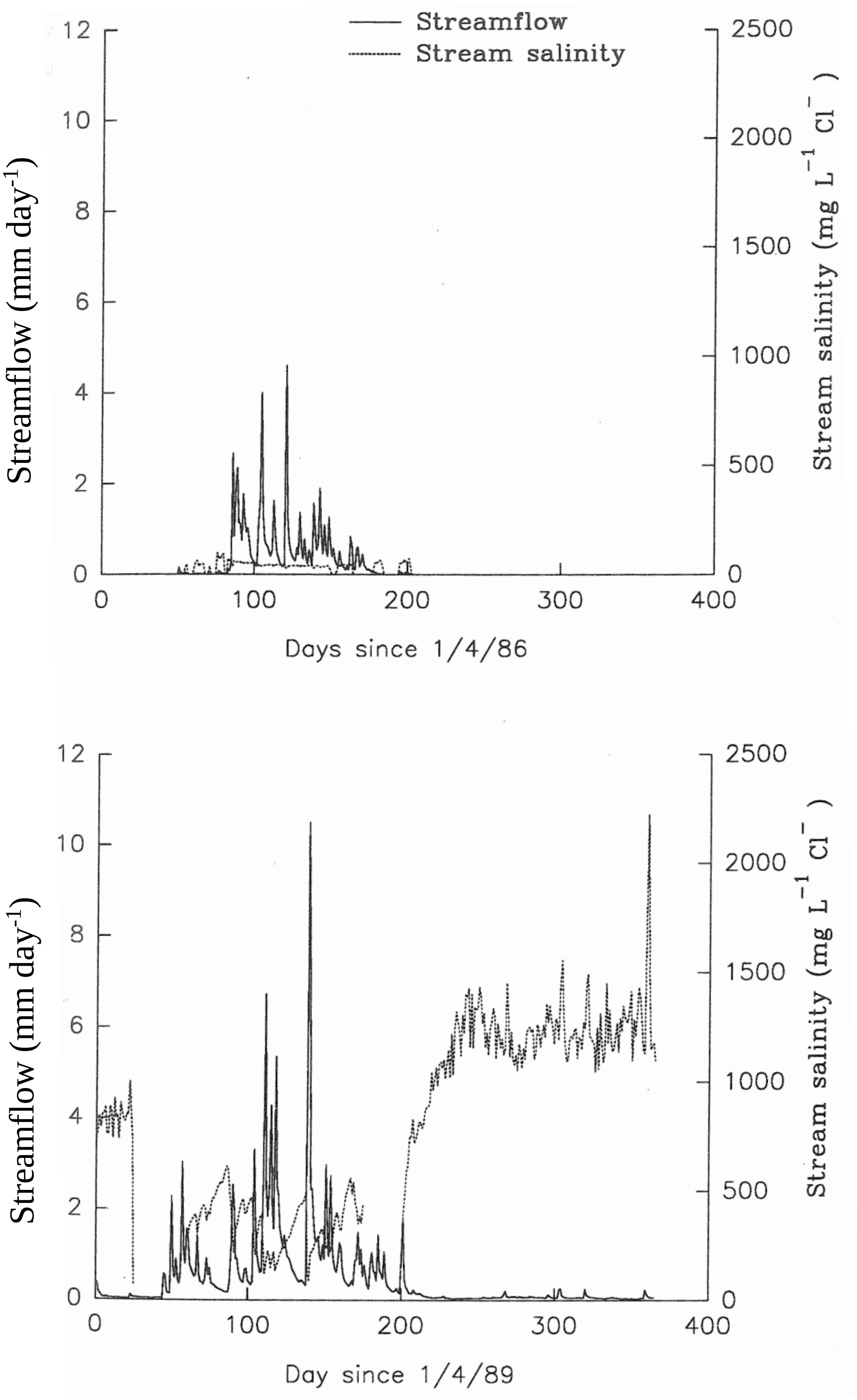

Figure 4.19 Daily streamflow and stream salinity hydrographs for Lemon catchment: (a) 1986; (b) 1989

4.2.6 Discussion

Groundwater rise due to clearing

Rates of rise in groundwater levels averaging 2.3 m yr^{-1} from 1986 have been observed in Lemon catchment on the cleared valley floor and lower slopes. These rates of rise are higher than those reported for Wights catchment (Peck and Williamson 1987a) in a higher rainfall area (more than 1200 mm yr^{-1}). The faster rates of rise at Lemon may be due to the fact that there was no groundwater discharge to the stream (as opposed to Wights) and consequently all the recharge (until 1988 at least) had been manifest in groundwater level rise.

Hookey (1987), using a two-dimensional, finite difference groundwater model, predicted that Lemon catchment would begin discharging groundwater to the stream 12 years after clearing (1989) and would not reach equilibrium for 30 years, by which time the seepage area would have grown to 19 ha. The rates of groundwater rise were not directly affected by the annual rainfall. Despite low rainfall in 1986 and 1987 (approximately 150 mm below average), the rate of groundwater level rise increased.

Groundwater recharge

Estimates of groundwater recharge from native forest catchments such as Ernies and Lemon prior to clearing, are in the order of 2 mm yr^{-1} (Johnston 1987b; Ruprecht *et al.* 1990). Turner *et al.* (1987b) found little vertical downward movement of unsaturated water below 6 m depth, based on tritium concentrations. Groundwater recharge has not been measured directly in this experiment. Indirect estimates of recharge based on rates of change of piezometric level have not been attempted either because observed values would result from a combination of local vertical recharge and downslope movement of groundwater.

Streamflow response to clearing

The streamflow response to clearing consisted of three identifiable stages. The initial stage (1977 to 1981) showed a small but variable increase in streamflow. This may reflect a time over which soil water equilibrated. (Sharma *et al.* 1987b) found that the unsaturated zone soil water had approached equilibrium after 2 years in the high rainfall catchment cleared for agriculture (Wights) and concluded that the equilibrium time for Lemon catchment would be longer. For the period 1982 to 1988 a second stage was evident when a reasonably stable relationship is reached between the annual streamflow and rainfall. In 1989 a third stage was apparent where the influence of the groundwater discharge area on streamflow generation becomes important. The large increase in streamflow for 1989 is considered to be a result of additional throughflow and saturation excess overland flow generated from the newly emergent impermeable groundwater table, as described by Ruprecht and Schofield (1989a) (Section 4.1).

Streamflow hydrograph response to clearing

The streamflow hydrographs of Lemon catchment after clearing but prior to groundwater discharge (1978 to 1988), show that Lemon catchment commenced flowing after only 100 mm of rainfall (April to June) while the forested Ernies catchment required approximately 400 mm to initiate streamflow. It is believed that this difference is due to the small soil water deficits created in Lemon catchment owing to the shallow-rooted pastures transpiring comparatively little water during summer and autumn (Schofield *et al.* 1989a). In 1989, when ground waters commenced discharge to the stream, Lemon streamflow became perennial.

Stream salinity

Stream salinities are increasing in many rivers within SWWA with significant amounts of agricultural clearing (Schofield *et al.* 1988). From basin studies of stream salinity (Collins and Barrett 1980; Collins and Fowlie 1981) a catchment with 600 to 900 mm rainfall and 50% cleared could be expected to reach an average annual salinity of approximately 800 mg L^{-1} Cl^{-}. However, similar catchments within the Wellington Reservoir catchment have average stream salinities of approximately 2700 mg L^{-1} Cl^{-}. Consequently there is the potential for further significant increases in stream salinity from Lemon catchment.

Salt balance

Prior to clearing, Lemon catchment had been accumulating salt in the soil profile for at least 7000 to 8000 years (Johnston 1987a). In the first 10 years after clearing, the O/I ratio increased from 0.12 to 0.36 which is probably attributable to the increase in streamflow. However, when the groundwater reached the surface there was a substantial increase in the salt export to 14 Cl^{-} g m^{-2} for 1989. The O/I ratio for Lemon catchment in 1989 was 5.3. This compares with an O/I ratio for Wights catchment of 10 to 15, 2 to 5 years after clearing (Williamson *et al.* 1987). The total salt storage of Lemon catchment is equivalent to 7000 to 8000 times the annual input and about 1200 times the 1989 salt output. Peck and Hurle (1973) determined that the characteristic time (ratio of salt storage to salt discharge) for equilibrium to a new salt balance following the partial clearing of Lemon catchment would be approximately 200 years.

4.2.7 Conclusions

Deforestation and establishment of pasture on the downstream 53% of Lemon catchment has caused groundwater levels to rise, a seep area to develop and stream salinity to increase. Groundwater levels in the valley floor have risen at an average rate of 2.3 m yr^{-1}, with a maximum of 4.8 m yr^{-1} for 1986 to 1989. The ground water intersected the ground surface 12 years after clearing and is continuing to rise beneath the lower slopes. A seep or groundwater discharge area of 1 ha had developed by 1989 (12 years after clearing).

Streamflow increases occurred in three distinct stages following clearing. The latter and most significant stage was clearly identified as corresponding to the development of the seep area. Stream salinity has increased dramatically since 1987 in association with the approach of the saline ground water to the soil surface and stream zone. The salt output to input ratio of the catchment has increased from 0.13 to 5.3; that is from salt accumulation to net salt export. The characteristic salt leaching time for the catchment has been estimated at 200 years. In 1989 the ground water level was rising, the seep area was growing, streamflow was increasing, stream salinity was increasing and salt export was increasing. This behaviour is likely to continue for some years.

The next section of this chapter examines the impact of targeted forest clearing for agriculture to evaluate the impact on the hydrology and salinity in comparison to the extensive clearing studied in this section.

4.3 Strip, soils and parkland clearing[7]

4.3.1 Abstract

A 300 ha experimental catchment in a high salt storage landscape in SWWA was subject to several different clearing treatments to determine their respective impacts on groundwater levels. Within subcatchments that were 60 to 70% cleared, groundwater level rises were observed of 7.8 to 10.2 m compared with a 5.8 m rise with 32% clearing and a 2.3 m fall in a native forest control, from 1977 to 1989. Under parkland clearing (leaving wide-spaced trees), the rate of groundwater level rise has declined in recent years in response to increasing crown leaf areas, and in contrast to other treatments (strips and block clearing) where groundwater rise has accelerated. The presence of a bedrock high at the valley invert within one subcatchment has led to substantial groundwater rises in recent years which could intercept the surface in the next 5 years if the current rates of rise are maintained.

The combination of clearing treatments affecting 38% of the catchment has resulted in a modest (13 mm) increase in catchment streamflow. This was over double the average forested streamflow. Stream salinity has remained low. Stream yield, stream salt load and stream salinity appear to be at equilibrium. However, as a result of considerable expanses of clearing located close to or on the streamline or valley invert, there is a significant risk of salt discharge to the stream in the future.

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[7] Published as: **Ruprecht, J.K**. and Schofield, N.J., 1991. Effects of partial deforestation on hydrology and salinity in high salt storage landscapes. II. Strip, soils and parkland clearing. *Journal of Hydrology*, 129: 39-55.

4.3.2 Introduction

The mechanisms of stream salinization following extensive forest clearing for agricultural development in Western Australia have been described in Section 4.2. This Section describes the effects of three alternative clearing strategies, termed 'strip', 'soils' and 'parkland' clearing and their combined effects. These treatments were carried out in Dons catchment, which is adjacent to Lemon and Ernies catchments in the eastern Collie basin (see Fig. 4.12). One of the original aims of the project was to retain a distribution of forest cover that would not significantly impact the water balance and stream salinity (Peck and Williamson 1987b).

4.3.3 Site Description

The general site characteristics were given by Bettenay *et al.* (1980) and Ruprecht and Schofield (1991). The native vegetation on Dons catchment prior to treatment was a jarrah - marri forest with a basal area of 19 to 24 m^2 ha^{-1} (Bettenay *et al*. 1980). The basal area was considerably less (8 to 16 m^2 ha^{-1}) in the valley floors, where only occasional flooded gum occurred, and in the wandoo woodlands. There were small areas that were treeless.

The average total soil salt storage was 25 kg m^{-2} Cl^- for Dons catchment (Turner *et al.* 1987a). The average profile maximum Cl^- solute concentration was approximately 8000 mg L^{-1} Cl^-. Measurement of salinity and salt storage are expressed in terms of Cl^- for this paper unless otherwise stated.

4.3.4 Instrumentation

The instrumentation of Dons catchment was similar to that of Lemon and Ernies catchments. Rainfall was continuously measured at a suitably exposed site close to the gauging station and at a second site just outside Dons catchment to the southwest (Fig. 4.20). Streamflow was measured at a sharp-crested V-notch weir at both Dons and Ernies catchments. The electrical conductivity of stream water was continuously monitored using a toroidal cell (Brown and Hamon 1961). The piezometric network on Dons catchment (Fig. 4.20) included bores located on a 200 m^2 grid and line transects. Additional piezometers were placed across the catchment outlet. Groundwater levels were monitored on a monthly to 3-monthly basis. Groundwater salinity was measured simultaneously with the groundwater level measurement until 1989; thereafter groundwater salinity was measured at 6-monthly intervals.

4.3.5 Treatments

The treatment strategies comprised three categories - strip clearing, soil clearing and parkland clearing (Fig. 4.20). The clearing of native forest for the different treatments commenced in November 1976 and was completed by March 1977. There was some regrowth control needed in the first few years after clearing. The cleared areas were mainly sown to grass and pasture although there was some initial cropping of oats and barley.

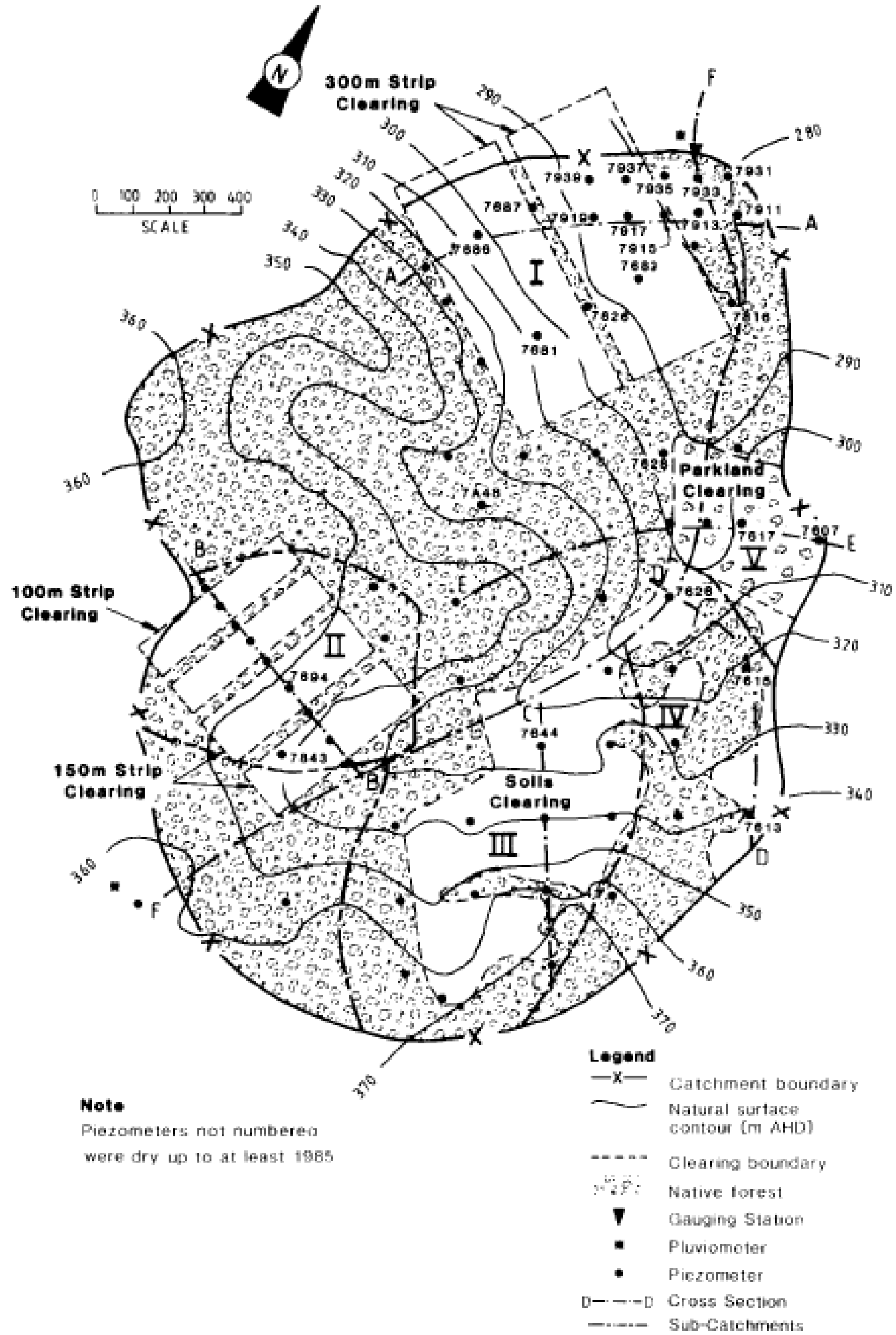


Figure 4.20 Dons catchment map

Strip clearing

The strip clearing was carried out parallel to surface contours and presumed normal to the expected groundwater flowlines. Following removal of the native vegetation, the land was developed for pasture. Two different treatments were used and are identified in Fig. 4.20 in subcatchments I and II.

These subcatchments have the following characteristics.

Subcatchment I

Two large strips of 300 m width (parallel to surface contours) were cleared, leaving a 30 m uncleared strip between them. The total area of the subcatchment is 53 ha, of which 35 ha (67%) was cleared. About 4 ha (7.5%) of native vegetation was left between the streamline and the first clearing line.

Subcatchment II

Two strips of 100 m width and two strips of 150 m width were cleared, leaving three 30 m uncleared strips between them. The total area of the subcatchment is 40 ha, of which 24 ha (60%) has been cleared.

Soils clearing

Two different clearing strategies based on soil type were tested. The first (subcatchment III, Fig. 4.20), termed 'downslope soils clearing', included leaving the native vegetation on the lateritic duricrust soil type on the ridges and spurs and clearing the middle and lower slope gravels and sands. The area cleared comprised 62% of the subcatchment. The second strategy (subcatchment IV), termed 'upslope soils clearing' involved leaving the lower slope gravels and sands uncleared and clearing the lateritic duricrust. The area cleared comprised 32% of the subcatchment.

Parkland clearing

The parkland clearing treatment (subcatchment V) reduced the forest to a 20 x 20 m grid of trees and to a basal area of 1/10th of the native forest (J.K. Marshall, personal communication, 1988). The total area of subcatchment V was 13.4 ha. The area of clearing was approximately 90% of the subcatchment.

4.3.5 Results

Rainfall and salt fall

The average annual rainfall for Don's catchment between 1974 and 1989 was 701 mm (based on the average of the two pluviometers) which was 3% below the long-term average (1926 to 1979) of 720 mm (based on interpolation of isohyets). Annual Cl^- precipitation averaged 3.0 g m^{-2} with a coefficient of variation of 43% over the study period.

Groundwater

Of the 67 piezometers installed within Dons catchment, only 26 (30%) were useful for analysis, the others remaining dry. As a result, there were typically only one or two non-dry piezometers within each treatment, except for subcatchment I which had ten non-dry piezometers. Within Dons catchment there was only one piezometer with adequate data for a control bore within remaining native forest. At this site (7A48, Fig. 4.20) the groundwater level reduced by 2.3 m from 1977 to 1989, corresponding to a rate of decline of 0.2 m $year^{-1}$. This compares with a 2.0 m reduction at a piezometer within the remaining native forest in Lemon catchment (Section 4.2) and a 1 to 2 m groundwater level reduction within the forested Ernies catchment over the same period.

Subcatchment I- strip clearing

The typical groundwater response from a bore within subcatchment I is shown in Fig. 4.21. The piezometric levels on a transect (A-A) within the 300 m strip clearings are shown in Fig. 4.22(a). The natural surface and bedrock information has been interpolated from the drilling information at the piezometer locations. Prior to clearing, the groundwater level was relatively flat with a groundwater gradient of 0.2%. By 1988, a slight groundwater mound had formed underneath the 300 m strip clearing. The average groundwater rise from 1977 to 1989 in the 300 m strip clearing was estimated at 8.3 m (Table 4.4) with a range of 5.7 to 10.8 m. Bores downslope of the 300 m strip clearing showed groundwater rises ranging from 5.1 m at the clearing boundary down to 1.1 m in the valley floor.

These rises are in contrast to groundwater reductions of 1 to 2 m in native forest unaffected by clearing. The initial average rate of groundwater rise for subcatchment I was 0.24 m yr^{-1} (Table 4.4). However, the rate of rise had increased to 0.96 m yr^{-1} in recent years. The depth to the groundwater table was still at least 14 m in 1989.

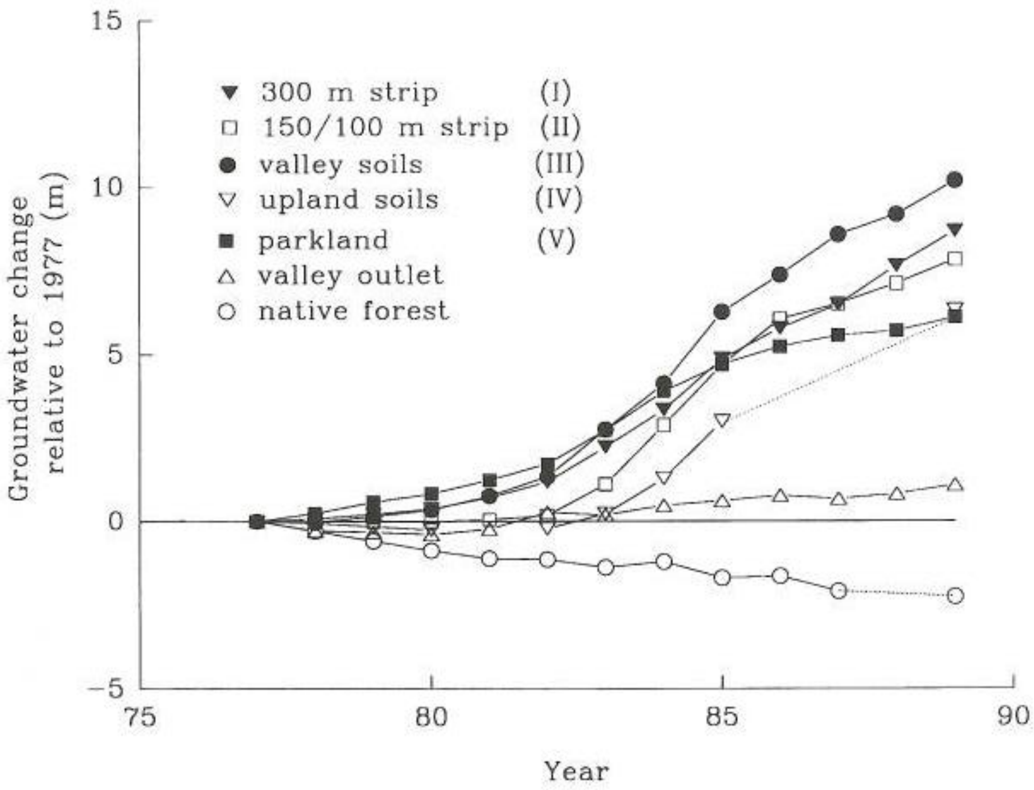


Figure 4.21 Groundwater rises of selected bores within each of the subcatchments

Table 4.4 Comparison of subcatchment groundwater response

Subcatchment	Treatment	Area cleared (%)	Initial depth to groundwater (m)	Total groundwater rise (m)	Rate of rise (m $year^{-1}$)	
					Initial(1)	Recent(2)
I	300 m strip	67	30.0	8.3	0.24	0.96
II	150 m strip	60	37.0	7.8	0.04	0.59
III	Lower slope soils	62	23.4	10.2	0.27	0.93
IV	Upland soils	32	35.5	5.8	-0.09	0.83(3)
V	Parkland	90(4)	22.4	6.1	0.35	0.28
Control	None	0	25.0	-2.3	-0.58	-0.20
Catchment outlet	Varied	38	13.5	1.1	-0.06	0.11

(1) Initial rate based on 1977 to 1981
(2) Recent rate based on 1985 to 1989
(3) Recent rate based on 1984 to 1989, no data for 1985 to 1989
(4) Initial area cleared based on parkland clearing to 1/10th of native forest basal area. The compensating leaf area growth of the remaining trees has been estimated to increase leaf cover to approximately 50%

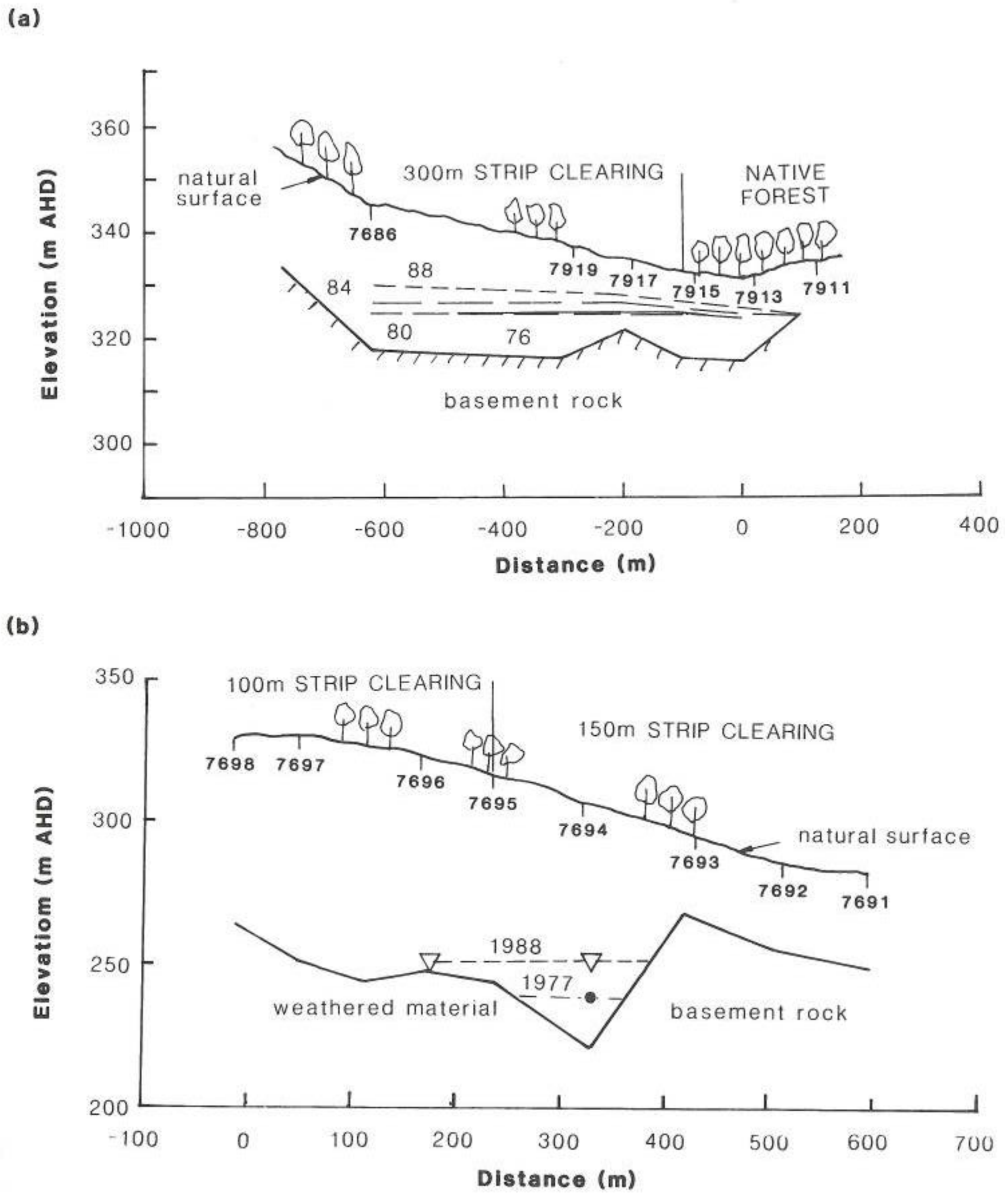


Figure 4.22 Groundwater response from different clearing treatments based on transects: (a) subcatchment I, 300 m strip clearings; (b) subcatchment II, 100 and 150 m strip clearings

Consequently, assuming that the rate of groundwater level increase remains constant, groundwater would not be expected to reach the ground surface for a further 15 to 20 years from 1989.

Subcatchment II strip clearing

Fig. 4.21 shows a typical groundwater response within the 100 to 150 m strip clearing treatment. Fig. 4.22(b) depicts transect B-B (Fig. 4.20) through the 100 and 150 m strip clearings. Soil profile depths downslope of bore 7693 are about 30 m while upslope of this bore, soil depths are of the order of 60 m. Prior to the clearing treatments there was only one bore in which ground water was observed (7694). From this it was inferred that there was no ground water immediately downslope of bore 7693. The groundwater levels had risen by 7.8 m at bore 7694 in the 11 years since clearing to 1988. By 1988 the groundwater level had risen to be close to the bottom of the bores upslope of bore 7694 (Fig. 4.22(b)). The current rate of groundwater rise at bore 7694 is 0.59 m $year^{-1}$ (Table 4.4). However, the depth to the groundwater table at this location is still 27 m and thus ground water is unlikely to reach the surface for many years.

Subcatchment III- soils clearing

Fig. 4.23(a) shows transect C-C (Fig. 4.20) through the lower slope soils clearing. The three bores upslope of bore 7644 had relatively shallow basement rock (15 to 28 m), while bore 7644 had a very deep soil profile (more than 40 m). Groundwater level responses were only observed in bore 7644 within the lower slope soil clearing (Fig. 4.21). At this location, groundwater level rises of 10.2 m were observed from 1976 to 1989 with a recent rate of rise of 0.93 m $year^{-1}$ (Table 4.4). As with subcatchment II, a groundwater system was not detected above the shallower bedrock and it would appear that any water which infiltrated to the bedrock upslope of bore 7643 flowed into the 'sink' surrounding bore 7644.

Subcatchment IV- soils clearing

Fig. 4.23(b) shows transect D-D through the upslope soils clearing. Bore 7613 which is within the upslope soil clearing showed a rise of 6.29 m from 1977 to 1989, corresponding to a rise of 0.5 m year^{-1}. However, the depth to ground water was still 27 m by 1989. The groundwater level at bore 7615, within native forest, increased by 5.3 m over the 12 years since clearing.

Subcatchment V - parkland clearing

The groundwater rise in response to the parkland clearing was less than in subcatchments I, II or III (Fig. 4.21). Fig. 4.24(a) shows the groundwater response for transect E-E. The transect was not aligned in the flow direction owing to the lack of groundwater observation bores. However, the transect alignment was still considered to be applicable. The transect would indicate that groundwater flow was away from the valley floor and across the catchment boundary. The groundwater levels rose by 6.1 m beneath the parkland clearing during 1976 to 1989. The initial rate of groundwater rise was 0.35 m yr^{-1}. There was then an increase in the rate of groundwater rise to 0.86 m yr $^{-1}$ from 1982 to 1986, followed by a decrease to 0.28 m yr^{-1} (Table 4.4) for 1986 to 1989.

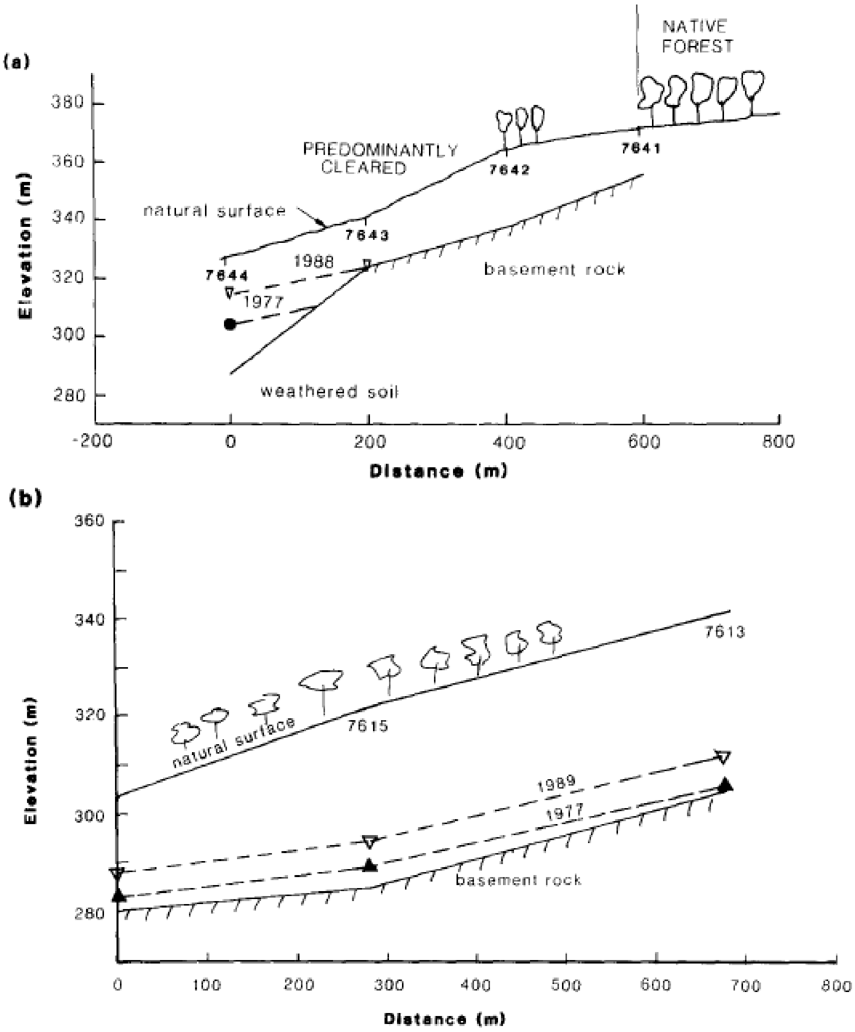


Fig. 4.23 Groundwater response from different clearing treatments based on transects: (a) subcatchment III, lower slope clearing; (b) subcatchment IV, upland soils clearing

Catchment transect

Fig. 4.24(b) shows a transect (F-F) along the catchment valley floor. This figure highlights the variation in bedrock topography and the influence of bedrock highs and lows. At the commencement of the treatments (1977), the ground water occurred beneath the downstream valley and in an upstream bedrock low. At the bedrock high, adjacent to a dolerite dyke, in the soils clearing area (bore 7635), ground water has been 'forced' closer to the ground surface than elsewhere. At the current rate of groundwater rise of over 1 m year^{-1}, at bore 7635, ground water will reach the surface within 5 years.

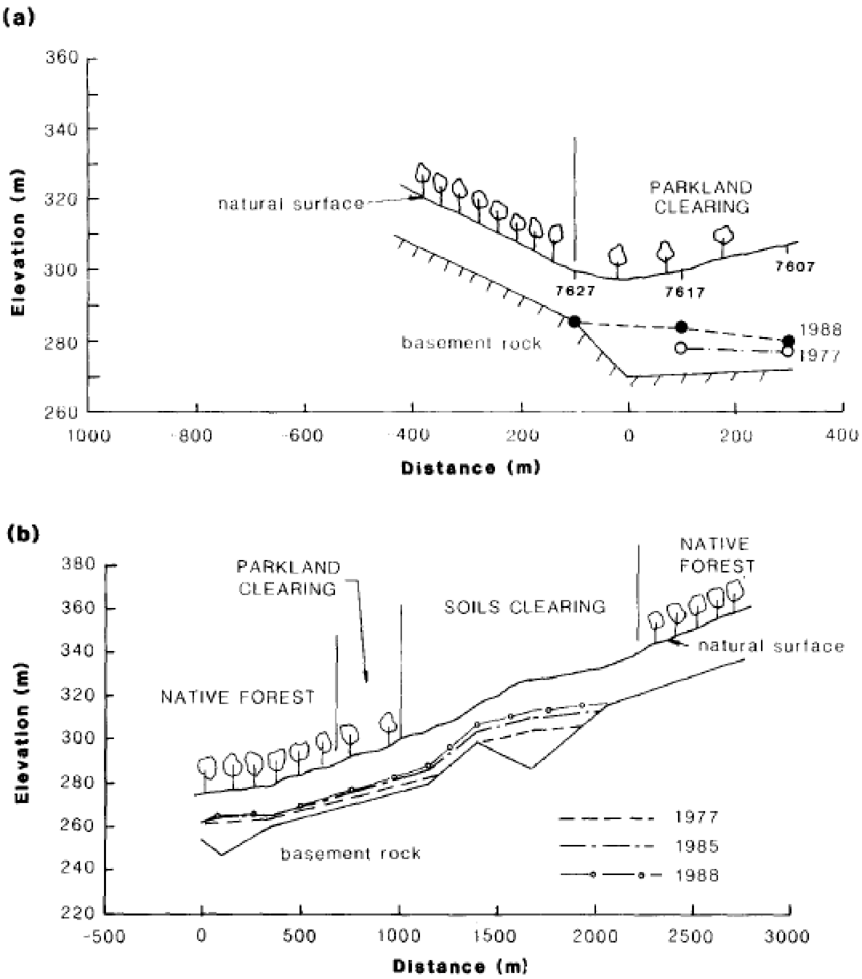
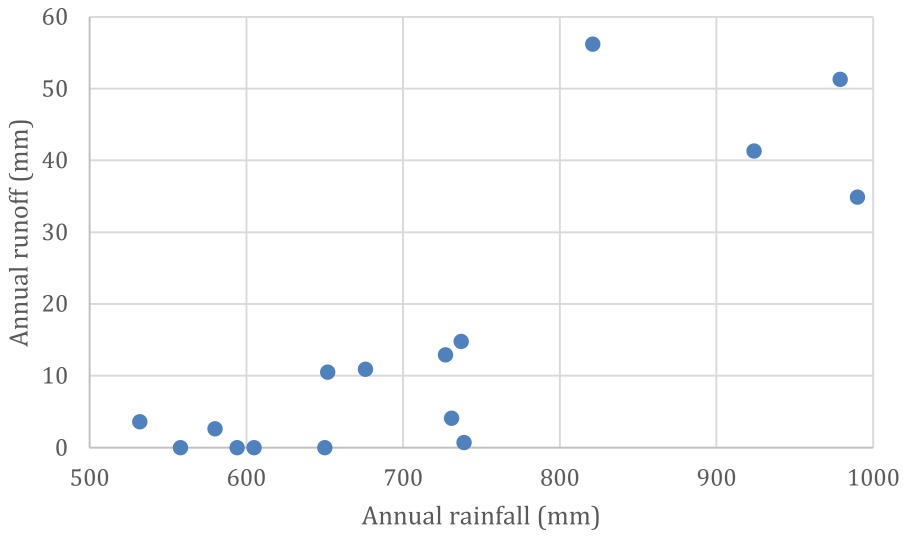


Fig. 4.24 Groundwater response from different clearing treatments based on transects: (a) subcatchment V, parkland clearing; (b) valley invert transect

Streamflow

The annual streamflow to rainfall relationship shown in Fig. 4.25(a) highlights some changes which have occurred with the partial clearing of Dons catchment, which in total corresponded to a clearing area of 38% of the catchment. Prior to clearing, streamflow did not occur for an annual rainfall of 739 mm (1975). After clearing, streamflow was observed with only 532 mm annual rainfall (1982). The additional streamflow generated following the clearing is shown in Fig. 4.25(b). The additional streamflow was determined by using the 3 year pre-treatment record of Dons and Ernies (control) catchments to derive a pre-treatment regression equation (Williamson et al., 1987). This regression equation was used to predict the annual streamflow from Dons with no clearing. This analysis confirms that additional streamflow was generated from Dons catchment as a result of clearing. However, there is still considerable variability in the data. The additional streamflow for an average rainfall year was approximately 13 mm which was 210% of the average forested streamflow for Dons and 1.8% of average annual rainfall. The comparison of streamflow hydrographs for a pre-treatment year (1974) and after clearing (1988) for Dons and Ernies are shown in Figs. 4.26(a) and 4.26(b). The pre-treatment hydrographs show very good agreement. After clearing, the major change is that Dons catchment starts to flow earlier than Ernies. Once both catchments have started to flow there is no significant differences in the streamflow hydrographs.



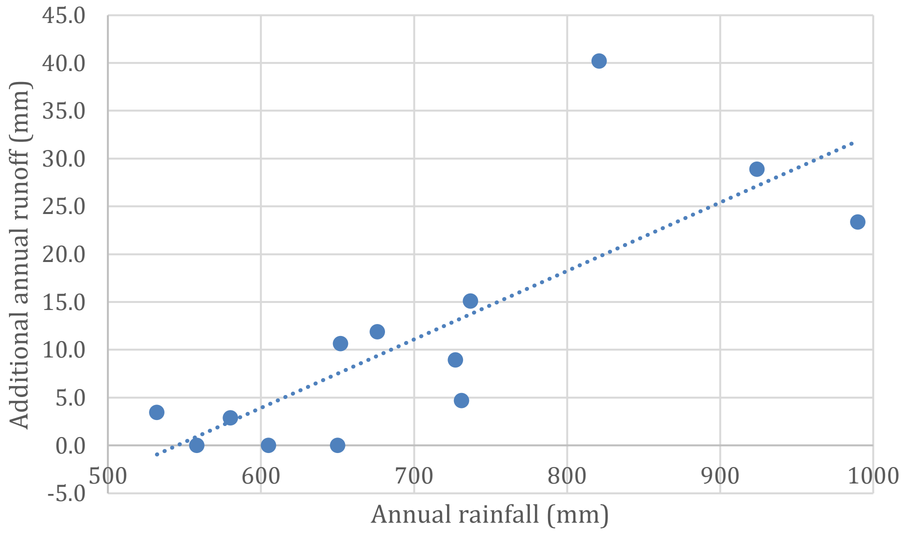


Figure 4.25 Streamflow relationships: (a) annual streamflow to rainfall relationship; (b) additional streamflow with respect to rainfall following clearing

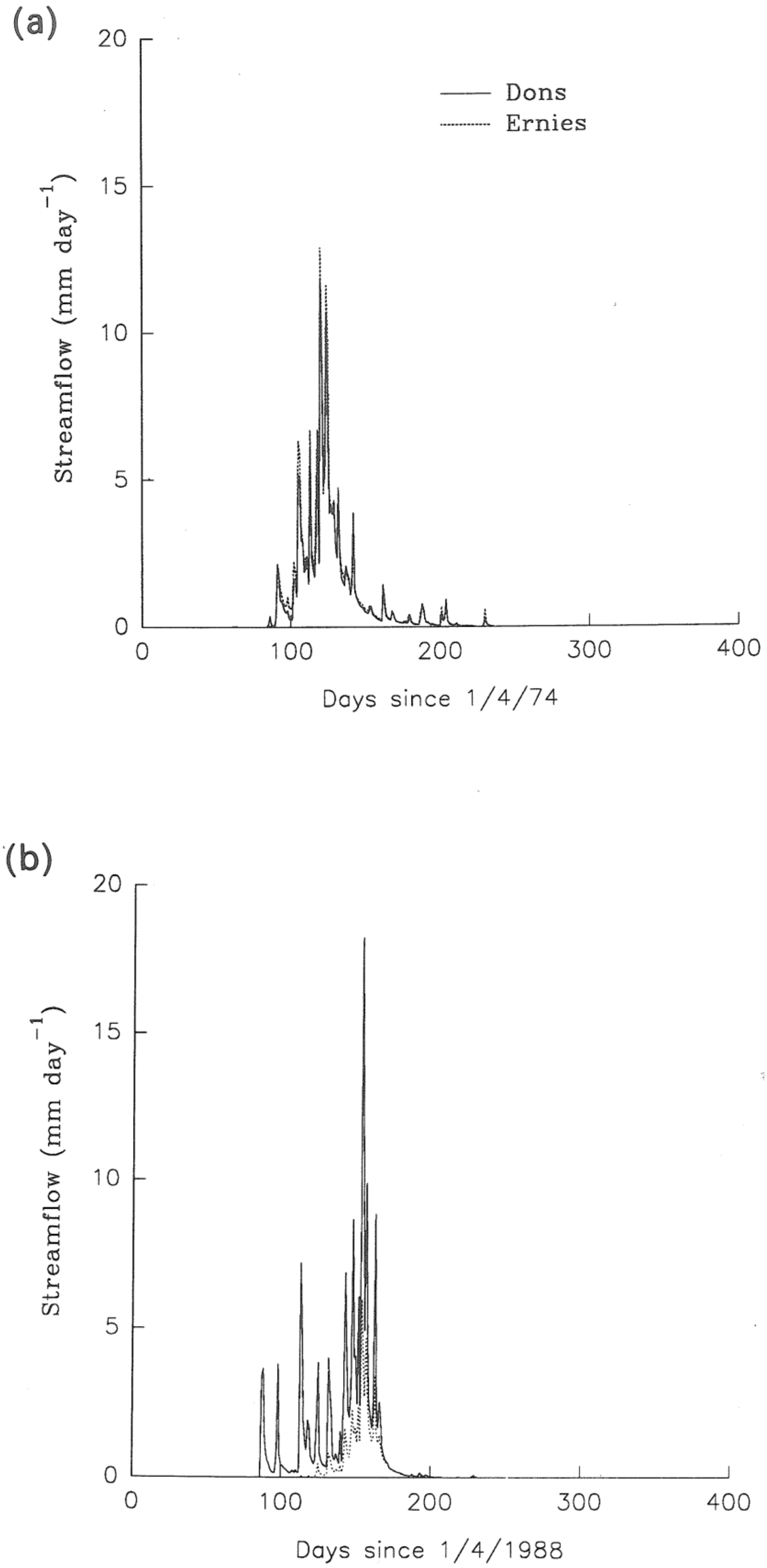


Figure 4.26 Streamflow hydrographs of Dons and Ernies catchments for: (a) 1974; (b) 1988

Stream salinity

The annual stream salinities for Dons catchment do not show any significant trend following partial clearing. They range from 20 to 40 mg L^{-1} Cl^-. The average daily stream salinity ranges from 21 to 48 mg L^{-1} Cl^- for pre-clearing and from 20 to 81 mg L^{-1} Cl^- for post-clearing.

Table 4.5 Dons catchment water and chloride balances

Year(1)	Rain	Streamflow		Saltfall(2)	Chloride flow		
	mm	mm	% rain	g m^{-2} Cl^-	g m^{-2}	mg L^{-1}	O/I(3)
1974	979	51.3	5.2	3.8	24	1.23	0.33
1975	739	0.7	0.1	2.5	38	0.02	0.02
1976	594	0	0	1.6	-	0	0
1977	650	0	0	2.4	-	0	0
1978	727	12.9	1.8	2.4	35	0.45	0.20
1979	605	0	0	1.7	-	0	0
1980	731	4.1	0.6	3.2	41	0.17	0.07
1981	990	34.9	3.5	4.3	29	1.01	0.25
1982	532	3.6	0.7	2.6	38	0.14	0.07
1983	821	56.2	6.8	6.8	26	1.44	0.22
1984	676	10.9	1.6	2.4	39	0.43	0.18
1985	737	14.8	2.0	2.8	18	0.27	0.10
1986	580	2.6	0.4	1.9	38	0.1	0.05
1987	558	0	0	2.3	-	0	0
1988	924	41.3	4.5	3.9	20	0.83	0.21
1989	652	10.5	1.6	2.4	23	0.24	0.10

(1) Water year commencing 1 April

(2) From 1974 to 1983 the input in saltfall was taken from observations reported in Williamson *et al.* (1987). From 1984 to 1989 saltfall was estimated by assuming a constant chloride concentration in saltfall of 4.1 mg L^{-1}

(3) Chloride output to input ratio

Salt balance

The atmospheric input of Cl^- (saltfall) and the chloride load in streamflow for Dons catchment are shown in Table 4.5. The mean saltfall was 3.1 (± 1.2) g m^{-2}, with a mean concentration of 4.4 mg L^{-1}. Prior to the clearing, the output to input ratio (O/I) for Cl^- had ranged from 0.33 to 0.00 with an average of 0.12. For the 10 years after clearing, the O/I ratio had ranged from 0.25 to 0.00 with an average of 0.11. Consequently there does not appear to be any significant increase in salt load as a result of the partial clearing of Dons catchment.

4.3.6 Discussion

Strip clearing

The two strip clearing treatments resulted in similar groundwater rises of 8.3 m (subcatchment I) and 7.8 m (subcatchment II). The percentages cleared for both subcatchments were 67 and 60%, respectively. At these levels of clearing, the groundwater response has been significant with respective rates of rise of 0.6 m yr^{-1} and 0.55 m yr^{-1}. There is also little evidence that these rates of rise are declining and, at current rates, ground waters will reach the ground surface in 15 to 20 years. The rates of rise at the valley floor are considerably lower and so the future impact on stream salinity is unclear.

Soil clearing

The upland soil clearing (subcatchment IV) had a substantially smaller groundwater rise than the valley soil clearing (subcatchment III). This was most probably due to the smaller level of clearing as percentage of forest cover has been closely correlated with groundwater level change (Bell *et al.* 1990; Schofield 1990). At a bedrock high within the downslope soil clearing, groundwater levels rose rapidly over the last 4 years of monitoring. The influence of bedrock highs and lows within the catchment is considered critical as to when ground water reaches the surface. The effect of the bedrock high may be similar to that of dolerite dykes, as reported by Engel *et al.* (1987) for catchments 100 km east of Dons, in that a reduction in groundwater flow occurs from a reduced transmissivity due to either a reduction in hydraulic conductivity or groundwater flow area.

Parkland clearing

The initial parkland clearing of subcatchment V resulted in a reduction in vegetation density to 10% of native forest. However, by 1984 (7 years after clearing), the leaf area of the remaining trees was nearly 5 times as great as the leaf area of comparable trees in the native forest (J.K. Marshall, personal communication, 1989). This increase in leaf area of the remaining trees within the parkland clearing is the probable cause of the reduction in the groundwater rise to 0.3 m $year^{-1}$ since 1986. This significant increase in crown leaf area can only occur under parkland clearing because trees are 'released' from competition. However, parkland clearing is not considered to be a viable long-term clearing strategy because the trees do not regenerate.

Comparison of the groundwater response between treatments

The percentage cleared of the subcatchments ranges from 32% for subcatchment IV to 90% initially for subcatchment V (see Table 4.4). Subcatchment IV had the lowest overall rise in groundwater level while the initial rate of groundwater rise was greatest within the parkland clearing. Three of the subcatchments had clearing percentages of approximately 60 to 70% (I, II and III). The current rates of groundwater rise are reasonably similar (Table 4.4) for these three subcatchments. The presence of remnant valley forest in subcatchment I has reduced the groundwater response at the streamline, compared with either subcatchments II or III.

In summary, the change in vegetation cover is considered to have the dominant effect on groundwater tables. The responses generally reflect magnitude, distribution and type of vegetation change. The bedrock topography appears to play a role in determining when and where groundwater seeps may develop.

Integrated effect of treatments on Dons catchment

The total area cleared for agriculture in Dons catchment was 38%. Although this clearing was distributed across the catchment, considerable expanses of clearing were located close to or on the streamline or valley invert. As a result of this, there is a significant risk of salt discharging to the stream in the future. The importance of replanting the stream zone and lower slopes to control salinity has been demonstrated by (Schofield *et al.* 1989a). Similarly, in the context of clearing, extensive stream buffers should be retained to prevent salt discharge. Unfortunately, this strategy was not adequately tested in the clearing scenarios.

Streamflow response to clearing

The streamflow response to the distributed clearing had two identifiable stages. The initial stage (1977 to 1981) showed a small but variable increase in streamflow. This was probably a time over which soil water equilibrated. From 1982 to 1989, a second stage is evident where the streamflow response is stable relative to rainfall. This was similar to Lemon catchment (see Section 4.2) until 1989 when substantial increases in streamflow were observed which coincided with groundwater levels intercepting the ground surface and a groundwater seep forming.

The change in hydrograph response between Dons catchment and Ernies catchment (Fig. 4.26(b)) was considered to be a result of the reduced soil water deficits experienced under the treated catchment compared with the native forest. Consequently, less water was required for streamflow to commence. However, once the soil water deficit had been reduced by winter rainfall in the native forest (Ernies catchment) then the hydrographs were very similar. This would indicate that the mechanisms of streamflow generation for Dons catchment were still similar to the native forest.

Salt balance

The salt balance for Dons catchment has not changed significantly as a result of the partial clearing. The O/I ratios have remained substantially below unity and this implies that salt is still being accumulated in the catchment. However, (Johnston 1987a) calculated that groundwater flow out of the catchment could be exporting a similar magnitude of salt as the input in saltfall.

4.3.7 Conclusions

Groundwater levels have risen substantially (6 to 10 m) in response to a range of vegetation treatments involving replacing 32 to 90% of native forest with pasture. Groundwater levels were still rising in 1990 under all treatments and are likely to intercept the ground surface at one point by 1995. Parkland clearing to 10% of original forest density resulted in the fastest initial rise in groundwater level but has more recently decreased to the slowest in response to increasing leaf area of the crowns. The retention of native forest on the valley floor in one treatment significantly depressed groundwater level rises beneath the streamline. Groundwater responses were dominated by the vegetation treatment in terms of its magnitude, distribution and type. Bedrock topography and geology were identified as important factors in determining when and where groundwater may intercept the ground surface. Soil type did not appear to be an important factor in the groundwater response to clearing. Clearing located close to or on the streamline or valley invert, has led to a significant risk of salt discharge to the stream in the future.

Streamflow increases as a result of clearing were estimated to be 13 mm compared with an estimated native forest average streamflow of 6 mm. There was a large variability in streamflow response from high rainfall years. Stream salinity has remained fresh and unchanged to date. The salt output to input ratio has remained relatively constant over the study period, varying from an average of 0.12 prior to clearing to 0.11 after clearing.

The next section of this chapter examines the impact of forest thinning on the hydrology and salinity in comparison to more permanent clearing for agriculture.

4.4 Early hydrological response to intense forest thinning in SWWA[8]

4.4.1 Abstract

A small forested catchment in SWWA was thinned to study the effect on hydrology, wood production and disease escalation. This paper deals primarily with the hydrological aspects. The uniform, intensive thinning treatment reduced crown cover from 60 to 14%, which resulted in an increase in streamflow of approximately 20% of annual rainfall (260 mm for an average year) after 3 years, compared with a streamflow yield of 6% of annual rainfall before thinning. The deep groundwater attained a new equilibrium after 2 years, rising by approximately 2 m in the area adjacent to the swamp, and by 5 m upslope. The ephemeral shallow groundwater system expanded in duration from 2 months per year pre-treatment to approximately 6 months per year after thinning. The major components of streamflow generation present before thinning remained as the major components after thinning. The expansion of the saturated source area and the presence of a shallow groundwater system for extended periods, as a result of an increase in available water from the reduction in interception and evaporation from the overstorey, were considered to be the major causes of increased streamflow.

[8] Published as: **Ruprecht, J.K**., Schofield, N.J., Crombie, D.S., Vertessy, R.A. and Stoneman, G.L., 1991. Early hydrological response to intense forest thinning in southwestern Australia. *Journal of Hydrology*, 127: 261-277.

4.4.2 Introduction

A major review of 94 catchment experiments to determine the effects of vegetation change on catchment yield (Bosch and Hewlett 1982) concluded that, in all cases except one, yield increased with decreasing forest cover or, conversely, yield decreased with increasing forest cover. The authors also suggested that the approximate magnitude of water yield change could be estimated; for example, eucalypt forest types yielded on average 40 mm yr^{-1} per 10% change in forest cover.

Attention was drawn, however, to the high variability of water yield responses, part of which could be explained by mean annual rainfall, rate of regeneration and location of cuttings in relation to the source area of streamflow. With such variability it is essential that the water yield-forest cover relationship be determined for a specific region with particular attention to annual rainfall, silvicultural methods and regrowth.

The effectiveness of forest reduction as a management tool to increase major water supplies has received widest discussion in the USA. Dortignac (1967) described forest water yield management opportunities for the USA, and more recent detailed regional reviews of the potential of water yield augmentation, ranging from pessimistic to optimistic, have been given by Douglass (1983) for the eastern USA, Troendle (1983) for the Rocky Mountain region, Hibbert (1983) for the western Rangelands, Harr (1983) for western Washington and western Oregon, and Kattelman *et al.* (1983) for Sierra Nevada. The potential of forest management for both timber and water yields was highlighted for the Colorado subalpine forests by Bowes *et al.* (1984). Ziemer (1986) focused on problems of regional implementation of forest management programmes to enhance water yield and in particular highlighted the dangers in extrapolating small-scale catchment experiments to a regional, operational scale. This problem was addressed by Cheng (1989), who showed that clear-cut logging over 30% of a 34 km^2 catchment in British Columbia produced clear and consistent streamflow increases in good agreement with findings of previous small-scale (<2.5 km^2) studies.

The interest in forest management for water production in Western Australia has arisen from a rapidly increasing demand (6% yr^{-1}) on limited water supplies in the Perth ~State capital, population 1.1 million) region. Sixty-seven per cent of Perth's public water supply is delivered from forest catchments[9]. Also, there is growing concern about a predicted drying climate for the region owing to the greenhouse effect (Pittock 1988), which could lead to stream yield reductions of 40% (Sadler *et al.* 1988). Although there is still the potential to develop new water resources in the forest, they will be at increasing distances from Perth, leading to greater cost. An economic analysis of long-term water supply options (WAWRC 1988) found that forest management was by far the least expensive.

Forest management, primarily involving thinning of dense regrowth stands, could also be expected to improve wood production significantly (Stoneman and Schofield 1989). The potential for silvicultural thinning of the jarrah forest to increase surface water supplies significantly is high. Estimates indicate that thinning 1000 km^2 of suitable forest within existing water supply catchments

[9] Currently as of 2016 only 7% of Perth public water supply is delivered from surface water catchments

would increase reservoir inflows by 47% or 127 x 10^6 m^3 yr^{-1}. These estimates, however, were based on indirect measures and modelling, as few appropriate experimental data were available. The existing experimental data, though, were able to identify the major features of water yield response. First, an analysis of streamflow generation after total deforestation of a catchment with moderately high rainfall (1200 mm yr^{-1}) produced an average streamflow increase of 239 mm over 10 years and a maximum streamflow increase of 359 mm (Ruprecht and Schofield 1989a). Stream yield was found to increase progressively for 7 years after deforestation before levelling off. This trend was closely correlated with a rise in groundwater levels and the expansion of the groundwater discharge area of the catchment. Second, in the low-rainfall region (mean average annual rainfall of 700 mm) of the jarrah forest, a selection cut and regeneration treatment reducing crown cover from 38% to 20% had a negligible impact on stream yield and groundwater level over 4 years (Stokes and Batini 1985). Third, experimental results from the southern jarrah/karri forests involving clearfelling and regeneration show an increasing yield for 3 years followed by a declining yield for 6 years (Borg *et al.* 1988). Thus the factors of annual rainfall and regeneration are prevalent, but also an apparently unique feature of the local hydrology - the response of the groundwater system, needs to be taken into account.

This section describes the groundwater, streamflow and vegetation response to an intensive, uniform thinning of a small (0.8 km^2) jarrah forest catchment using the paired catchment approach.

4.4.3 Site Description

The two experimental catchments - Hansen (treated) and Lewis (control) catchments - are located approximately 100 km south of Perth and 8 km north-northwest of Dwellingup (Fig. 4.27), within the high-rainfall zone (>1100 mm yr^{-1}) of the northern jarrah forest. The climate of the region is characterised by high winter rainfall and hot, dry summers. The catchment mean annual rainfall (1978 to 1988) was 1179 mm. However, the long-term mean annual rainfall was 1300 mm. The long-term mean annual pan evaporation was 1600 mm (Luke *et al.* 1988).

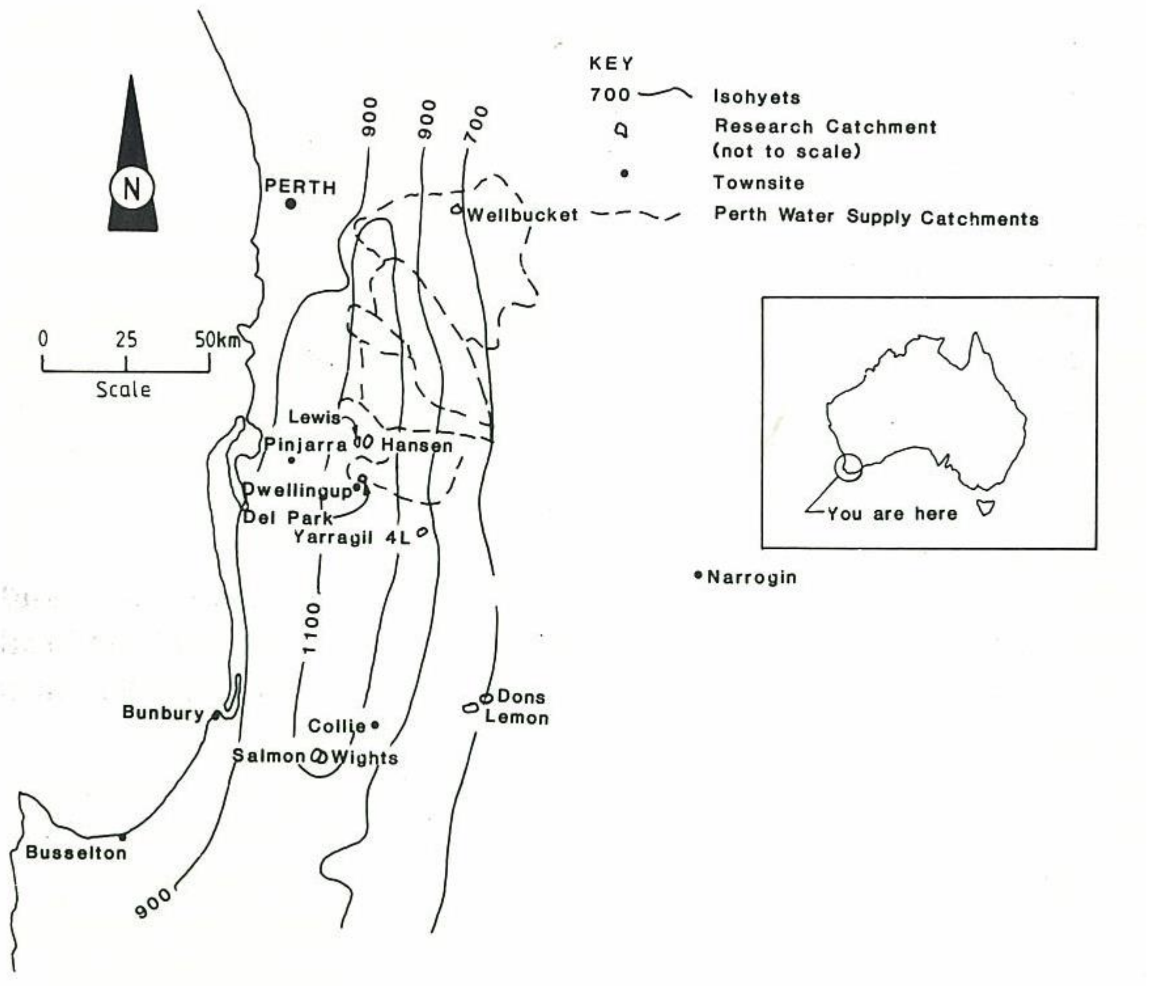


Figure 4.27 Location map

Geomorphology and soils

The hydrology of Hansen catchment is strongly affected by variations in the soil profile across the catchment. Four principal soil associations may be recognised, and these are mapped in Fig. 4.28 as (I) ridge top, (II) side-slope, (III) valley and (IV) low-lying swamp areas. In the ridge-top unit, a continuous sheet of duricrust, known locally as caprock, is exposed at the surface or overlain by less than 10 cm of coarse surficial gravels. The surficial gravels consist of fine pisolitic gravels in a loamy sand matrix (W.M. McArthur, personal communication, 1988). Saturated hydraulic conductivity (K_s) of these surficial materials varies from 1 to 10 m day^{-1} (R. Bell, personal communication, 1989). The side-slope unit is characterised by fractured caprock overlain by medium-fine gravels varying in thickness between 20 and 70 cm. In the valley unit the surficial materials are fine gravels up to 2 m in depth, and the caprock is highly fractured. In the low-lying swamp area, no caprock is present and the surficial materials are sandy loams of about 50 cm thickness.

The surface materials and caprock of the catchment are underlain by a finer-textured saprolite layer comprising mottled light to medium clays with some sand and fine gravel present, overlying pallid silt loams to light clays. The saprolitic layer varies in thickness from about 15 m at the ridge to about 6 m beneath the swamp and is located on bedrock.

Vegetation

Hansen and Lewis catchments were initially covered by open forest dominated by jarrah and marri regenerating from selective logging which occurred between

1940 and 1950. The understorey was patchy, generally less than 1 m tall and was dominated by various small-leaved sclerophyllous shrubs. Dieback caused by *Phytophthora cinnamomi* (a soil-borne pathogen) had killed susceptible understorey species over 28 ha (35% of the catchment) in Hansen catchment, mainly in the central, low-lying parts of the catchment. This area has become occupied by sedge-like plants. Few dieback related tree deaths were observed.

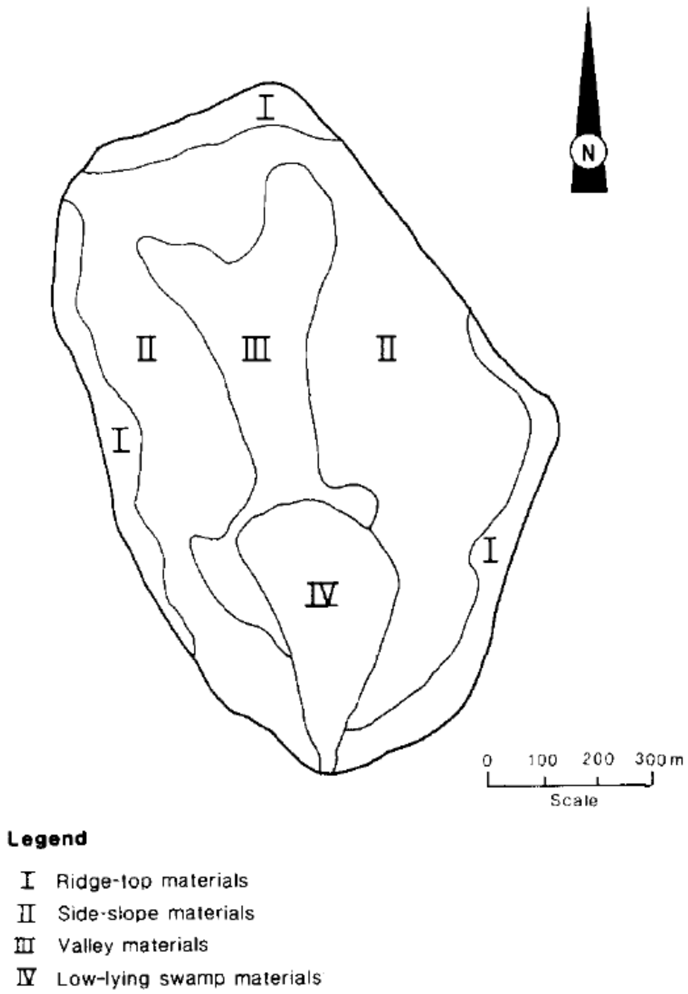


Figure 4.28 Principal soil provinces in Hansen catchment

4.4.4 Experimental Method and Instrumentation

Hansen catchment was surveyed for forest basal area (cross-sectional areas of stems), stand density (number of stems) and understorey biomass 4 years after thinning. The survey encompassed 136 points spread uniformly over the catchment and 125 points distributed through unthinned forest adjacent to it.

Tree basal area (stem cross-sectional area (m^2) per unit ground area) was measured using angle gauges. Stand density of trees and of large seedlings was measured using a point sampling technique based on the area of the triangle enclosing the sample point and formed by the three closest trees or coppice to that point (C. Ward, personal communication, 1989).

Stump coppice density and basal area were obtained by recording number and size of all stems within a 10 m radius of the sample point. Understorey biomass was estimated by complete harvests of 15 (1 m^2) quadrats on and off the thinned catchment. In the summer of 1985 to 1986 an intensive, uniform thinning treatment was applied across the catchment excluding the swamp and a 50 m buffer surrounding the swamp and stream (Fig. 4.29).

Hydrological measurements carried out at Hansen and Lewis catchments included rainfall, groundwater level and streamflow. Rainfall measurements commenced at Hansen and Lewis catchments in late 1977 using a free-standing pluviometer located in forest clearings to meet >30° exposure. Streamflow was measured using a sharp-crested 90 ° V-notch weir located on clay with deep (approximately 2 m) cut-off walls. In April 1984, 13 deep piezometers were installed, together with 12 shallow piezometers. The locations of these piezometers are shown in Fig. 4.29.

The deep piezometers were monitored on a monthly basis, and the shallow piezometers were monitored on a weekly basis during winter. In 1988 most of the shallow piezometers were instrumented with continuous recording water-level probes.

The deep piezometers at Hansen catchment were compared with control piezometers in an adjacent catchment (Del Park) where groundwater had been monitored over the same period.

The analysis of the annual and monthly streamflow was based on determining regression equations for the pre-treatment period between Hansen and Lewis catchments. These regressions were applied to the post-thinning data for Lewis to obtain predicted streamflow for Hansen catchment. This approach provided an experimental control over climatic variations within and between years.

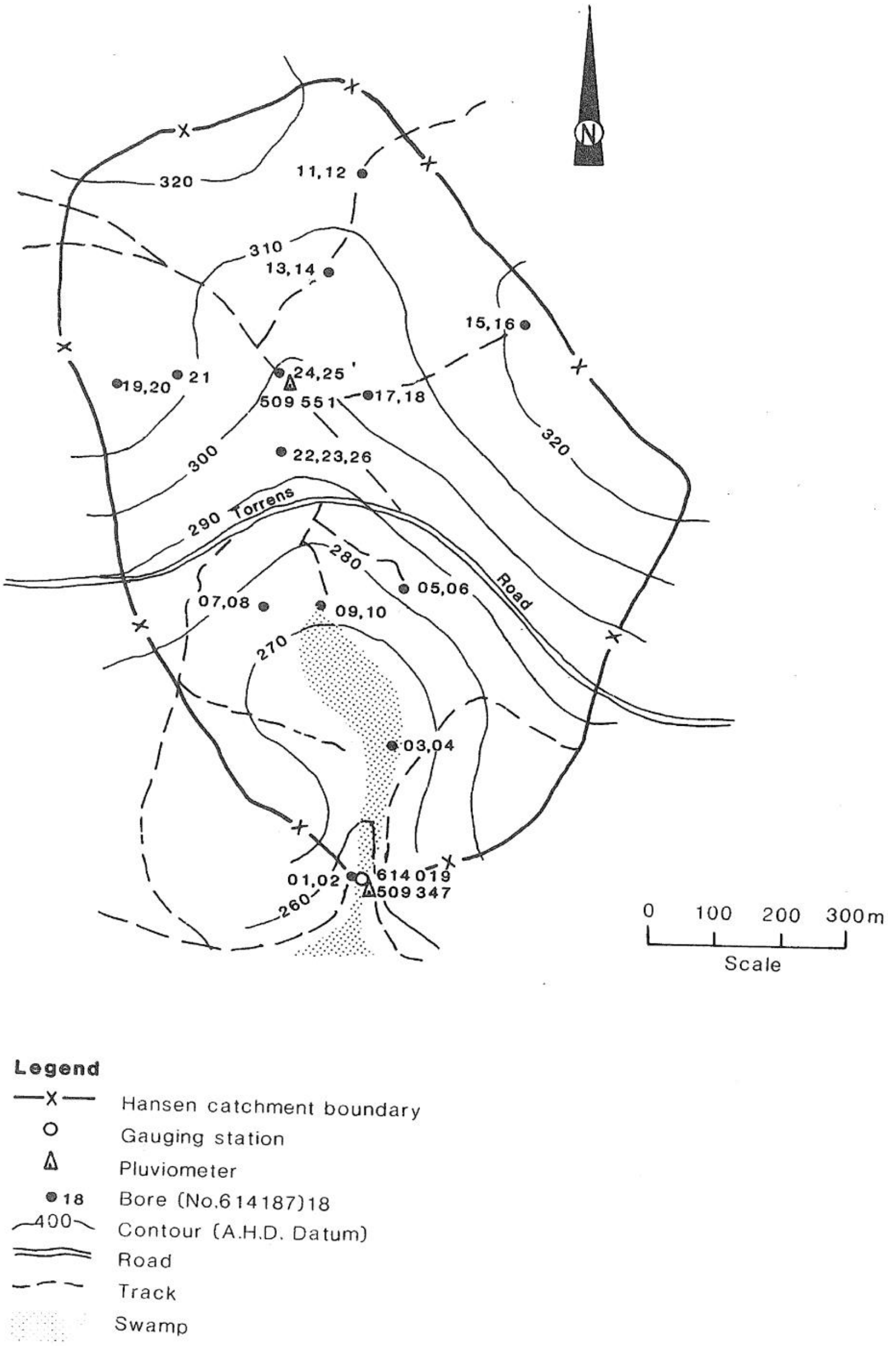


Figure 4.29 Catchment map with instrumentation

4.4.6 Vegetation Response

Thinning reduced the basal area of trees from 27.1 to 7 m^2 ha^{-1}, stand density from 700 to 110 tree stems ha^{-1} and canopy cover from approximately 60 to 14% (Table 4.6). Crowns of individual trees had become denser by 1989, but had not increased in size, since thinning. Stump coppice after 4 years was abundant and vigorous on the thinned catchment but was virtually absent from unthinned forest. However, stump coppice (646 stems ha^{-1}) contributed only 0.54 $m^2\ ha^{-1}$ to catchment basal area. Relatively little ground coppice (large seedlings) was removed in thinning and lack of substantial new growth indicated that ground coppice had not grown significantly since thinning. Understorey biomass was the same (around 1.4 t ha^{-1} by dry weight) in both thinned and unthinned forest.

4.4.7 Hydrologic Response

Rainfall

The average rainfall for the 8 years of pre-treatment analysis was 1200 mm yr^{-1}, which was 100 mm below the long-term average (1934 to 1988) for Hansen catchment. The rainfall, post-treatment, consisted of 2 years (1986 and 1987) of substantially below-average rainfall followed by 1 year (1988) above average.

Table 4.6 Understorey biomass, number and basal area of ground coppice (GC), stump coppice (SC) and trees inside and immediately adjacent to the thinned catchment; measurements were made between September 1989 and February 1990

	Understorey biomass (t ha^{-1})	GC		SC		Trees	
		stems ha^{-1}	$m^2\ ha^{-1}$	stems ha^{-1}	$m^2\ ha^{-1}$	stems ha^{-1}	$m^2\ ha^{-1}$
Unthinned	1.5	1984	0.2	0	0.01	700	27.1
Thinned	1.4	1399	0.2	646	0.54	110	7.0

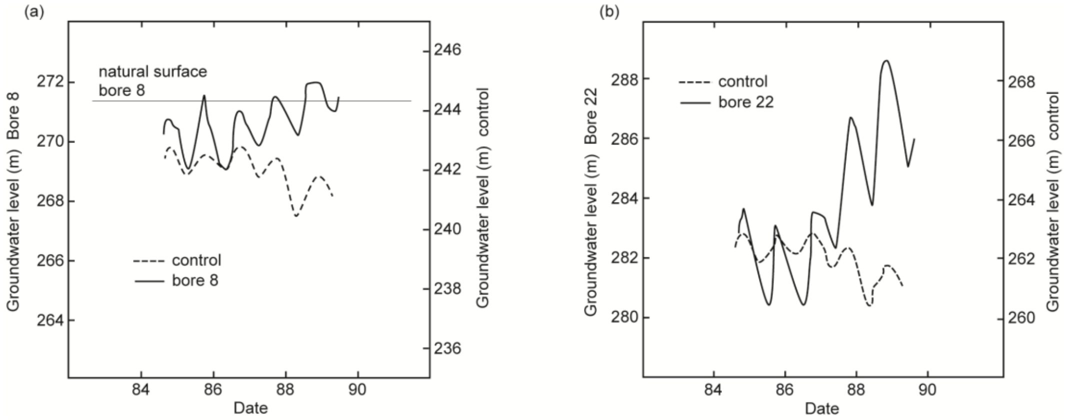


Figure 4.30 Deep groundwater: (a) downslope location; (b) upslope location

Deep groundwater aquifer

A 'permanent' unconfined groundwater system located on bedrock exists throughout the catchment and discharges through the swamp area (Fig. 4.28). Temporal changes in groundwater levels are shown in Fig. 4.30(a) for a downslope location and in Fig. 4.30(b) for an upslope location relative to a control.

The groundwater rise relative to a control bore was approximately 2.5 to 3.0 m (Fig. 4.30(a)) surrounding the groundwater discharge area. In the upslope areas (bores 13, 17, 22 and 24; Fig. 4.29) the groundwater rise was approximately 5.0 m (Fig. 4.30(b)). The response was smaller downslope because of the proximity of the water table to the ground surface. In both areas the rise slowed considerably in the last three annual minima (1988 to 1990). This would imply that the groundwater system had approached a new equilibrium.

Shallow ground water

A shallow groundwater system, perched on the caprock or underlying clay was observed for approximately 2 months in the year (Fig. 4.31) at a lower-slope location (bore 10) before thinning. In the first year after thinning the shallow groundwater system was observed over 3.5 months and in the second year over 6 months. The depth to the shallow groundwater system was reduced from a minimum of 2.2 to 1.6 m after thinning.

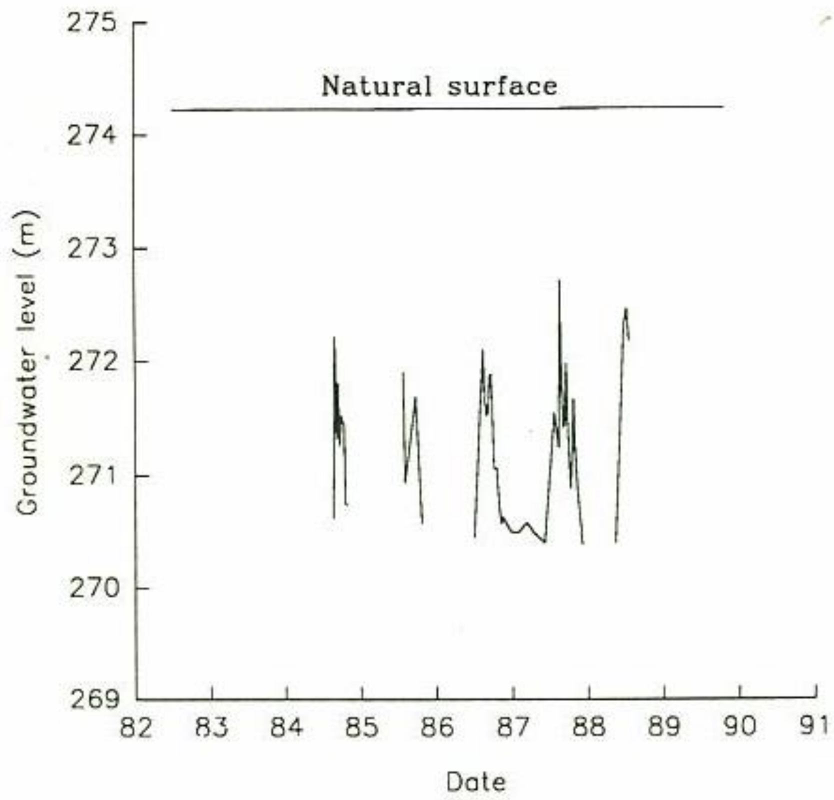


Figure 4.31 Shallow groundwater response over 5 years

Streamflow

To account for the effect of climatic variation on streamflow changes, a paired catchment approach was used. An 8 year pre-treatment period was allowed for calibration. The streamflow relationship between Hansen and Lewis is shown in Fig. 4.32. A simple linear regression was used to relate the annual streamflow for Hansen to Lewis catchment for the pre-treatment period (1978 to 1984). Because of a spring (September to November) burn to reduce forest litter in the Hansen catchment in 1984, the streamflow for 1985 was omitted from the regression analysis for the pre-treatment period. The resultant equation was

$$H = 2.16 + 0.94L \ (r^2 = 0.74, P < 0.01) \qquad (1)$$

where H is the Hansen catchment streamflow (mm) and L is the Lewis catchment streamflow (mm).

This relationship was used to predict streamflow for the period 1986 to 1988 as if there had been no thinning on Hansen catchment (see Table 4.7). The rate of increase in annual streamflow has been approximately linear, increasing from 7% of the average long-term rainfall 1 year after thinning to about 20% of the average long-term rainfall or nearly 400% of the average streamflow after 3 years.

The annual streamflow vs. rainfall for the Hansen catchment is plotted in Fig. 4.33. Before the thinning treatment there was an approximately linear relationship between streamflow and rainfall. However, 2 years, 1980 and 1981, are substantially below this relationship. This may be attributable to the sequence of low rainfall for 4 years before 1980. After the thinning treatment there was a significant increase in the streamflow response relative to annual rainfall. In particular, 1988 was substantially above what would be predicted based on no treatment.

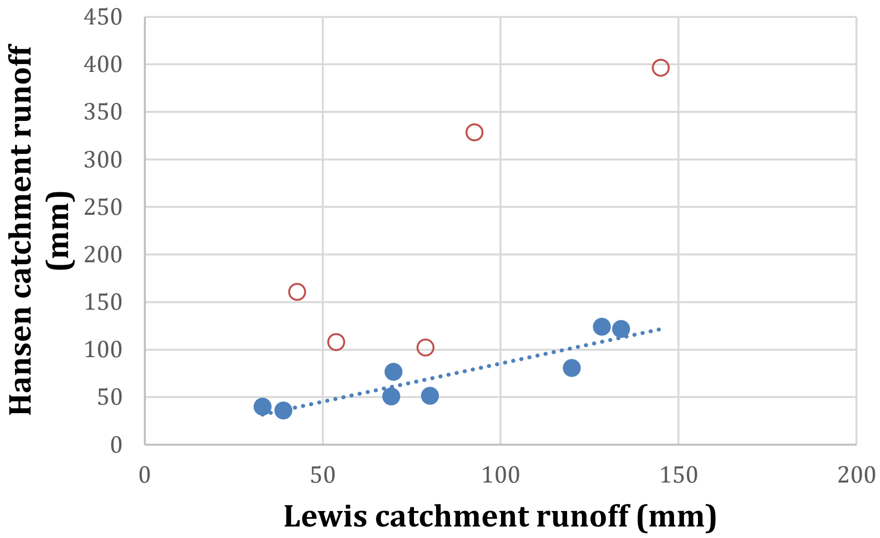


Figure 4.32 Hansen catchment annual streamflow vs. Lewis catchment annual streamflow (blue dots represent pre-treatment period and open red dots are post-treatment)

Table 4.7 Streamflow increase for Hansen catchment as a result of forest thinning

Year	Rainfall (mm)	Measured flow (mm)	Predicted flow (mm)	Measured-predicted flow	
				(mm)	(% rainfall)
1978	1098	52	66	-14	-1
1979	1004	28	28	0	0
1980	1324	54	58	-4	0
1981	1487	71	102	-31	-2
1982	1098	71	54	17	2
1983	1294	123	110	13	1
1984	1330	123	104	19	1
1985	1157	104	63	40	3
1986	965	109	44	65	7
1987	1034	165	36	129	12
1988	1572	423	119	304	19

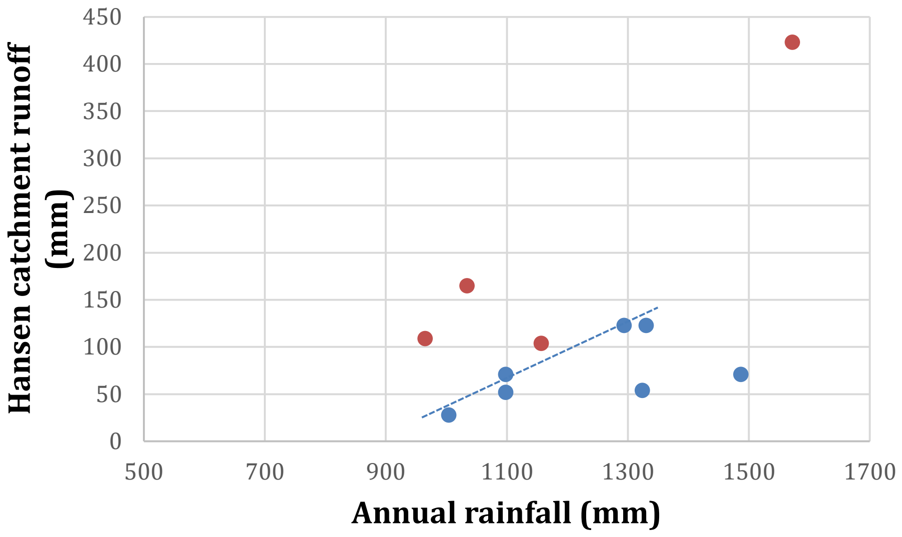


Figure 4.33 Annual streamflow vs. rainfall for Hansen catchment (blue dots are pre-treatment data and red dots are post-treatment)

Regression equations between the monthly streamflow for Hansen catchment and Lewis catchment were evaluated for each month starting in April (see Table 4.8). The r^2 for the regression equations ranged from 0.62 to 0.99. However, for October there was considered to be two outliers (1983 and 1984) in the monthly streamflow at Hansen catchment compared with Lewis catchment and these were not included in the analysis. The monthly regressions were used to predict the monthly streamflow from Hansen catchment for 1988 assuming there was no thinning. The major predicted increase in streamflow occurred from July to October (Fig. 4.34). However, there were still significant increases in streamflow for June, November and December. Before thinning there was only very small streamflow from February to May. However, after thinning there was a significant flow over summer and early autumn.

Table 4.8 Monthly regression equations for Hansen catchment

Month	a	b	r^2	n
April	0	0	-	6
May	-1.45	0.59	0.98	6
June	3.02	1.00	0.85	6
July	5.88	0.57	0.62	6
August	1.21	0.84	0.98	6
September	2.39	0.76	0.96	6
October	1.58	0.57	0.99	4
November	0.43	1.25	0.85	6
December	-1.41	2.01	0.71	6
January	-0.14	1.43	0.91	6
February	-0.09	1.16	0.61	6
March	-0.015	0.71	0.72	6

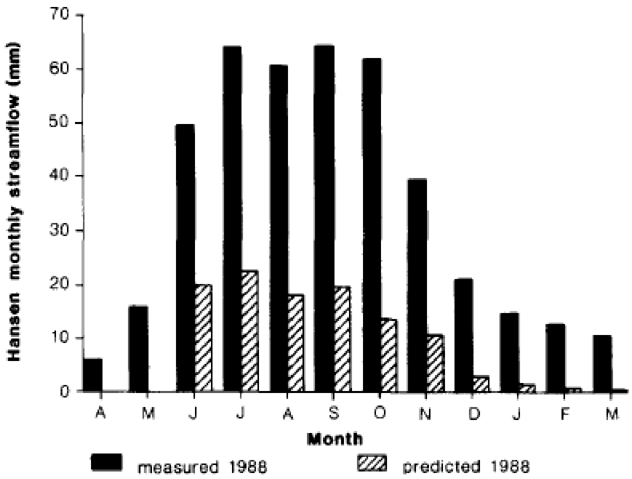


Figure 4.34 Monthly streamflow for Hansen catchment compared with predicted monthly streamflow

Comparison of streamflow hydrographs

A comparison of the pre-treatment streamflow hydrographs of Hansen and Lewis catchments (Fig. 4.35(a)) shows that there is little difference in the hydrographs apart from a larger streamflow response to rainfall events at the commencement of the streamflow by Lewis catchment for 1983 (before thinning). The streamflow hydrographs post-treatment (Fig. 4.35(b)) highlight the changes that have occurred in Hansen catchment as a result of forest thinning. In the post-treatment period the peak flows for Hansen catchment were approximately 50% greater than those at Lewis, whereas in the pre-treatment period there was no discernible difference. Using the baseflow separation algorithm of Lynne and Hollick (1979), the quick flow response, despite being quantitatively higher, had reduced as a percentage of flow from 15 (1983) to 11% (1988).

The winter baseflow (shallow subsurface flow and groundwater flow) of Hansen catchment compared with Lewis catchment had increased from approximately equal pre-treatment, to approximately four times greater post treatment. Before treatment the streamflow duration was approximately 10 months. However, after treatment, streamflow in Hansen catchment became perennial (Figs. 4.34 and 4.35).

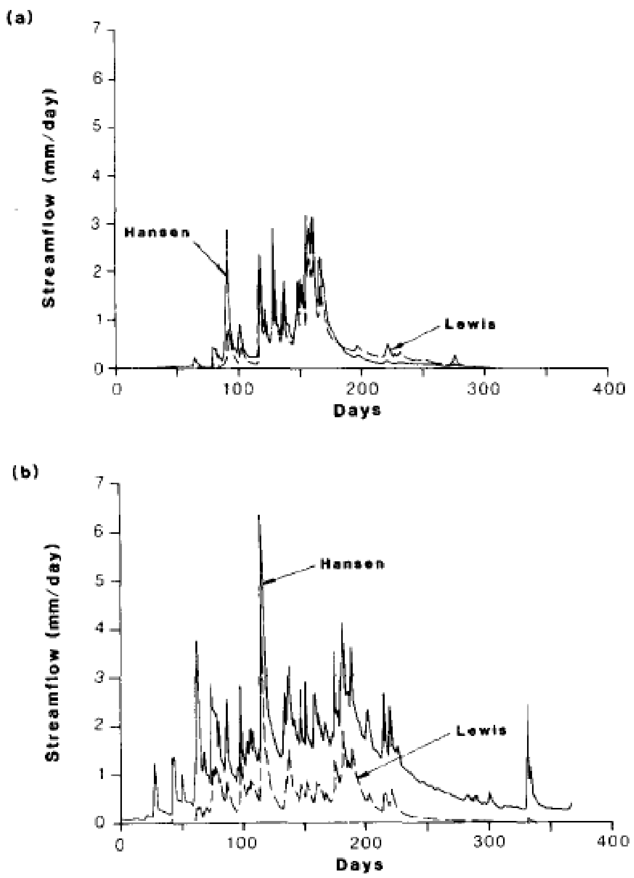


Figure 4.35 Streamflow hydrographs for Hansen catchment for (a) 1983 and (b) 1988 (day 0 represents 1 April)

4.4.8 Discussion

Vegetation response and need for regrowth control

Naturally seeded regeneration and understorey showed little response to thinning, suggesting that increased growth by these plants will not rapidly replace leaf area caused by thinning. However, rapidly growing *E. marginata* stump coppice does have the potential to negate the benefits of thinning in the medium term.

Although many of the small coppice stems will be suppressed and die in coming years, the larger coppice (>50 mm diameter) may continue to grow and reach 100 mm in diameter 10 years after thinning (Abbott and Loneragan 1982). Stump coppice basal area would then be around 4.6 m^2 ha^{-1}. Over half the basal area of trees was left after thinning but it was still <20% of the pre-thinning basal area. Coppice regeneration of this order would be expected to make a significant contribution to catchment transpiration and interception and its removal could be considered as a means of maximising gains in water yield (Stoneman and Schofield 1989).

Prediction of streamflow increases

The 75% reduction in forest cover on Hansen catchment resulted in an increase in streamflow of approximately 304 mm for an above-average rainfall year after 3 years. Bosch and Hewlett (1982) predicted an average increase of 40 mm per 10% change in forest cover for eucalypt forest types. Consequently, the Bosch and Hewlett prediction for Hansen catchment would be 300 mm, which compares well with the observed increase.

Sources of additional streamflow

The cause of additional streamflow following forest thinning is the reduction in evapotranspiration brought about by a reduction in overstorey (Stoneman and Schofield 1989). The dominant streamflow generation mechanism in forested catchments in SWWA has been found to be throughflow from perched, shallow groundwater systems, whereas overland flow and groundwater flow make only minor contributions (Loh *et al.* 1984; Ruprecht and Schofield 1989a; Stokes 1985;

Stokes and Loh 1982; Turner *et al.* 1987b). Insufficient data are available to describe quantitatively the components of streamflow generation on Hansen catchment both pre- and post-treatment. However, the increased persistence of perched water tables indicates that throughflow has increased substantially after the forest thinning treatment.

Comparison of streamflow increases after forest reduction

The streamflow response to forest thinning within SWWA has varied with a number of factors including mean annual rainfall and level of vegetation reduction. Hansen catchment, which has had an intensive thinning in a high-rainfall zone of the jarrah forest, is likely to give an extreme yield increase. Lower levels of vegetation reduction in lower rainfall areas have resulted in little or no streamflow response (Stokes and Batini 1985; Stoneman 1986) and only small rises in groundwater levels.

Role of the permanent groundwater system

A study of streamflow generation after deforestation (see Chapter 4,1 and (Ruprecht and Schofield 1989a)) showed that the water yield increases were closely related to an increasing groundwater discharge area over a 7 year period. In the present study the groundwater system appears to have 'equilibrated' after 2 to 3 years.

This quicker response may be the result of a higher saturated hydraulic conductivity of the deep aquifer. The observation of a new groundwater equilibrium may mean that further streamflow increases may not occur.

4.4.9 Conclusions

Streamflow increases of approximately 20% of annual rainfall were recorded 3 years after thinning. Streamflow increases after thinning occurred throughout the year but particularly during the months July to October. The previously ephemeral stream became perennial after thinning.

Winter baseflow increased four-fold after thinning and was the major component of increased streamflow. Groundwater level rises were observed only for 2 years, thus appearing to reach a new equilibrium. The shallow ephemeral groundwater system became longer in duration and greater in areal extent after thinning.

Regeneration did not rapidly replace reduce stand leaf area after thinning. However, regrowth control could be required in a further 10 years to maintain significant water yield increases.

The magnitude of streamflow increase after thinning of Hansen catchment is in line with most studies reported elsewhere. The major local factors controlling this magnitude are mean annual rainfall and level of vegetation reduction.

Chapter 5 Forest disturbance and water values

5.1 Introduction

This chapter critically examines the paired catchment studies, across SWWA, into the impact of forest disturbances on the hydrology and water resources of these forest areas. This chapter expands on the Ruprecht and Stoneman (1993)[10] paper with new data and learnings.

The jarrah forest of SWWA generates small amounts of streamflow from mean annual rainfall ranging from 600 – 1300 mm (Gentilli 1989). The low water yields are attributed to the high infiltration rates of surface soils and the large soil water storage available for continuous use by the forest vegetation. These attributes have made assessing the impact of forest disturbance more difficult and have required long periods of post-treatment monitoring.

Local studies into forest disturbance have focused on clearing for agricultural development, forest harvesting and regeneration, forest thinning and the consequences of bauxite mining. The research level paired catchments that have been used in these studies were summarised in Chapter 2 with a total of 44 catchments (treated and control) across SWWA (Fig. 2.4).

[10] Based on: **Ruprecht, J.K**. and Stoneman, G. L., 1993. Water yield issues in the jarrah forest of south-western Australia, *Journal of Hydrology* 150: 369-391.

5.2 Forest disturbance and water yield

5.2.1 Impact of clearing of forest for agriculture

Forest clearing generally results in an increase in streamflow and a rise in groundwater levels (Chapter 4). For example, agricultural development (100% clearing) of Wights catchment in a high rainfall zone (1200 mm annual rainfall) resulted in a 400% increase in water yield over the 10 years after clearing. This increase was attributed primarily to decreases in transpiration and interception loss (Williamson *et al.* 1987). Streamflow increased markedly in the first year after clearing (by about 10% of annual rainfall) and continued to increase at a slower rate for a further 5 years, when a new streamflow equilibrium was reached. Section 4.1 found that the nature of the water yield increase after clearing was controlled largely by the expanding permanent groundwater discharge area. By 1983, 7 years after clearing, the groundwater discharge area had expanded to about 18% of the catchment area. It was considered that this would significantly increase throughflow and overland flow.

Lemon catchment, a small experimental catchment in the low rainfall zone (800 mm annual rainfall) was partially (53%) cleared in 1976. The impact on the groundwater system was dramatic. Initial rates of rise were only 0.11 m yr $^{-1}$, but this increased after 10 years to average 2.3 m yr $^{-1}$ (Section 4.2). After 13 years, the groundwater had risen 15 m in the valley and 20 to 25 m in the lower slopes were observed over 13 years. Clearing of 53% of Lemon catchment also resulted in some fundamental changes in the annual streamflow to rainfall relationship (Fig. 5.1). Before clearing, approximately 700 mm of annual rainfall was required for streamflow to occur, equivalent to 400 mm in the first 4 months of the water year (commencing on 1 April).

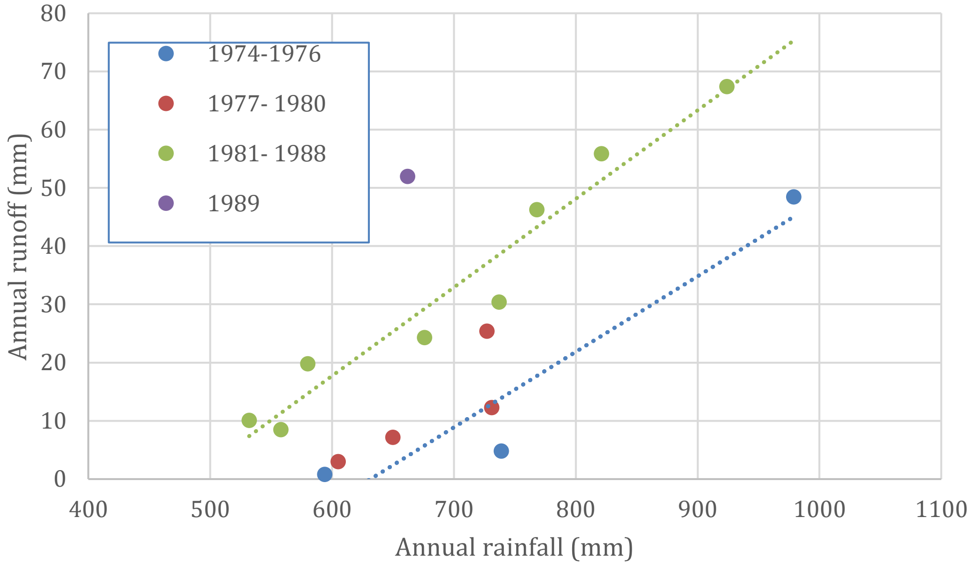


Figure 5.1 Streamflow to rainfall relationship response to clearing for Lemon catchment for different time periods including pre-clearing (1974 to 1976)

A new "quasi-equilibrium" developed from 1994 as the groundwater discharge area stabilised and subsequently the Lemon catchment became perennial where it was ephemeral both within a year and from year to year. The Wights catchment took approximately 8 years to reach a streamflow equilibrium, whilst the Lemon catchment took approximately 20 years (Fig. 5.2).

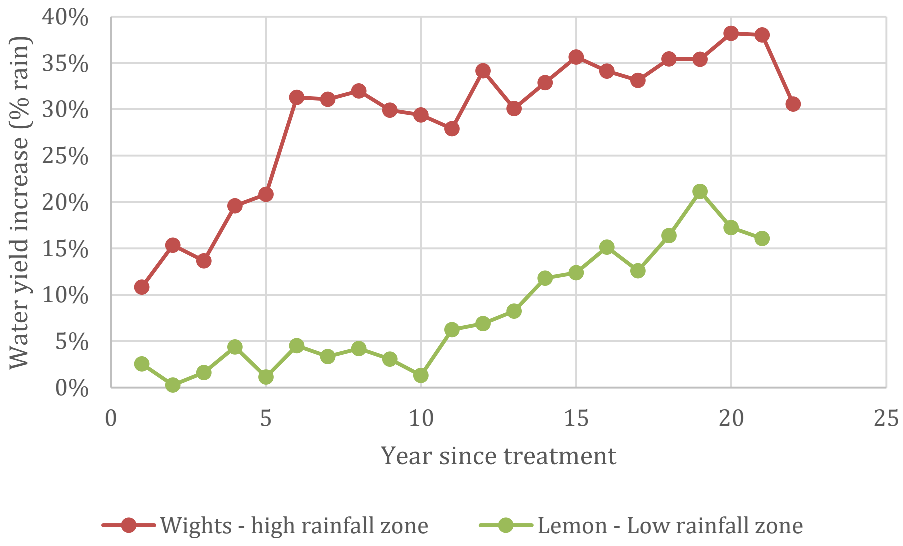


Figure 5.2 Comparison of water yield responses to agricultural development for different rainfall zones (adapted from Bari and Ruprecht, 2003)

The comparison of paired catchment studies from conversion of forest to agriculture across Australia (Table 5.1), although geographically limited, does show a significantly greater increase in runoff for the SWWA sites as a percentage of annual rainfall. The SWWA results are significantly larger than summarised in Sahin and Hall (1996) which showed that, for eucalypt forests a 10% reduction in cover resulted in a water yield increase of only 6 mm. However there are limited studies into the conversion of forest into agriculture (Brown et al. 2005) to provide greater confidence in generalising the impact.. The influence of the groundwater discharge area in streamflow generation in SWWA catchments as demonstrated in Chapter 4 is considered a critical reason why increases are more pronounced in these catchments.

Table 5.1 Comparison of streamflow increase from conversion of forest to agriculture from paired catchment studies across Australia

Study	MAR[1] (mm)	Streamflow increase	
		(mm)	(% of annual rainfall)
Wights, WA	1200	420	35
Lemon, WA	720	140	20
Babinda, Qld	4300	293	7
Brigalow, Qld	700	28-42	4-6

(1) MAR is mean annual rainfall

Analysis at the research catchment scale shows a clear impact of agricultural clearing on floodflows (Fig. 5.3). Even for greater than a 1-in-10 Average Recurrence Interval (ARI) (10% probability on non-exceedance) and close to a 1-in-100 ARI (1% probability of non-exceedance) there is no clear convergence between the flood frequency curve for the cleared research catchment compared to the forested research catchment. This large increase is considered to be due to the increase in groundwater discharge area creating a means for signficant additional quick response runoff from the saturated area.

Silberstein *et al.* (2004) concluded from analysis of the Collie paired catchments that runoff coefficients rise after clearing, with greater impact on low flows than high flows after clearing. Analysis of small catchments (Leanne Pearce unpublished) found a relationship between increased peak flow with increasing area of catchment cleared of forest. However, there is uncertainty as to whether these relationships are still valid at a larger catchment scale (Calder and Aylward 2006).

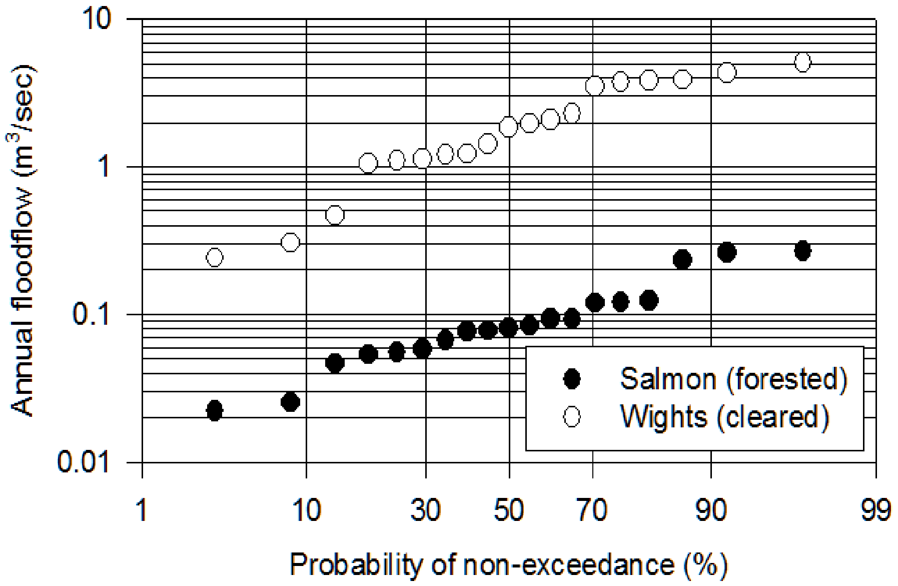


Figure 5.3 Flood frequency for cleared and forested research catchments

5.2.2 Impact of timber harvesting on forest hydrology

Forest harvesting and regeneration generally led to initial water yield increases followed by a gradual return to pre-disturbance values (Fig. 5.4). The maximum streamflow increase ranged from 150 to 200 mm for high rainfall research catchments (Fig. 5.5) and to close to zero at 700 mm mean annual rainfall in the Wellbucket catchment (Stokes and Batini 1985).

The streamflow increase from logging and regeneration for the low rainfall Wellbucket research catchment peaked at 3 mm or 0.4% of the annual rain (Fig 5.4) (Stokes and Batini 1985). The two low rainfall research catchments (Yerraminnup and Wellbucket) had significantly less increase in streamflow.

For the high rainfall research catchment, Lewin South, water yield initially increased by 15% of annual rainfall 3 years after harvesting, whereas by 12 years post-harvest the water yield decreased to below pre-treatment levels. For the low rainfall catchment (Yerraminnup South) the initial increase in water yield was 4.4% of annual rainfall, and had diminished to no increase 10 years after harvesting (Fig. 5.4). In comparison to Eastern Australian studies summarised in Chapter 2 (Fig. 5.6), the SWWA studies show a significantly lower peak increase in streamflow following timber harvesting based on area treated (mean for Eastern Australia is 174 mm and for SWWA the mean is 114 mm). However, some of the paired catchment studies from eastern Australia have higher mean annual rainfall so in part the increased streamflows can be explained by higher rainfall.

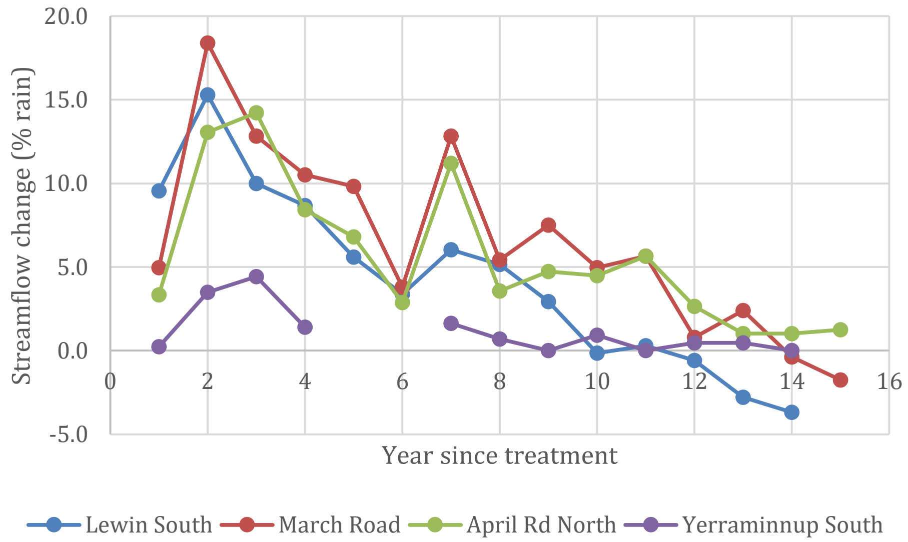


Figure 5.4 Streamflow response to logging experimental catchments in the Southern Forest (Bari and Ruprecht 2003).

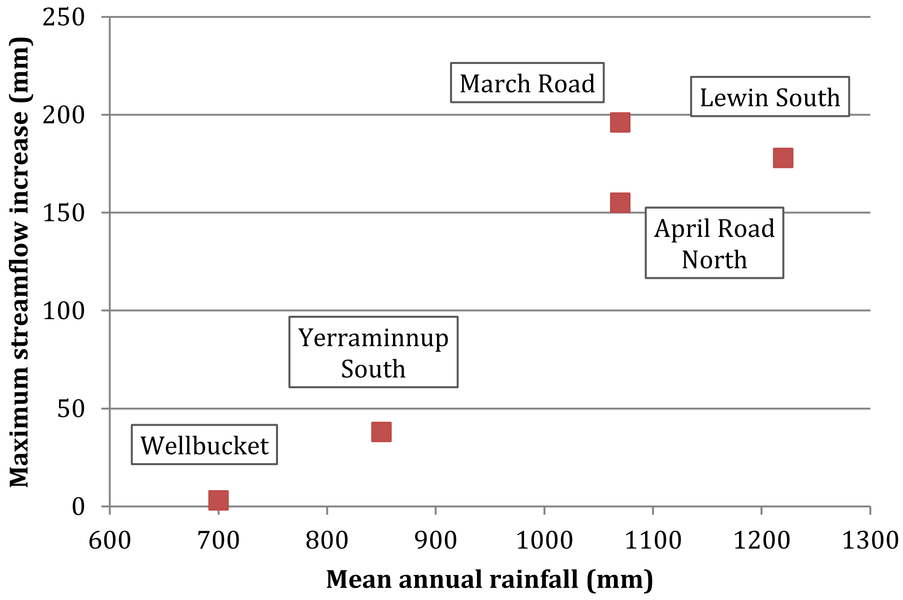


Figure 5.5 Maximum streamflow increases from logging and regeneration

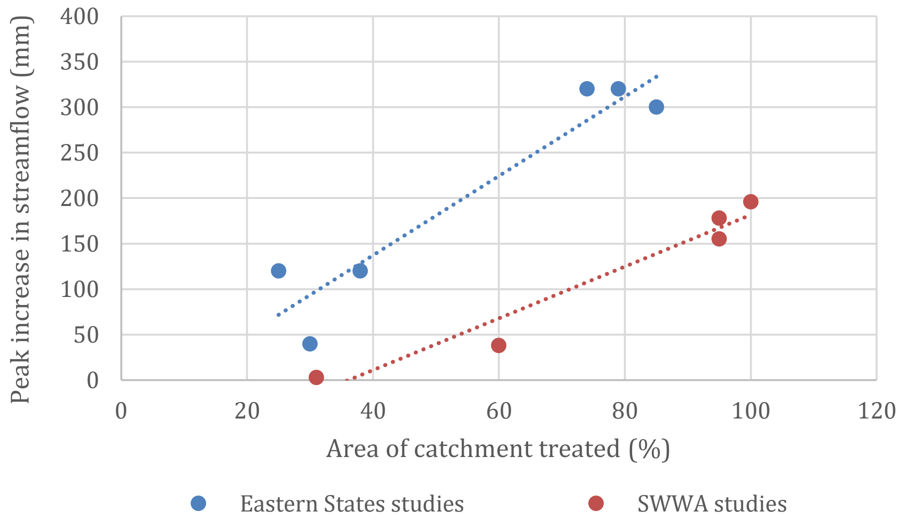


Figure 5.6 Comparison of peak increase in streamflow for SWWA compared to Eastern States studies from timber harvesting. The Eastern States studies have higher mean rainfall (900 to 1500 mm yr^{-1}) compared to those from SWWA (700 to 1220 mm yr^{-1}).

The longer term assessment of water yield indicates that streamflow has returned to pre-treatment levels and at the point where monitoring ceased, streamflows appeared to decrease to below pre-treatment levels (Fig. 5.4) (Bari and Ruprecht 2003). Macfarlane *et al.* (2010) found that regrowth eucalypt forests used significantly more water than old-growth eucalypt forests. Borg and Stoneman (1991) predicted that streamflow from clearfelling in southern forests is likely to lead to lower streamflows persisting for at least 60 years.

Studies in eastern Australia have considered that the longer term decrease in streamflow is due to the vigorous nature of the regrowth, which has a greater transpiration rate compared with old growth forests (Cornish and Vertessy 2001; Roberts *et al.* 2001; Vertessy *et al.* 2001).

5.2.3 Impact of forest thinning

Forest thinning generally results in an initial increase in streamflow followed by a return to pre-disturbance levels (Table 5.2 and Fig. 5.7). However, the return to pre-disturbance levels can be delayed by regrowth suppression at the initial thinning or by later regrowth control. For example, the uniform, intensive thinning treatment of Hansen catchment (annual rainfall 1200 mm) reduced crown cover from 60 to 14%, which resulted in an increase in streamflow of approximately 20% of annual rainfall (260 mm for an average year) after 3 years, compared with a streamflow yield of 6% of annual rainfall before thinning. The major components of streamflow generation present before thinning remained as the major components after thinning. The expansion of the saturated source area and the presence of a shallow groundwater system for extended periods, as a result of an increase in available water from the reduction in interception and evaporation from the overstorey, were considered to be the major causes of increased water yield.

Substantial changes to monthly water yield occurred from July to October as a result of the thinning treatment, based on monthly regressions with the untreated control catchment. However, there were still significant increases in streamflow for June, November and December. Stoneman (1993) found an increase did occur of 86 mm (7.1% of annual rainfall) in water yield, 9 years after the forest thinning of Yarragil 4L catchment, which is on the border of the high and intermediate rainfall zones. In the low rainfall zone (700 mm annual rainfall) of the jarrah forest, a selection cut and regeneration treatment reducing crown cover from 38% to 20% had a negligible impact on water yield over 4 years (Stokes and Batini 1985). Reed *et al.* (2012) reported no significant streamflow change from the thinning of the Chandler Road catchment, but a 15 to 50 mm increase in runoff at Cobiac (Fig. 5.8). The reasons for lack of streamflow increase at the Chandler Road catchment were considered due to the small percentage of the catchment treated, a large part of the catchment was internally draining due to mining rehabilitation, and significant area of lower slopes were excluded from thinning by the Department of Environment and Conservation (Reed *et al.* 2012).

Research scale catchment studies have shown significant increases in water yield from forest thinning, with a reduction in basal area of 60% is likely to be required to generate measurable increases in streamflow. However, the experience from Wungong Catchment Trial (Reed *et al.* 2012) found it difficult to observe increases at a larger scale. Part of the reason is due to the difficulty in treating whole forests to the level undertaken at the research catchment scale. Another reason is that given the drying climate it is difficult to determine the un-treated or control level of water yield.

Table 5.2 Impact of forest thinning on water yield for research catchments

Catchment	Treatment (reduction in basal area)	Rainfall Zone	Peak increase in water yield		Increase in water yield after 10 years	
	(%)		(% rain)	(mm)	(% rain)	(mm)
Hansen	80	High	19	260	7.5	90
Higgens	60	High	12	156	0	0
Jones	60	High	9	103	0	0
Yarragil 4L	60	Intermediate	12	131	11	97
Cobiac	40	Intermediate	~5	40	na[(1)]	na
Chandler Rd	39	Intermediate	-	-	-	-

(1) Has not reached ten years from treatment

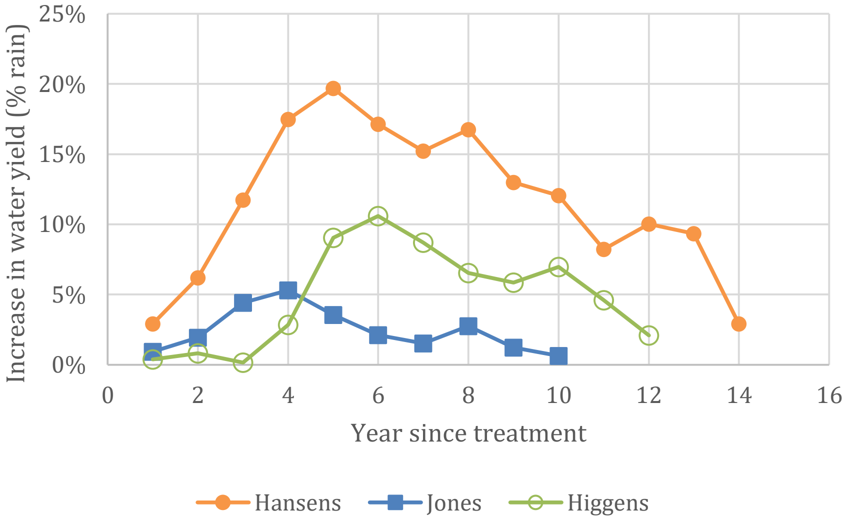


Figure 5.7 Water yield increase (%) from HRZ forest thinning research catchments

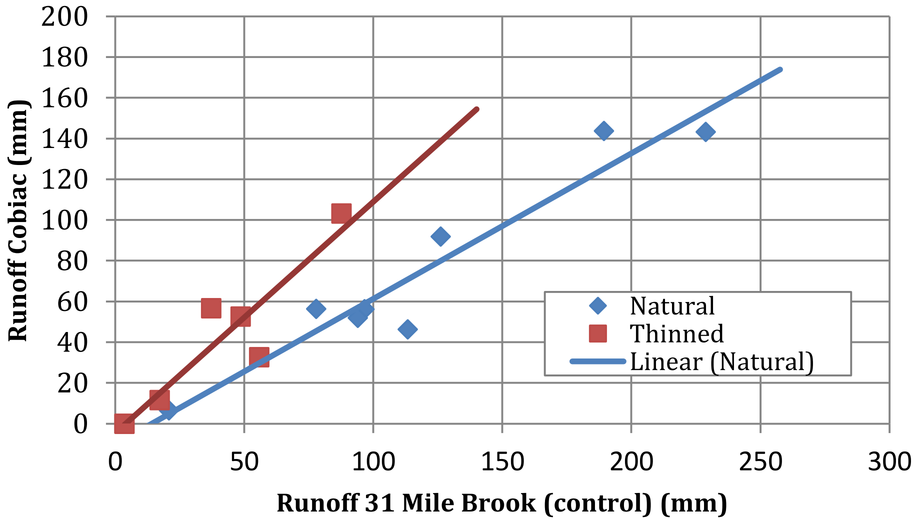


Figure 5.8 Impact of forest thinning on streamflow for Cobiac catchment

The issue of observing increases in water yield from forest management at large catchment scales has been well documented (Huff *et al.* 2000; Ziemer 1986). Cheng (1989) in a study into the clear-cut logging of over 30% of a 34 km^2 area in British Columbia, Canada found clear and consistent water yield increases similar to those observed in research scale catchments. Troendle *et al.* (2001) in a paired catchment study, treated a 16.7 km^2 catchment and an increase of 76 mm in runoff (17%) was observed. The extrapolation from small or medium scale paired catchment studies to water supply or regional scales has required catchment modelling or statistical approaches (Robles *et al.* 2014; Watson *et al.* 1999). Hawthorne *et al.* (2013) in a study into the long term effects of forest thinning found an initial significant increase in water yield followed by a decreasing trend in water yield. However, this result was confounded by the decade long drought experienced in south-eastern Australia from 1997 to 2008 (Hawthorne *et al.* 2013).

Macfarlane *et al.* (2010) concluded that forest thinning to promote “old-forest-like” attributes could be a viable means to provide both water and conservation benefits.

5.2.4 Impact of bauxite mining on forest hydrology

The hydrologic impact of bauxite mining has been monitored at a range of scales; from hillslope, to research catchment, water resource catchment, and water supply catchment scale. The changes in the catchment hydrological processes and responses as a result of mining and rehabilitation are complex. The major initial physical changes owing to mining are considered to be clearing of all vegetation; removal of surface sand/gravels and caprock layer (typically 4 to 6 m in thickness), albeit with the replacement with shallow surface soils; deep ripping of the soil profile; grading and contouring. Over longer time periods the major impacts are caused by an increase in vegetation cover, changing components of vegetation and changing surface topography and soil characteristics.

At the Del Park catchment 16% was mined over a period of 4 years (1975 to 1979) and then a further 10% was mined in 1990. Hydrological processes have been studied at this catchment in Chapter 3 and 4 (Ruprecht and Schofield 1990b; c; 1993) during the native forest, mining and rehabilitation sequences. The major observations included:

(1) Infiltration capacities of native jarrah forest surface soils are high, with surface runoff rarely being observed.

(2) Infiltration capacities of mined soils are similar to those of native forest soils except in contour troughs, where extensive ponding was observed, which resulted in the generation of significant overland flow.

(3) Although mining removes several metres of soil profile, the overall soil water holding capacity of the sites is not majorly impacted, due to the pre-existing deep weathering profiles.

(4) A shallow ephemeral saturation zone developed downslope of mining.

(5) Groundwater levels within minepits increased by approximately 3 m over 2 to 3 years, whereas 8 to 9 years after mining the groundwater increase had declined to only 0.4 m. Downslope of mining, initial groundwater response was an increase of approximately 2 m, followed by a gradual return to pre-mining levels.

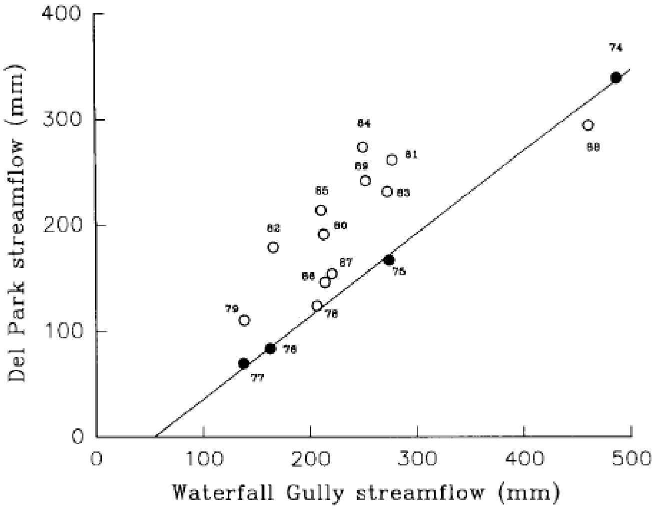
Mining at the Del Park catchment led to an increase in water yield of a maximum of 8% of annual rainfall (Fig. 5.9). However, with rehabilitation of the mined area, the water yield returned to the pre-mining level after 12 years. By 1988 water yield reduced to below pre-mining levels, and by 2001 the decline had reached 20% of streamflow (Croton 2004).

The two larger research catchments - Seldom Seen and More Seldom Seen were mined from 1969 to 1994 and have been subject to a number of studies (Croton *et al.* 2005; Croton and Reed 2007; Loh *et al.* 1984; Ruprecht 1991). The streamflow increase for Seldom Seen and More Seldom Seen peaked at 62 and 90 mm respectively (Croton *et al.* 2005). This was followed by a reduction to 49 mm and 71 mm in runoff compared to untreated catchments (Fig. 5.10 for More Seldom Seen catchment) (Croton *et al.* 2005). However, some of the reduction in streamflow is considered to be due to forest growth, not solely mining rehabilitation. The aim with the paired catchment approach means that changes due to climate are accounted for and only land use changes can explain the difference.

Monitoring of the Cameron catchments was established to evaluate the impact of mining the higher salt risk intermediate rainfall zone of the northern jarrah forest. In the time period of observation (7 years, 2004 to 2011) following mining of Cameron West and Central, no real effects on streamflows due to mining are observable. As well, the streamflows declined further due to the historical-low rainfall of 2010 combined with no following years of above-average rainfall (James Croton personnel communication, 2013).

A comparison of the studies (Fig. 5.12) into bauxite mining and rehabilitation shows an increase in flow peaking at 50 to 140 mm, with a clear relationship with area mined. The decline in streamflow then ranged from zero to over 50 mm. Loh *et al.* (1984) found that peak flow ratios increased by about 2 to 3 times for the Seldom Seen and More Seldom Seen research catchments relative to the Waterfall Gully control catchment. The longer term analysis shows a sustained relative increase in flood peak for the mined More Seldom Seen catchment, however peak flows have reduced in real terms for both control and mined catchments (Fig. 5.12). Despite a reduction in streamflow to below pre-treatment water yield values compared to the control, the peak flow response has remained above pre-treatment values.

(a)



(b)

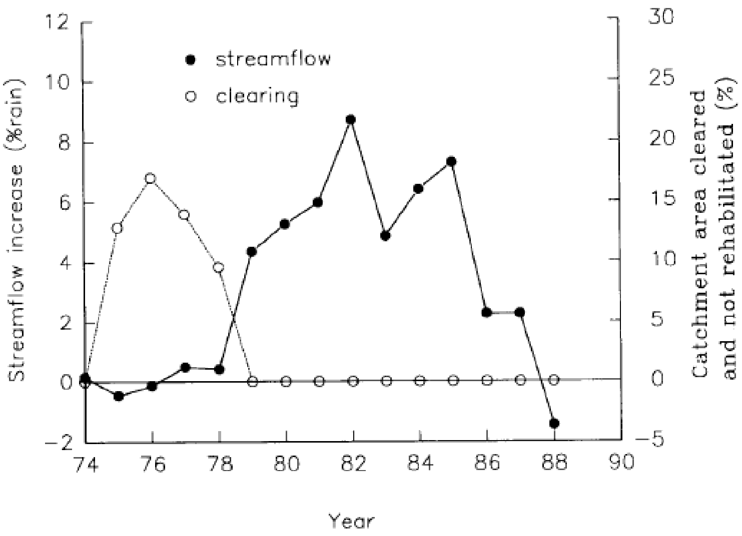


Figure 5.9 Water yield response at the Del Park catchment following bauxite mining – (a) annual streamflow relationship, (b) streamflow increase and clearing for mining (Ruprecht 1991)

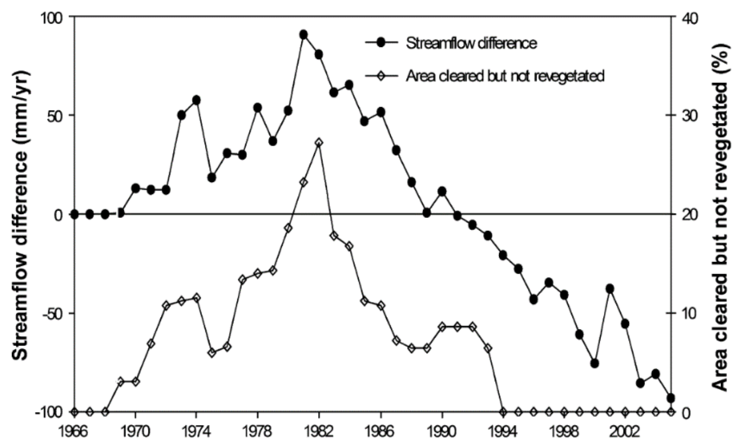


Figure 5.10 Streamflow change with mining at More Seldom Seen research catchment (Croton *et al.* 2005)

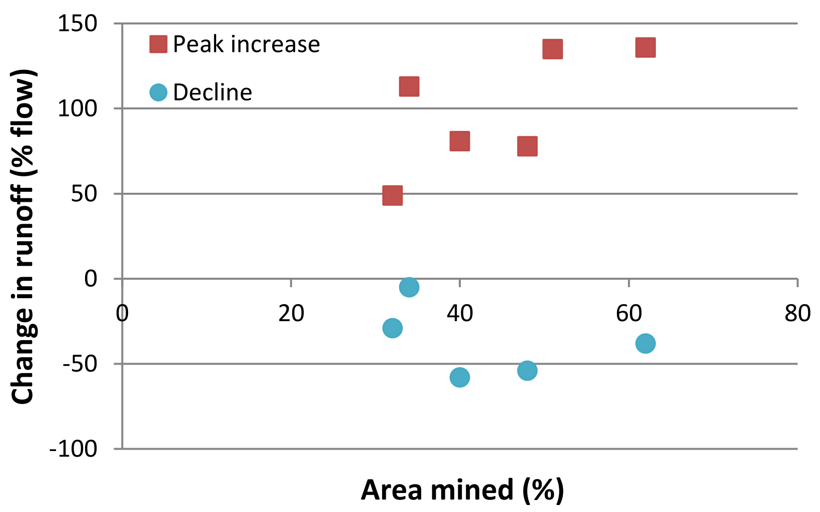


Figure 5.11 Runoff change with proportion of catchment (%) mined for bauxite (peak increase in maximum observed change in runoff, and decline is reduction in runoff observed post rehabilitation)

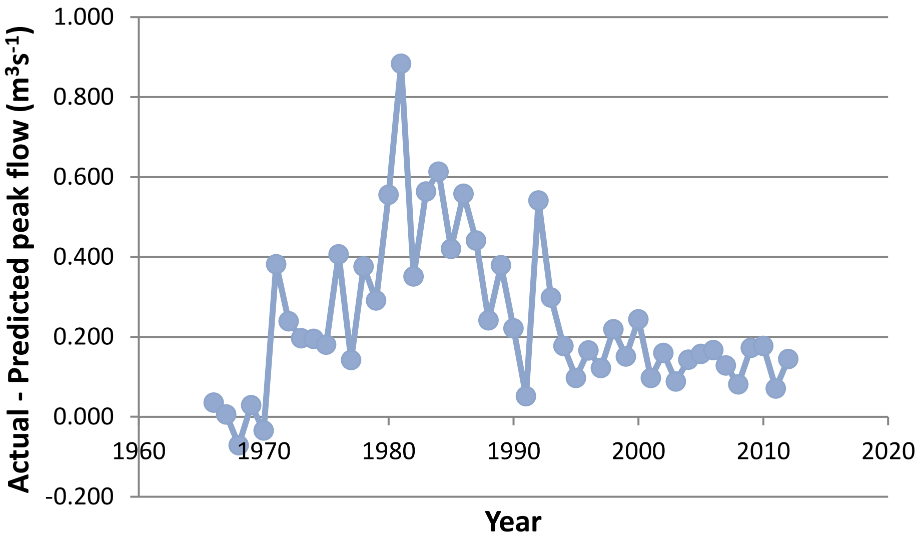


Figure 5.12 Comparison of peak flow for mined catchment (More Seldom Seen) compared to control catchment (Waterfall Gully)

5.2.5 Impact of bushfires on forest hydrology

Forest disturbance due to wildfires or planned fires[11] reduces vegetation cover and evapotranspiration followed by rigorous regrowth. Fire also removes surface litter which intercepts rainfall. Although overseas and eastern Australian studies, have reported increases in water repellency and overland flow following forest fires (Cawson *et al.* 2013; Scott 1997; Scott and van Wyk 1990), this effect has not been reported in the forests of SWWA. The impact of fire on water yield can be transient or longer term. Batini and Barrett (2007) reported on the initial impact of a bushfire on the hydrology within the Mundaring weir catchment. In the first winter after the fire the burnt catchments yielded approximately 2.2 times what would be expected based on pre-fire conditions (Batini and Barrett 2007).

[11] the term "bushfire" is used to encompass both types of fire.

Studies in eastern Australia into effects of severe, uncontrolled bushfires on forest hydrology concluded that increases in streamflow, erosion and sediment load occurred (Brown 1972; Lane *et al.* 2006; Langford 1976). The effect was observed to last up to five years after the fire. In one south-eastern Australian study (Lane *et al.* 2006) the initial increase in streamflow was followed by a decrease due to the evapotranspiration of the regrowth. The impact on forest hydrology will depend on the severity of the fire, with a severe fire killing mature trees compared to a mild, regular burning.

Kuczera (1987) modelled the long-term impact of bushfire on water yield in an Ash-Mixed species eucalypt forest and developed an equation describing the reduction in water yield following a bushfire, based on forest age. Mannik *et al.* (2013) estimated the change in streamflow resulting from the 2003 and 2006/2007 bushfires in southeastern Australia. They found that a significant reduction in streamflow is expected to occur in most of the impacted catchment, with the largest reductions to occur 15 to 25 years after the fires.

Doerr and Shakesby (2006) summarised international studies into the impact of forest fire on catchment hydrology and found that the post-fire hydrological responses can be increased by up to two orders of magnitude, with increases in peak discharge being often considerably greater than those in total flow. They also found that the time for enhanced hydrological response to return to pre-fire levels was highly variable.

5.2.6 Impact of dieback disease

As described in Chapter 1, jarrah dieback (*Phytophthora cinnamomi*) is a soil-borne fungus that can lead to tree and understorey death of susceptible species such as jarrah and banksia. Because it can lead to substantial jarrah tree death, dieback disease is considered to have a major impact on water yield of the northern forest catchments (Schofield *et al.* 1989b). For example, studies have found an increasing trend in streamflow to be associated with an increasing area affected by dieback in the Wungong water supply catchment (Batini *et al.* 1980). The reduced vegetation cover owing to dieback led to reduced transpiration and interception components. However, dieback sites are associated with soils with poor vertical drainage and greater throughflow (Kinal 1986).

5.2.7 Reforestation

Reforestation generally leads to a decrease in water yield due to increased transpiration; however, the magnitude can vary with the level and location of reforestation. The effects of reforestation techniques on groundwater level, streamflow and salt load have been reported (Bari and Boyd 1994; Bari and Schofield 1992; Bell *et al.* 1990; Schofield *et al.* 1989a). The effects of partial reforestation on streamflow and salinity were monitored at the Maringee Farm, Batalling Creek, Padbury Road, Pardellup and Barrama catchments. Through catchment modelling or paired catchment studies, the impact of reforestation ranged from negligible at 17% of cleared area reforested to 24% for 40% cleared area reforested (Table 5.3; Fig. 5.13).

Table 5.3 Summary of catchment studies into reforestation

Catchment	Cleared area reforested (%)	Streamflow change (%)	Salt load change (%)
Maringee	18	-10	-15
Batalling Creek	40	-16	-14
Padbury Road	75	-20	-20
Pardellup (first 3 years only)	40	-24	-29
Barrama (first 3 years only)	17	2	-1

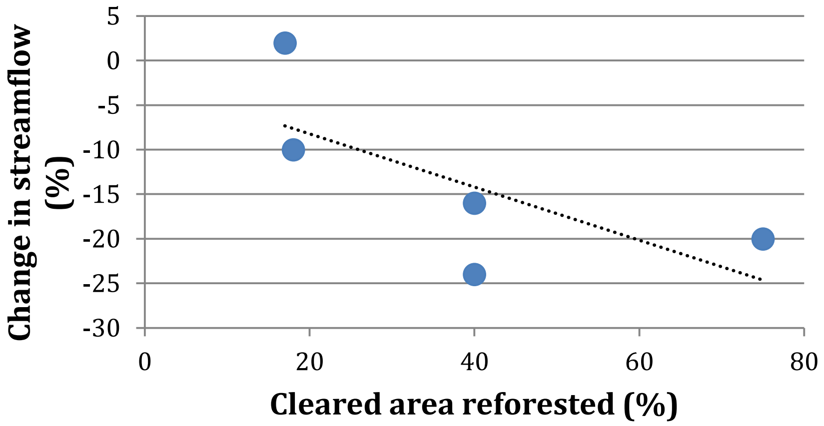


Figure 5.13 Impact of change in proportion of catchment reforested on streamflow (% change)

5.2.8 Comparison of land use practices

The comparison of water yield change after agricultural clearing in different rainfall zones (Fig. 5.2) emphasises the varying magnitude of responses to different types of treatments. The higher average annual rainfall and the presence of a groundwater discharge area were considered crucial for the significant increase in water yield from the low rainfall to the high rainfall case (Chapter 4; Ruprecht and Schofield (1989a)). A comparison of the impact of land use practices on water yield (Table 5.4) within the jarrah forest shows the difference between transient and permanent land use change (Fig. 5.14). For forest harvesting and regeneration, and bauxite mining, the water yield returned to levels either below or approaching pre-treatment levels after 10 to 12 years. The long-term prognosis is uncertain, and further reductions in water yield are possible. In general, the greater the reduction in forest density, the greater the initial water yield increase. In addition, a more sustained forest density reduction led to more sustained water yield increase. Whilst this is consistent with the Budkyo and Zhang (Zhang *et al.* 2001) concepts the Zhang model has been modified to suit the vegetation and soil profile characteristics of south west Western Australian catchments (Department of Water 2007).

The treatments illustrated in Fig. 5.14 show a lag between treatment and peak water yield change. This lag was 3 years for forest logging, 5 years for forest thinning, 6 years for bauxite mining and 8 years for agricultural development. These lags are typically longer than reported elsewhere (Bosch and Hewlett 1982; O'Shaughnessy and Jayasuriya 1991) and can be attributed to the deep soil profiles and large soil water storage capacities of the jarrah forest catchments. From Fig. 5.14, the time for the water yield to return to pre-disturbance levels after forest harvesting and regeneration, and bauxite mining and rehabilitation, is of the order of 12 to 15 years.

Table 5.4 Comparison of water yield increases from various land uses in high rainfall areas of south-west Australia

Catchment	Forest disturbance	Cover before (%)	Cover after (%)	Maximum increase in water yield (% rain)	Permanent increase in water yield
Wights	Clearing	80	0	30	Yes
Lewin South	Timber harvesting	90	20	14.5	No
Hansen	Thinning	60	14	20	Yes
Del Park	Mining	80	66	9	No
More Seldom Seen	Mining	n/a	n/a	19	No

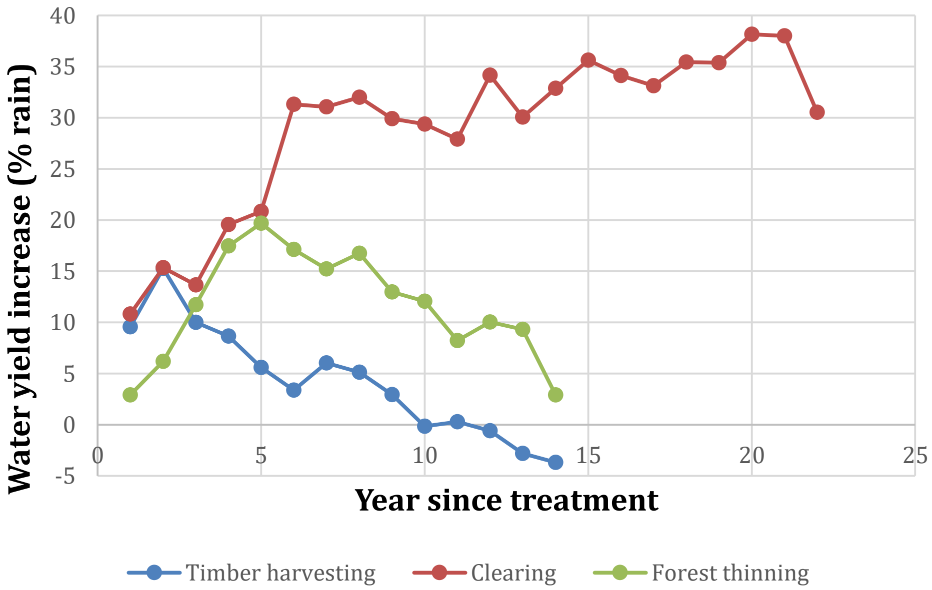


Figure 5.14 Comparison of water yield responses to land use practices within the high-rainfall zone (after Bari and Ruprecht, 2003)

5.2.9 Long-term implications for management of water yield

The long-term prognosis for water yield from areas subject to forest harvesting, forest thinning and bauxite mining is complex due to the interrelationships between vegetation cover, tree height and age, and catchment evapotranspiration. This is particularly the case when more recent forest disturbances are overlaid on historical forest disturbances, coupled with the impact of a drying climate in the south-west.

The increase in water yield associated with a reduction in vegetation cover is clearly shown in the research catchment scale studies (Section 4.1, 4.2 and 5.2). The reduction in water yield with increasing vegetation cover is also clear. However, quantifying the long-term equilibrium water yield from bauxite mining, forest thinning or even reforestation is difficult. For bauxite mining, the changes to the soil profile, catchment topography and nutrient status mean that to determine the long term prognosis with confidence requires further investigation.

Studies in eastern Australia have shown that long-term annual water yields can decline by up to 50% as a result of forest disturbance (O'Shaughnessy and Jayasuriya 1991). Stoneman *et al.* (1988) found that vegetation cover from jarrah forest areas subject to logging recovered to levels similar to, but not exceeding, pre-disturbance levels of vegetation cover. However, Borg and Stoneman (1988) discussed the possibility that younger trees have the potential to transpire more per leaf area and this was also reported for regrowth forests by Macfarlane *et al.* (2010). The prognosis on long-term water yield for areas rehabilitated after bauxite mining is even less certain, as this involves a changed understorey, different nutrient balance and changed soil profile.

The jarrah forest is now a complex mosaic of different forest types and ages. Management of the forest for water yield needs to acknowledge this complexity and evaluate forest management strategies both at the large catchment scale and at short and long time-scales. The historical extensive network of small catchment experiments, regional studies, process studies and catchment modelling studies were integral to the understanding of the hydrologic impact of various forest disturbances. Targeted research programmes are critical to assist with management decisions regarding forest management to enhance water values for current and future generations.

5.3 Forest disturbance and salinity

The salinity of un-disturbed forest catchments typically range from 80 to 350 mg L^{-1} TDS (Mayer *et al.* 2005). The salt that falls on the catchment from rainfall and dryfall is considered the salt input to the catchment and the salt exported through streamflow is termed the output. Salt accumulation in catchments can occur, with salt stored in the soil profile or in the groundwater. A salt output/input ratio of >1 means that more salt is being exported from the catchment that is coming in with the rainfall. A salt output/input ratio of <1 means that salt is accumulating in the soil profile or groundwater. The salt output/input ratios in forested catchments have been observed to range from 2.1 (mean annual rainfall of 1300 mm) to 0.17 (mean annual rainfall of 740 mm) (Mayer *et al.* 2005). Salt output/input ratios of forested catchments have been observed to decrease with decreasing mean annual rainfall (Loh *et al.* 1984).

5.3.1 Clearing for agriculture and salinity

Forest clearing for agriculture typically changed the salt balance of a catchment from a state of equilibrium or accumulation to a state of net salt export (Williamson *et al.* 1987). After forest clearing, output/input ratios exceeded one when mean annual rainfall was below 1000 mm. The high rate of salt exported from cleared catchments means that eventually stream salinity will drop to levels equivalent to the salt input from rain and dry fallout. However, the time required for the salt to leach is significant (this is further explained in later paragraphs).

The annual stream salinity at the Wights catchment increased immediately after clearing from an average of 360 to 515 mg L^{-1} TDS (Fig. 5.15). The average annual salt load increase at the Wights catchment was 14-fold (compared with the control catchment). At the Lemon catchment, from 1977 to 1987 (before the groundwater system was connected to the streambed), the average annual stream salinity rose from 80 to 127 mg L^{-1} TDS. After 1987 when the groundwater system reached the streambed, the stream salinity generation process changed significantly and annual average salinity increased systematically to 1700 mg L^{-1} TDS. The annual stream salt load increased 180-fold compared with the control catchment. Modelling by Croton and Bari (2001) demonstrated that under current land use and climate, the annual stream salinity of the Lemon catchment will be stable for at least the next 50 years.

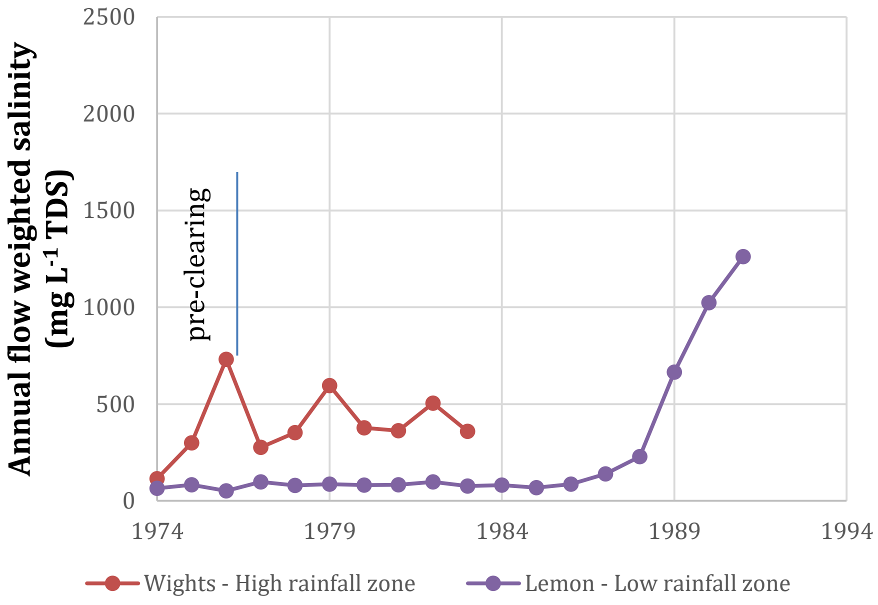


Figure 5.15 Increase in stream salinity following clearing for agriculture at Wights and Lemon catchments (adapted from Bari and Ruprecht 2003)

Two conceptual methods have been applied to estimate the time for the salt outputs of a cleared catchments to return to a situation where these equal inputs. These are the characteristic time approach (Peck and Hurle 1973) and the salt storage export method (Fig. 5.16) (Mayer *et al.* 2005). Another method applied was the exponential decay, which Hatton *et al.* (2002) used to estimate salt leaching. However, Hatton *et al.* (2002) also showed that if the target is freshwater then the time required may be significantly shorter for streams in higher rainfall areas to become fresh (<500 mg L^{-1} TDS) again, reducing to decades rather than centuries. In addition, the time to achieve a target of 500 mg L^{-1} TDS at downstream locations which receive significant fresh water contributions may also be much shorter.

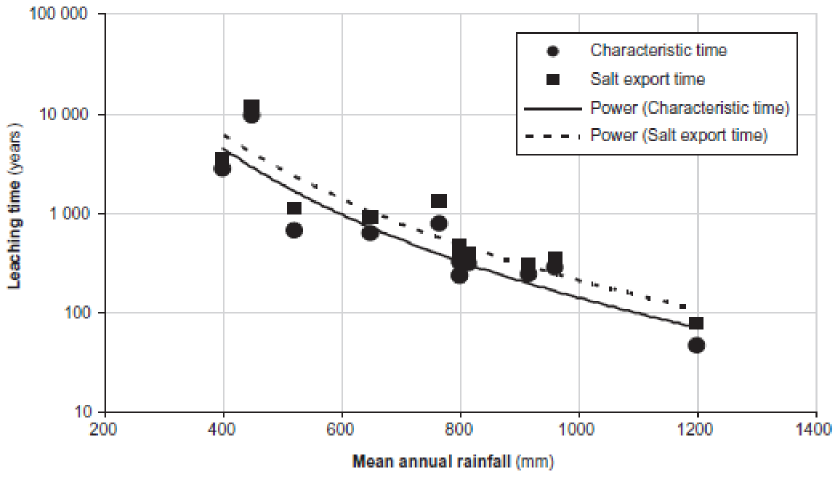


Figure 5.16 Estimated time for salt to be leached from catchments (from Mayer et al. 2005)

5.3.2 Timber harvesting and salinity

After timber harvesting, stream salinity increased at all the treated paired catchments in the southern forest research trial. The annual stream salinity at three treated catchments, but not the March Road catchment (no buffer), remained below 500 mg L^{-1} TDS (the upper limit for fresh water resources set by the Western Australian Water Resources Council (1986)) (Fig. 5.17). In the dry-year of 1987, highest annual stream salinity (780 mg L^{-1} TDS) was recorded at the March Road catchment — the only record exceeding 500 mg L^{-1} TDS (Fig. 5.17). Most of the time the annual salinity was less than 200 mg L^{-1} TDS.

After logging, the annual stream salinity at Lewin South increased until 1985 and then levelled off. At March Road, stream salinity decreased in 1982 and then began to rise. The maximum increase in annual salinity (320 mg L^{-1} TDS) occurred in 1987. Since then there has been a decreasing trend (Fig. 5.17). If there had been no logging of the March Road catchment, annual stream salinity in 1987 would have been 460 mg L^{-1} TDS (Bari and Boyd 1993). At April Road North, stream salinity increased until 1987 and then began to fall. The maximum increase was 115 mg L^{-1} TDS. In 1994 the stream salinity was slightly higher than it would have been without logging. At the Yerraminnup South catchment there were no significant changes in stream salinity following logging (Fig. 5.17). After logging, the stream salt load increased at all treated catchments as a result of higher flow and salinity.

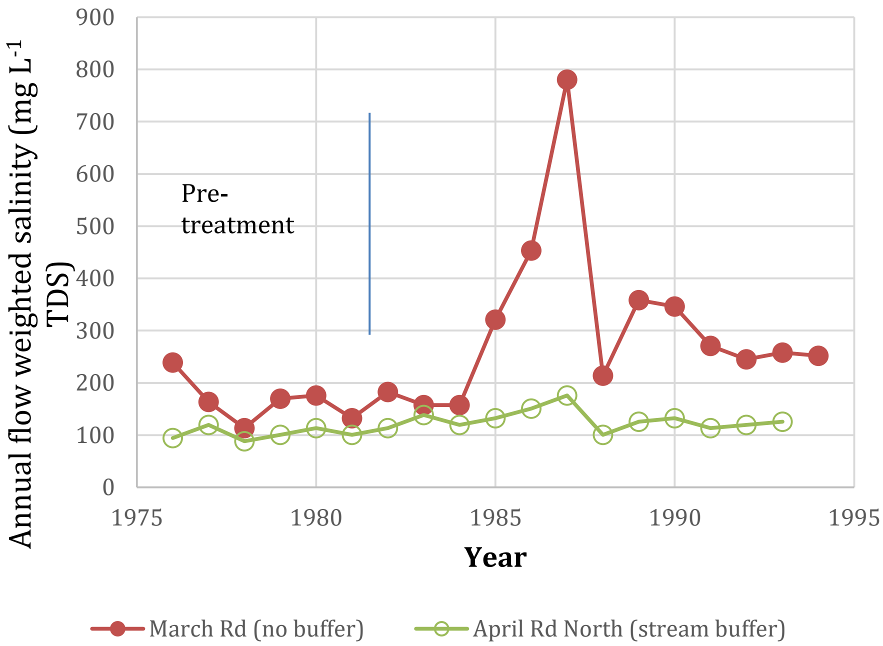


Figure 5.17 Impact of timber harvesting on stream salinity (from Bari and Ruprecht, 2003)

At the Yarragil 4X (standard harvest treatment) and Yarragil 6C (intensive harvest treatment) research catchments annual stream salinity did not increase in response to either treatment because saline deep groundwater did not rise following the treatments (Kinal and Stoneman 2011).

Historic records of intensive logging within the karri (*E. diversicolor*) forests in the SWWA suggest that salinity has remained low and clearfelling in karri and heavy selection cutting in jarrah forests have minor effects on salinity (Johnston *et al.* 1980). In summary, there is no evidence to suggest that harvesting of jarrah and karri leads to long-term or severe increases in salinity, as long as regeneration occurs soon after completion of logging (Borg *et al.* 1987).

5.3.3 Forest thinning and salinity

The annual flow-weighted stream salinities measured through grab samples are in the range of 110 to 120 mg L^{-1} TDS at the Hansens and Lewis catchments and 120 to 130 mg L^{-1} TDS at Higgens and Jones. These annual stream salinities are about the same as the groundwater salinities (Robinson *et al.* 1997). For the high rainfall forest thinning catchments, the stream salinity post-treatment was not discernibly different from pre-treatment salinities. The streamflow on all catchments was fresh (well below 500 mg L^{-1} TDS) for the entire study period. These results are typical for streamflow in the High Rainfall Zone because of low salt storage (Robinson *et al.* 1997). At the Yarragil 4L catchment, in the Intermediate Rainfall Zone, the annual stream salinity was highly variable (from 75 to 170 mg L^{-1} TDS. At this catchment, there was no apparent increase in stream salinity following forest thinning, despite the relatively higher salt storage in the landscape, because there was no groundwater discharge to the stream.

Moulds and Bari (1994) studied the effects of forest thinning on streamflow and salinity associated with rainfall at an intermediate rainfall catchment. The catchment studied had an average annual catchment rainfall of 1100 mm and had low salt storage more typical of the high rainfall zone. Results of the study indicated that salt flow increased, but increases in streamflow diluted additional salt load so that stream salinity was generally lower than pre-treatment levels. In effect forest thinning produced a significant increase in water quantity without reducing quality.

5.3.4 Bauxite mining and salinity

Stream salinity was measured by manual sampling at the Del Park, Seldom Seen and More Seldom Seen catchments. Annual stream salinities at the Seldom Seen and More Seldom Seen catchments were higher in the 1960s and late 1970s, but these results may be influenced by the small number of samples taken. Between 1976 and 2000 the annual flow-weighted salinities of these catchments were stable. Annual flow-weighted stream salinity at the Del Park catchment varied between 100 to 150 mg L^{-1} TDS and during most of the year was lower than at the other two catchments. The streamflow of all three catchments remained fresh (well below 500 mg L^{-1} TDS) during the entire study period (Croton 2004; Croton *et al.* 2005).

5.3.5 Reforestation and salinity

The effects of partial reforestation on salinity were monitored at the Upper Hay research catchments, Maringee Farm, Batalling Creek and Padbury Road catchments. Catchment modelling of the Maringee Farm and Batalling Creek catchments indicated that 18% reforestation at Maringee Farm, predominantly along the stream zone, reduced streamflow by 10% and salt load by 15% (Bari and Croton 2000). However, salinity increased due to the disproportionate reduction in streamflow. Ten years after reforestation, the catchment showed a new stability and no further reduction in streamflow or load is to be expected (Bari and Croton 2000). Modelling of the Batalling Creek catchment showed similar results (Bari and Croton 2002).

At the Padbury Road catchment, where 75% of the cleared area was replanted between 1977–1983, streamflow and salt load fell to approximately 20% of the pre-planting values and stabilised in the 1990s, 10 years after planting (Fig. 5.18). The average annual stream salinity increased from a pre-planting value of 1070 mg L^{-1} TDS to more than 2000 mg L^{-1} TDS in 1986. After that, there was a decreasing trend in stream salinity which stabilised at an average 1020 mg L^{-1} TDS in the mid-1990s.

Reforestation at the small catchment scale clearly demonstrates a reduction in salt load. However, this is in many cases associated with an even greater reduction in streamflow. van Dijk and Keenan (2007) outlined this issue for the Murray Darling Basin. However, the importance of reducing salt load in salt hazard areas as part of a larger water supply catchment scale has been demonstrated for the Denmark River, where stream salinity has reduced significantly since major reforestation in the upper catchment (Ward *et al.* 2011). The reduction in streamflow was not as significant as the reduced salt load, despite higher local stream salinities.

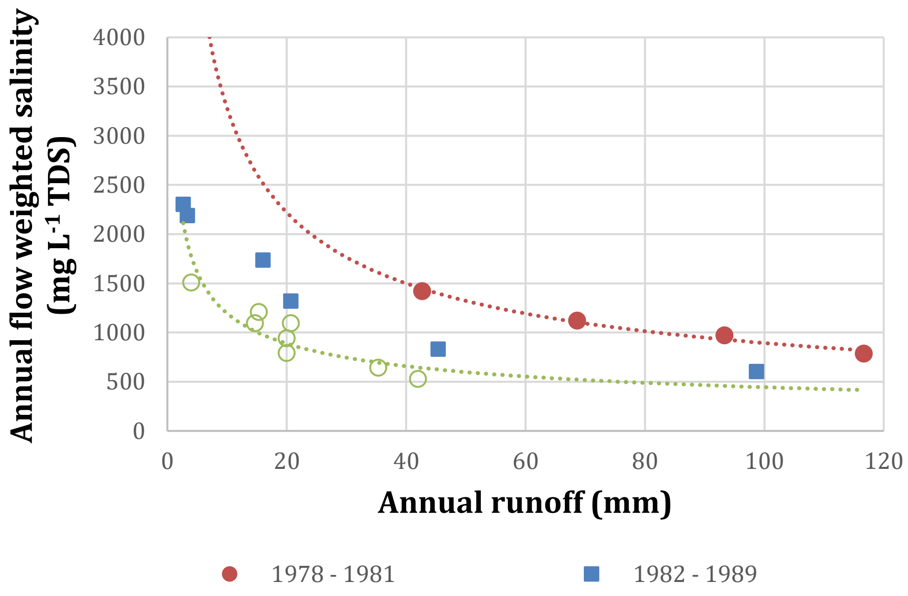


Figure 5.18 Salinity to streamflow relationship for the Padbury Road catchment (Bari and Ruprecht, 2003)

5.3.6 Regional salinity implications

The estimated average annual discharge from the rivers of SWWA is 4700 GL and with this discharge is about 7.5 million tonnes of salt (Mayer *et al.* 2005). Of these rivers in SWWA, 44% of the flow out to sea each year is fresh, 10% marginal, 21% brackish, about 20% moderately saline, 3% saline and 2% highly saline.

The major source of additional salt in streamflow comes from cleared areas where salt has accumulated in the subsoil prior to clearing. Salts have accumulated in the deep subsoils of the Darling Range and the Leeuwin-Naturaliste Ridge, particularly in areas where the annual rainfall is <900 mm.

For any stream east of the Darling scarp and the 900 mm isohyet, clearing as little as 10% of its catchment is sufficient to cause it to become either brackish or saline (Fig. 5.19). Without higher rainfall areas to dilute the additional salt discharge from the small cleared area, average stream salinities are commonly >2000 mg L^{-1} TDS. Stream salinities frequently exceed 10 000 mg L^{-1} TDS at low flows.

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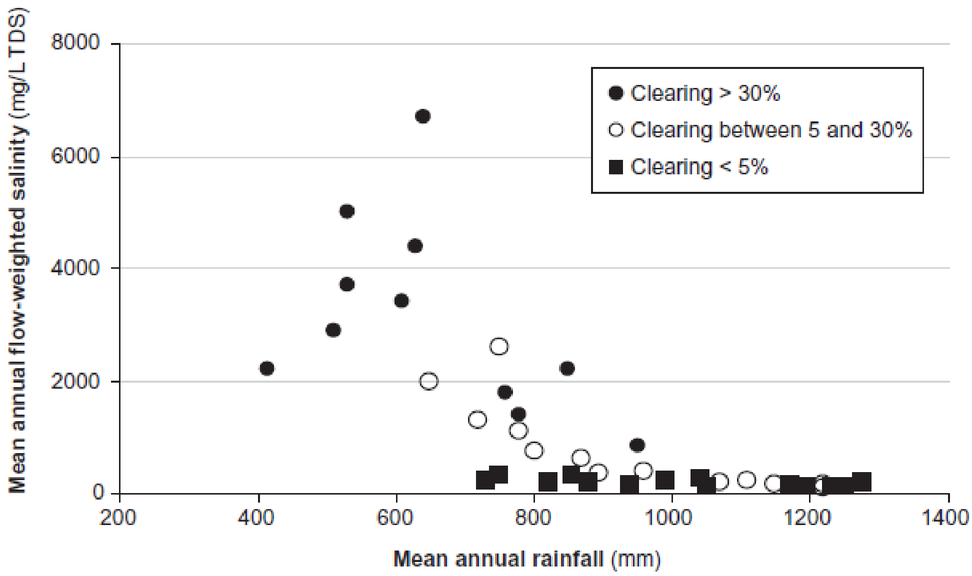

Figure 5.19 Salinity relationship with rainfall and clearing (from Mayer *et al.* 2005)

The permanent clearing of native vegetation in the 900 to 1100 mm rainfall zone can lead to average stream salinities of marginal quality. Streams draining these areas commonly have average salinities between 500 and 1000 mg L^{-1} TDS (Schofield *et al.* 1988). The salinity risk is low in areas where mean annual rainfall exceeds 1100 mm.

Table 5.5 Salinity of water resource recovery catchments

River	Gauging station	Mean annual salinity (mg L^{-1} TDS)		Reduction
		1993 to 2002	2010 to 2013(1)	
Denmark River	603136	750	460	-39%
Kent River	604053	1700	1075	-37%
Warren River	607220	990	870	-12%
Collie River	612002	1500	1290	-14%
Helena River	616216	1300	1000	-23%

(1) Based on mean annual flow

Five river systems in SWWA have been designated as Water Resource Recovery Catchments (State Salinity Council 2000). These catchments, which either have the potential to provide future water supplies (Denmark, Kent, Warren), or are already dammed (Collie, Helena), are all affected by salinity caused by land clearing to some degree.

The Collie, Warren, Kent and Hay Rivers (Table 5.5) are the most highly salt-affected river systems which drain the study region. While diluted by fresh inflows from the forested higher rainfall parts of their catchments, the salinities of all four catchments are either of marginal salinity or brackish. The degree of dilution varies between catchments. The proportion of the total catchment cleared of the Warren and Kent river catchments is similar (36 and 40% respectively). The salinity of the Kent is significantly higher as it has less fresh streamflow from the higher rainfall portion of its catchment than does the Warren River. All the water resource recovery catchments have shown a reduction in stream salinity in recent years (Table 5.5). This is considered due to a combination of lower rainfall, and substantial plantations established (in the Denmark and Kent River catchments in particular)(Bari *et al.* 2004; De Silva *et al.* 2006).

Chapter 6 Impact of a changing climate on forest hydrology

The forested water supply catchments in SWWA have seen a reduction in streamflow of between 36 and 52% in the last ten years compared to the previous 35 years (Petrone *et al.* 2010; Silberstein *et al.* 2012). However, rainfall across the northern jarrah forest has only declined by 4 to 6% during this period. This section will explore the relationship between the drying climate and the forest hydrology of the SWWA, including the possible causes of the changing relationship between rainfall and runoff, and the likely implications of recent climate change scenarios (Hope *et al.* 2015).

6.1 Hydrologic impact of observed climate

As outlined in the Background and thesis aims (Chapter 1) the climate of SWWA is considered Mediterranean-type with approximately 80% of annual rainfall occurring between May and October. The mean annual rainfall in the forest areas varies from over 1200 mm in the western area to about 500 mm on the eastern edge (see Fig. 2.4).

6.1.1 Observed changes in rainfall

In SWWA a substantial drying trend has been observed over the last 40 years, with significant decreases in rainfall over this period (Fig. 6.1) (IOCI 2012). This drying trend in rainfall has strengthened and extended. For example, since 2000 annual rainfall has continued to be below average and in addition early winter

(May to July) rainfall at many sites was significantly below the average for 1969 to 1999 (IOCI 2012).

Winter rainfall (May to October) at the Jarrahdale rainfall station shows the typical decline in rainfall since the 1970s (Fig. 6.2). Rainfall in the period from 1889 to 1974 was 18% higher than that of 1975 to 2000. From 2001 to 2012 the mean has been 9% lower than the 1975 to 2000 period. Since 2012 the trend has been variable with Jarrahdale stablising at the 2001 to 2012 period, whilst Collie and Darkan observed continued decline in annual rainfall.

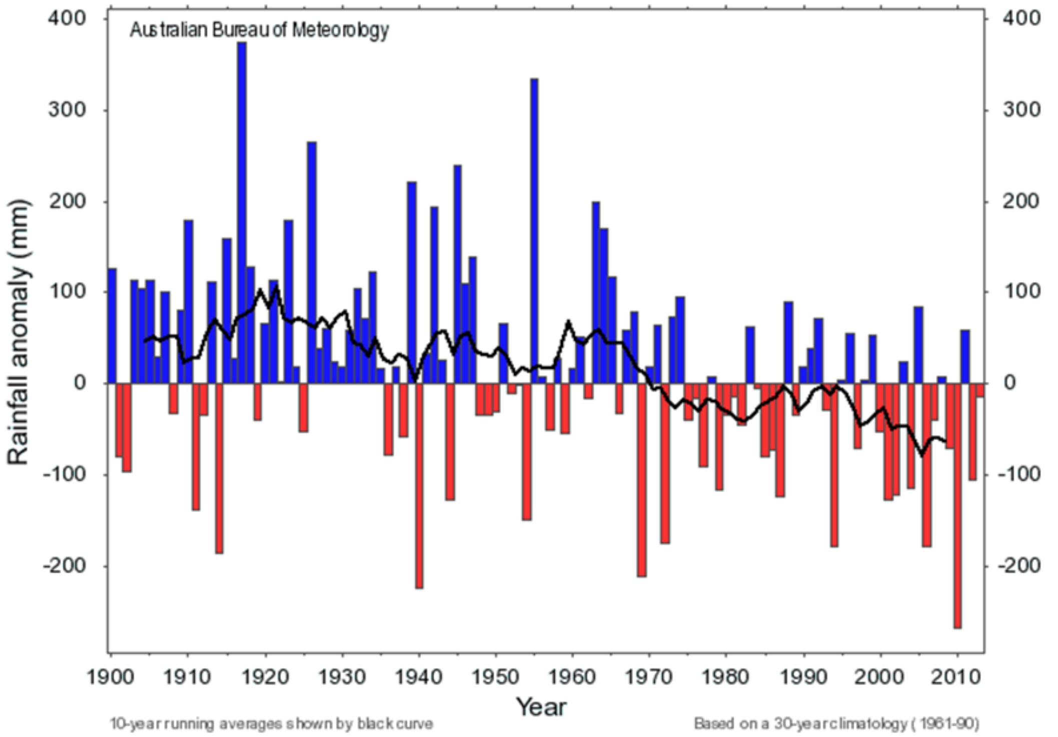

Figure 6.1 Rainfall anomaly for south-west Western Australia (data from Bureau of Meteorology)

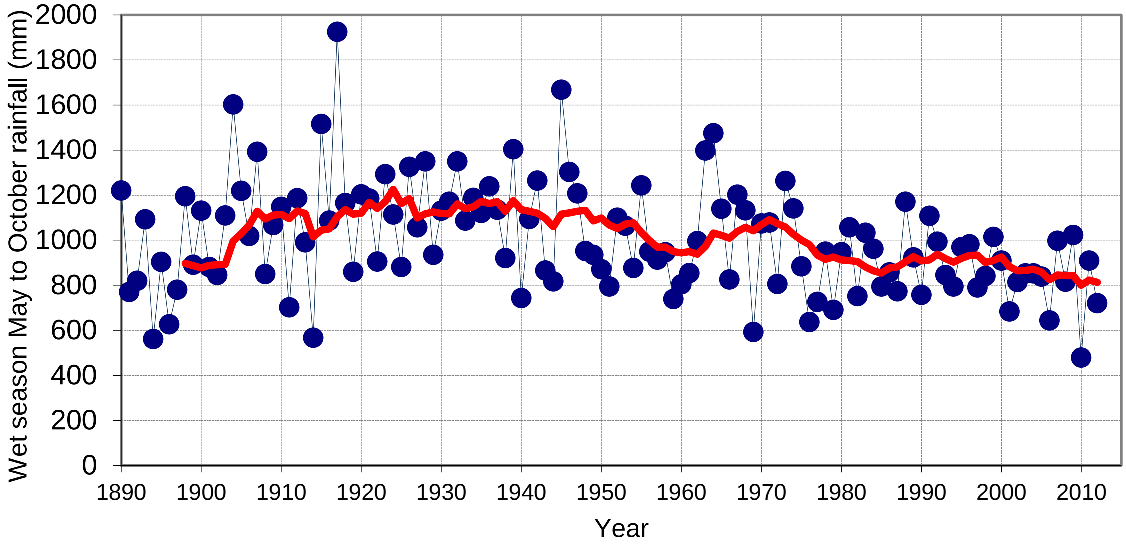

Figure 6.2 May to October rainfall for Jarrahdale (009023) rainfall station (red line is 10 year backward moving average)

The change in mean annual rainfall across the northern jarrah forest shows a range from +2 to -7% for five representative sites (Table 6.1). Given that most of the runoff is generated in the higher rainfall parts of the catchments, the decline in the two higher rainfall sites (Jarrahdale and Dwellingup) is considered to be more relevant.

Although mean annual rainfall may have only declined 6 to 7% in the high rainfall areas, the below average years have been significant. The winter rainfall years of 2001, 2006 and 2010 show deficits of between 300 and 500 mm (Fig. 6.3). The documented drying trend (Bates *et al.* 2008a; IOCI 2012) in rainfall since the mid-1970s has been considered due to a shift in the global climate system which has influenced sea surface and atmospheric temperatures and ocean currents (Giese *et* *al.* 2002; Wainwright *et al.* 2008). The Indian Ocean Climate Initiative (IOCI) produced station-scale climate projections indicate that SWWA rainfall is likely to continue to decline through the middle and end of this century (IOCI 2012). The climate modelling by the IOCI used the results from the IPCC Coupled Model Intercomparison Project Phase 3 (CMIP3). More recent results (Hope *et al.* 2015) have not diverged from this assessment.

Table 6.1 Change in mean annual rainfall for selected forested locations within SWWA

Site	Mean (1975 to 2012)	Mean (2001 to 2012)	Change (%)
Dwellingup	1157	1074	-7
Jarrahdale	1048	984	-6
Collie	830	802	-3
Darkan	492	501	+2

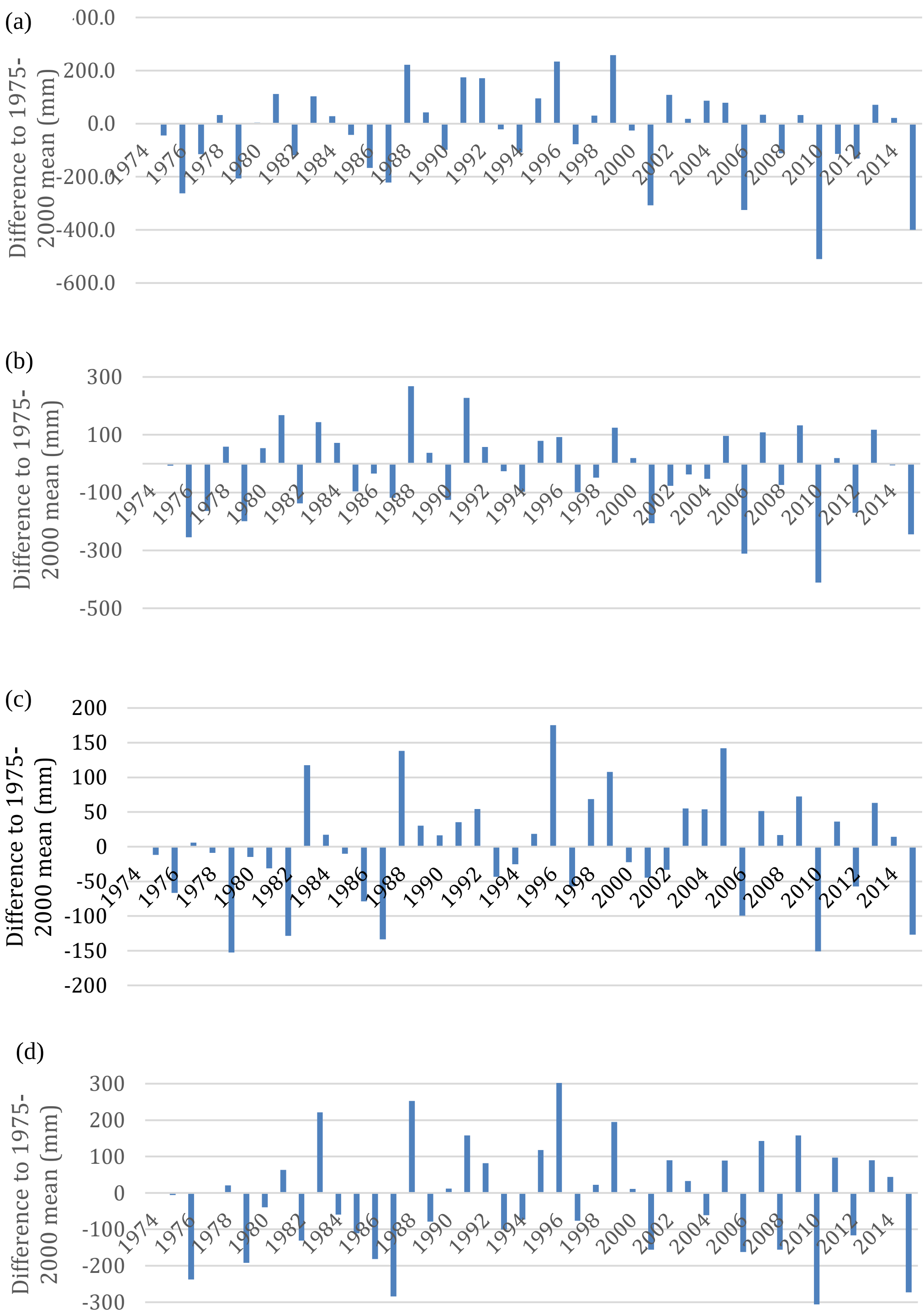

Figure 6.3 Annual winter rainfall compared to the 1975 to 2000 mean for selected stations across the jarrah forest (a) Dwellingup, (b) Jarrahdale, (c) Collie, and (d) Darkan.

6.1.2 Observed changes in forested groundwater levels

An important aspect of the water balance in the jarrah forest is groundwater. The importance of groundwater to historical streamflow generation has been clearly shown in Chapters 4 and 5. The presence of a groundwater discharge area significantly increases runoff generation, streamflow volumes, and flow duration (Chapters 4 and 5).

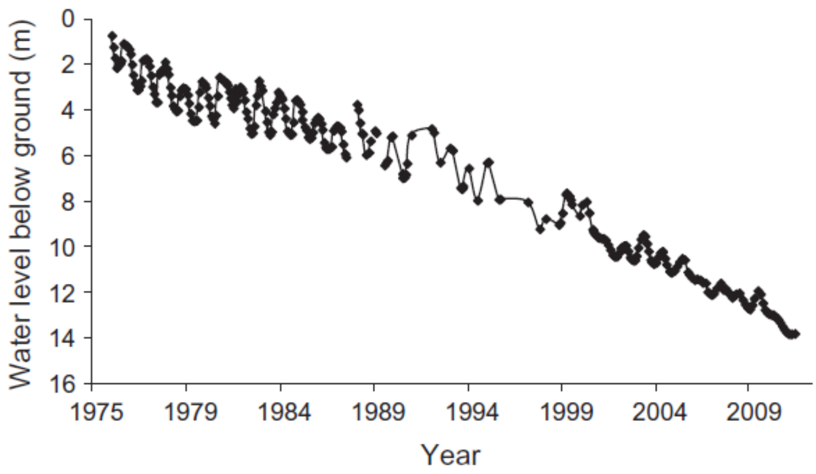

Figure 6.4 Valley piezometer in Yarragil 4X catchment (Kinal and Stoneman, 2012) showing clear decline in groundwater levels.

In addition to the declines identified in rainfall, groundwater levels have declined (Fig. 6.4; Fig. 6.5). The data from the research catchment Yarragil 4X (Kinal and Stoneman 2012) shows a nearly continuous decline in groundwater from 1975. Croton *et al.* (2013) also reported groundwater declines in the Gordon catchment (untreated forested control) from the commencement of monitoring in 1993 (Fig. 6.5). If the current rainfall trend continues Croton *et al.* (2013) forecasts that the regolith groundwater in the Gordon research catchment (which was 10 GL in 1996) could disappear in 13 years.

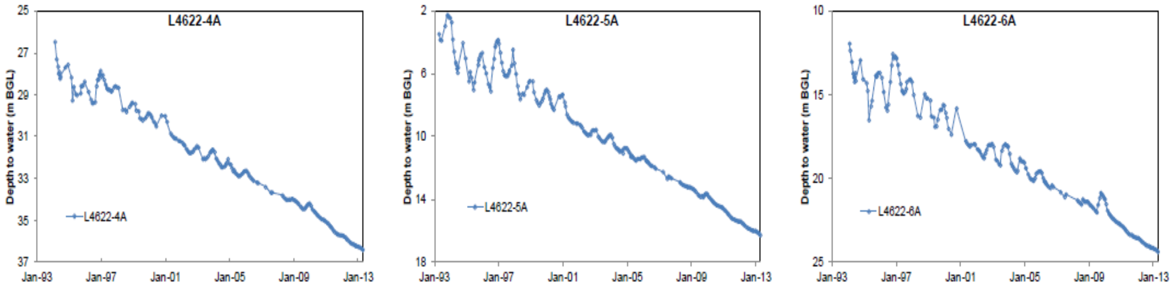

Figure 6.5 Groundwater hydrographs for Gordon Catchment in the IRZ (Croton *et al*. 2013)

The potential for an areal expansion in the disconnection of groundwater from the valley floor is considered greatest in the high rainfall zone of the northern jarrah forest and the southern forest areas. In the high rainfall area of the northern jarrah forest, groundwater levels have been historically higher than the lower rainfall areas of the jarrah forest (Schofield *et al.* 1989b). Therefore, the extent of connection with the surface has also been greater in the high rainfall areas (Kinal and Stoneman 2012). In the southern forests of SWWA, the rate of decline in groundwater levels has been relatively slower and hence groundwater levels have remained comparatively high (Kinal and Stoneman 2012).

Modelling has also demonstrated the contraction of the groundwater systems in different catchments, and has indicated that under the expected continuing drying trend, groundwater will continue to decline to such a level that there is likely to be wide scale disconnection of the groundwater from the streamzone, resulting in ongoing streamflow declines (Croton *et al.* 2013).

6.1.3 Observed changes in forested streamflows

Forested streamflows have also shown a decline since the 1970s. For example, annual runoff at Yarragil Brook (Fig. 6.6) had an observed decline of greater than 60% for 2001 to 2012 (9 mm) compared to 1975 to 2000 (23.3 mm), with the decline continuing beyond 2012. The majority of the decline has been observed in the winter period from June to August (Fig. 6.7).

In addition to the decline in annual water yield, many forested streams are decreasing in daily flow or changing from perennial to ephemeral, or becoming increasingly ephemeral (Fig. 6.8). This is considered to be linked to the declining groundwater levels and a disconnection of the groundwater system from the valleys (Petrone *et al*. 2010). This disconnection means that the saturated area for streamflow generation is significantly smaller at the beginning of winter. A saturated area (perched aquifer) may occur where the groundwater is not at the surface, but it will require substantially greater winter rainfall to develop.

Table 6.2 summarises the declining water yields for major water supply and irrigation sources across SWWA, based on the long-term average for 1975 to 2012, with the summary also outlined in Fig. 6.9. The decline ranges from 28 to 58% for urban water supplies and from 18 to 42% for irrigation water supplies. Over all sources, the total mean annual flow has reduced from 435 to 265 GL yr^{-1}, or a 39% reduction. In addition to the reduction in mean annual flow the streamflow has become more variable with an average coefficients of variation increasing from 0.42 to 0.55.

In comparison to the 1961 to 1990 reference period, the IWSS sites for 2001 to 2012 have reduced by 67% and the irrigation dam inflows by 41% (Fig. 6.10) compared to rainfall which has reduced by 15% at Dwellingup, 16% at Jarrahdale, and 10% at Darkan and Collie for the same time periods.

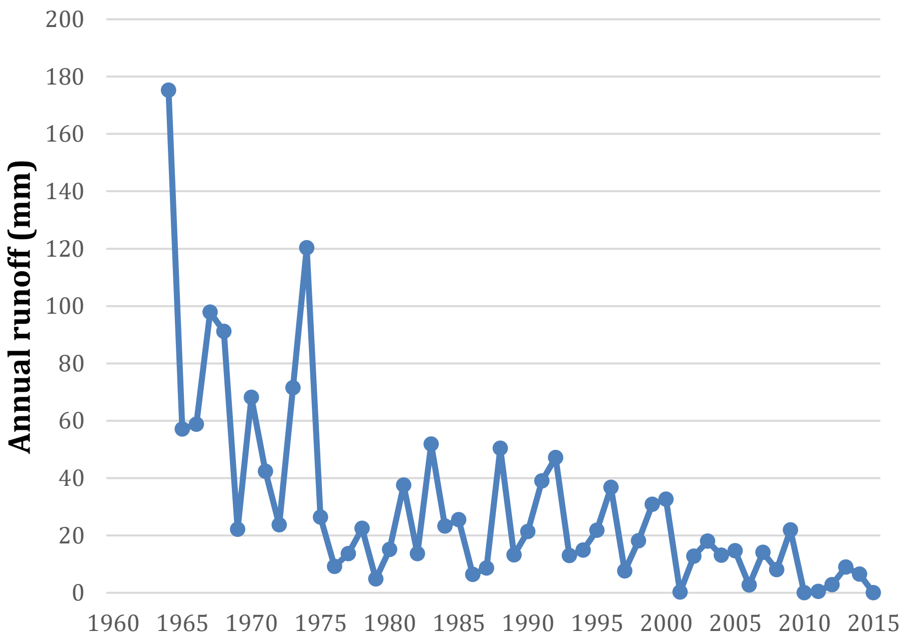

Figure 6.6 Annual runoff for Yarragil Brook showing declining runoff since the 1970s (Department of Water – Water Information Reporting)

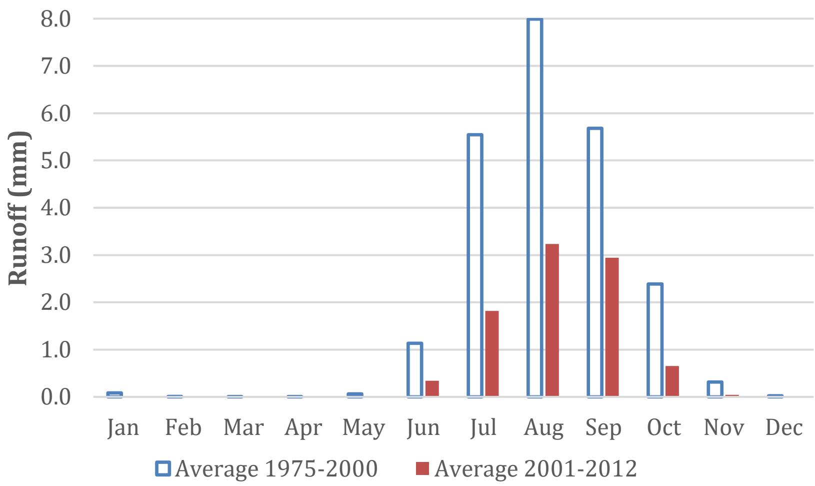

Figure 6.7 Monthly streamflow for Yarragil Brook for the periods 1975 to 2000 and 2001 to 2012 (Department of Water – Water Information Reporting)

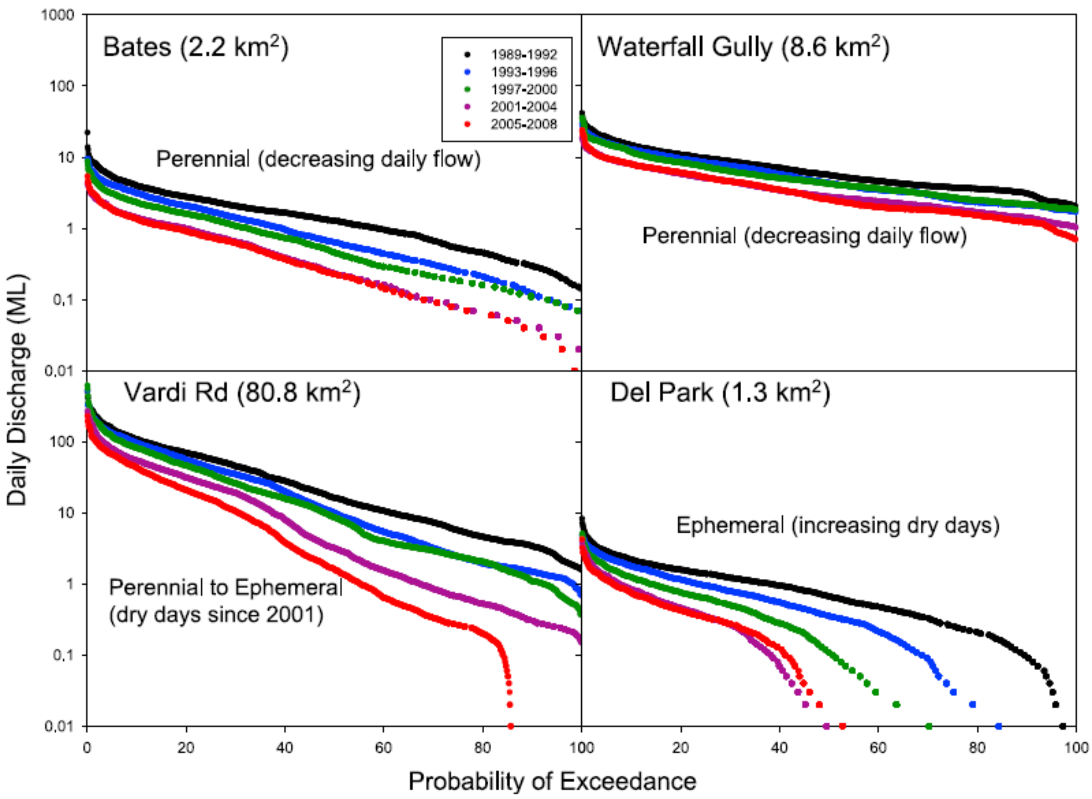

Figure 6.8 Flow duration curves for representative forested catchments (Petrone e*t al*. 2010)

The probability distribution for streamflow for 2001 to 2012 (Fig. 6.11) shows a greater reduction for the lower flows compared to the median and above. The probability distribution is similar to the Lemon research catchment when groundwater was disconnected from the ground surface (Fig. 6.12).

Figure 6.9 Change in streamflow at south-west dams (2001 to 2012 compared to 1975 to 2000)

Table 6.2 Summary of mean annual flow (GL) into forested catchments developed for urban water supply or irrigation for the periods 1975 to 2000 and 2001 to 2012 (based on Water Corporation data)

Dam	Catchment Area (km^2)	Forested area (%)	Mean annual flow 1975 to 2000 (GL)	Mean annual flow 2001 to 2012 (GL)	Change (%)	Use
Mundaring	1456	95	18.5	7.8	-58	GAWS & IWSS(1)
Lower Helena	114	79	12.8	5.0	-61	GAWS & IWSS
Canning	732	100	27.9	12.8	-54	IWSS
Wungong	132	100	19.3	8.1	-58	IWSS
Bickley	50	100	2.0	1.0	-51	IWSS
Churchman	16	100	2.9	1.4	-50	IWSS
Victoria	37	93	3.0	1.2	-59	
Serpentine	647	100	37.2	17.0	-54	IWSS
Serpentine Pipehead	28	100	3.6	1.6	-55	IWSS
North Dandalup	148	100	18.7	9.8	-47	IWSS
South Dandalup	310	100	19.6	10.7	-46	IWSS
Stirling(2)	251	100	48.7	32.6	-33	IWSS
Samson(2)	65	100	16.7	12.1	-28	IWSS
Harris(2)	321	100	20.5	16.1	-21	GSTWS and IWSS
Harvey(2)	376	65	39.3	25.9	-34	Irrigation
Logue(2)	64	99	12.6	7.6	-40	Irrigation
Waroona(2)	42	100	10.4	8.2	-21	Irrigation
Drakesbrook(2)	13	60	2.5	2.1	-18	Irrigation
Wellington	2829	76-83(3)	122.5	84.8	-31	Irrigation

(1) IWSS is the Integrated Water Supply System for Perth and Mandurah, GAWS is the Goldfields and Agricultural Water Supply, GSTWS is the Great Southern Towns Water Supply
(2) Based on data to 2011 from Department of Water.
(3) By 1977 76% was forested, but with reforestation this had been increased to 83% forested or plantation.

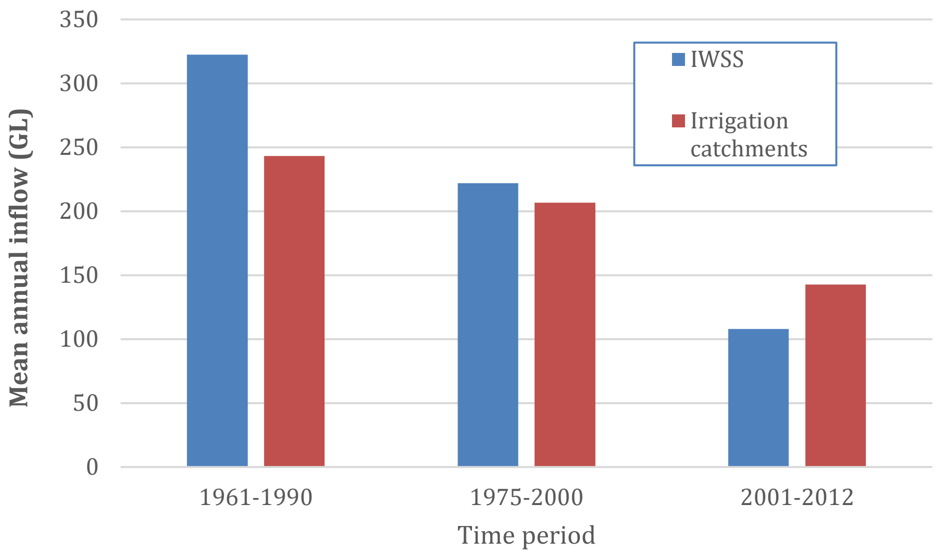

Figure 6.10 Trend in mean annual inflow for Integrated Water Supply Scheme and irrigation catchments for three time periods

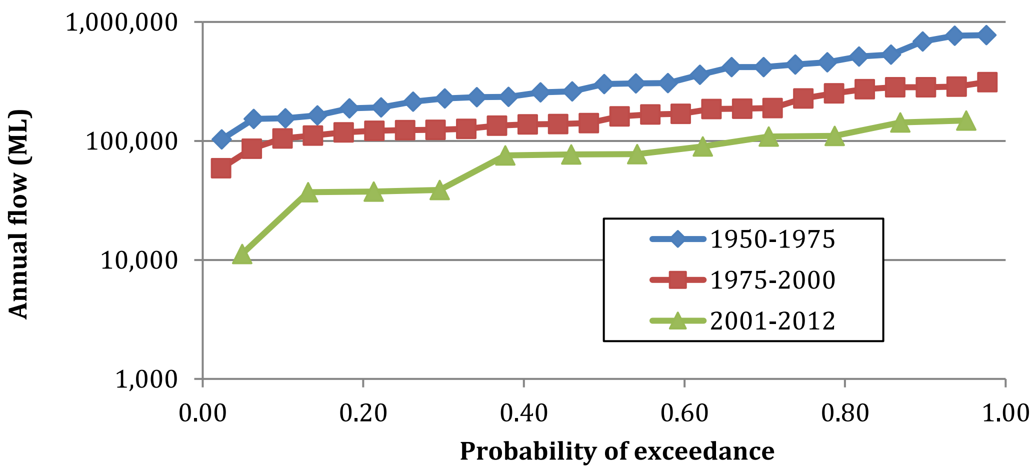

Figure 6.11 Probability distribution for Perth metropolitan dam streamflow for three periods (1950 to 1975, 1975 to 2000, 2001 to 2012)

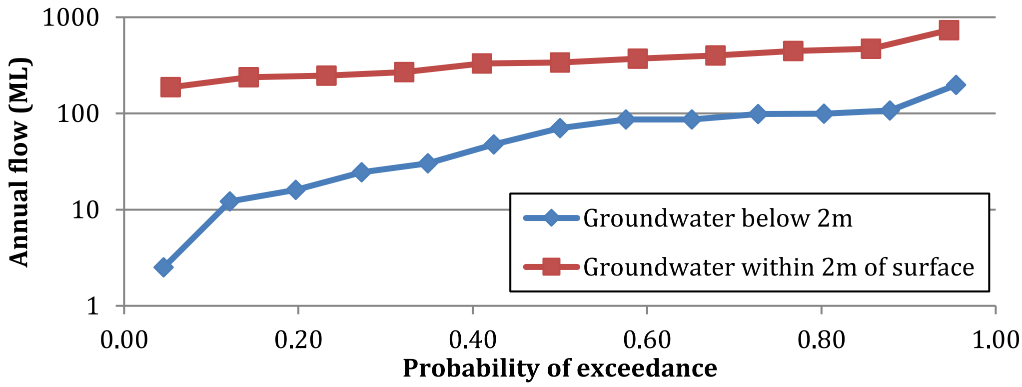

Figure 6.12 Probability distribution for streamflow for Lemons catchment pre and post groundwater discharge

Changing rainfall to runoff relationship

Chapter 4 identified the importance of a groundwater discharge area to increased streamflow generation following forest clearing. Bari *et al.* (1996) also found a similar relationship with increasing water yield following timber harvesting in the southern forests of SWWA. Subsequent studies (Kinal and Stoneman 2012; Petrone *et al.* 2010) have identified the influence of the groundwater discharge area on the declining streamflows observed with reduced rainfall. Kinal and Stoneman (2012) found that for the Yarragil catchment there was not only a change in the streamflow to rainfall relationship, but also a more variable relationship. This is consistent with the data of the water supply catchments (highlighted on page 264).

The importance of the groundwater discharge area on streamflow generation in the high rainfall zone means that the drying climate has led to a greater reduction in streamflow than expected for many catchments. For example the relationship between rainfall and runoff for the Yarragil research catchment has changed significantly since 2002 (see Fig. 6.13). However the changing rainfall to streamflow relationship is not consistent across the jarrah forest. The Salmon catchment, which was gauged from 1974 to 1998, and then from 2009, does not show a change in relationship between rainfall and runoff for these two periods (Fig. 6.14). Part of this can be attributed to the lack of a significant groundwater discharge area contributing to streamflow generation, in this case estimated at only 1% of the catchment area. Therefore, streamflow generation is related to the "wetting up" of the lower slope soil profile rather than generated from a saturated groundwater discharge area. However the 31 Mile Brook catchment (Fig. 6.15) shows a similar change in relationship between rainfall and runoff to that of Yarragil 4X (Fig. 6.9).

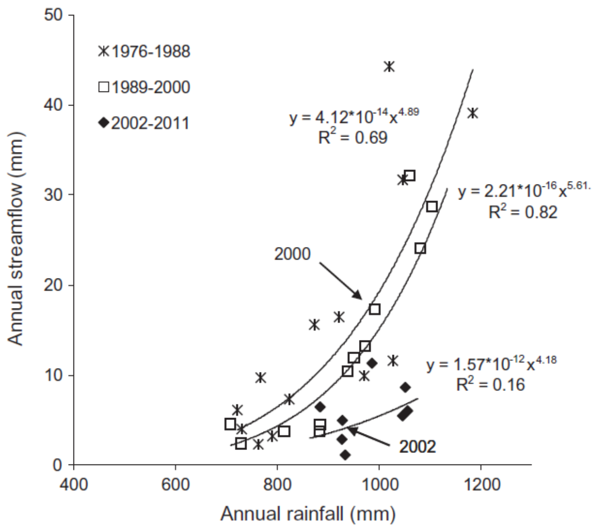

Figure 6.13 Annual streamflow to annual rainfall relationship for Yarragil catchment for 1976 to 1988, 1989 to 2000, and 2002 to 2011 (Kinal and Stoneman, 2012)

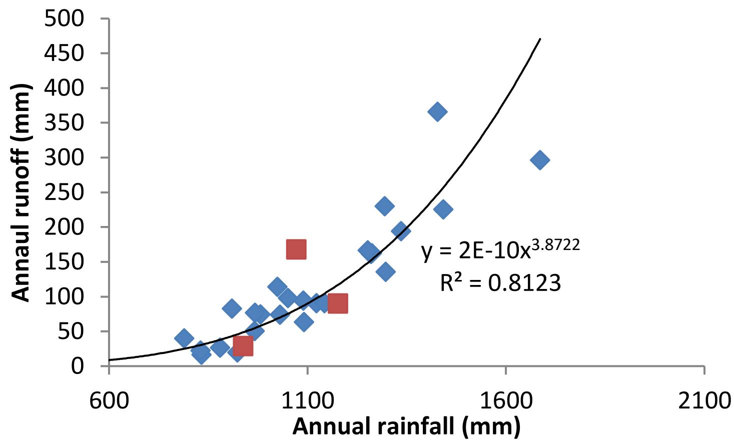

Figure 6.14 Annual streamflow in relation to annual rainfall for Salmon

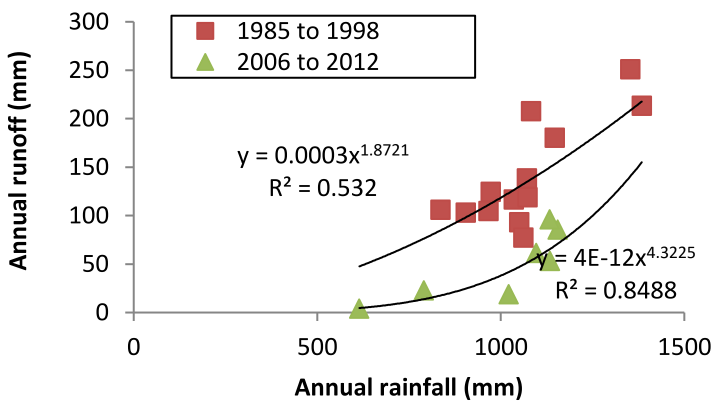

Figure 6.15 Annual streamflow in relation to annual rainfall for 31 Mile Brook with fitted power curves (Croton *et al.* 2013)

The data from the water supply catchments supports the concept of a changing relationship between rainfall and streamflow (Fig. 6.16). The significance of the changing relationship is that for specific annual rainfall the predicted annual streamflow is much lower since 2001. This leads to the question as to whether the relationship will continue to change or whether this represents a new equilibrium relationship

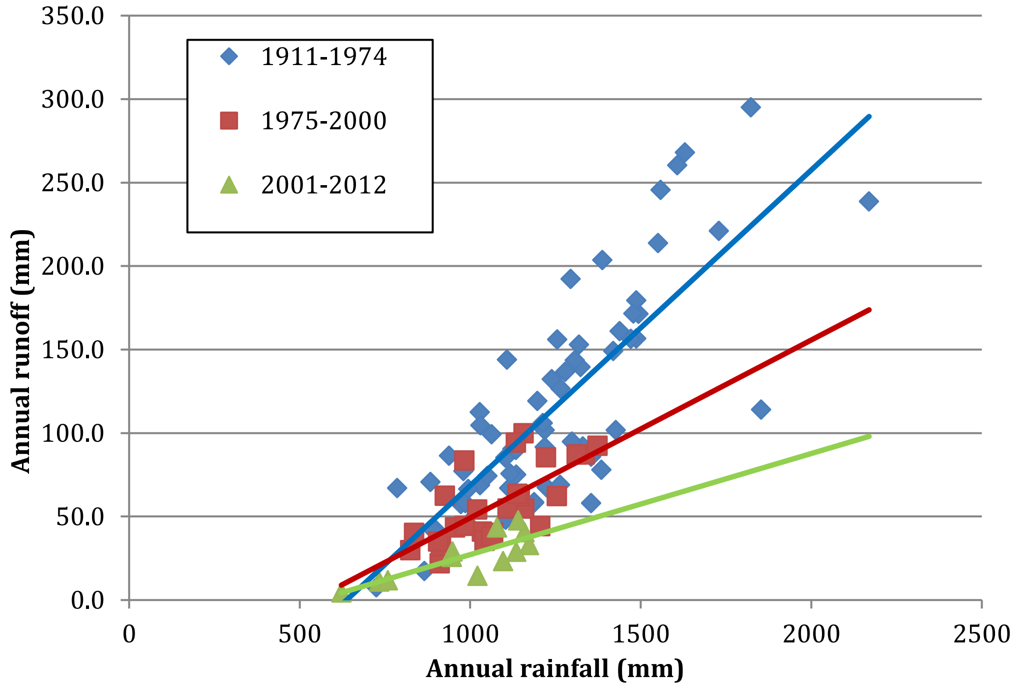

Figure 6.16 Changing rainfall to runoff relationship for Serpentine Catchment

The reduction (based on total volume) in mean annual streamflow from 1975 to 2000 compared to 2000 to 2012 was 51%. However the mean annual rainfall only declined 9%. From the linear regression the reduction is 30% from the change in mean annual rainfall and 70% from the change in the rainfall to runoff relationship. CSIRO (2009) indicate that the median future climate rainfall may decline by another 10%. Based on the existing rainfall to runoff relationship this equates to a further runoff decline of 30%. Petrone *et al.* (2010) considered "rainfall variability superimposed on falling water tables as an important cause of streamflow decline in SWWA observed as a threshold response in a changing climate". In addition the changes in flow duration and monthly flow distribution for forest streams are considered to be related to falling groundwater levels and loss of groundwater-surface water connectivity, contributing to lower annual runoff (Petrone *et al.* 2010). The observed current declines in catchment runoff and Perth water reservoir inflows brings into question the reliance on surface water catchments for future water supply as well as the ecology of jarrah forest streams (Petrone *et al.* 2010).

However the continued change to the rainfall to streamflow relationship since 2001 needs further investigation. Whilst rainfall in higher rainfall areas has continued to decline since 2000 by 6 to 7%, streamflow at the water supply catchment scale has declined by 45 to 55%. One explanation for the decline is that a lag in the groundwater response to the lower rainfall experienced since 1975 delayed the change in the rainfall to streamflow relationship. Alternatively, the changing approach to jarrah forest management (Conservation Commission of Western Australia 2013) over the last 30 to 40 years may be a contributing factor (Burrows *et al.* 2011).

6.1.4 Impact of changing forest hydrology on environment

Water availability within the forest is considered a key driver of vegetation patterns and the types and abundance of fauna. As seen, groundwater monitoring has shown that water tables in the northern jarrah forest have fallen by approximately 0.2 m yr^{-1} over the last 35 years (Croton *et al.* 2012; Kinal and

Stoneman 2012), and streamflow has reduced by more than 50% over the same period. This has resulted in many perennial streams becoming ephemeral, most ephemeral streams becoming even more variable and flowing in less years and projections that more perennial streams will become ephemeral.

6.1.5 Impact of climate change on forest hydrology

The drying climate observed in the southwest has not only reduced streamflows, but has impacted on the forest environment. Although MTF ecosystems such as the jarrah forest are understood to be resilient to drought and other disturbances, Matusick *et al.* (2013) observed a sudden and unparalleled forest collapse in small parts of the jarrah forest corresponding with record dry and heat conditions in 2010/2011.

Climate change scenarios show further reductions in mean annual rainfall in SWWA in the decades to come (Bates *et al.* 2008a). The key findings of the last IOCI report (IOCI 2012) were that the winter (May to July) drying trend in SWWA rainfall has intensified and autumn rainfall has declined 15% since 2000. In addition maximum summer temperature on the west coast of SWWA have increased, as has the hot spell frequency.

Based on large-scale climate modelling, the prognosis is that as greenhouse gas concentrations continue to increase, additional rainfall reductions are likely for SWWA in all months from May to October, with reductions ranging from 8 to 26% compared to the present day (IOCI 2012). IOCI3 projections suggest that hot spells will become more intense over all of SWWA by the end of the century

(2070 to 2099), and that the frequency of hot spells is projected to increase in south-west parts of SWWA (IOCI 2012).

A consequence of the reduction in average annual rainfall, is that the SWWA has experienced and is experiencing more years with well below average rainfall. Cullen and Grierson (2009) in a reconstruction based on tree ring analysis at the eastern edge of the SWWA revealed that rainfall naturally varies from relatively dry periods lasting to 20–30 years to 15-year long periods of above average rainfall. Cullen and Grierson (2009) also consider that although the drivers remain unclear, there is increasing evidence that sustained dry periods are a natural part of the rainfall regime for SWWA.

Previous modelling studies at the large water supply catchment scale found that for a unit change in rainfall there was a threefold change in runoff for unit change in rainfall (Berti *et al.* 2004; Kitsios *et al.* 2009; Silberstein *et al.* 2012; Smith *et al.* 2009). However the data from the water supply catchments (see Table 6.2) indicate that this change may be greater than threefold.

The impact of climate change scenarios on the six[12] major water supply catchments (Silberstein *et al.* 2012) highlight the significant reductions possible with further declining rainfall (Fig. 6.17). In that analysis, the historical period was from 1975 to 2007, the recent period is from 1997 to 2007, and the three future scenarios (wet -2%, median -8% and dry -14% compared to historical rainfall 1975 to 2007) are from rescaling the historical rainfall sequence based on

[12] Mundaring, Wungong, Canning, North Dandalup, South Dandalup, and Serpentine

45 Global Circulation Models (GCMs). This analysis suggests a reduction in mean annual inflow from +7% for the wet scenario to -43% for the dry scenario. However, in the 12 years from 2001, for the same six reservoirs, the inflow has averaged 94.5 GL which represents a reduction of 36% which is close to the dry scenario. This may mean that the climate impacts on inflow to major water supply reservoirs for Perth are under predicted.

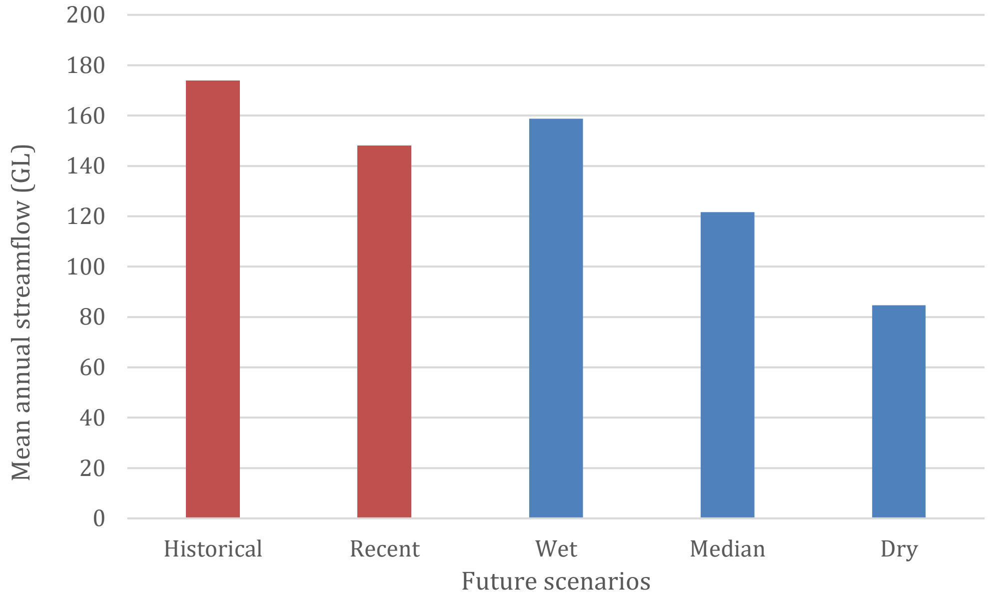

Figure 6.17 Impact of climate change scenarios on water supply catchments (adapted from Silberstein *et al.* 2012)

6.2 Impact of observed climate on stream salinity

For a given salinity to streamflow relationship, a lower streamflow is likely to result in a higher salinity. Salinity is discharged into streams from groundwater (Chapter 1), thus any changes in groundwater will also affect stream salinity. Given the influence of declining rainfall on reducing groundwater discharge area there will be lower rates of saline groundwater discharge. This means that the salinity to streamflow relationship changes resulted in lower salinities overall.

The enduring below average rainfall observed in the jarrah forest over the last 35 years has meant a reduction in stream salinity as the volume of the more saline groundwater reduced. Schofield and Ruprecht (1989) reported on declining stream salinity in the Mt Saddleback catchments, while Kinal and Stoneman (2012) showed a reducing stream salinity from 1976 to 2011 for a forested research catchment (Fig. 6.18). For both studies the reduction in salt load was greater than the reduction in streamflow, therefore the salinity concentration was lower for the period with lower streamflow. The reduced salt load was due to reduced groundwater discharge which carries higher salinity water. The lower groundwater levels reduced the groundwater discharge to the valley floor.

Thus, it is likely that stream salinity will continue to reduce in the forested catchments under the predicted climate change scenarios, albeit with a reduced streamflow. The impacts of this reduction in salinity on stream ecology has not been considered in the literature. Some consequences could be the reestablishment of fresh-water species.

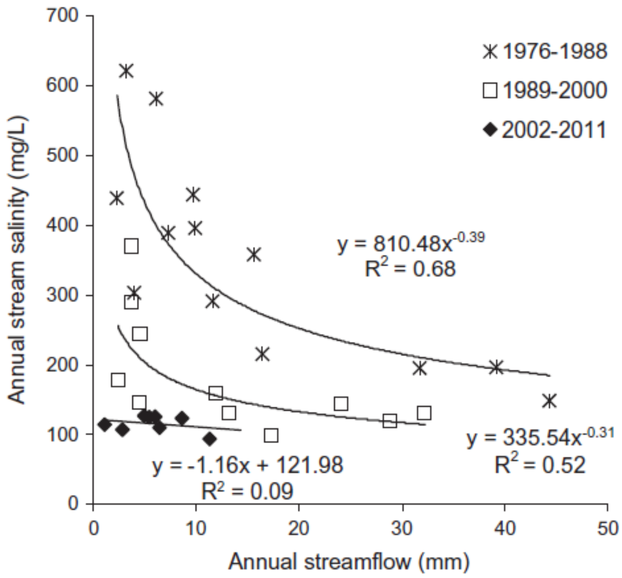

Figure 6.18 Flow-weighted annual stream salinity to annual streamflow relationship for 1976 to 1988, 1989 to 2000, 2002 to 2011 (Kinal and Stoneman 2012)

6.3 Concluding remarks

The drying climate in the SWWA is leading to reduced water yields, reduced groundwater tables, a change in the relationship between rainfall and streamflow and reductions in stream salinity. The lower streamflow for a similar rainfall has been found to be caused by the reduction in groundwater discharge areas in the valley floors with this resulting in lower streamflow generation.

The impact of the drying climate in many cases has led to a greater impact on the hydrology of the jarrah forest than transient forest disturbances outlined in Chapter 5 such as forest thinning, timber harvesting and bauxite mining. However, much of the complex management occurs with the interaction of forest disturbance and drying climate impacts. The next chapter discusses management approaches to mitigate or adapt to the complex interaction of climate and forest disturbances.

Chapter 7 General Discussion

7.1 Introduction

The hillslope chapter (Chapter 3) confirmed the importance of soil water storage, an ephemeral saturated zone, and high infiltration rates in jarrah forest soils to runoff generation. Additionally, the paired catchment chapter (Chapter 4) confirmed the critical importance of groundwater discharge to both runoff generation and salinization. Building on the understanding from the hillslope and paired catchment studies, Chapters 5 and 6 critically examined the impact of forest disturbances and climate on water yield and stream salinity, respectively.

These studies have shown the complex relationship between water, forests and climate. They have also shown the critical requirement of long-term studies. Monitoring in some of the bauxite mining paired catchments commenced in 1966 and in the Collie experimental catchments in 1974. Both the long-term nature of the responses to forest disturbance and the response to changing rainfall means developing an understanding of forest-hydrological responses requires longer term monitoring.

The impact of forest disturbance on forest hydrology in SWWA environment is now much better understood because of this research and in particular the link between streamflow generation and groundwater discharge areas. Despite the critical nature of the understanding from hillslope and paired catchment studies, management responses are needed in shorter time frames. The following sections outline the management response to several forest-hydrological management issues: clearing for agriculture, mining, and timber harvesting and forest regeneration. This is followed by a critical evaluation of the evolving approach to forest management for water values and future research needs.

7.2 Forest disturbance and management responses

7.2.1 Forest disturbance - agriculture

The paired catchment studies (Chapter 4) demonstrated that stream salinity increases occur when native vegetation is cleared for agriculture, resulting in reduced evapotranspiration and the resultant increase in groundwater recharge and watertable rise which mobilises salts stored in the deep soil profile.

Prior to 1960s there was a period of substantial land clearing in this region with observed increases in stream salinity. In addition, the increasing population was resulting in increased water demand. In response the Western Australian State Government legislated to control land alienation (release of Crown land to private ownership) on water catchments in higher rainfall areas across SWWA (Schofield *et al.* 1988). This response was complemented in the late 1970s with clearing control legislation (*Country Areas Water Supply Act Amendment Act* 1976) in those surface water catchments with substantial potential water resources that either had currently marginal water quality or were at risk of increasing stream salinity (Helena, Collie, Warren, Kent and Denmark). In each case it was considered that water quality could be improved through land-use interventions (Government of Western Australia 1996).

Even with these early interventions, stream salinities continued to increase. In response a reforestation program was commenced in 1980 in the Collie catchment and to-date 6740 ha of land has been reforested by the Government. In addition, 9500 ha has been reforested by private companies. The potential effectiveness of reforestation strategies to control stream salinity in the south-west was initially summarised by Schofield *et al.* (1989a) with many subsequent reviews of actual performance (DEWHA 2009; Hatton 2001; Hatton *et al.* 2003). While it is clear from these reviews, and indeed from the results presented in Chapter 4, that there is a link between the removal of forest cover and catchment hydrology, arguments continued around the scale of reforestation required to reverse stream salinity (George *et al.* 1999; Hatton 2001). In particular, it had been posited that catchment hydrology could be restored with partial reforestation (Mauger *et al.* 2001b) of catchments (Bartle and Shea 1989; Shea and Bartle 1988), whereas subsequent experience suggested that was not the case and more extensive reforestation would be required (George *et al.* 1999). Assessment through catchment modelling: MAGIC (Mauger 1996); WEC-C (Bari and Croton 2002); and LUCICAT (Bari and Smettem 2006) has confirmed that extensive reforestation is needed to restore stream salinity to pre-clearing levels. This modelling was based in part on the work presented in Chapters 4 and 5.

Specific studies have reviewed the hydrology and potential salinity recovery options for the Helena, Collie, Warren, Kent and Denmark river catchments (Bari *et al.* 2004; De Silva *et al.* 2006; Mauger *et al.* 2001a; Smith *et al.* 2006; Smith *et al.* 2007).

The Western Australian Salinity Action Plan (Government of Western Australia 1996) set targets to achieve potable salinity levels for the Helena, Collie, Warren, Kent and Denmark rivers. The aim of the program was to work with the farming communities to implement recovery programs in the catchments of the Collie, Warren, Kent and Denmark rivers. The Salinity Action Plan also defined a series of catchments where biodiversity and rural towns were at risk of salinization and proscribed catchment assessments for rural areas. A key component of the recovery actions has been to encourage privately funded reforestation with this predominantly comprised of pulpwood eucalypts (Harper *et al.* 2009a).

The Denmark River catchment is particularly illuminating as an example of this program. Here the trend of previously increasing salinizing stream salinity has been reversed (Ruprecht *et al.* 2014). This catchment has an area of 525 km^2 to the Mt Lindesay gauging station, with mean annual rainfall ranging from 650 mm in the upper catchment to 1100 mm by the coast (1975– 2003 average). Clearing native vegetation for agriculture in the upper part of the Denmark River catchment resulted in stream salinity exceeding potable levels (500 mg L^{-1} TDS) and peaking at an annual flow-weighted salinity of 1500 mg L^{-1} TDS in 1997 at the Mt Lindesay gauging station. Clearing began in 1870 and continued at a steady rate until it rapidly expanded after World War II when heavy machinery became more widely available. The native, deep-rooted perennial vegetation was replaced by annual shallow-rooted pasture and crops changing the water balance. The lower evapotranspiration rate of the new vegetation and the consequent increased infiltration of rainfall to groundwater stores resulted in higher groundwater levels, saline valley floors, and increased saline discharge into rivers and streams.

The clearing of land peaked at 19% of the catchment (to the Mt Lindesay GS). Reforestation in the catchment began in the early 1990s with the Integrated Catchment Management – Upper Catchment Project promoted by the then Water Authority and the Department of Agriculture (Ferdowsian and Greenham 1991). This project helped farmers prepare farm plans identifying areas suitable for reforestation and constructing fences and drains, using the prevailing assumption that catchment recovery could be achieved with partial reforestation. The Water Authority supplied investment capital and the Department of Conservation and Land Management's Timberbelt Sharefarming Scheme acted as a vehicle for managing the plantations (Bartle 1991; Schofield *et al.* 1989a). Some farmers also used their own capital to plant additional trees. By 2010 around 5200 ha of *Eucalyptus globulus* plantations had been established, leaving 7% of the catchment cleared.

Annual flow weighted stream salinity by 2010 was below 400 mg L^{-1} TDS, this thus representing one of the only examples of landscape scale catchment recovery. The Denmark River is now a viable water supply for the south coast of Western Australia, given the main alternative was the development of a sea-water desalination plant at a conceptual cost of $70 to 100 million (Ward *et al.* 2011).

In 2000 a State Salinity Strategy (State Salinity Council 2000) reinforced the water resource recovery catchment approach. However, since 2010 there has been a reduction in on-ground activity within the water resource recovery catchments. This is considered due to the magnitude of the intervention required identified in the salinity situation statements and the change of focus to climate independent water supplies (e.g. desalination) for the SWWA (Water Corporation 2011).

Additionally, new private plantation investment has effectively stopped, particularly due to the collapse in the managed investment scheme model (Harper *et al.* 2017). The plantations within the Denmark catchment on their own are currently marginally profitable due to the low world woodchip prices, but they are considered to have an economic water benefit of up to \$7000 ha^{-1} (URS 2002). Removal of the plantations will result in a future deterioration of water quality and a key issue will be valuing the reforestation in terms of its ongoing water management role.

Although the work in the Denmark River and other salinity recovery catchments had a range of actors, its scientific basis was substantially underpinned by the paired catchment studies into forest clearing for agriculture and of reforestation (Chapters 4 and 5). In short, the work reported in this thesis has assisted in understanding the causes of salinity and in developing effective salinity mitigation or management options.

7.2.2 Forest disturbance - mining

Alcoa Australia operates the Huntly and Willowdale mines in the high to intermediate rainfall jarrah forest, and BHP operates the Boddington mine in the low rainfall jarrah forest. Originally research was focused on the intermediate rainfall zone of Alcoa's mining lease due to concerns about increasing stream salinity in the water supply reservoirs for Perth (Loh *et al.* 1984). As part of the environmental review for the Wagerup Alumina Project, a commitment by Government and industry was made that mining would not take place in the eastern, lower rainfall portion of the Alcoa lease until research showed that mining could be undertaken without significantly increasing the salinity of the water resources (Environmental Protection Authority 1978). Over time, the mining rehabilitation prescription has changed leading to a mosaic of rehabilitation types with associated complex management (Koch and Hobbs 2007).

Alcoa's bauxite mining is overseen by the Mining and Management Program Liaison Group (MMPLG) and is chaired by the Department of State Development. As of June 2015 the Bauxite Hydrology Committee (BHC) is a sub-committee of the MMPLG and provides advice on the influence of mining operations on the implications of salinity and water yield. The BHC also oversees a research program into the impact of mining of the intermediate rainfall zone on salinity (the Intermediate Rainfall Zone Research Programme) (Croton and Reed 2007) that incorporates hillslope studies, process-based catchment modelling, paired catchments, trial mining, and water resource catchment studies. The trial mining research transitioned into phased mining as research results suggested that the risk to salinization of water supply reservoirs was low (Croton *et al.* 2013).

As the forest areas experienced a declining rainfall the impact of mining on water yield became a more significant issue and the scope of the research into bauxite mining expanded to focus on water yield as much as salinity (Croton and Reed 2007; Mauger *et al.* 1998). The paired catchment studies of Del Park, Seldom Seen, and More Seldom Seen have shown the decline in water yield to below pre-mining or control catchment levels (Croton 2004; Croton *et al.* 2005; Ruprecht 1991). The research programme has expanded more recently to consider hydrologically sensitive rehabilitation that includes vegetation density, fire management, soil nutrient levels, rehabilitation topography, and soil structure.

7.2.3 Forest disturbance - timber harvesting and regeneration

The Conservation Commission, in which the State forests and timber are vested in, produced the Forest Management Plan 2014-2023 (FMP) which is a management plan for the State forest and timber reserves (Conservation Commission of Western Australia 2013). A previous plan (The Forest Management Plan 1994-2003) (Conservation Commission of Western Australia 2004) revised the informal reserve system on road, river and streams and reduced maximum coupe size to minimise the transient stream salinity impacts (Borg *et al.* 1987).

The current FMP adopts the Montreal Criteria as the framework (Conservation Commission of Western Australia 2013) within which to set the goals and management activities. One of the Montreal Criteria is the conservation and maintenance of soil and water. The FMP outlines operations and key performance indicators for soil and water. One of the performance targets for water is no decline in groundwater level as a result of management activities. Given the significant declines observed by Kinal and Stoneman (2012) and Croton *et al.* (2013) in groundwater levels in forested catchments it may be difficult to achieve this performance target. Moreover further work will be required to distinguish climate impacts from management impacts.

Streamflow data of logging within the karri forests in the SWWA indicate that annual flow weighted salinities have remained below 200 mg L^{-1} TDS across all rainfall zones (HRZ, IRZ, IRZ) except where no stream buffer was in place (Bari and Boyd 1993; Borg *et al.* 1988; Western Australian Steering Committee for Research on Land Use and Water Supply 1987). Clearfelling of karri and heavy selection cutting of jarrah forests, where appropriate buffers are retained, has had only minor impacts on salinity and sediment loads (Bari and Boyd 1993) as suggested by earlier studies (Johnston *et al.* 1980). For example, mean annual sediment concentrations were observed to be higher during logging and for two to three years after harvesting and then return to pre-treatment levels (Borg *et al.* 1987). The greatest sediment loads were related with logging during wet periods and where stream buffers were absent. Borg *et al.* (1987) found that there was no information to indicate that harvesting of jarrah and karri led to persistent or significant increases in sediment loads and salinity, provided effective regeneration occurs soon after logging is completed. The important management approaches to reduce sediment loads and maintain low stream salinity levels include the retention of vegetation buffers, particularly along stream zones, and appropriately designed roads (Borg *et al.* 1987).

7.3 Impact of climate on water availability

The Water Corporation in their Water Forever strategic planning (Water Corporation 2011) are now predicting no contribution from forested catchments to the Perth water supply in ten years. This contrasts with over 70% supplied from forested catchments in the 1970s and a current (2006 to 2012) mean annual inflow of 100 GL. The rapid transition to water supplies being derived from forested catchments and shallow groundwater to more climate independent sources such as sea water desalination and deep groundwater reduces the uncertainty and complexity of resource management, but comes at a substantially increased financial cost.

In contrast to Perth, and despite the extended drought in the late 2000s, forested catchments still provide 77% of urban water supplies in the other capital cities of Australia (Fig. 7.1). Perth accounts for 88% of the groundwater supplied and 83% of the desalination provided to major urban areas in 2014 to 2015.

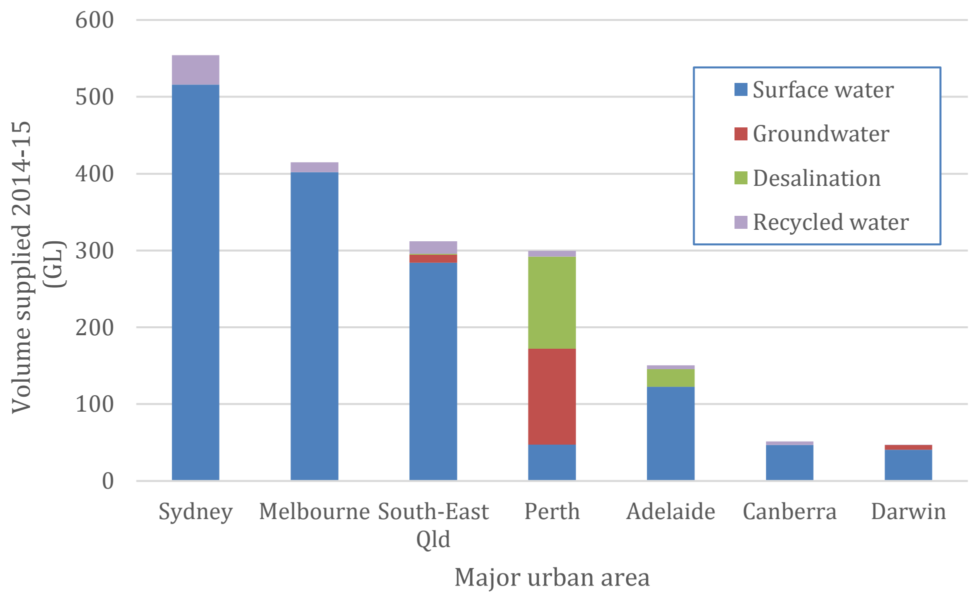

Figure 7.1 Water sources for major urban areas (from Bureau of Meteorology (2016)

Webb (2012) in a study, based in south eastern New South Wales, into co-developing timber harvesting and water resources, concluded that timber harvesting, if spatially and temporally constrained, can occur without compromising water supplies and may assist with improving forest heterogeneity and resilience. Ellison and Daily in Oregon State University (2008) argue that the greatest value of forests is sustainable water supply given the drying climate and threats to existing water supplies.

Forest thinning to improve water yield has the potential to increase streamflow from water supply catchments for Perth by more than 50% (Fig. 7.2) (Reed *et al.* 2012). The treatable area of 65 000 ha would be distributed across 8 key water supply catchments. Individual water supply catchment streamflow increases were estimated by multiplying the area of the catchment suitable for silviculture for water production by the estimated streamflow increase for each rainfall zone.

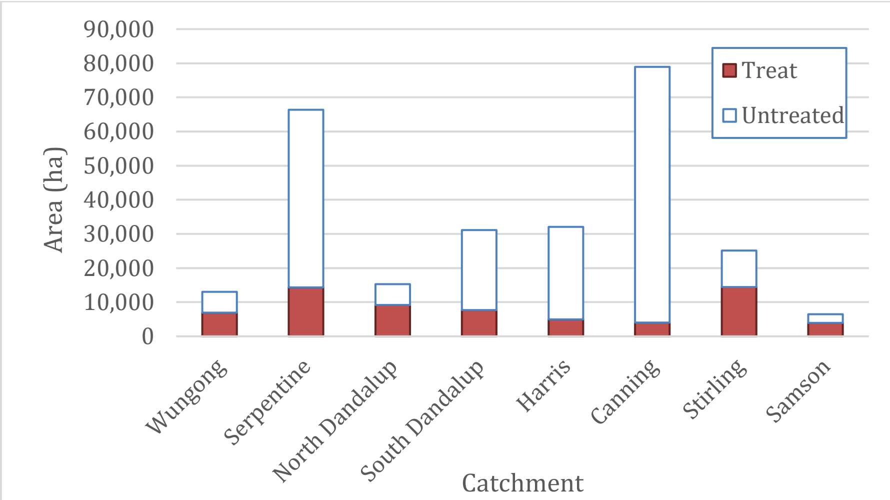

Figure 7.2 Potential treatable area for catchment thinning from surface water supply catchments for Perth, Western Australia (adapted from Reed et al. 2012)

Again, this analysis was based on the scientific understanding developed in Chapter 5 and associated studies (Stoneman and Schofield 1989).

However, both forests and climate in Western Australia and Australia are changing rapidly and additional research will be required to find new approaches to maintain and improve forested water supplies. This approach is clearly articulated in other regions. For example, McLaughlin *et al.* (2013) highlighted the potential of regional water yield benefits from managing forests in southeastern USA. The authors evaluated a seven-year cycle of forest thinning to basal areas of 8 to 10 m^2 ha^{-1}. McLaughlin *et al.* (2013) argue that forest management approaches that increase and improve water availability will be critical to maintaining water resource values and protecting natural systems.

Similarly, Bales *et al.* (2011) estimated from modelling that reducing forest cover by 40% of maximum levels across a catchment in the Sierra Nevada (California, USA) could increase water yields by about 9%. Bales *et al*. (2011) also contended that forest management for water yield provides other benefits such as mitigating wildfire impacts and consequently reduced sediment loads. These characteristics are similar to what was recommended by Reed *et al.* (2012) in a review of the Wungong Catchment trial in SWWA.

Croton *et al.* (2014) identified areas within the Northern Jarrah Forest that would require a Leaf Area Index to ensure a mean annual streamflow of 100 mm. If this level of streamflow was achieved groundwater would be kept close to the surface and this in turn would lead to environmental benefits to the jarrah forest streamzones.

Croton *et al.* (2014) also confirmed the work of Smettem *et al.* (2013) that there was a buffered response of LAI to changes in annual rainfall. This buffered response was attributed to the availability of deep soil water and/or groundwater. However, as seen groundwater levels are continually reducing within the regolith (Kinal and Stoneman 2012) and will eventually reach a situation where this will not exist and the vegetation responses will no longer be buffered.

7.4 Impact of climate on forest ecology

Allen *et al.* (2010) reported 88 worldwide examples of forest damage occurring from drought and/or heat related events, demonstrating the susceptibility of forests to extreme climate conditions. The studies range from Scots pine in the Swiss

Alps to tree mortality and defoliation in Spain, to tropical rainforests. Choat *et al.* (2012), in evaluating the safety margins of the water-transporting xylem tissue across 226 forest species, concluded that there is a global convergence in the vulnerability of forests to drought.

Matusick *et al.* (2013) observed a "sudden and unprecedented forest collapse" (74% mortality from 236 independent forest patches) in the Northern Jarrah Forest corresponding with the record dry and heat conditions in 2010/2011. Figure 6.3 highlighted the magnitude of the reduced rainfall in 2010. In total, tree deaths represented 1.5% of forest in the surveyed area. The long-term consequences of large-scale drought induced mortality events are poorly understood in the northern jarrah forest and in other MTFs (Matusick *et al.* 2013), and in particular whether these losses are a harbinger of future more extensive mortality.

The impacts in the Northern Jarrah Forest reported by Matusick *et al.* (2013) indicate that MTFs may significantly change through drought-driven processes in the coming decades with a changing climate. They also suggest that MTFs, once thought to be resilient to climate change are susceptible to sudden and severe forest collapse when key thresholds are reached. Laurance *et al.* (2011) discussed ecosystems reaching tipping points and it may be these forest ecosystems, such as the MTFs, may be reaching a tipping point.

As groundwater disconnects from the valley floor, streamflow is reduced and this causes many once perennial streams to become ephemeral and ephemeral streams to become dry. Wetter parts of the landscape including winter wetlands, peat swamps, riparian zones and other wetter habitats embedded in the forest complex are drying. This is considered to be having acute impacts on the stream and riparian zone biota as well as predisposing them to damaging bushfires (Burrows *et al.* 2011).

Grant *et al.* (2013) argue that management approaches such as forest thinning, focusing on planting drought-tolerant species, and irrigating and harvesting more water for transpiration in forest landscapes may need to be applied. The concept of ecological forest thinning, in response to extreme climate conditions has been assessed globally with examples from Victoria's Box—Ironbark forests (Pigott *et al.* 2010), Norway's spruce forests (Kohler *et al.* 2010), and southwestern USA forests (Robles *et al.* 2014).

Verschuyl *et al.* (2011) in a summary of 33 forest thinning treatments and forest ecological studies undertaken in North America found that forest thinning treatments had mostly positive or neutral effects on diversity and abundance across all taxa. D'Amato *et al.* (2013), from a long-term replicated thinning experiment, confirmed the potential of forest thinning (density management) to mitigate some drought impacts.

Martínez-Vilalta *et al.* (2012) argue that the historical changes to forest management and land use need to be accounted for to understand the timing and distribution of current forest mortality. In Western Europe tree mortality appears to be associated with densification of forests owing to extensive ceasing of intensive forest management (Vila-Carbrera *et al.* 2011).

The concept of ecological thinning (Verschuyl *et al.* 2011) is to hasten the development of older growth conditions of ecosystem function, forest structure and habitat diversity, rather than for timber production or for water supply. This ecological thinning can assist the resilience of forests to climate change or provide improved forest ecosystem.

The current forest management plan (Conservation Commission of Western Australia 2013) provides for 'silviculture for ecosystem health' and 'silviculture for water production'. These approaches are focused on water values and would benefit from considering broader ecological health of the overall forest ecosystem.

7.5 Managing forests for water values in a climate of uncertainty

Forests of SWWA are managed according to the principles of ecological sustainable forest management (Conservation Commission of Western Australia 2013) and for a diverse range of values including: nature conservation, tourism and recreation, water catchment protection, and timber production.

For many years forest managers have relied on the paradigm of ecological sustainability and ecological sustainable forest management to set goals and establish management plans (Millar *et al.* 2007). This concept uses historical forest conditions as a means of gaining information about how forests should be managed into the future.

Historical data have been of considerable worth in developing an understanding of ecosystem responses to land use and environmental changes and in setting management objectives. However, many forest managers also use the range of historical ecosystem conditions as a management target, assuming that by restoring and maintaining historical conditions they are maximising ecosystem sustainability into the future (Millar *et al.* 2007). This approach is often taken even as ongoing climate change pushes global and regional climates beyond the bounds of the last several centuries (IPCC 2007; 2014; Millar *et al.* 2007).

Some of the world's forested ecosystems may already be responding to climate change and becoming increasingly vulnerable to higher background tree mortality rates and die-off in response to future warming (Allen *et al.* 2010). SWWA was not represented in this study, but subsequently Matusick *et al.* (2013) have reported on tree mortality in the jarrah forest of SWWA. The reality is that climate variability – both naturally caused and anthropogenic from a range of land-use practices create new ecological conditions never before experienced by the forest environments.

Table 7.1 Adaptive strategies for forest ecosystems (after Millar *et al.* 2007)

Option	**Strategies**
Resistance	Anticipate impacts and protect highly valued areas
Resilience	Improve capacity of ecosystem to return to desired conditions after disturbance
Response	Assist the transition of ecosystems from current to new condition

Millar *et al.* (2007) suggested adaptive strategies (outlined in Table 7.1) to assist forest ecosystems to accommodate change. Developing and testing various tactics for reducing forest vulnerability to drought stress is a clear need for the SWWA forests. Adaptation in forestry includes a climate change focus (Spittlehouse and Stewart 2003) and includes:

- Create objectives for the forest under climate change.
- Awareness and training within the forestry community about adaptation to climate change.
- Assess the vulnerability of forest ecosystems, forest communities, and society.
- Develop effective approaches to adaptation.
- Manage the forest to reduce vulnerability and enhance recovery.
- Monitor to determine the state of the forest and identify when critical thresholds are reached.
- Respond to the impact when it occurs, speed recovery, and reduce vulnerability to further climate change.

Given the decline in streamflows and groundwater it could be argued that critical thresholds or tipping points have already been reached for the forests of SWWA. There are a range of possible strategies available to achieve forest ecosystem or water supply objectives (Table 7.2).

The current water management approaches for the SWWA forests were developed for the climatic conditions prevailing in the 1960s and 1970s. The changing climate, the impacts this is having on streamflow and forest health, and the reduced importance on the northern forests to produce water for urban consumption mean that there needs to be a rebalancing of water management aims and approaches.

Table 7.2 Possible strategies to meet forest ecosystem and water supply objectives

Component	Strategy
Forest ecosystem	Forest management – reducing forest density
	Prescribed burning to reduce understorey water use
	Selection of tree-species for a more climate resilient forest
	Matching plant water demand to plant water availability
	Reducing soil evaporation
	Hillslope water capture and re-distribution
Water supply	Silvicultural management to lower density
	Suppressing understorey

To meet the concerns of resource managers of forests and water, forest hydrology research needs to move from principles to prediction (National Research Council 2008). This has been partially achieved in the Western Australian context, with the development of forest hydrological models (e.g. MAGIC, WEC-C and LUCICAT), as described in Section 7.2.1, however these models are not used to guide forest management. Predictions are needed to understand indirect and interacting hydrologic responses to changes in forest landscape associated with climate change, forest disturbance, forest management and mining. How these changes will affect water quantity and quality over extended spatial and temporal scales is critical.

A landscape perspective on forest hydrology links principles from processes and paired catchment experiments with hydrologic responses at larger spatial scales that are changing over longer time scales. This landscape perspective needs to not only understand the current hydrology but be able to predict future hydrology given the current forest structure and the range of climate scenarios.

7.6 ***Future research needs***

Research into forest disturbance and a drying climate has led to a greater understanding of the hydrology of the SWWA forests. However given the observed collapse in 2010 to 2011 in parts of the jarrah forest, the declining water yields from water supply catchments, the change from perennial flow to ephemeral streams, and continual reduction in groundwater levels across the jarrah forest, it is clear ongoing research is needed to assist improved management.

The major forest water issues that have been identified in this thesis are the declining water values in forested areas, such as less water volumes, shorter flow periods, and declining groundwater levels. Likely drivers of these changes include:

- A drying climate with direct and indirect impacts on the overall water balance. Direct impacts are through reduced rainfall, indirect impacts through changes in the survival and growth in vegetation.

- Responses to historical forest management, such as forest harvest and thinning.
- Long term impacts of bauxite mining and subsequent forest rehabilitation. In particular the relationship between declining water values and increased vegetation density, higher nutrient levels, and landscape changes.
- Quantifying the hydrological impacts of reforestation.
- Interaction of the range of forest disturbances at a catchment scale.

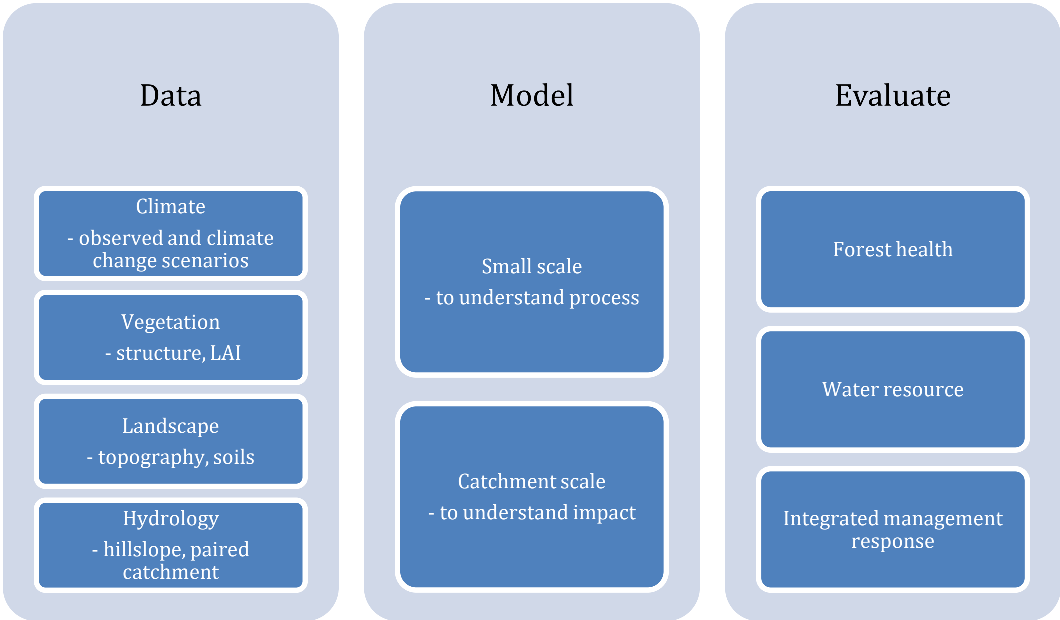

Figure 7.3 Understanding required for forest disturbance hydrology

Research is needed across data, modelling and predictions (Fig. 7.3) to guide forest management given the issues facing the forests of SWWA. With respect to data, the historical long-term paired catchment studies have provided invaluable insight into SWWA forest hydrology and need to be an integral part of a research programme into the future.

The main research needs can be categorised as:

- Understanding the impact of climate variability and change,
- Understanding forest disturbance at a landscape scale,
- Understanding the cumulative and long-term impacts of forest disturbance and climate change,
- Developing management approaches to develop resilient forests under conditions of climate uncertainty.

7.6.1 Understanding the impact of climate variability and change

The forest collapse observed in the SWWA and worldwide means that understanding the relationship between climate and hydrology still remains critical. In addition, although the water supply for Perth and environs is planned to be predominantly from manufactured water (desalination) and deep groundwater into the future, the agricultural sector uses irrigation water from dams across the SWWA.

Specific research needs include how the hydrology and water processes change in response to the change in rainfall and temperature characteristics, and indeed the effects of changes in forest cover as a result of climate change on forest hydrology. It is not only the volume of rainfall, but characteristics such as extended periods of below average rainfall at multi-year scales and within season distribution which are critical for both water supply and forest ecology. Paired catchments will assist with calibrating conceptual frameworks and models that are necessary to understand the impacts of future climates.

7.6.2 Understanding forest disturbance at a landscape scale

The paired catchment studies across the SWWA forests have shown the initial and longer-term impacts of forest disturbance. However, the impact across larger scales, such as water supply catchments or ecological zones, has not been clearly shown. Further research at larger scales is critical to understand not only the implications of both forest disturbance and changing climate, but also the impact of treatments to prevent or restore forest values.

The interaction of forest disturbance and a drying climate is difficult to distinguish. For example the Denmark River with a catchment area of 505 km^2 to the Mount Lindesay gauging station has experienced significant afforestation over the last twenty years, with 25 km^2 planted. These plantings are considered to have significantly reduced stream salinity by lowering groundwater levels. At the same time mean annual runoff has reduced from 63 to 34 mm, which is similar to what has been observed in the northern jarrah forest just with drying climate. It is difficult to distinguish how much of the reduced runoff is due to afforestation and how much is due to the drying climate. Catchment modelling is critical to understanding the causes of reduced streamflow but it requires long term hydrologic data to calibrate and validate.

7.6.3 Understanding the cumulative and long-term impacts of forest disturbance and climate change

The long-term water yield implications from mining forest and its rehabilitation needs further research. As described in this thesis, the impact of changing landscape, soil profile, nutrient levels, and tree density may all have an impact on the water balance. However, it is not clear which aspect is more dominant and what response needs to occur to develop a sustainable and resilient water balance from rehabilitated catchments. Due to the loss of 4 to 6 m of the soil profile, rehabilitated forest areas from mining are potentially less resilient than jarrah forest, which is already showing signs of mortality from drought conditions.

This uncertainty with respect to the impact of long-term drying climate on undisturbed and disturbed forest environments requires research at a range of scales – hillslope, research catchment, and water resource catchment scale. The research should involve process and catchment understanding to assist modelling to review a range of scenarios. The benefit will be improved management protocols for environmental and water supply outcomes.

7.6.4 Management approaches to develop resilient forests given climate uncertainty

Forests have unique characteristics of stability and longevity, but their structure and function can be altered by management or natural disturbance. Depending on the intensity of the management or disturbance, the structural and functional changes can be temporary or long term.

The need for further research into forest thinning relates to both the environmental concerns of further and widespread forest collapse in the jarrah and karri forests of SWWA and the potential water supply benefits. The main gaps in understanding that would benefit from research include examining innovative thinning treatments that could increase water yield and develop a more resilient forest ecosystem. In particular, investigation is needed of the benefits and risks of treating stream zones. The stream zones are areas of greater risk to a drying climate and also the areas which are a primary source of streamflow generation.

The need to understand mitigation options and their benefit and impact on forest hydrology and ecology is important. Research into mitigation is now urgent given the long term monitoring required to understand the response to any action in SWWA forests.

7.7 In conclusion

Research into the hydrologic impact of a changing climate and with forest disturbance has led to improved understanding and to improved forest and water management. The research has included developing a process understanding (Chapter 3) and paired catchment studies (Chapter 4). This research has given a solid foundation of general principles that describe how water is connected to and travels through forests and how hydrologic processes respond to forest disturbances. This research has occurred over an extensive period of time, with some of the studies commencing over 40 years ago.

The challenge for the next forty years is for forest hydrology research to influence current and future forest management to improve environmental and water supply outcomes for the forests of not only SWWA, but globally.

Understanding the causes of the drying up of forest rivers and streams in the south-west is a key challenge for forest hydrology research. But the challenge is to not only understand the causes, but to provide viable options to improve management and ultimately forest outcomes, including for water supply and water environments.

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