

**Modelling
Climate – Surface Hydrology
Interactions in Data Sparse Areas**

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The IHACRES rainfall-runoff model, which formed the basis of the rainfall-runoff model developed within, was originally developed by Prof. Anthony Jakeman. This work has not previously been submitted for a degree or diploma in any institution of higher education.

Some of the work presented in Chapter 4 has previously been published in Evans and Jakeman (1997, 1998); some of the work from Chapter 8 has been published in Evans et al. (1999); and some of the work from Chapter 9 has been published in Evans et al. (2000). Any contributions made by co-authors are indicated in the text.

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Abstract

The interaction between climate and land-surface hydrology is extremely important in relation to long term water resource planning. This is especially so in the presence of global warming and massive land use change, issues which seem likely to have a disproportionate impact on developing countries. This thesis develops tools aimed at the study and prediction of climate effects on land-surface hydrology (in particular streamflow), which require a minimum amount of site specific data. This minimum data requirement allows studies to be performed in areas that are data sparse, such as the developing world.

A simple lumped dynamics-encapsulating conceptual rainfall-runoff model, which explicitly calculates the evaporative feedback to the atmosphere, was developed. It uses the linear streamflow routing module of the rainfall-runoff model IHACRES, with a new non-linear loss module based on the Catchment Moisture Deficit accounting scheme, and is referred to as CMD-IHACRES. In this model, evaporation can be calculated using a number of techniques depending on the data available, as a minimum, one to two years of precipitation, temperature and streamflow data are required. The model was tested on catchments covering a large range of hydroclimatologies and shown to estimate streamflow well. When tested against evaporation data the simplest technique was found to capture the medium to long term average well but had difficulty reproducing the short-term variations.

A comparison of the performance of three limited area climate models (MM5/BATS, MM5/SHEELS and RegCM2) was conducted in order to quantify their ability to reproduce near surface variables. Components of the energy and water balance over the land surface display considerable variation among the models, with no model performing consistently better than the other two. However, several conclusions can be made. The MM5 longwave radiation scheme performed worse than the scheme implemented in RegCM2. Estimates of runoff displayed the largest variations and differed from observations by as much as 100%. The climate models exhibited greater variance than the observations for almost all the energy and water related fluxes investigated.

An investigation into improving these streamflow predictions by utilizing CMD-IHACRES was conducted. Using CMD-IHACRES in an “offline” mode greatly improved the streamflow estimates while the simplest evaporation technique reproduced the evaporative time series to an accuracy comparable to that obtained from the limited area models alone. The ability to conduct a climate change impact study using CMD-IHACRES and a stochastic weather generator is also demonstrated. These results warrant further investigation into incorporating the rainfall-runoff model CMD-IHACRES into a limited area climate model in a fully coupled “online” approach.

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List of Symbols

<i>Symbol</i>	<i>Meaning</i>	Units
$\sim$	denotes an environmental value	-
α	represents an arbitrary variable; also absorptivity	-
α_c	cloud water radiation absorption coefficient	m^2g^{-1}
α_{cd}	cloud water downward radiation absorption coefficient	m^2g^{-1}
α_{cu}	cloud water upward radiation absorption coefficient	m^2g^{-1}
α_i	cloud ice radiation absorption coefficient	m^2g^{-1}
α_p	radiation absorption coefficient for precipitation	m^2g^{-1}
β	fraction of updraft condensate that re-evaporates in the downdraft	-
$\delta\theta$	soil moisture	mm
$\Delta = \frac{de^*}{dT}$	slope of the saturation water vapour pressure curve	mbK^{-1}
Δs	model gridpoint spacing	m
Δt	model time step	s
ϵ_d	downward emissivity function	-
ϵ_{tot}	total emissivity	-
ϵ_u	upward emissivity function	-
ϵ_{vapour}	water vapour emissivity	-
ϕ_0	soil water suction for saturated soil	m
ϕ_m	wind profile function	-
γ	psychrometric constant	mbK^{-1}
γ_ξ	correction to the local gradient that incorporates the contribution of large scale eddies to the total flux	-

η	fraction of transpiration from the top soil layer	-
μ	cosine of the zenith angle	-
$\tilde{\nu}$	wavenumber	m^{-1}
ν_a	annual frequency	s^{-1}
ν_d	diurnal frequency	s^{-1}
θ	potential temperature	K
θ_s	appropriate near surface temperature	K
θ_v	virtual potential temperature	K
θ_{va}	virtual potential temperature at the lowest model level	K
ρ	density	kgm^{-3}
ρ_a	density of surface air	kgm^{-3}
ρ_r	particle density	kgm^{-3}
ρ_s	density of the subsurface soil layer	kgkg^{-1}
ρ_w	soil water density	kgkg^{-1}
ρ_{wsat}	saturated soil water density	kgkg^{-1}
σ	terrain following vertical coordinate	
σ_f	fractional foliage cover for each grid point	-
σ_{SB}	Stefan-Boltzmann constant	$\text{Wm}^{-2}\text{K}^{-4}$
τ_c	cloud extinction optical depth	m
τ_q	quickflow time constant	days
τ_s	slowflow time constant	days
ω	particle single scattering albedo	
Ω_{r0}	maximum transpiration that can be sustained	$\text{kgm}^{-2}\text{s}^{-1}$
ξ	represents a prognostic variable	-
ψ_{w1}	diffusion of water from rooting zone to surface soil layer	mms^{-1}
ψ_{w2}	diffusion of water from total column to rooting zone	mms^{-1}
Ψ_w	rate of transfer of water by diffusion to the upper soil layer	mms^{-1}
a	Marshall-Palmer distribution parameter	-
A	surface area	m^2

A_{γ}	absorptivity due to a given gas	-
ABE	buoyant energy available	J
ABE''	production of available buoyant energy by large scale motions during the time Δt	J
b	Marshall-Palmer distribution parameter	-
B	Plank function; also Clapp and Hornberger exponent	-
B	bias	mm/day
B_0	Bowen ration	-
C_D	aerodynamic drag coefficient over land	-
C_{DN}	drag coefficient for neutral stability	-
CMD	catchment moisture deficit	mm
C_p	specific heat of air	$\text{Jkg}^{-1}\text{K}^{-1}$
C_s	specific heat of the subsurface soil layer	$\text{Jkg}^{-1}\text{K}^{-1}$
C_{SOILC}	transfer coefficient between canopy air and underlying soil	-
D	diameter of droplet; relative drying power; also discharge from catchment	m; - ; mm
D_{db}	diurnal penetration depth	m
D_f	characteristic dimension of the leaves in the direction of wind flow	-
D_s	water diffusivity in the soil	m^2s^{-1}
D_w	rate of excess water dripping from leaves per unit land area	mm m^{-2}
E	evapotranspiration	mm
e_a	vapour pressure in near surface atmosphere	mb
E_a	total evaporative flux from the surface to the atmosphere	$\text{kgm}^{-2}\text{s}^{-1}$
E_A	drying power of the air	mm
E_f	evaporative flux from foliage	$\text{kgm}^{-2}\text{s}^{-1}$
E_f^{WET}	evaporation from wet foliage per unit wetted area	$\text{kgm}^{-2}\text{s}^{-1}$
E_g	evaporative flux from the ground	$\text{kgm}^{-2}\text{s}^{-1}$
E_p	potential evapotranspiration	mm
E_{p0}	evapotranspiration rate at which $E_p=E$	mm

E_{PT}	Priestly & Taylor potential evapotranspiration	mm
e_s^*	saturation vapour pressure at the surface temperature	mb
E_{tr}	transpiration	mm
E_{trmax}	maximum transpiration	$\text{kgm}^{-2}\text{s}^{-1}$
$F^\uparrow$	upward long wave radiation flux	Wm^{-2}
$F^\downarrow$	downward long wave radiation flux	Wm^{-2}
f_{cld}	probability of a cloud existing in a given atmospheric layer	-
f_{clear}	clear sky fraction of atmospheric column	-
$F_{clr}^\downarrow$	clear sky downward longwave radiation	Wm^{-2}
$F_{clr}^\uparrow$	clear sky upward longwave radiation	Wm^{-2}
f_g	wetness factor	-
F_H	horizontal diffusion effects	
F_q	moisture flux from ground to atmosphere	$\text{kgm}^{-2}\text{s}^{-1}$
F_{qm}	maximum moisture flux through the wet surface that the soil can sustain	$\text{kgm}^{-2}\text{s}^{-1}$
F_{qp}	potential evaporation	$\text{kgm}^{-2}\text{s}^{-1}$
F_{rr}	the unfrozen soil water	mm
F_s	atmospheric sensible heat flux	Wm^{-2}
F_V	vertical turbulent mixing effects	
g	gravity (ms^{-2}); also asymmetry parameter	-
G	net water applied to the surface in the absence of vegetation; also specific flux of heat into the ground	mm; Wm^{-2}
h	moist static energy; also height of PBL (m)	-
H_a	total sensible heat flux from the surface to the atmosphere	Wm^{-2}
H_f	sensible heat flux from foliage	Wm^{-2}
H_g	sensible heat flux from the ground	Wm^{-2}
h_l	meridionally varying, empirically derived local liquid water scale	-
h_s	surface heating	Wm^{-2}
I_1	amount of condensation integrated over the whole depth	-

	of the updraft normalised by the updraft mass flux	
I_2	evaporation in the downdraft normalised by the downdraft mass flux	-
J	relative evaporation; also Thornthwaite heat index	-
k	von Karman constant	-
K_ξ	the eddy diffusivity coefficient	m^2s^{-1}
k_{sb}	thermal diffusivity of soil for diurnal wave	m^2s^{-1}
K_{w0}	saturated hydraulic conductivity	ms^{-1}
K_{zm}	momentum diffusivity coefficient at height z above the surface	m^2s^{-1}
K_{zm}	eddy diffusivity for moisture at height z above the surface	m^2s^{-1}
K_{zt}	eddy diffusivity for temperature at height z above the surface	m^2s^{-1}
L	latent heat; Monin-Obukhov length scale; also daytime hours	Wm^{-2} ; - ; 12 hrs
L_{AI}	leaf area index	-
L_d	fraction of foliage surface free to transpire	-
L_e	latent heat of evaporation	Wm^{-2}
L_f	latent heat of fusion	Jkg^{-1}
L_s	latent heat of sublimation	Jkg^{-1}
LS	denotes the large scale tendency	-
L_v	latent heat of vaporisation	Jkg^{-1}
L_w	fractional area of leaves and stems covered by water	-
M	moisture availability parameter	-
m_0	mass flux at the downdraft originating level	$\text{kgm}^{-2}\text{s}^{-1}$
m_b	mass flux at the updraft originating level	$\text{kgm}^{-2}\text{s}^{-1}$
MC	denotes the model calculated tendency	-
m_d	mass flux of the downdraft	$\text{kgm}^{-2}\text{s}^{-1}$
M_f	stomatal resistance dependence on soil moisture	-
m_u	mass flux of the updraft	$\text{kgm}^{-2}\text{s}^{-1}$
n	the displacement from the nearest boundary	grid points
N_0	Marshall-Palmer distribution parameter	m^{-4}

NA	rate of change of available buoyant energy per unit of mass flux	Js^{-1}
NSE	Nash-Sutcliffe efficiency	-
p	pressure	Pa
P	precipitation	mm
p^*	$p_s - p_{top}$	Pa
p_{clbk}	pressure level of the cloud base at k	Pa
P_{CON}	condensation of water vapour into cloud	$\text{kgkg}^{-1}\text{s}^{-1}$
P_{ID}	sublimation/deposition of cloud ice	$\text{kgkg}^{-1}\text{s}^{-1}$
P_{II}	initiation of ice crystals	$\text{kgkg}^{-1}\text{s}^{-1}$
P_{MF}	melting (freezing) of snow or ice (rain or cloud) due to atmospheric advection	$\text{kgkg}^{-1}\text{s}^{-1}$
Pr	Prandtl number	-
P_r	precipitation falling as rain	mm
P_{RA}	accretion of cloud by rain	$\text{kgkg}^{-1}\text{s}^{-1}$
P_{RC}	conversion of cloud to rain	$\text{kgkg}^{-1}\text{s}^{-1}$
P_{RE}	evaporation of rain	$\text{kgkg}^{-1}\text{s}^{-1}$
P_{RM}	snow melting to become rain	$\text{kgkg}^{-1}\text{s}^{-1}$
p_s	prognostic surface pressure	Pa
p_{top}	pressure specified to be the model top	Pa
q	specific humidity; also streamflow (observed)	kgkg^{-1} ; mm
$\hat{q}$	modelled streamflow	mm
q_a	specific humidity of the lowest model level	-
q_{af}	water vapour specific humidity of the air within the foliage	-
q_c	mixing ratio of cloud water	-
q_g	saturated specific humidity at the temperature of the surface	-
Q_g	net outflow of ground water	mm
$q_{g,s}$	saturated specific humidity at soil surface temperatures	-
Q_n	available energy flux density	Wm^{-2}

q_r	mixing ratio of rain water	-
Q_s	net outflow of surface water	mm
q_v	mixing ratio of water vapour	-
$\mathcal{R}$	residual of water balance	mm
r_e	effective of cloud droplets	m
R_f	stomatal resistance dependence on solar radiation	-
R_g	groundwater runoff	mm
Ri_B	surface bulk Richardson number	-
Rib_{cr}	critical bulk Richardson number	-
r_{la}	aerodynamic resistance to moisture and heat flux	sm^{-1}
R_{net}	net incident radiation at the surface	Wm^{-2}
r_s	stomatal resistance	sm^{-1}
R_s	surface runoff	mm
r_{smin}	minimum stomatal resistance	sm^{-1}
R_{ti}	fraction of roots in soil layer i	-
s	volume of water divided by volume of water at saturation	-
S	sources and sinks; also water volume stored in the system	- ; mm
S_0	solar constant	Wm^{-2}
S_a	clear air absorption of shortwave radiation flux	Wm^{-2}
S_{AI}	stem area index	-
S_{ca}	cloud absorption of shortwave radiation flux	Wm^{-2}
S_{cs}	cloud scattering of shortwave radiation flux	Wm^{-2}
S_d	downward shortwave radiation flux	Wm^{-2}
S_f	stomatal resistance dependence on temperature	-
S_g	solar flux absorbed over bare ground	Wm^{-2}
s_i	soil water in layer i	-
S_m	rate of snow melt	$\text{kgm}^{-2}\text{s}^{-1}$
SM	soil moisture	mm
S_{rw}	rooting zone soil water	mm
S_{rwmax}	maximum rooting zone soil water	mm
S_s	clear air scattering of shortwave radiation flux	Wm^{-2}
S_{sw}	surface soil water (upper layer)	mm

S_{swmax}	maximum upper soil water	mm
S_{tw}	total water in the soil	mm
S_{twmax}	maximum water in total soil column	mm
s_w	soil water for which transpiration essentially goes to zero	-
t	time	s
T	Temperature	K
T_a	air temperature of lowest model layer	K
T_{af}	temperature within the foliage layer	K
T_c	cloud water transmissivity	-
T_f	temperature of foliage	K
T_{g1}	surface soil temperature	K
T_{g2}	subsurface temperature	K
T_{g3}	deep soil temperature	K
T_p	precipitation transmissivity	-
T_v	water vapour transmissivity	-
u	cross front wind velocity	ms^{-1}
U	horizontal wind speed; also effective rainfall	ms^{-1} ; mm
$\bar{u}$	mean wind speed	ms^{-1}
u_*	surface frictional velocity scale	ms^{-1}
U_a	horizontal wind above the canopy	ms^{-1}
U_{af}	wind velocity within foliage layer	ms^{-1}
u_c	cloud water path	gm^{-2}
u_g	geostrophic wind	ms^{-1}
u_p	effective water path	gm^{-2}
v	along front wind velocity	ms^{-1}
V_f	fall speed of rain or snow (ms^{-1}); also stomatal resistance dependence on vapour pressure deficit	-
V_q	relative volume of flow that travels through as quickflow	-
V_s	relative volume of flow that travels through as slowflow	-
w	vertically integrated cloud water path length; also a weighting function for the lateral boundary conditions	gm^{-2}

W_{dew}	total water stored by canopy per unit land area	mm m ⁻²
W_{DMAX}	maximum water the canopy can hold	mm m ⁻²
W_{LT}	soil dryness (or plant wilting) factor	-
w_s	mixed layer velocity scale	-
$x^{(q)}$	quickflow	mm
$x^{(s)}$	slowflow	mm
z	height above the surface	m
z_0	originating level of downdraft; also the roughness length	m
z_1	height of lowest model level	m
z_b	originating level of updraft	m
Z_r	depth of soil rooting layer	m

List of Acronyms

<i>Acronyms</i>	<i>Meaning</i>
ABRACOS	Anglo-Brazilian Climate Observation Study
AMIP	Atmospheric Model Intercomparison Project
ARI	Average Recurrence Interval
ARPE	Average Relative Parameter Error
ASCE	American Society of Civil Engineers
BALTEX	Baltic Sea Experiment
BATS	Biosphere-Atmosphere Transfer Scheme
BOREAS	Boreal Ecosystem-Atmosphere Study
CMD	Catchment Moisture Deficit
CSIRO	Commonwealth Scientific and Industrial Research Organisation
ECMWF	European Centre for Medium Range Weather Forecasts
EF	Evaporative Fraction
EFEDA	ECHIVAL Field Experiment in Desertification threatened Areas
ET	evapotranspiration
FIFE	First ISLSCP Field Experiment
GARP	Global Atmospheric Research Program
GCM	Global Climate Model
GFDL	Geophysical Fluid Dynamics Laboratories
GISS	Goddard Institute for Space Studies
GMT	Greenwich Mean Time
HAPEX	Hydrological and Atmospheric Pilot Experiment
HEIFE	Hei Ho River basin Field Experiment
IFC	Intensive Field Campaign

IHACRES	Identification of Hydrographs And Components from Rainfall, Evaporation and Streamflow data
IPCC	Intergovernmental Panel on Climate Change
ISLSCP	International Satellite Land Surface Climatology Project
IUH	Instantaneous Unit Hydrograph
LAM	Limited Area (climate) Model
LTER	Long Term Ecological Reserve
MM5	Penn State/NCAR Mesoscale Model version 5
MOBILHY	Modelisation du Bilan Hydrique
NASA	National Aeronautics and Space Administration
NCAR	National Center for Atmospheric Research, USA
NCEP	National Center for Environmental Prediction, USA.
NOAA	National Oceanic and Atmospheric Administration
NOPEX	Northern Hemisphere Climate Processes Land Surface Experiment
NSE	Nash-Sutcliffe Efficiency
PAM	Portable Automatic Mesonet stations
PBL	Planetary Boundary Layer
PET	Potential evapotranspiration
PILPS	Project for Intercomparison of Land-surface Parameterisation Schemes
PIRCS	Project to Intercompare Regional Climate Models
RegCM2	NCAR Regional Climate Model version 2
SHEELS	Simulator for Hydrology & Energy Exchange at the Land Surface
SRIV	Simple Refined Instrumental Variable technique
SVAT	Soil-Vegetation-Atmosphere Transfer scheme
UTC	Coordinated Universal Time (GMT)

Chapter 1

Introduction

The main objective of this dissertation is the development of tools and methodology to perform assessments of the potential impact of climate change on the surface hydrological regime in data sparse areas. This involves the development and implementation of models, which require the minimum of site specific data, allowing the impact of climate change on streamflows to be assessed with reasonable confidence. This problem is addressed in several stages:

1. Identification and development of a parsimonious hydrological model allowing determination of streamflow and evapotranspiration behaviour given historical precipitation, temperature and stream discharge data.
2. An assessment of the ability of three regional climate models to reproduce current climate and an intercomparison of their simulation results. Particular attention is given to the simulation of energy and water budgets at the land surface and the associated uncertainties.
3. The runoff simulated by a regional climate model is then improved upon by the incorporation of the previously developed hydrological model.
4. A combination of a climate model, a stochastic weather generator and the hydrological model are used in a case study to assess the likely impact of global warming on the surface hydrology regime.

1.1 Rationale

The hydrological cycle is one of the most visible of the planetary cycles, with clouds, rain, rivers and oceans, and is critical to the existence of life. Improving our understanding of the hydrological cycle is gaining more and more importance as the demands on the fresh water supply increase worldwide every day. Population and economic pressures are driving agriculture to be more productive through the use of irrigation, which in places like much of the Murray-Darling basin in Australia is leading to salinization of soils. Human activities in drought prone areas can lead to desertification. In 1977 UNESCO produced a map "showing the extent and degree of desertification hazards by bioclimatic climates". On this map 30% of the land area of the Earth was depicted as at threat from desertification. Large cities are also consuming

huge amounts of water. Once major rivers, like the Colorado river in the US, now reach the ocean as small streams, and entire countries, like Spain, have effectively run out of water. The need to plan and manage water resources has never been as great and is likely to increase in the foreseeable future. The need for accurate modelling of the hydrological cycle is obvious. This model development activity is driven not just by scientific curiosity but by the requirements of countries to manage this most precious, and increasingly scarce, resource in a sustainable fashion for future generations.

Sustainable water resource systems are those designed and managed to fully contribute to the objectives of society, now and in the future, while maintaining their ecological, environmental and hydrological integrity.

(UNESCO 1999)

Sustainable water resource planning therefore requires knowledge (or educated guesses) of the future water demand and supply. The water demand is affected by many economic and social factors, most notably population growth, while water supply is primarily affected by changes in land use and climate. Regions which are experiencing the greatest population growth tend to also be the regions with the largest increases in water demand. Information about the future of the water supply is absolutely critical for sustainable water resource planning in these regions. High rates of population increase is almost solely a characteristic of developing nations and hence there is added importance in studying the water supply in these areas.

Climate related changes in the water supply are less influenced by human activity than either the water demand or changes in land use. As such changes in land use and water demand may be used to counter or complement any climate related changes. In any water resource planning then, climate related changes to the water supply can be considered the principal change, which then constrains changes related to land use or demand.

Quantifying these climate related changes is of the foremost importance in water resource planning, particularly in the developing world. The current state of climate science however, does not allow true quantitative predictions about future climate

without considerable associated uncertainties. Frederick and Gleick (1999) note that ‘Global climatic changes will have major effects on precipitation, evapotranspiration and runoff. But estimating the nature, timing, and even the direction of the impacts at the regional and local scales of primary interest to water planners involves many uncertainties.’

As the presence of global warming becomes increasingly certain, it seems likely that at least in some regions the hydrological regime will be altered to such an extent that it will no longer correspond to any regime found in the historical records. The need to manage water resources into this uncertain future, along with a desire to reduce this uncertainty are major motivating factors behind this thesis.

1.2 Background

Climate models have been developed predominantly by atmospheric scientists, and as a result, historically little attention was given to processes occurring at the land surface. A simple model of land surface hydrology was first developed in the mid-60s (Manabe et al. 1965), adapting Budyko's "bucket" model (Budyko 1956) from observations to models. This simple scheme uses the idea that the soil acts like a bucket which is filled by rainfall and overflows as runoff. It does not reproduce observed processes at the land surface (Wood et al. 1991, Chen et al. 1996, Timbal and Henderson-Sellers 1998).

Over the last several decades, the importance of improving land surface schemes used in climate models has become more apparent. Predictions of the energy and water balance at the surface, including the atmospheric feedbacks and the potential impacts of climate and/or land use change on streamflows, all need to be investigated. As a result, the development of land surface hydrology components inside climate models has been receiving considerably more attention. Very complicated Soil-Vegetation-Atmosphere Transfer (SVAT) schemes for dealing with these land surface processes (Dickinson et

al. 1986, Sellers et al. 1986, Abramopoulos et al. 1988) have been developed. The use of these complex models has brought new problems, including over-parameterization and substantial computational demand when included as part of a climate model (Pitman et al. 1999).

Hydrologists over the years have developed a plethora of surface hydrology models, spanning a wide range of complexity, which are able to reproduce observations well. Hydrological models can be broadly classified into three types (Wheater et al. 1993): empirical, conceptual and physically-based. Empirical models use a ‘*black box*’ paradigm to model the rainfall-runoff processes. That is, they use mathematical functions to transform the rainfall input time series into the runoff output time series. No attempt is made to ascribe physical meaning to the mathematical functions. These models cannot reliably predict under different conditions than existed in the historical record used for calibration.

Both the conceptual and physically-based models fall into the ‘*grey box*’ paradigm. That is, they also use mathematical functions to transform the rainfall into runoff. However they explicitly ascribe physical meaning to the mathematical functions. Often this physical meaning gives rise to various interpretations of the internal states of the model which are unverifiable. Physically-based models in particular have drastically increased in complexity due to the attempt to include within the grey box all processes that are thought to be involved in the rainfall-runoff transformation. Using a physical interpretation allows the potential measurement of internal variables thereby increasing the falsifiability of the model. For example, hydrological models will often predict evapotranspiration and soil moisture as internal variables, these can be measured and the model validated against them. Often though the physical meaning ascribed to a parameter or mathematical function is somewhat vague and does not correspond to a measurable quantity as was found in Shao et al. (1994). Observations of internal variables or states of a model can be thought of as splitting the grey box into a series of smaller (lighter) grey boxes. The complexity contained within the grey box of a physically-based model, along with the inability to measure physical parameters severely reduces the predictive capability of these models (Beven 1989, Grayson et al. 1992a, Lettenmaier 1995).

Conceptual type models display a wide range of complexities, with the most complex conceptual models containing a similar level of complexity to that found in physically-based models. The more complex schemes typically contain too many parameters to allow identification from rainfall-runoff times series (Jakeman and Hornberger 1993). In fact Kirkby (1976) suggests a maximum of around 10 parameters, while Jakeman and Hornberger (1993) demonstrate that only 6 – 7 parameters can be justified when identifying parameters based on daily rainfall-runoff time series. This lack of identifiability means that significant uncertainty exists in the parameter values themselves and thus reduces the predictive capability of these models.

The model chosen for further development in this study, IHACRES (Identification of Hydrographs and Components from Rainfall, Evapotranspiration and Streamflow data), is a lumped parameter, conceptual rainfall-runoff model containing only seven parameters. The parameters of the model are not ascribed direct correspondence with a particular physical process, as would be the case in physically-based models, instead the parameters represent a number of physical processes lumped together.

While several attempts have been made to modify existing hydrological models for use within climate models (Dumenil and Todini 1992, Wood et al. 1992, Bobinski et al. 1993) they still have difficulty reproducing observations. Numerous studies have investigated land surface schemes in climate models in terms of their atmospheric feedbacks (latent and sensible heat etc.) and surface processes such as runoff (Entekhabi and Eagleson 1989, Wood et al. 1991, Dumenil and Todini 1992, Kuhl and Miller 1992, Wood et al. 1992, Garratt et al. 1993, Polcher et al. 1996, Timbal et al. 1997, Franks 1998). It was generally found that the climate models perform reasonably well in terms of atmospheric feedbacks but they tend to perform poorly in terms of runoff.

1.3 Chapter Outline

Part I addresses the streamflow modelling component of the dissertation, it contains appropriate review material and the development of a rainfall-runoff model. Chapter 2 presents a review of the current state of rainfall-runoff modelling. It includes a discussion of the various types of rainfall-runoff model, and issues such as those related to scale and over-parameterisation. It also presents the reasons for the choice of rainfall-runoff model used in this study. Chapter 3 reviews the current knowledge related to modelling evapotranspiration. It summarises the various techniques currently used in rainfall-runoff and climate models. Issues involved with the measurement and validation of evapotranspiration model results are also discussed. In Chapter 4, knowledge from Chapters two and three are brought together in the development of the rainfall-runoff model CMD-IHACRES. The model-simulated streamflow is validated on several catchments covering a range of hydroclimatologies. The impact of various evapotranspiration formulations is also considered.

Part II addresses the climate modelling component of the dissertation, it contains appropriate review material and a study comparing three regional climate models. Chapter 5 presents a review of the climate system and climate models. It summarises the major components of global climate models and goes on to look at issues related to regional climate modelling. It provides a reasonably detailed description of the regional climate models (limited area models) used in the dissertation: MM5/BATS, MM5/SHEELS and RegCM2. The Chapter finishes with a discussion of climate model validation and a review of previous intercomparison studies. Chapter 6 presents a comparison of the surface energy budget simulated by the regional climate models and observations taken during the First ISLSCP (International Satellite Land Surface Climatology Project) Field Experiment (FIFE). It focuses on a comparison of the models and measurements with respect to incident radiation, latent heat, sensible heat, net surface heating, near surface temperature and wind, as well as the overall surface energy balance. Chapter 7 presents a comparison of the surface water budget simulated by the regional climate models and observations taken during FIFE. It focuses on a comparison of the models and measurements with respect to precipitation, evapotranspiration, runoff, soil moisture, as well as the overall surface water balance.

Part III brings the first two parts together. It contains a study of the combined use of a regional climate model and a rainfall-runoff model as well as a case study of the potential hydrological impact of global warming. Chapter 8 analyses the daily streamflow simulation, including characteristics, produced by the three different climate models alone and by the combination of the climate models with the rainfall-runoff model. Chapter 9 presents a practical case study of the potential climate change impact on the hydrology of a region outside Perth, Western Australia. It uses climate change scenarios, developed through the use of a global climate model and a stochastic weather generator, to drive the rainfall-runoff model.

Finally, the main conclusions of this dissertation are presented in Chapter 10.

Part I: Streamflow Modelling



Before an attempt is made to model the effects of a change in climate on the streamflow produced in a catchment, an adequate rainfall-runoff model is required. In particular the model is required to perform satisfactorily in a range of hydroclimatologies, to have minimal data requirements, and to be linked explicitly to the atmosphere through evaporation. This part presents reviews of the current state of rainfall-runoff and evaporation modelling. This is followed by the development of a rainfall-runoff model, CMD-IHACRES, which satisfies all of the above criteria.

Chapter 2

Rainfall-Runoff Modelling Review

The primary aim of catchment-scale rainfall-runoff modelling is to estimate runoff exiting a catchment given the rainfall entering it. Research has demonstrated, however, that additional variables are required to estimate runoff with any measure of skill. These data requirements can be as simple as using surface air temperatures (Jakeman et al. 1990) or as complex as requiring detailed spatial distributions of various soil and vegetation characteristics along with a series of atmospheric variables (Abbott et al. 1986b). As well as this considerable range in data requirements, there is also a considerable range in the complexity of current rainfall-runoff models. The earliest attempts to mathematically model the rainfall-runoff process, occurring around the 1930s, were relatively simple approaches. The subsequent increase in complexity and data requirements closely followed increases in computing power, which essentially allowed a more detailed description of the processes involved to be modelled explicitly.

In order to understand why such a plethora of rainfall-runoff models exist we first look at the complexity involved in the rainfall-runoff process itself.

2.1 What physical processes are involved in converting rainfall to runoff?

A comprehensive study of all the processes involved in modelling the conversion from rainfall to runoff is beyond the scope of this review. Presented here instead is an overview of the most significant processes, with the goal of providing a good indication of the complexity involved in rainfall-runoff modelling

2.1.1 *Non-streamflow losses from a catchment*

Not all of the rain that falls within a catchment reaches the outlet as streamflow. A significant amount, or in some cases even all, of the rainfall will leave the catchment via mechanisms other than streamflow. These losses are primarily evapotranspiration to the atmosphere, and groundwater seepage into deep aquifers.

Losses to aquifers depend to a great extent on the geological structure of the subsurface. The amount of water loss to the aquifer is affected by the presence and extent of recharge zones as well as the velocity of water movement within the aquifer. The velocity is largely determined by the slope of the aquifer, if any, as well as the medium through which the water moves, i.e. the soil types and distributions. It can also be influenced by human activities, especially the pumping of water out of the aquifer for irrigation of crops etc.; for large aquifers this may occur even outside the catchment of interest. Aquifers are not an infinite sink though, and the water must eventually leave the aquifer itself, thus aquifers are also capable of supplying water to a catchment which may have originated as rainfall in another catchment. In general, movement of water through aquifers is orders of magnitude slower than the near surface movement and as such it can be ignored in many practical applications of rainfall-runoff modelling.

Losses due to evapotranspiration, on the other hand, act on time scales similar to water movement through a catchment, and thus can have a major impact on the amount of water that eventually reaches the stream. A more thorough analysis of the evapotranspirative process is given in Chapter 3, so only a brief description is included here. Essentially evapotranspiration consists of the transpiration from all the vegetation within a catchment as well as actual evaporation occurring from any surfaces, including that of the vegetation and bare ground. Transpiration itself is a complex process involving many biological, atmospheric and soil properties. It can differ substantially between plant species and between seasons. Evaporation is somewhat simpler, depending only on atmospheric conditions and soil properties. In arid and semi-arid areas evapotranspiration will often account for up to 100% of the rainfall and as such it is a very important component of rainfall-runoff modelling.

2.1.2 Movement of water through a catchment

Of the water that does exit the catchment via the stream there are many pathways by which it may move through the catchment. Rain may land directly in the stream, travel as overland flow, infiltrate into the soil and then travel laterally or even infiltrate all the way to a groundwater table and travel via that. Infiltration may occur as capillary flow in the soil matrix or flow via natural pipes or macropores. Chorley (1978) defines a spectrum of these runoff generation mechanisms. In reality, complex combinations of all possible pathways are utilized to generate the observed runoff.

2.1.3 Snow processes

In regions affected seasonally by snow, development of the snowpack often dominates the local hydrology. Snow accumulation and snow melt become a major influence on streamflow, and are dependent on many topological factors such as aspect and slope as well as meteorological factors such as wind direction, incident solar radiation and precipitation. The majority of the work presented in this thesis has been performed in snow free areas and as such little attention will be given to snow related processes.

2.1.4 Spatial and temporal heterogeneity

The extreme heterogeneity inherent in many of the key inputs and landscape characteristics is also important in determining streamflow. Rainfall in particular can display significant heterogeneity both in spatial extent and in overall duration. Indeed, within any one rainfall event large variations in spatial and temporal intensity can exist. Traditional point measurements are incapable of capturing this heterogeneity, though more recent small-scale radar measurements show promise in this regard.

Landscape characteristics such as slope, aspect, soil type and vegetation cover all display significant spatial heterogeneity. For vegetation there is the added complication of (sometimes extreme) seasonal changes as well. This heterogeneity creates many problems when it comes to measurement. Either point measurements are taken and efforts are made to scale the observations to an appropriate level for modelling, or remote sensing techniques may be employed which, although showing much promise, have several limitations of their own.

2.2 Modelling approaches

When attempting to model rainfall-runoff processes it is clear that simplifications must be made. This simplifying remains valid only if the dominant processes in each particular situation are modelled. Unfortunately it is often impossible to know a priori which physical processes and pathways dominate, as was demonstrated in a study by Pilgrim et al. (1978). It is also often difficult to distinguish the dominant processes after the fact, with models using different simplifications (that is, models based on different dominant processes) performing comparably in the same places and for the same events.

These difficulties have led to an abundance of rainfall-runoff models. Several attempts have been made to classify these models into generic types though all classifications seem to have fuzzy borders. Presented here is an outline of various model types using the classification proposed by Wheater et al. (1993). This study split rainfall-runoff models into metric (empirical), conceptual and physically-based. An additional often used and occasionally useful distinction is the split between lumped and distributed models.

2.2.1 Metric (Empirical) Models

According to Wheater et al. (1993) ‘The essential characteristic of metric models is that they are based primarily on observations and seek to characterize system response from those data.’ Metric models are essentially empirical or ‘black box’ models. They make little if any attempt at explaining internal processes, instead merely using mathematical techniques to perform an input-output transformation. These models cannot be used to make predictions on periods independent of the calibration period, unless the periods have very similar characteristics.

This approach was developed primarily in the early days of hydrological modelling and can be traced to the development of unit hydrograph theory for catchment scale stormflow simulation (Sherman 1932). The unit hydrograph concept considers only event response. The unit hydrograph $h(t)$ is the stormflow response at time t to a unit input of effective rainfall $u(t)$. Effective rainfall is that part of the rainfall that eventually becomes stormflow, that is rainfall minus losses (evapotranspirative and aquifer). Stormflow is modelled explicitly as a linear, time invariant function of effective rainfall, i.e. its convolution with the unit hydrograph.

$$x^{(q)}(t) = \int_0^t h(t-s)u(s)ds \quad (2.1)$$

The unit hydrograph concept has been extended to include modelling of baseflow and hence allow simulation between events when stormflow has waned. One such model is the CLS model of Natale and Todini (1977).

2.2.2 Conceptual Models

Conceptual rainfall-runoff models are designed to approximate within their structure the general physical mechanisms which govern the hydrologic cycle (Sorooshian 1991). Development of this class of model is driven almost solely by a hydrologist’s individual

perception of the rainfall-runoff processes operating. The level of complexity incorporated in such a model can vary considerably, but generally they are based on a representation of internal storages and the flows between these stores.

Wheater et al. (1993) claim the ‘essential features of these models are that (i) the model structure is specified a priori, according to the perception of the important component processes... and (ii) parameter adjustment, through calibration against observed data, is required to optimise parameter values to represent the system of interest.’

Examples of conceptual models include the following: STANFORD IV (Crawford and Linsley 1966), SACRAMENTO (Burnash et al. 1973), TANK (Sugawara et al. 1983), SFB (Boughton 1984), MODHYDROLOG (Chiew and McMahon 1994), HBV (Bergstrom 1995) and LASCAM (Sivapalan et al. 1996a, Sivapalan et al. 1996b, Sivapalan et al. 1996c).

2.2.3 *Physically-based Models*

According to Wheeler et al. (1993), physically-based models represent the component hydrological processes ‘in a more classical mathematical-physics form, based on continuum mechanics, through numerical solution of the relevant equations of motion using a finite difference or finite element spatial discretisation.’

Examples of physically-based models include the Institute of Hydrology Distributed Model (IHDM) (Beven et al. 1987), the Système Hydrologique Européen (SHE) (Abbott et al. 1986a, b), WSHS (Al-Soufi 1987), THALES (Grayson et al. 1992b) and TOPOG (Dawes and Short 1994). Typically, these models use the Richards equation for subsurface flows and a kinematic wave equation for surface flow. In theory, the parameters of these models have explicit physical meanings and can be measured a priori.

2.2.4 Lumped vs Distributed

Two approaches to the overall spatial scale can be distinguished. *Lumped* generally refers to a model which deals with the catchment in its entirety, while *distributed* refers to a model which subdivides the catchment into smaller spatial units often using a grid system. Whole catchment models are referred to as *lumped* since they take all the subcatchment variability and ‘lump’ that into a set of effective parameters for the entire catchment. *Distributed* models on the other hand, have a set of effective parameters for each cell or spatial unit.

This distinction is somewhat artificial when you consider that within each cell of a *distributed* model the parameters are effectively *lumped* parameters for that cell. Also a *lumped* model may be applied to several neighbouring catchments or sub-catchments and hence give a distribution of parameter values for the basin, e.g. Schreider et al. (1996).

2.3 Previous reviews and model comparisons

Several previous reviews of the progress in hydrologic research have been published in the last decade. Presented here is a summary of the findings of these reviews followed by a review of rainfall-runoff model intercomparison studies.

Goodrich and Woolhiser (1991) provide a review of catchment hydrology focused largely on distributed catchment modelling. They discuss a number of model developments, improvements, comparisons and assessments and concluded that “uncertainty in the effective rainfall distribution dominated runoff prediction uncertainty... Those model evaluations which utilized observed data, do not paint an encouraging picture of our ability to model catchment response.” They also found several innovative techniques for the treatment of catchment variability including the

concept of a Representative Elementary Area (REA) put forth by Wood et al. (1988). For scales smaller than the REA, spatially detailed representations of relevant variables must be used. For scales larger than the REA, such a spatially detailed representation is not required, instead average or lumped relationships may be used. They emphasise the importance of experimental catchment hydrology, stating that “Experimental studies are essential if we are to attain a truly scientific understanding of hydrologic phenomena.”

Hornberger and Boyer (1995) reviewed advances in watershed modelling and drew particular attention to approaches that employ Geographic Information Systems (GIS), remotely sensed data and environmental tracers. The use of digital terrain data, embodied within a GIS, has been facilitated in part by the introduction of a modelling approach based on topographic indices ((Beven and Kirkby 1979, O'Loughlin 1981, Moore et al. 1991, Wolock 1993)). Examples of topographically based models are TOPMODEL (Beven and Kirkby 1979) and TOPOG (Dawes and Short 1988). These models calculate an index to catchment wetness based on topography for each grid cell, grid cells possessing the same value of the topographic index are assumed to behave the same hydrologically, regardless of their location in the landscape. Hornberger and Boyer (1995) note that although this is an area of active and ongoing research “Attempts to correlate these indices with measures of soil wetness have met with limited success (Jackson 1991, Barling et al. 1994).”

Hornberger and Boyer (1995) also found that the increasing availability of remotely sensed data has led to new modelling approaches that use this information as well as facilitating characterization of the landscape heterogeneity. Uses of remotely sensed data included using satellite derived land cover classes (Duchon et al. 1992, Kite and Kouwen 1992), using radar derived rainfall data (Nicks and Schiebe 1992) or even using spatially distributed radiation data to improve the modelling of snow cover and snow melt (Martinec et al. 1983, Blöschl et al. 1991, Marks and Dozier 1992).

Hornberger and Boyer (1995) consider “The most serious problems relative to the establishment of a truly accurate, useful, and tenable theory of catchment functioning as embodied in mathematical models are those of 1) the identifiability of model parameters, ... and 2) the incorporation of relatively small-scale heterogeneity into

models applied at relatively larger scales. These continue to be areas of active research today.

In a review of advances related to large-scale field experiments in hydrology, Kustas (1995) emphasises the role these experiments play in studying the land surface – atmosphere interactions from the local to the regional scale. He suggests they might facilitate the integration of hydrologic models with climate models. A list of these large-scale field experiments can be found in Table 5.3. These experimental results suggest “that variations of less than 20% in the (energy) fluxes cannot be easily distinguished from measurement errors.” However, Kustas still considers the data from these experiments to be “the most comprehensive and the best available to improve parameterisation schemes in models coupling hydrologic and atmospheric processes.”

A review of catchment scale hydrologic modelling approaches is provided by Jayatilaka and Connell (1995). They considered two model types in this review: (1) physically-based distributed models: SHE (Abbott et al. 1986a, b), TOPOG (Dawes and Hatton 1993), IHDM (Beven et al. 1987), THALES (Grayson et al. 1992b), TOPMODEL (Beven and Kirkby 1979), WSHS (Al-Soufi 1987) and (2) lumped conceptual models: SDI (Kuczera 1988), CATPRO (Kuczera et al. 1993) and MIDASS (Nathan 1993). Some general conclusions found in this review are that lumped conceptual models are suitable for the prediction of spatially averaged catchment response. This is particularly the case in the absence of spatially distributed input data. One problem with physically-based models was that the detailed information required as input was rarely available making calibration difficult. However physically-based models can provide valuable understanding of the physical processes at the scale where the required input can be adequately defined.

There have also been studies which perform comparisons of the performance of different hydrological models on the same catchment. Some of these include the study by the World Meteorological Organisation (1975), which conducted a study of ten models applied in six catchments ranging in size from 1445 km² to 131,500 km². Unfortunately only two models were applied to all six catchments and with the

catchment size raising significant questions over the representativeness of the precipitation, no clear conclusions could be drawn.

Loague and Freeze (1985) compared three rainfall-runoff models: a regression model, a unit hydrograph model and a quasi-physically based model. Comparison was made for 269 events occurring in three experimental catchments in the United States. While the study found that all models produced poor forecasting efficiencies, the simpler models performed at least as well as the more complex quasi-physically based model. They state that “The fact that simpler, less data intensive models provided as good or better predictions than a physically based model is food for thought.” Hetrick et al. (1986) and Troutman et al. (1989) also found that simulations from a simple model compared reasonably well with more complex model process representations.

Ward et al. (1989) compared seven models on 26 South African catchments ranging in size from a few hectares to 100 km². The catchments embodied many different land uses. The models were applied without optimisation of parameters and the performance in general was found to be poor. Not surprisingly each model worked best when applied over the particular land use condition it was designed for.

In contrast to the majority of these comparison studies, Chiew et al. (1993) found that a complex conceptual model significantly outperformed simpler models. They compared six rainfall-runoff models, including IHACRES, applied on eight catchments in Australia. They found that the complex conceptual model, MODHYDROLOG, with 19 tunable parameters provided by far the best simulation of daily flows, with simpler conceptual and time series models only being adequate for estimating monthly and annual yields. However in a later study, Ye (1996) compared the performance of IHACRES and MODHYDROLOG in three Australian catchments and found that the complex conceptual model produced marginally better results on a daily basis.

Michaud and Sorooshian (1994) compared a simple model (SCS) to a complex distributed model (KINEROS) applied on a mid-sized semi-arid catchment in the United States. They found that the performance of all the models was disappointing.

However the simple model proved to be as accurate as the complex distributed model, provided calibration was performed.

Refsgaard (1994) provided a comparison of the performance of a lumped conceptual model, a distributed model of moderate complexity, and a distributed model of very high complexity. He applied the models to three African catchments ranging in size from 254 km² to 1040 km². He concluded that the lumped conceptual model, being easiest to apply and most accurate, should be chosen for dry catchments when calibration data are available.

Ye et al. (1997) compared three rainfall-runoff models applied to three low-yielding ephemeral catchments. The models included a simple conceptual model (GSFB: 8 parameters), a metric/conceptual model (IHACRES: 6 parameters) and a complex conceptual model (LASCAM: 22 parameters). The catchments are all found in Western Australia and range in size from 0.82 km² to 517 km². They found that the metric/conceptual model performed as well as the more complex conceptual model and concluded that for semi-arid ephemeral catchments “a model of about six parameters, albeit in an appropriate model structure, is sufficient to characterize the information in rainfall-discharge time series over a wide range of catchment sizes.”

2.4 Scale Issues in Hydrological Modelling

Issues of scale are by no means unique to the hydrological sciences, however the scientific study of water occurs over a very large range of both space and time scales, as pointed out by Dooge (1997) and shown in Figure 2.1. The spatial scales span 18 orders of magnitude and the temporal scales span 20 orders of magnitude. Attempting to deal with this massive range of scales is an area of active research. Reviews of scale issues in relation to hydrological research can be found in Dooge (1982), Klemes (1983), Dooge (1986), Wood et al. (1990), Beven (1991), Mackey and Riley (1991).

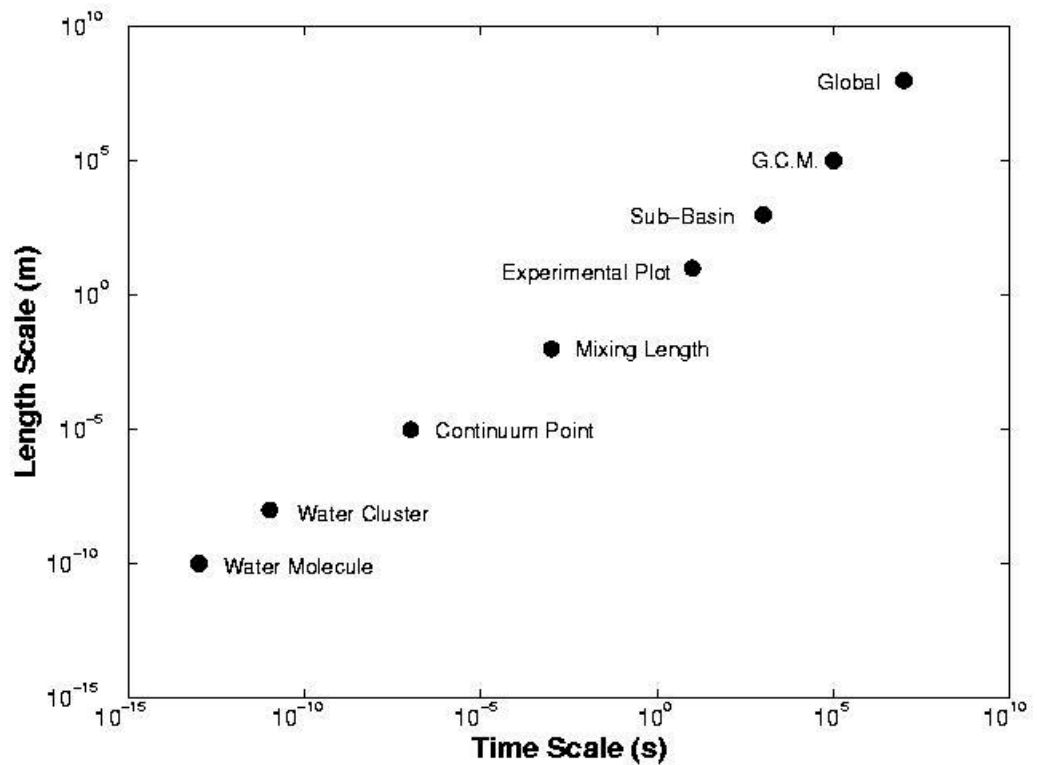


Figure 2.1 Scales in hydrology

A review of the scale issues in hydrological modelling can be found in Blöschl and Sivapalan (1995). They define the term scale as a characteristic time or distance of a process, observation or model. The process of information transfer from one scale to another is referred to as *scaling*. Scaling is required if there exists a mismatch between the process scale, the observation scale and/or the modelling scale. Many of the problems encountered when dealing with multiscale phenomena are due to the high degree of heterogeneity exhibited by natural catchments. This heterogeneity (or variability) manifests itself at a range of scales.

Blöschl and Sivapalan (1995) emphasise the potential for the use of dimensional techniques in catchment hydrology. These techniques have proved extremely useful in hydraulic applications but have only been applied sparingly in catchment hydrology.

The concept of similarity is the foundation of all dimensional techniques. Similarity exists between two systems whenever the characteristics of one system can be related to those of another via a simple conversion or scale factor (Langhaar 1951). They group the methods used to determine scale factors into three main classes: dimensional analysis, similarity analysis and functional normalization.

Dimensional analysis is a process in which the dimensional quantities used to describe a system are reduced to a fewer number of non-dimensional quantities. The procedure by which dimensional analysis should be carried out can be found in Fischer et al. (1979) and Stull (1988), it does not require the laws or equations governing the system to be known. An example of a dimensional analysis study is that of Strahler (1964). He related the normalized drainage density of catchments to three physically based similarity parameters.

Similarity analysis seeks to organize variables into non-dimensional groups and determine relationships between the groups. The main difference between this and dimensional analysis is the requirement that the physical laws or equations governing the system of interest are known. An example of a similarity analysis study is that of Sivapalan et al. (1987) who assumed that the required physical laws relating to runoff generation were embodied in the hydrological model TOPMODEL. By recasting the model equations into non-dimensional form they identified five non-dimensional similarity parameters and three auxiliary variables. Later work by Larsen et al. (1994) resulted in a number of similarity relationships between these variables.

Unlike the previous two methods, functional normalization is an empirical method (Tillotson and Nielsen 1984). This method begins with a set of empirical relationships between two or more variables and then attempts to combine all such empirical relationships into a reference curve which describes the set as a whole. Examples of the use of functional normalization in studies demonstrating empirical regionalisation of flood frequency can be found in McKerchar and Pearson (1989) and Pearson (1993).

Often hydrological models have been developed with little concern being displayed about appropriate scales and subsequently these models are applied at scales

incommensurate with the concepts embodied in the model. Beven (1995) argues that “the aggregation approach towards macroscale hydrological modelling, in which it is assumed that a model applicable at small scales can be applied at larger scales using ‘effective’ parameter values, is an inadequate approach to the scale problem.”

In a study questioning the appropriateness of the effective parameter concept, Wood et al. (1988) introduced a critical spatial scale below which accurate spatial representation of variables is required for the modelling of runoff generation, while above the critical scale statistical knowledge of this small scale heterogeneity is enough for the modelling of runoff. They refer to this scale as the Representative Elementary Area (REA). The REA is defined in terms of the variable contributing area as modelled by the hydrological model TOPMODEL. While this concept shows promise in determining an appropriate grid size for hydrological modelling, analysis of TOPMODEL raises concerns over process representation and physical credibility (Franchini et al. 1996). It should be noted that other models, such as the ARNO model (Todini 1996), predict different contributing areas, hence these must be regarded as model dependant and not necessarily representative of the real contributing areas of a catchment. Thus O'Connell and Todini (1996) note that “an attempt to define the size of a ‘representative elementary area’ based on TOPMODEL simulation experiments, must be regarded as rather speculative.”

2.5 Regionalisation

Hydrological sciences have been seeking the ability to accurately model the hydrologic response of ungauged catchments for some time. This is particularly important for planning and management purposes where decisions must often be made without sufficient observations. For example, the estimation of design flood hydrographs for sizing hydraulic structures such as dams and bridges (NERC 1975). A review of the regionalisation of hydrographic data can be found in Bates (1994).

Here the focus is on efforts to model ungauged catchments through the regionalisation of rainfall-runoff model parameters. This regionalisation is generally attempted by regressing model parameter estimates obtained from gauged catchments on the physical characteristics of the catchments within an assumed hydrologically homogenous region. As such, these derived relationships are empirical and apply only to the region for which they have been derived. Bates (1994) considers the successful regionalisation of model parameters to depend on

- Accurate estimation of model parameters for gauged catchments;
- The selection of catchment characteristics that affect catchment response to rainfall;
- The delineation of homogenous regions;
- The degree to which the model parameters are correlated with catchment characteristics; and
- Correct specification of the regression model for each region.

While Bates (1994) acknowledges that there is abundant literature on the regionalisation of parameters (Hughes 1985, Pirt and Bramley 1985, Weeks and Ashkanasy 1985, Srikanthan et al. 1989, Litchy and Karlinger 1990) he also notes that “The accuracy and precision of the predictions obtained from the resulting parameter – catchment characteristics relationships varies from good to very poor. This raises questions about the utility of the procedures used, the impacts of model and data error, and the extent of our knowledge of and capability to represent the link between catchment geomorphology and catchment response to rainfall.” Regionalisation is an area of ongoing research.

Several studies have attempted to relate the parameters of complex models (ie. models which contain many parameters) to landscape attributes. These include the studies of James (1972), Weeks and Ashkanasy (1985), Srikanthan et al. (1989), Braun and Renner (1992), and Chiew and McMahon (1994). While these studies display some success at deriving relationships for a limited subset of the model parameters, they generally found it impossible to derive relationships for all of the models parameters.

Similar regionalisation studies using simpler models have demonstrated a greater level of success (Jarboe and Haan 1974, Magette et al. 1976, Boughton 1984, Servat and Dezetter 1993, Post 1996, Post and Jakeman 1996). Thus it seems that simpler models (with fewer parameters) demonstrate much more potential, in terms of the regionalisation of model parameters, than complex models.

2.6 Selection of a rainfall-runoff model

In order to address the central problem of this thesis, namely, how climate change influences streamflow, a suitable rainfall-runoff model must be selected. The criteria which the required model should satisfy include:

1. the ability to perform satisfactorily in various locations and hydroclimatologies;
2. that data required by the model be widely available; and
3. linkages to the atmosphere, and hence the climate system, be explicit within the model.

While none of these requirements are individually restrictive, in combination they rule out the vast majority of currently available rainfall-runoff models because of parameter identifiability problems, as well as data collection and application problems.

The parameter identifiability problem typically applies to complex conceptual and physically-based models where the model complexity exceeds the information content of the available data (Beck et al. 1990, Jakeman and Hornberger 1993). This problem manifests itself in the inability to obtain a unique optimal set of model parameters. Hence, given that alternative combinations of parameter values can yield equivalent model performance, there can be little confidence in the process representations embodied in any particular parameter set.

Given that precipitation, streamflow and temperature are generally the only widely available data, how many parameters can a model contain without being considered

overparameterised? Beven (1989) comments “There is a great danger of overparameterisation if it is attempted to simulate all hydrological processes thought to be relevant, and fit those parameters by optimisation against an observed discharge record... It appears that three to five parameters should be sufficient to reproduce most of the information in a hydrological record.” Jakeman and Hornberger (1993) state that permissible model complexity is generally low, “containing around half a dozen parameters”.

Physically-based models attempt to avoid much of this parameter identifiability problem by including parameters that relate directly to physical quantities that at least in principle can be measured. Unfortunately, almost all observations of relevant variables are point observations, and hence need to be scaled-up to an appropriate effective grid value to be used as a parameter in these models. This scaling-up requirement is extremely problematical due to the spatial heterogeneity described in section 2.1.4. It is rarely the case that all required measurements for a physically based model are actually taken and in practice physically-based models must still be calibrated. Abbott et al. (1986a), in reference to the SHE physically-based rainfall-runoff model, note that “Problems such as inadequate representation of the hydrological processes and the possible difference in scale between the measurement and the model grid scale mean that some calibration is likely to continue to be required.”

It seems clear that an appropriate model for the work in this thesis cannot be solely physically-based and should contain only a small number of parameters. The hybrid metric-conceptual rainfall-runoff model, IHACRES, presented in Jakeman et al. (1990) was chosen for this work and is described in section 2.7. It does not, however, deal explicitly with atmospheric linkages and further model development was undertaken to incorporate these, as described in Chapter 4.

2.7 The IHACRES rainfall-runoff model

The IHACRES model is a lumped conceptual model, which uses observations (the metric paradigm) to examine hypotheses about the structure of component hydrological stores and hence can be considered a hybrid metric-conceptual rainfall-runoff model. IHACRES undertakes identification of hydrographs and component flows purely from rainfall, temperature and streamflow data (Jakeman et al. 1990, Jakeman et al. 1993, Jakeman et al. 1994). The IHACRES module structure consists of a non-linear loss module, which converts observed rainfall to effective rainfall, and a linear streamflow routing module, which extends the concept from unit hydrograph theory that the relationship between rainfall excess and total streamflow (not just quick flow) is conservative and linear. This structure is shown in Figure 2.2.

The linear module allows any configuration of stores in parallel or series. From the application of IHACRES to many catchments it has been found that the best configuration is generally two stores in parallel, except in semi-arid regions or for ephemeral streams where often one store is sufficient (Ye et al. 1997). In the two-store configuration, at time step k , quickflow, $x_k^{(q)}$, and slowflow, $x_k^{(s)}$, combine additively to yield streamflow (discharge), q_k :

$$q_k = x_k^{(q)} + x_k^{(s)} \quad (2.2)$$

with

$$x_k^{(q)} = -\alpha_q x_{k-1}^{(q)} + \beta_q U_k \quad (2.3)$$

$$x_k^{(s)} = -\alpha_s x_{k-1}^{(s)} + \beta_s U_k \quad (2.4)$$

where U_k is the effective rainfall. The parameters α_q and α_s can be expressed as time constants for the quick and slow flow stores, respectively:

$$\tau_q = -\Delta / \ln(-\alpha_q) \quad (2.5)$$

$$\tau_s = -\Delta / \ln(-\alpha_s) \quad (2.6)$$

where Δ is the time step (taken to be daily for this work).

Parameters expressing the relative volumes of quick and slow flow, V_q and V_s , can also be calculated:

$$V_q = 1 - V_s = \frac{\beta_q}{1 + \alpha_q} = 1 - \frac{\beta_s}{1 + \alpha_s} \quad (2.7)$$

The non-linear loss module accounts in principle for all losses from the catchment as described in section 2.1.1. It provides the effective rainfall, U_k , which is input into equations 2.3 and 2.4. The basic form of the non-linear module is:

$$\tau_w(t_k) = \tau_w \exp[(20 - t_k) f] \quad (2.8)$$

$$s_k = \frac{r_k}{c} + \left(1 - \frac{1}{\tau_w(t_k)}\right) s_{k-1} \quad (2.9)$$

$$U_k = s_k r_k \quad (2.10)$$

where r_k and t_k are the rainfall and temperature time series respectively; $\tau_w(t_k)$ is a time constant which is inversely related to the rate at which catchment wetness declines at temperature t_k ; the parameter τ_w is the value of $\tau_w(t_k)$ at 20°C; the parameter f is a temperature modulation factor on the rate of evapotranspiration, with units of °C⁻¹; s_k is a catchment storage (or wetness) index, which is dimensionless; and c is a parameter chosen to constrain the volume of the effective rainfall to equal runoff over the calibration period and has the same units as rainfall.

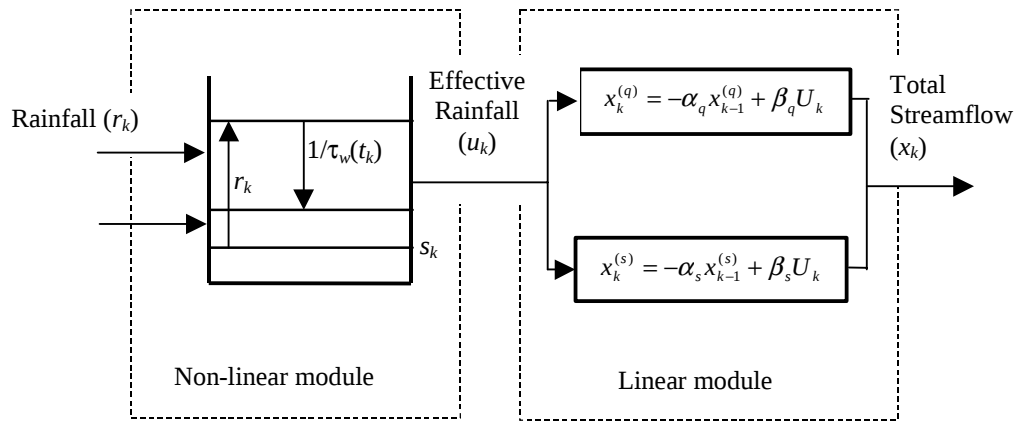


Figure 2.2 Structure of the IHACRES model

Chapter 3

Evapotranspiration: Modelling and Measurement

3.1 Introduction

Evaporation is one of the most important processes occurring at the land-surface and effecting the atmosphere and hence, climate. It plays the dual roles of providing water vapour to the atmosphere as well as absorbing incident radiation in the processes. These roles make it a key processes in both the water budget and the energy budget as well as providing a strong link between them.

In order to perform studies of the hydrologic effects of climate change in data sparse areas it is clear that models with specific characteristics are required. In particular a land surface hydrology model which explicitly predicts streamflow and evaporation on appropriate time scales, and which requires only widely available data is desirable. Evaporation here refers to the total water within a catchment that evaporates from surfaces or is transpired by vegetation and is often referred to as evapotranspiration (ET). Additionally, in order to keep the parameter identification problem to a minimum the total number of parameters in the model must be relatively small. These two requirements eliminate the vast majority (if not all) of the currently available land surface hydrology models.

Appropriate time scales for streamflow and ET prediction can be determined by a combination of the time scale of the available observational data and the characteristics of the particular research question being asked. Streamflow data are frequently available on a daily basis, less so on a finer time step, and this is the maximum time step required to capture the water balance dynamics of a catchment, as such streamflow is dealt with almost exclusively on a daily basis throughout this thesis. Atmospheric variables (including ET) are often measured on much shorter time scales, since diurnal changes in these variables can be quite large. While the majority of this thesis deals with ET prediction on a daily basis, an extra requirement placed on the land surface hydrology model is that it be able to perform prediction of streamflow and ET using different time steps. This allows the utilisation of sub-daily atmospheric observations when available.

It was necessary to create a new land surface hydrology model in order to fulfil all of the above requirements. This new model, presented in chapter 4, uses the linear streamflow routing module of the IHACRES rainfall-runoff model presented in section 2.6. A non-linear loss module that deals explicitly with ET was then developed (Evans and Jakeman 1998). This Chapter presents a review of ET modelling and measurement techniques, and provides perspective for the model development given in the subsequent chapter.

3.2 Modelling evapotranspiration

Human beings have long observed evaporation of water in their surroundings, and speculated on the nature of this phenomenon. Various writings about evaporation can be found going as far back as Greek antiquity (~8th century BC). In this section a review of modern ET modelling techniques is presented. Much of the development of these techniques has been driven/hindered by the availability of field data, as such section 3.3 gives an overview of current ET measurement techniques. Both the modelling and measurement techniques are subject to ongoing research with even the most advanced current techniques having recognised limitations. In fact, ET is generally modelled and measured indirectly. As commented by Oliver (1985), ET “is the most difficult component of the water balance to determine with any accuracy”.

3.2.1 *The water budget approach*

Estimation of ET can be done in several ways. Two rather obvious ways are to consider it the unknown term in either the water budget equation or the energy budget equation. In any given system at the Earth’s surface, ET is the connecting link between the water budget and the energy budget.

The water budget equation, which expresses the conservation of mass in a lumped or averaged hydrological system, can be written as

$$(P - E)A - Q_s - Q_g = dS/dt \quad (3.1)$$

where P is the mean rate of precipitation on the system, E the rate of evapotranspiration, A the surface area, Q_s the net outflow of surface water, Q_g the net outflow of ground water and S the water volume stored in the system. Solving equation 3.1 for E is not generally practical as unavoidable errors in measuring precipitation and runoff, and the difficulty in measuring groundwater contributions or a change in the storage, can often produce large errors in the resulting evaporation.

3.2.2 The energy budget approach

For a simple lumped system, when effects of unsteadiness in the atmosphere, ice melt, photosynthesis and lateral advection can be neglected, the energy budget is defined by

$$R_{net} = L_e E + H_a + G \quad (3.2)$$

where R_{net} is the net incoming radiation, L_e is the latent heat of evaporation, H_a the specific flux of sensible heat into the atmosphere and G the specific flux of heat conducted into the earth. The energy of the incoming radiation is partitioned at the earth's surface into net long wave back radiation, upward thermal conduction and convection of sensible heat, evaporation of water, and downward conduction of heat into the earth (Brutsaert 1982). ET as a latent heat flux plays a crucial role in heating the atmosphere and hence governing the weather and climate. Unfortunately, using equation 3.2 to determine E is not often done directly, since the assumptions made above mean it often cannot be applied and measurement of the other terms involve significant problems. For example, G depends on land cover, soil type and albedo, and the extrapolation from point measurements to areal values is very difficult.

Since the main objective here is the determination of the rate of evaporation E , it is convenient to rewrite equation 3.2 as

$$L_e E + H_a = Q_n \quad (3.3)$$

where Q_n is defined as the available energy flux density. In hydrological applications it is common practice to express the specific energy fluxes as equivalent rates of evaporation; equation 3.3 is then written as

$$E + H_e = Q_{ne} \quad (3.4)$$

where $H_e = H_a/L_e$ and $Q_{ne} = Q_n/L_e$.

The sensible heat flux H is closely related to the rate of evapotranspiration E , and these variables are often treated together. The ratio of these two flux terms, called the Bowen ratio,

$$B_o = H_a / L_e E \quad (3.5)$$

is a useful meteorological and climatological parameter providing a measure of the distribution of the available radiant energy between sensible and latent heat.

3.2.3 The aerodynamic approach

Aerodynamic approaches consider turbulence of the lower atmosphere to be the prime driving force in the removal of water vapour from the surface. The foundations of present aerodynamic theories probably began with the publication of Dalton's paper in 1802. Dalton (1802) explained his findings, "...in short, the evaporating force must be universally equal to that of the water, diminished by that already existing in the

atmosphere.” When the evaporating force is the same in different cases, the different rates of evaporation are “...regulated solely by the force of the wind.” That is, the ET rate is determined only by the vapour pressure deficit of the air and the ability of the air to carry the water vapour away from the surface, given as a function of wind speed. From these comments, it follows that Dalton’s results can be written in present day notation as

$$E = f(\bar{u})(e_s^* - e_a) \quad (3.6)$$

where e_s^* is the saturation vapour pressure at the temperature of the water surface, e_a the vapour pressure in the near surface atmosphere and $f(\bar{u})$ is a function of the mean wind speed $\bar{u}$. To this day there is no generally accepted wind function, though many have been proposed.

A summary of various aerodynamic based ET equations can be found in Singh and Xu (1997), this includes various versions of $f(\bar{u})$ found in the literature. They refer to these equations as mass-transfer based and present a generalised equation that encompasses the 13 other approaches presented. They find that these equations are particularly sensitive to errors in vapour pressure data. Other applications of aerodynamic ET equations can be found in Flerchinger et al. (1996).

A further generalisation of the aerodynamic approach involves modelling of the PBL. Here the lower atmosphere convective processes (PBL) are considered to interact with the surface fluxes in an intimate fashion. Further discussion along these lines can be found in de Bruin (1987), Cleugh (1991), Garratt (1992), and Raupach and Finnigan (1995).

3.2.4 Methods for potential evapotranspiration

The concept of potential evaporation (or potential ET) is widely used as a hydrological parameter even though a multiplicity of “definitions” exist as pointed out by Granger

(1989b). It was originally introduced by Thornthwaite (1948) as a climate index, but it has been used more widely as an index from which the actual evaporation could be estimated. In general, it is used to represent the *maximum* ET that would occur under current atmospheric conditions given no water supply limitations at the surface. This definition is somewhat vague though since it says nothing about the surface variables, which have a significant impact on evaporation such as surface temperature and energy supply. Granger (1989b) specified five definitions of potential evaporation (PE) that are outlined in Table 3.1

Table 3.1 Definitions of potential evaporation

Name	Definition	Calculation	Common usage
<i>PE1</i>	Evaporation rate which would occur if the surface was brought to saturation	Indeterminate	“definition” of potential evaporation
<i>PE2</i>	Evaporation rate which would occur if the surface was brought to saturation and the energy supply to the surface was held constant	Indeterminate; regression with solar radiation (Priestley and Taylor 1972)	Wet-environment or equilibrium evaporation; advection free evaporation
<i>PE3</i>	Evaporation rate which would occur if the surface was brought to saturation and the atmospheric parameters and the energy supply to the surface were held constant	Energy balance, vapour transfer equations	Penman potential; wet-surface evaporation
<i>PE4</i>	Evaporation rate which would occur if the surface was brought to saturation and the energy supply and the surface temperature were held constant	Indeterminate	Not used
<i>EP5</i>	Evaporation rate which would occur if the surface was brought to saturation and the atmospheric parameters and the surface temperature were held constant	Vapour transfer equation (if surface temperature is known)	Potential (Van Bavel 1966)

3.2.4.1 The Penman combination equation

Many techniques for calculating evaporation from the land surface are based on an energy budget approach, often using the Bowen ratio, or they are based on an aerodynamic approach. A major drawback of either of these methods is that they involve taking measurements of various atmospheric variables such as vapour pressure at a number of heights in the atmosphere. This is obviously impractical outside specific experimental situations.

Penman (1948) combined these approaches into a well-accepted model for the evaporation from open water.

$$E = \frac{\Delta}{\Delta + \gamma} Q_{ne} + \frac{\gamma}{\Delta + \gamma} E_A \quad (3.7)$$

where $\Delta = (de^*/dT)$ is the slope of the saturation water vapour pressure curve, $e^* = e^*(T)$ and is given by the Clausius-Clapeyron equation, γ is the psychrometric constant (which actually varies with temperature and pressure) and is given by

$$\gamma = \frac{c_p p}{0.622 L_e} \quad (3.8)$$

where c_p is the specific heat capacity for constant pressure and p is the pressure.

In addition, E_A , a drying power of the air, is defined by

$$E_A = f(\bar{u})(e_a^* - \bar{e}_a) \quad (3.9)$$

where $\bar{e}_a$ is the mean vapour pressure in the air, T_a is the air temperature, $e_a^* = e^*(T_a)$ the corresponding saturation vapour pressure.

Equation 3.7 has been widely used, but there is still no generally accepted way to formulate $f(\bar{u})$ the wind function in E_A . The simplest approach consists of using an empirical wind function. Penman (1948) originally proposed the equation

$$f(\bar{u}) = 0.26(1 + 0.54\bar{u}) \quad (3.10)$$

where $\bar{u}$ is the mean wind speed at 2 m above the surface in ms^{-1} , and the constant requires that E_A be in mm day^{-1} and the vapour pressure in mb.

From a practical point of view, the main feature of equation 3.7 is that it requires measurements of mean vapour pressure, wind speed and temperature at one level only. For this reason, it is very useful when measurements at several levels are unavailable or impractical (i.e. most of the time). It is important to remember that the equation derived by Penman was intended for an open water surface. It has also been applied over well-watered land where it is considered to represent the *potential evaporation*.

Penman assumed that in order to go from potential ET to actual ET some account must be taken of the factors limiting water availability at the surface, i.e. factors affecting the transpiration of vegetation as well as the capillary rise of water through the soil.

The Penman relationship for potential evaporation has also been modified to include the influence of atmospheric stability on turbulent transport of water vapour (Federer 1970, Mahrt and Ek 1984).

3.2.4.2 Equilibrium evaporation

The two-term structure of equation 3.7 suggests an interpretation that may serve as an aid in understanding the effect of regional or large-scale advection. When the air has been in contact with a wet surface over a very long fetch, it may tend to become vapour saturated, so that E_A should tend to zero. Accordingly, Slayter and McIlroy (1967) reasoned that the first term of equation 3.7 may be considered to represent a lower limit to evaporation from moist surfaces, which they referred to as *equilibrium evaporation*. Thus, by definition it is written as

$$E_e = \frac{\Delta Q_{ne}}{\Delta + \gamma} \quad (3.11)$$

The second term of equation 3.7 can be interpreted as a measure of the departure from equilibrium in the atmosphere. In the absence of cloud condensation or radiative divergence, this departure would stem from large-scale or regional advection effects, involving horizontal variation of surface or atmospheric conditions (i.e. strong winds, weather fronts etc.).

3.2.4.3 Average conditions of minimal advection

In reality equilibrium conditions practically never occur. This idea though has led to further developments. Priestley and Taylor (1972) have taken equilibrium evaporation as the basis for an empirical relationship giving evaporation from a wet surface under conditions of minimal advection, E_{PT} . They analysed data obtained over ocean and saturated land surfaces in terms of a constant quantity α , defined by

$$E_{PT} = \alpha \frac{\Delta Q_{ne}}{\Delta + \gamma} = \alpha E_e \quad (3.12)$$

Priestley and Taylor (1972) reported an average value for α (over pasture) of 1.26. Many others have since reported various values for α . Over pasture α can range from just over 1 up to 1.3, while for forest it can range from around 0.7 up to just over 1. A table of various values for α which have been reported in the literature can be found in

Spittlehouse (1989). Applications of the Priestly-Taylor approach can be found in McNaughton and Spriggs (1987).

3.2.5 Methods for Actual Evapotranspiration

Field studies have established that ET from bare soil or grassed surfaces proceeds in at least two distinct stages (Idso et al. 1974, Brutsaert and Chen 1995). In the first-stage, ET rate (often referred to as energy limited, atmosphere limited or potential) is governed by the ability of the atmosphere to supply enough energy to vaporise the water and to diffuse the vapour away from the surface. The second stage (often referred to as soil-moisture limited) is characterised by falling ET rates and is limited by moisture diffusion within the soil. As such, actual ET is often approximated by using an estimate of potential ET for the first stage, and then adjusting this with a soil-moisture dependent term for the second stage. This was done for instance, in Eagleson (1978), Kotoda (1989), and Salvucci (1997).

3.2.5.1 Penman's approach extended to nonsaturated surfaces

Granger and Gray (1989) use a derivation similar to that used by Penman to obtain a general combination equation which describes evaporation from nonsaturated surfaces. Penman's derivation relied on the fact that the relationship between surface temperature and surface vapour pressure is unique and known for a saturated surface. This is not the case for unsaturated surfaces. To overcome this problem they define the 'relative evaporation', J , as the ratio of actual to potential evaporation using aerodynamic considerations (equation 3.6)

$$J = \frac{E}{E_p} = \frac{f(u)(e_s - e_a)}{f(u)(e_s^* - e_a)} \quad (3.13)$$

The equation for evaporation is then given in terms of the relative evaporation as

$$E = \frac{\Delta J Q}{\Delta J + \gamma} + \frac{\gamma E_a}{\Delta J + \gamma} \quad (3.14)$$

Equation 3.14 is very similar to Penman's equation (equation 3.7) but differs due to the inclusion of the relative evaporation.

In order to remove the dependence of J on surface variables, Granger and Gray relate J to another ratio they call the 'relative drying power', D , which is given by

$$D = \frac{E_a}{E_a + Q} \quad (3.15)$$

They go on to claim that the relationship between J and D is single valued and independent of land use. This claim seems somewhat dubious given the scatter of results they report and the limited number of sites that were used for verification. For the two Canadian sites which they used the relationship was reported as

$$J = \frac{1}{1 + 0.028 \exp(8.045D)} \quad (3.16)$$

3.2.5.2 Adjustment of Penman's approach with bulk stomatal resistance

Even when well supplied with water near the soil surface or at the roots, transpiring vegetation cannot, in general, be considered wet, except after rainfall or dew formation. Therefore, equations such as Penman's are no longer applicable. To remedy this Penman and Schofield (1951) and later, more formally, Monteith (1965), Thom (1972) and others, introduced various resistance parameters to characterise the transfer between the supposedly vapour-saturated stomatal cavities and the atmosphere.

Monteith used two resistance parameters. The *aerodynamic resistance* r_{la} , which represents the aerodynamic resistance to the transfer of sensible heat and water vapour

from the surface to a reference level, and the *surface resistance* or *bulk stomatal resistance* r_s , which describes the resistance to the flow of water vapour from the evaporating surface to the immediate atmosphere. It includes such processes as the capillary action of water through soil, moisture stress at the roots of vegetation, biological and physical factors controlling the opening and closing of stomata and the spatial distribution of vegetation species within the area under consideration. In practice, the surface resistance is often determined in an empirical fashion due to the complexity contained within this one parameter.

These resistances can be incorporated into the Penman approach as follows. When the vegetation is not actually wetted, the surface vapour pressure e_s is not equal to e_s^* . However, they can be related using

$$(\bar{e}_s - \bar{e}_a) = \left(\frac{r_{la}}{r_s + r_{la}} \right) (\bar{e}_s^* - \bar{e}_a) \quad (3.17)$$

and the Penman-Monteith equation is given by

$$E = \frac{\Delta Q_{ne} + \rho c_p (e_a^* - \bar{e}_a) / (L_e r_{la})}{\Delta + \gamma(1 + r_s / r_{la})} \quad (3.18)$$

where ρ is the density of the air. Clearly, when $r_s=0$, equation 3.18 reduces to an expression which is equivalent with Penman's.

This approach when used to calculate areal ET assumes the postulate by (Monteith 1965) that a canopy acts as a 'big leaf' with a canopy conductance equal to the parallel sum of the leaf stomatal conductances. This postulate and the scaling assumptions it invokes have been an issue for several decades. Raupach (1995) provides further investigation into this scaling issue, providing a flux-matching scheme for scaling from leaf to canopy.

While applying the Penman-Monteith equation, Stewart (1989) notes that “if a sub-model of the response of surface resistance to environmental conditions is to be biologically realistic, the time interval should not be greater than one hour. As a result of this constraint the Penman-Monteith equation is preferably used either with hourly or half-hourly meteorological data.” Further applications of the Penman-Monteith equation can be found in Stewart and Gay (1989).

Further developments of this approach include explicitly modelling two canopy layers as well as including canopy interception and subsequent evaporation from the wet canopy. Famiglietti and Wood (1995) provide an example which explicitly models evaporation from the bare ground, evaporation from vegetation with a wet canopy as well as evaporation from vegetation with a dry canopy (transpiration). Biftu and Gan (2000) used their two-source model, which modelled the bare ground and the canopy above, along with below canopy interactions, in order to calculate the evaporation.

Some studies have attempted to estimate areal evaporation by combining a Penman-Monteith canopy evaporation model with a boundary layer growth model (or PBL). Possibly the first such model was developed by Brutsaert and Mawdsley (1976) who created regional evaporation predictions for the United States based on upper atmospheric data from the U.S. radiosonde network. Other studies along these lines have been conducted by de Bruin (1983), McNaughton and Spriggs (1986) and Cleugh (1991). A common finding of these studies is that errors in modelling the mixed layer depth are unimportant for modelling evaporation. Despite this, de Bruin (1987) goes so far as to suggest that a PBL parameterisation is necessary for the prediction of areal evaporation.

3.2.5.3 Complementary relationship between actual and potential evaporation

In a move away from the Penman belief that when the water supply is not abundant actual evaporation would be proportional to the potential evaporation and a function of the water availability, Bouchet (1963) uses the same measure of potential evaporation as a negative indication of the evaporation which is actually occurring. He essentially

views potential evaporation in two ways, first, similarly to Penman, it is the energy available for evaporation under conditions of an abundant supply of water, and second, unlike Penman, as a negative index of actual evaporation when supply is limiting.

Bouchet arrived at the complementary relationship , shown in Figure 3.1, between the potential evaporation E_p and the actual regional evaporation E :

$$E_p + E = 2E_{p0} \quad (3.19)$$

The actual regional evaporation rate E is the average value from a large uniform surface of regional size. The potential evaporation E_p is the evaporation that would take place under the prevailing atmospheric conditions if only the available energy were the limiting factor. Under conditions when E equals E_p , it is denoted by E_{p0} . The derivation of equation 3.19 may be given as follows.

If for one or another reason, independent from the available energy, E decreases below E_{p0} , a certain amount of energy becomes available, that is

$$E_{p0} - E = q \quad (3.20)$$

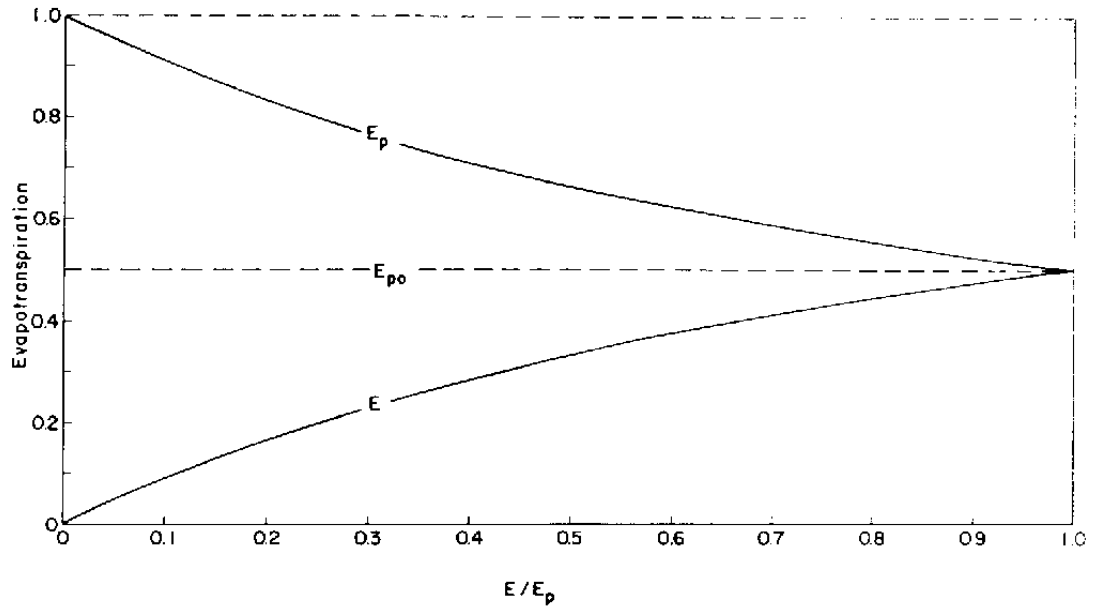


Figure 3.1 Sketch illustrating Bouchet's (1963) hypothesis; E and E_p are plotted versus E/E_p such that $E + E_p = \text{constant}$ as given in 3.19. From Brutsaert (1982).

At the scale of typical regions, this decrease of E , with respect to E_{p0} , should have a relatively small impact on the net radiation, affecting primarily the temperature, the humidity and the turbulence of air near the ground. As a result, this available energy flux q causes an increase in E_p . The main hypothesis of Bouchet (1963) was that in the absence of local oasis effects the energy budget remains otherwise unaffected, and the potential evaporation is increased by exactly that amount, or

$$E_p = E_{p0} + q \quad (3.21)$$

The combination of equations 3.20 and 3.21 produces the complementary relationship shown in equation 3.19. So using the complementary relationship, if E_p is known (say using Penman's equation) and E_{p0} is known (Morton (1976, 1983) used E_{PT} of Priestley and Taylor) then the actual evaporation can be calculated

Granger (1989a) used the Penman equation to represent E_{p0} , and he defined E_p in a manner similar to that of Van Bavel (1966). This choice of definitions allows both parameters (E_{p0} and E_p) to be derived from energy balance and mass transfer equations and to be expressed in terms of appropriate vapour pressure gradients. Using a

development which parallels Bouchet (1963), the general form of the relationship between E , E_p and E_{p0} is then derived. The resulting equation, which replaces equation 3.19, is

$$E_p \frac{\gamma}{\Delta} + E = E_{p0} \left(1 + \frac{\gamma}{\Delta} \right) \quad (3.22)$$

Equation 3.22 is not symmetrical, the changes in actual and potential evaporation are given by $\partial E / \partial E_p = -\gamma / \Delta$, whereas Bouchet assumed $\partial E / \partial E_p = -1$. Substituting the Penman equation into this new complementary relationship and introducing Granger's relative evaporation ($J = E / E_p$) results in a general equation for evaporation from nonsaturated surfaces; this equation is identical to equation 3.14 which was obtained using a development similar to that of Penman (1948). Granger (1989a) and Granger and Gray (1989) thus concludes that both Penman's approach and the complementary relationship are indeed compatible, as is also asserted by Nash (1989).

Brutsaert and Stricker (1979) proposed the 'advection-aridity approach' which uses the complementary relationship in a fashion similar to Morton (1990). Qualls and Gultekin (1997) tested this advection-aridity approach using data from a flat, semi-arid grassland and found a large bias between measured and predicted ET. They claim that 'external energy sources other than net radiation are a potential source of this bias'. Parlange and Katul (1992) extended the advection-aridity approach to include effects of atmospheric stability and found that their method agreed well with observations from a large, sensitive lysimeter.

A study by McNaughton and Spriggs (1987) found that "a basic assumption of complementary relationship theory, that the interaction between the convective boundary layer (or PBL) and the rest of the atmosphere does not depend on the surface energy balance, is not even approximately satisfied in the real atmosphere." This and similar concerns have limited the use of the complementary relationship.

3.2.6 Empirical approaches

Due to the data requirements of the methods presented in the previous sections many, less data intensive, empirical methods have been developed for practical use. These approaches almost invariably employ some temperature function. One of the most well known methods is that developed by Thornthwaite (1948). He correlated mean monthly temperature with ET as determined by the water balance for well-watered valleys. His equation for obtaining 30-day month values in centimetres is

$$E = 1.6L \left(\frac{10T}{J} \right)^a \quad (3.22)$$

where L is the daytime hours expressed in units of 12 hours, T is the mean monthly temperature ($^{\circ}\text{C}$), J is a heat index obtained by the summation of twelve monthly indices

$$J = \sum_{i=1}^{12} (T_i/5)^{1.514} \quad (3.21)$$

and the value of a is obtained from a cubic function of the heat index (J) as

$$a = 0.000000675J^3 - 0.0000771J^2 + 0.01792J + 0.49239 \quad (3.24)$$

Thornthwaite's method has been employed widely due to the advantage of only needing daylight hours, which can be calculated given latitude and time of year, and air temperature, even though the calculation is reasonably complex. Application of the Thornthwaite method can be seen in Alley (1984).

Simpler empirical equations have also been developed such as by Hamon (1961). He gives the formula

$$E = CL^2 P_T \quad (3.25)$$

where C is a constant chosen to be 0.55 and P_T is the saturated water vapour density (absolute humidity) at the daily mean temperature. Here, again, only daylight hours and mean temperature data (converted to saturation water vapour density) are required.

Other empirical methods that have been used widely include the Blaney-Criddle formula, the Christiansen method and Hargreaves method. Details of these ET formulations can be found in Jensen et al. (1990).

3.2.7 Comparisons of ET formulations

There have been several studies that compare the effectiveness of various ET estimation techniques. Rampisela et al. (1990) compared the performance of the Thornthwaite, Hamon, Penman and Penman-Monteith equations using a monthly time step, for a forested catchment in Japan. They concluded that the Penman-Monteith equation provided the best estimates of ET.

Crago and Brutsaert (1992) compared the Penman equation, equilibrium evaporation, minimal advection evaporation, advection-aridity evaporation and the Penman-Monteith approach. The comparison was performed over the grassland FIFE region of Kansas, USA. They found that using the equilibrium evaporation or the advection-aridity approach, which do not require information on the soil moisture availability, produced acceptable results. The other three techniques required soil moisture data and all performed significantly better.

Panu and Nguyen (1994) compared the performance of the Hargreaves, the Christiansen-Mehta and the Morton methods over four forested sites in Northwestern Ontario. They found that the Hargreaves method provided the best results.

Vardavas and Fountoulakis (1996) estimated lake evaporation using both the Penman and the Priestly-Taylor formulations for four Australian lakes. The lakes are in semi-arid tropical, warm temperate, Mediterranean and alpine climatic regions. They found both formulations worked well.

Barr et al. (1997) compared the Morton method, a modification of Penman's method and the Spittlehouse energy-limited (minimal advection) versus soil moisture-limited method within the SLURP hydrological model. The study area was the Kootenay Basin of eastern British Columbia, Canada. They found that the Spittlehouse method performed best and concluded that including "a soil moisture limitation to ET produced a worthwhile improvement in hydrological performance."

Xu and Singh (1998) compared four ET estimation methods with pan evaporation data collected in Changines, Switzerland. The ET methods compared were those developed by Thornthwaite (1948), Romanenko (1961), Penman (1948) and Turc (1961). They found that the Penman equation provided the best estimate of pan evaporation.

Biftu and Gan (2000) compared the Penman-Monteith (equation 3.18) and the modified Penman for non-saturated surfaces (equation 3.14) over a few land cover types. The study site was the Paddle River Basin, Alberta, Canada. The land cover types investigated were coniferous forest, mixed forest, agricultural and pasture. They found that the closed-canopy assumption of the Penman-Monteith approach is not valid for the mixed forest and pasture land covers and that the modified Penman model was preferred.

From these comparison studies two conclusions can be reached. First, ET estimation is significantly improved if soil moisture limitations are explicitly included in the ET formulation. Secondly, combination equation approaches such as Penman and Penman-Monteith, consistently produce the best ET estimates followed by related approaches such as that by Priestley and Taylor (1972).

3.3 Measuring Evapotranspiration

Measuring ET is itself a difficult problem with the currently best accepted methods (eddy correlation, Bowen ratio energy balance) being both expensive and labour intensive (Tapper 1996). Because of this problem, these data sets tend to have only small temporal extent.

The eddy correlation method allows the direct measurement of latent heat flux (and hence ET) by the correlation of fluctuations of vertical wind speed with fluctuations of vapour density. The main advantage of this method is that it measures the fluxes directly, i.e. the assumptions of similarity pertaining to the flux gradient approaches (e.g. Bowen ratio) are not required. The major disadvantages include fetch considerations, the high cost of the instruments and the fact that long-term *unattended* measurement is not viable due to the delicate sensors involved.

The Bowen ratio approach relies on the *similarity* principle. This holds that the atmospheric diffusion coefficients for heat, water vapour, momentum and carbon dioxide are equivalent. The approach then requires an accurate determination of temperature and vapour pressure at two levels in the atmosphere in order to determine the Bowen ratio (Tapper 1996). The net radiation and ground heat flux must also be measured so that the energy budget equation 3.2 can be utilised to calculate ET. This can be seen more readily if we combine equations 3.2 and 3.5 to give

$$L_e E = \frac{R_{net} - G}{1 + B_o} \quad (3.22)$$

While the Bowen ratio technique allows a direct measure of ET over relatively long-term measurement periods (compared to eddy correlation) it has several disadvantages. These include fetch considerations, an assumption of steady state conditions (known to be incorrect under certain conditions of atmospheric stability) and the fact that the accuracy is dependent upon the accuracy of the measurements of net radiation and

ground heat flux, which can lead to significant errors especially during low energy conditions.

The water balance approach can be applied at various scales. When applied to evaporation from small containers, such as pans or lysimeters, the measurements taken are of limited value due to two factors that make the evaporation unrepresentative. The first is the container itself, which has significant effects on radiation absorption and the production of turbulence. The second is that, at best, only low plant cover is included which is rarely representative of the surrounding area. On a larger scale, when looking at the water balance for lakes or river basins, realistic evaporation estimates for environmentally significant areas can be obtained, though this must be done on a longer timescale, e.g. monthly, seasonally or annually (Morton 1994).

When attempting to verify ET models, one is therefore forced to use either the short term data given by the eddy correlation or Bowen ratio methods above (e.g. days of hourly data) or longer term data obtained from a (catchment-scale) water balance approach (e.g. decades of yearly data). The type of data used for verification is usually determined by the time scale of interest, though it is worth noting that methods such as eddy correlation are point measurements, whereas the longer term water balance methods obtained using streamflow measurements are an integration of outputs from spatially varying runoff generation and transmission processes within a catchment.

Chapter 4

Development of CMD-IHACRES

This chapter describes the development of the rainfall-runoff model CMD-IHACRES. The model uses a semi-physical approach to produce estimates of streamflow and evapotranspiration at an appropriate time-step (usually daily in this dissertation). It was developed for use with limited data, that is the input data requirements were limited to precipitation, temperature and streamflow. The number of parameters was also kept to a minimum in order to maintain a sufficient level of parsimony within the model.

4.1 Model Description

The rainfall-ET-runoff model proposed here is based on the structure of the IHACRES metric/conceptual rainfall-runoff model shown in section 2.7. This model undertakes identification of hydrographs and component flows purely from rainfall, temperature and streamflow data (Jakeman et al. 1990, Jakeman and Hornberger 1993, Jakeman et al. 1994). The IHACRES module structure consists of a non-linear loss module, which converts observed rainfall to effective rainfall or rainfall excess (that share of rainfall that reaches the discharge measurement site), and a linear streamflow routing module, which extends the concept from unit hydrograph theory that the relationship between rainfall excess and total streamflow (not just quick flow) is conservative and linear.

The linear module used here is the same as that outlined in section 2.7; the main equations will be repeated here for completeness. The linear module allows any configuration of stores in parallel or series. From the application of IHACRES to many catchments it has been found that the best configuration is generally two stores in parallel, except in semi-arid regions or for ephemeral streams where often one store is sufficient (Ye et al. 1997). In the two-store configuration, at time step k , quickflow, $x_k^{(q)}$, and slowflow, $x_k^{(s)}$, combine additively to yield streamflow (discharge), q_k :

$$q_k = x_k^{(q)} + x_k^{(s)} \quad (4.1)$$

with

$$x_k^{(q)} = -\alpha_q x_{k-1}^{(q)} + \beta_q U_k \quad (4.2)$$

$$x_k^{(s)} = -\alpha_s x_{k-1}^{(s)} + \beta_s U_k \quad (4.3)$$

where U_k is the effective rainfall. The parameters α_q and α_s can be expressed as time constants for the quick and slow flow stores, respectively:

$$\tau_q = -\Delta / \ln(-\alpha_q) \quad (4.4)$$

$$\tau_s = -\Delta / \ln(-\alpha_s) \quad (4.5)$$

where Δ is the time step (daily here).

Parameters expressing the relative volumes of quick and slow flow can also be calculated:

$$V_q = 1 - V_s = \frac{\beta_q}{1 + \alpha_q} = 1 - \frac{\beta_s}{1 + \alpha_s} \quad (4.6)$$

The previous IHACRES loss module used a statistical approach to account for antecedent soil moisture conditions and ET losses. Here this module is replaced by a more physically based catchment moisture store accounting scheme which uses rainfall and temperature as inputs and provides ET and rainfall excess as outputs, creating the overall model structure shown in Figure 4.1.

The catchment moisture store accounting scheme calculates a Catchment Moisture Deficit at time step k , CMD_k , according to

$$CMD_k = CMD_{k-1} - P_k + E_k + D_k \quad (4.7)$$

where P is the precipitation, E is the ET loss and D is the discharge. CMD is zero when the catchment is saturated and increases as the catchment becomes progressively drier.

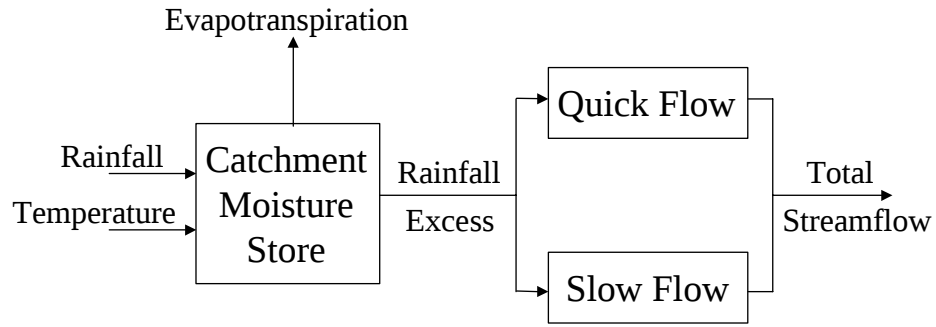


Figure 4.1 Structure of the rainfall-ET-runoff model.

Effective rainfall is calculated from

$$U_k = \begin{cases} D_k & CMD_k \geq 0 \\ D_k - CMD_k & CMD_k < 0 \end{cases} \quad (4.8)$$

Suitable parameterisations for both E_k and D_k were sought that minimised the number of parameters needed and for which the only data requirements are temperature, rainfall and streamflow. Several parameterisations were tried and the results for the simplest successful model are presented below.

Keeping in mind the need for a model that requires only widely available data, an attempt has been made to model ET using only temperature and precipitation. For ET modelling, techniques vary considerably in their relationship between ET and temperature, T . The effects of vegetation on ET have been represented in various ways, most notably by the incorporation of a 'surface resistance' in the Penman-Monteith equation. This surface resistance has itself been estimated in many ways, commonly with a dependence on the available soil moisture such as that given by Stewart (1989) where $ET = \exp[K(\delta\theta - \delta\theta_{\max})]$ in which $\delta\theta$ is the soil moisture and K is a constant. The parameterisation here is defined by

$$E_k = c_1 T_k \exp(-c_2 CMD_k) \quad (4.9)$$

Equation 4.9 has ET directly proportional to temperature and decreasing exponentially as CMD increases as can be seen in Figure 4.2. So higher temperatures will result in larger ET losses provided sufficient soil moisture is available. Using other atmospheric variables in its forcing when they are available may enhance the accuracy of the E_k term. Several techniques for this can be found in chapter two and further testing of the ET formulation can be found in section 4.2. It has been noted that “natural” controls on evaporation may constrain the modelled values to be reasonable even for simplistic evaporation functions (Wood et al. 1992).

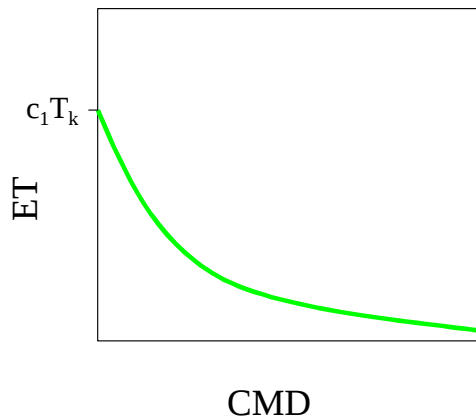


Figure 4.2 Illustration of the modelled relationship between evapotranspiration (ET) and catchment moisture deficit (CMD).

It was assumed that discharge is not temperature dependent. As with the modelling of ET, several parameterisations of discharge were investigated. One of the simplest investigated was

$$D_k = \begin{cases} \frac{-c_3}{c_4} CMD_k + c_3 & CMD_k < c_4 \\ 0 & CMD_k \geq c_4 \end{cases} \quad (4.10)$$

where c_3 and c_4 are non-negative constants. The discharge equation allows water to escape to the stream even when a moisture deficit exists within the catchment.

Parameter c_3 represents the maximum discharge that can occur whilst a moisture deficit exists. Parameter c_4 represents the maximum CMD that can occur before water ceases draining to the stream. This relationship is shown in Figure 4.3. To maximise simplicity and minimise computing time required, $[CMD_{k-1} - P_k]$ was used in place of CMD_k in the ET and discharge equations, along with the extra requirement that if $[CMD_{k-1} - P_k] < 0$ then it is considered as equal to zero.

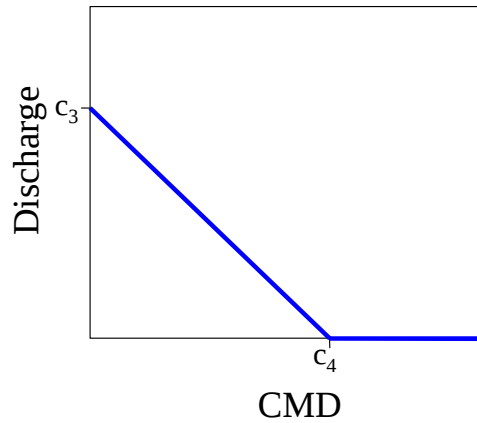


Figure 4.3 Illustration of the modelled relationship between Discharge (D) and catchment moisture deficit (CMD).

4.1.1 Performance assessment criteria and calibration procedure

Many statistical measures have been used to evaluate the performance of hydrological models in the past. Often these statistics are correlation based measures such as the Pearson's product correlation coefficient or its square, the coefficient of determination. However these measures have been found to be oversensitive to extreme values and insensitive to additive and proportional differences between model predictions and observations. In order to address this problem several studies have recommended the use of other statistical measures (ASCE 1993, Ewen and Parkin 1996, Refsgaard and Knudsen 1996, Legates and McCabe 1999). The recommendations include the use of graphical plots, the use of a relative error or goodness-of-fit measure (here the Nash-Sutcliffe efficiency (Nash and Sutcliffe 1970) is used) and the use of a measure of the absolute error (here the bias is used).

To measure the performance of the model estimate of streamflow, $\hat{q}_i$, two performance statistics are used: the bias (B) and the Nash-Sutcliffe efficiency (NSE). These are defined as

$$B = \frac{1}{n} \sum_{i=1}^n (q_i - \hat{q}_i) \quad (4.11)$$

$$NSE = 1 - \frac{\sum_{i=1}^N (q_i - \hat{q}_i)^2}{\sum_{i=1}^N (q_i - q_{mean})^2} \quad (4.12)$$

where q_i is the observed streamflow and q_{mean} is the mean of the observed streamflow.

Another assessment criteria that has been used previously is the average relative parameter error (ARPE), which is a by-product of the SRIV algorithm. A high ARPE

indicates that there is less confidence in the uniqueness of the estimated values of the linear parameters. More details about the calculation of the ARPE can be found in (Jakeman et al. 1990). In addition, cross correlation at lag 1 between the residuals and the modelled streamflow is considered. Large cross correlation values imply that there is a systematic mismatch between observed and model discharge recessions. Models with such characteristics are eliminated in our parameter estimation exercise.

Unfortunately, adequate ET data for the catchments investigated were generally not available so no direct performance measure of the ET estimate could be made. Only the FIFE site contained adequate data to analyse the ET formulation, this is given in section 4.2. It should be noted though that low bias in predicting streamflow indicates that losses from the catchment were overall accounted for well. The distribution of these losses from day-to-day is what remains in question.

The calibration procedure of CMD-IHACRES involves finding the global minimum of multivariate parameters. However, conceptual model objective functions often containing extraneous local minima which make it difficult, if not impossible, to obtain a unique set of optimal parameters for a conceptual model using an automatic parameter estimation procedure (Sorooshian and Gupta 1983, Sorooshian 1991). In this study, the entire sample space for the non-linear parameters, c_1 , c_2 , c_3 and c_4 , is sampled at fine scale manually. For each sample the effective rainfall time-series is calculated and SRIV estimation (Young 1985) is then applied to find the linear model parameters. Model output statistics are generated for each sample of the non-linear parameters. The optimum parameter set is then chosen as a trade-off between the assessment criteria with NSE being the major objective function to maximise during the calibration procedure.

4.2 Testing the ET formulation

The ET formulation given in equation 4.9 requires only temperature as input data, as well as continuous accounting of the catchment moisture store, in order to provide

estimates of ET. The question remains, is this formulation adequate to represent actual ET? It is clear from chapter two that many more variables and a much more complicated formulation is generally thought to be required. Here the performance of this formulation is tested against the performance of the widely used formulations of Penman (equation 3.7) and Priestly & Taylor (equation 3.12).

4.2.1 Site description

Due to the difficulties involved with measuring ET data (section 3.3) and the need for a range of atmospheric variables to be measured in order to apply the more complex ET formulations, only one site is used for this comparison. This is the First ISLSCP (International Satellite Land Surface Climatology Project) Field Experiment (FIFE) site. The FIFE project was designed to improve understanding of carbon and water cycles; to coordinate data collected by satellites, aircraft, and ground instruments; and to use satellites to measure these cycles. The FIFE site is located in the Konza prairie, south of Manhattan, Kansas. FIFE observations were made on a $15\text{km} \times 15\text{km}$ domain. (Betts and Ball 1998) averaged the surface meteorological and flux data to give a single time series representative of the FIFE site for the period May-October 1987. Within the FIFE site is the Kings Creek catchment, which is approximately 12km^2 in extent. Further description of the site can be found in section 6.2.2.



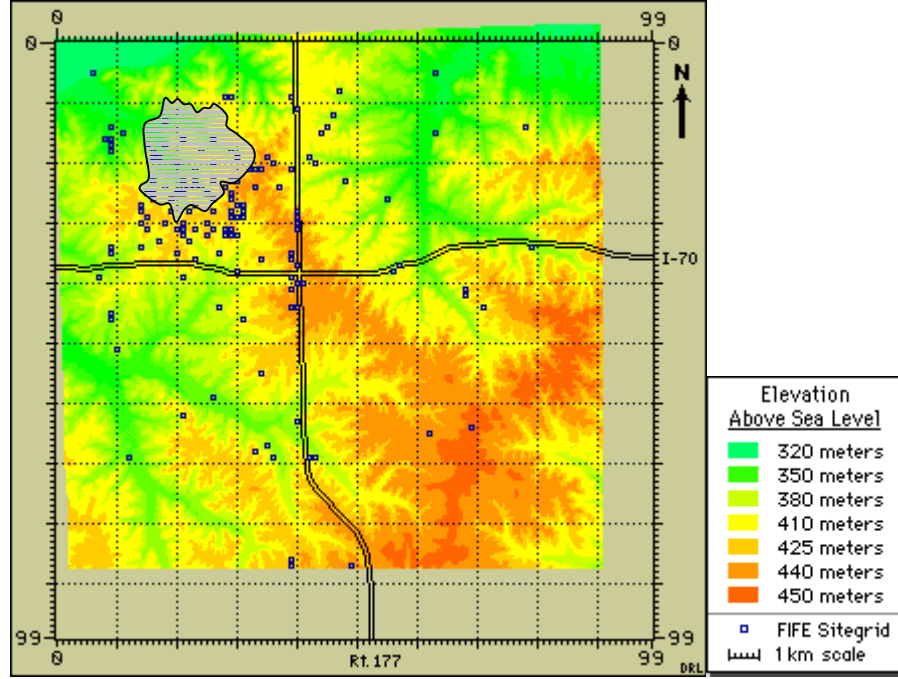


Figure 4.4 FIFE site showing approximate location of King's Creek catchment in the North-West corner (shaded area), ground measurement stations and the elevation.

4.2.2 Application of the *ET* formulations

Equation 4.9 can be thought of as an estimate of potential evaporation modified by a function of available soil moisture. Here temperature provides the surrogate for potential evaporation and the negative exponential function of CMD provides the soil moisture dependency. This perspective provides a simple way to incorporate the more complex formulations of potential evaporation and we can rewrite equations 3.31, 3.12 and 3.7 as

$$E_k = \Omega_k T_k \quad (4.13)$$

$$E_k = \Omega_k \frac{\Delta}{\Delta + \gamma} (R_n - G) \quad (4.14)$$

$$E_k = \Omega_k \left\{ \frac{\Delta}{\Delta + \gamma} (R_n - G) + \frac{\gamma}{\Delta + \gamma} f(u) (e_a^* - e_a) \right\} \quad (4.15)$$

where

$$f(u) = 0.26(1 + 0.54u) \quad (4.16)$$

$$\Omega_k = c_1 \exp(-c_2 CMD_{k-1}) \quad (4.17)$$

A version of CMD-IHACRES containing each of the ET formulations above was applied using FIFE data over the Kings Creek catchment for the 4.5 month period covering 28th May to the 15th October 1987. Here model calibration utilised both ET and streamflow data. The formulation given by the modified temperature approach of equation 4.13 requires only temperature data. The Priestly-Taylor formulation of equation 4.14 requires the temperature, net incident radiation and ground heat flux data. The Penman formulation of equation 4.15 requires all the same data as the Priestly-Taylor formulation plus the vapour pressure and the horizontal wind speed.

Since the model will be applied, in most cases, when no ET data are available, a short investigation into the implications of model calibration in the absence of this data was performed. This model, using the modified temperature approach of equation 4.13, was calibrated using both ET and streamflow data and then was calibrated again using only streamflow data. When calibrating using ET data the parameters in equation 4.17 are chosen by optimising the predicted ET time series while the rest of the parameters in the model are optimised against the streamflow data. The effects of this on both ET prediction and streamflow prediction were examined.

4.2.3 Results and discussion

Figure 4.5 shows the performance of each of the ET formulations given above. The overall worst performer is, not surprisingly, the modified temperature approach. This

simplest of approaches manages to give us the correct overall trend in the data though much of the detail is not captured, especially the day-to-day variance of ET. The Priestly-Taylor approach seems to over-estimate the variance in the ET data with consistent under and/or over prediction. However, it does come much closer than the modified temperature approach to the magnitude of this variance. The Penman approach seems to perform best most of the time although it does overestimate the ET late in the record when a few days of unusually strong winds occurred.

These results indicate that representing the fluctuations in ET on a daily basis requires knowledge of more atmospheric variables than just temperature. Thus, it would be wise to utilise as many of these variables as are measured at any particular site by choosing an appropriately complex ET formulation. Given that temperature is often the only atmospheric variable measured, using the modified temperature approach

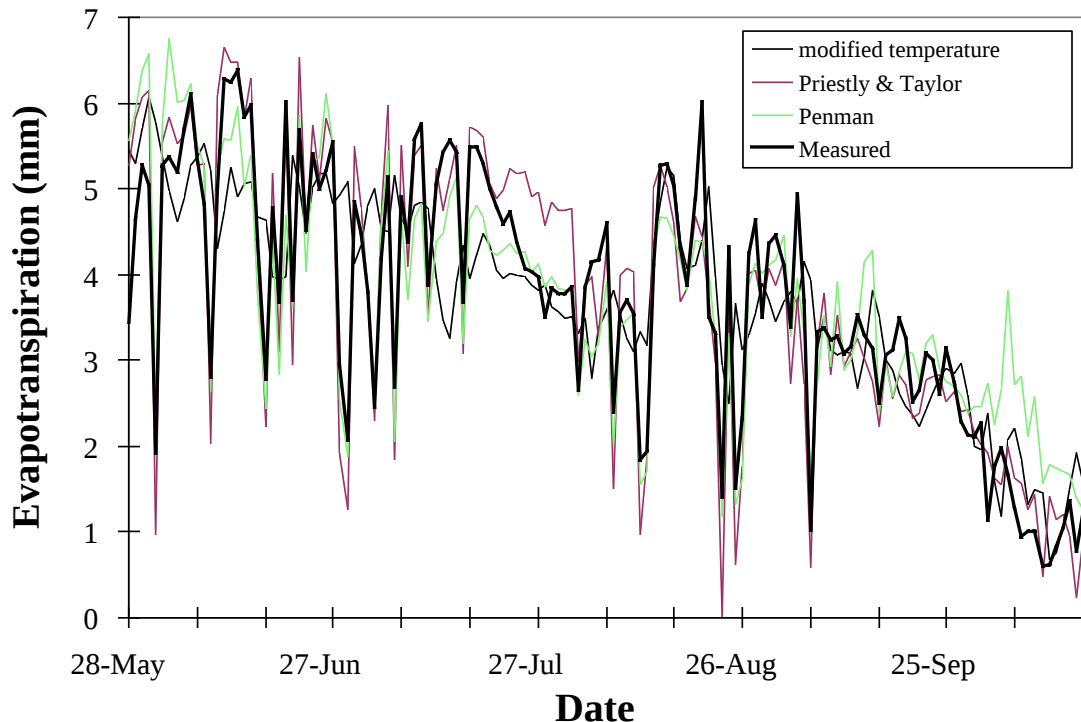


Figure 4.5 Performance of alternate formulations of ET over FIFE, 1987.

may be the only available option. It is important to be aware that while the trend and longer-term means (weekly to monthly) may be captured very well there are limitations in the daily values themselves.

The most common applications of the model will occur when no ET data are available. Figures 4.6 and 4.7 show the effects of this lack of data during the calibration process on ET and streamflow prediction respectively. Figure 4.6 displays large discrepancies between the ET predictions at the start of the record. The model calibrated using ET data performs best overall with the only major differences between ET predictions occurring during this early period. These discrepancies have all but vanished after the first 40 days. This could be explained, at least in part by the assumption that the catchment is saturated at the beginning (ie. $CMD = 0$) and the model takes a short while to reach an equilibrium. This assumption is unavoidable since the antecedent conditions are unknown.

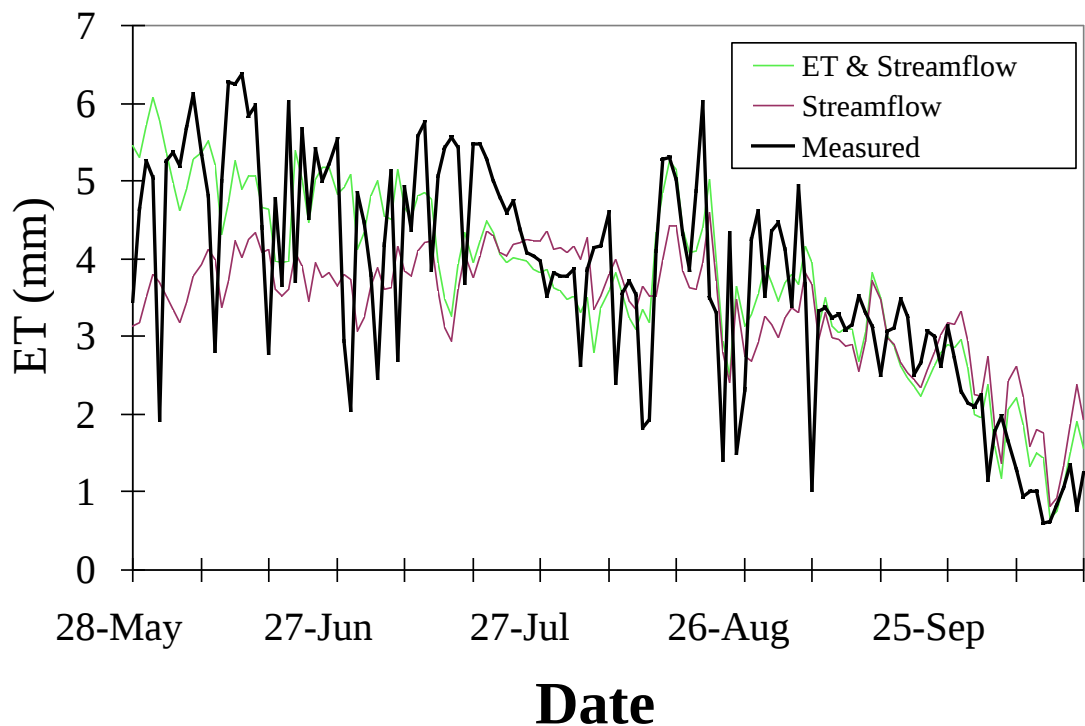


Figure 4.6 ET predicted by CMD-IHACRES using the modified temperature approach, when calibrated using: both ET and streamflow data; streamflow data alone.

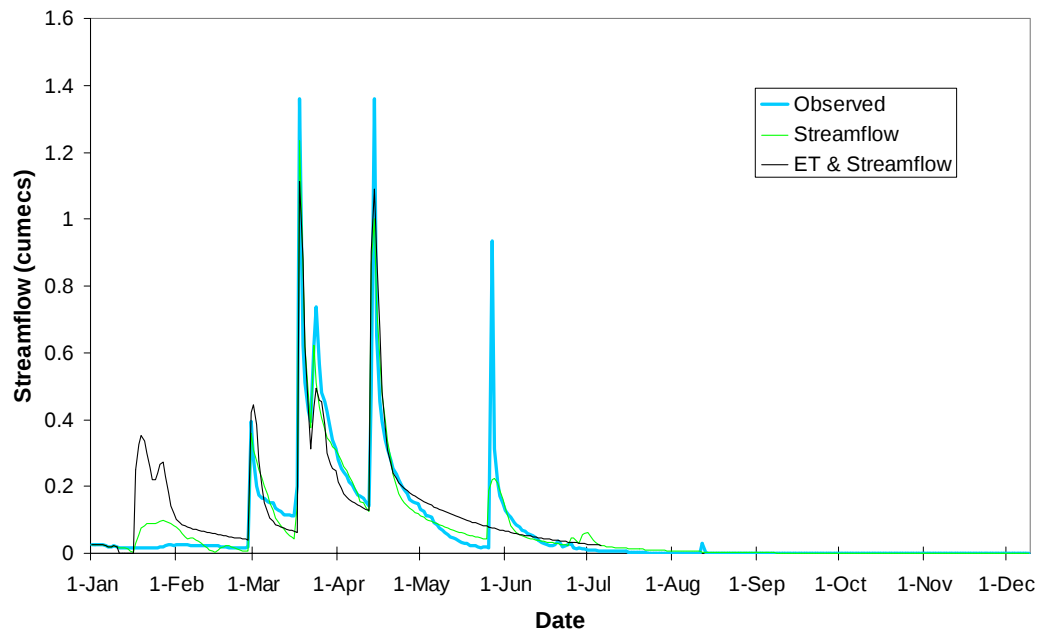


Figure 4.7 Streamflow predicted by CMD-IHACRES using the modified temperature approach, when calibrated using: streamflow data alone; both ET and streamflow data.

While this is by no means a comprehensive study it does imply that calibration using only streamflow will not bias the ET estimates substantially after some initial period, in this case around 40 days.

Calibration against streamflow data alone produces the best streamflow estimates as can be seen in Figure 4.7. Similarly to the case for ET, the major discrepancies occur at the beginning of the record though this time they last for about the first 20 days instead of 40. Also the model calibrated against both ET & streamflow misses the last peak altogether. Again, some of the early discrepancy may be related to the models reaching equilibrium with the antecedent conditions at different rates. Once this equilibrium is reached however, the models predict very similar streamflow.

In summary, accuracy of the ET estimation increases with both increases in the variety of data available and increases in the complexity of the formulation used. However, even the simplest formulation, equation 4.13, provides reasonable estimates of ET,

which may prove very reliable on weekly or longer time-scales. Calibrating the model using streamflow data only, rather than ET and streamflow data, creates bias in the ET estimate at the beginning of the record, after which the estimates appear quite robust, while overall performing better in terms of the streamflow estimates. This implies, at least in this case, that calibrating without ET data does not have a major negative effect on the model's ability to predict either streamflow or ET.

4.3 Description of catchments

To investigate the applicability of the new non-linear module for ET loss and discharge prediction, it was applied on a daily time step to three catchments covering a range of space scales and climatic conditions. The hydrometeorological characteristics of the selected catchments are given in Table 4.1.

Table 4.2: Hydrometeorological characteristics of catchments investigated

Catchment	Area, km ²	Precipitation, mm/yr	Average Daily Mean Temperature, °C	Annual Yield, %
Watershed 34 (Coweeta, North Carolina)	0.33	2019	12.7	46
Watershed 36 (Coweeta, North Carolina)	0.49	2232	12.7	64
Scott Creek @ Scotts Bottom	27	1090	14.8	20.5

Two small catchments were selected from the Coweeta Hydrological Laboratory in the United States. Coweeta is located in the Nantahala Mountains of western North Carolina. Coweeta is a relatively humid area and both streams are perennial. Rainfall occurs throughout the year. Watershed 36 is a high-elevation, steeply sloping catchment with shallow soils, a high annual yield and a large proportion of quick flow (Swift et al.

1988). Watershed 34 is a mid-elevation catchment with somewhat deeper soils and, consequently, substantially more delayed flow. The Coweeta site is covered predominantly by hardwood forest. Details of the physical characteristics of the Coweeta catchments are given by Swank and Crossley (1988).

The final catchment used in this investigation is Scott Creek at Scotts Bottom. It is situated near the city of Adelaide in South Australia. Scott Creek is an ephemeral catchment, it has a warm climate with hot, dry summers. The catchment is dominated by winter rainfall and the streamflow can cease during summer. Soils in the catchment are made up of a sandy loam top layer with a clay subsoil. More than 80% of the catchment is covered by grass. It is part of a network of benchmark catchments in Australia. More details can be found in Chiew and McMahon (1992).

4.4 Streamflow modelling results

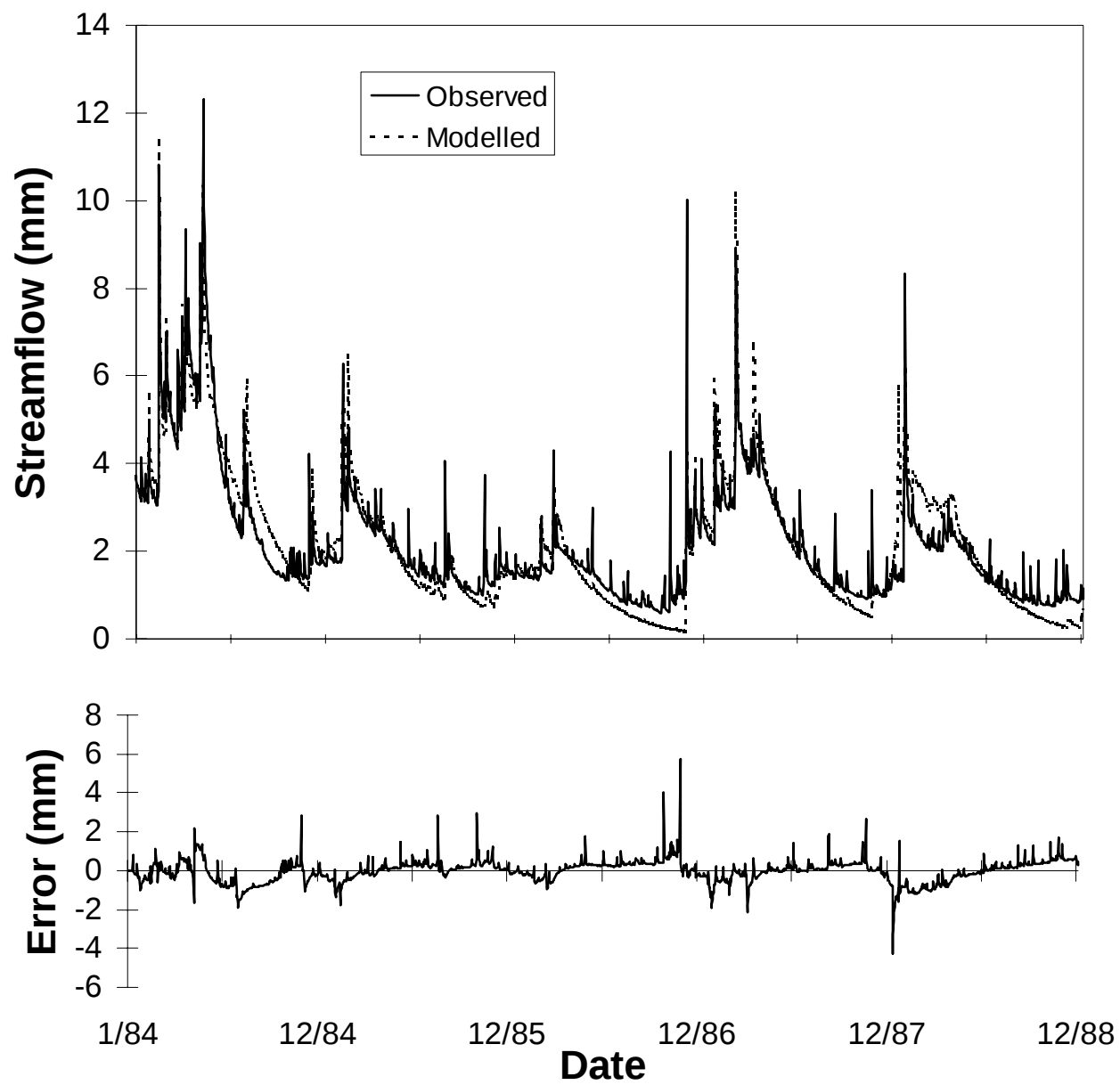
The model was calibrated on a daily time step over periods of two years on all three catchments and the results are given in Table 4.2. The performance in terms of goodness of fit criteria, *NSE* and *B*, is quiet than reasonable. The models were then used to simulate the daily streamflow over the subsequent 5 years and the results are given in Table 4.3. Figure 4.8(a-c) compares the modelled streamflow with observed streamflow for the five year simulation periods. Figure 4.8a indicates that the model had some difficulty reproducing hydrograph recessions when simulating streamflow in Watershed 34 even though the performance statistics remain quite good. Model residuals for watershed 34 show a slowly varying component due to under-prediction of baseflow at times. Figure 4.9 shows the modelled catchment moisture deficit changing over the simulation period in the Scott Creek catchment. It clearly demonstrates the seasonal nature of the catchment moisture store. Each year CMD builds over summer, a period of little rain and higher temperatures, and then it is reduced quickly when the rains return

Table 4.2: Calibration results for the three catchments.

<i>Catchment</i>	<i>Period</i>	c_1	c_2	c_3	c_4	τ_q days	τ_s days	V_s	<i>NSE</i>	<i>B</i> mm/d
Watershed 34, Coweeta	1/1/82 - 31/12/83	0.2	0.00	4	55	2.16	85.75	0.857	0.90	0.00
Watershed 36, Coweeta	1/1/82 - 31/12/83	0.15	0.00	5	95	1.59	43.79	0.324	0.87	-0.05
Scott Creek	24/2/70 - 23/2/72	0.24	0.01	2	8	0.71	53.07	0.475	0.88	0.00

Table 4.3: Simulation results for the three catchments.

<i>Catchment</i>	<i>Period</i>	<i>NSE</i>	<i>B</i> mm/d
Watershed 34, Coweeta	1/1/84 - 31/12/88	0.84	0.23
Watershed 36, Coweeta	1/1/84 - 31/12/88	0.76	-0.26
Scott Creek	24/2/72 - 23/2/77	0.79	0.00



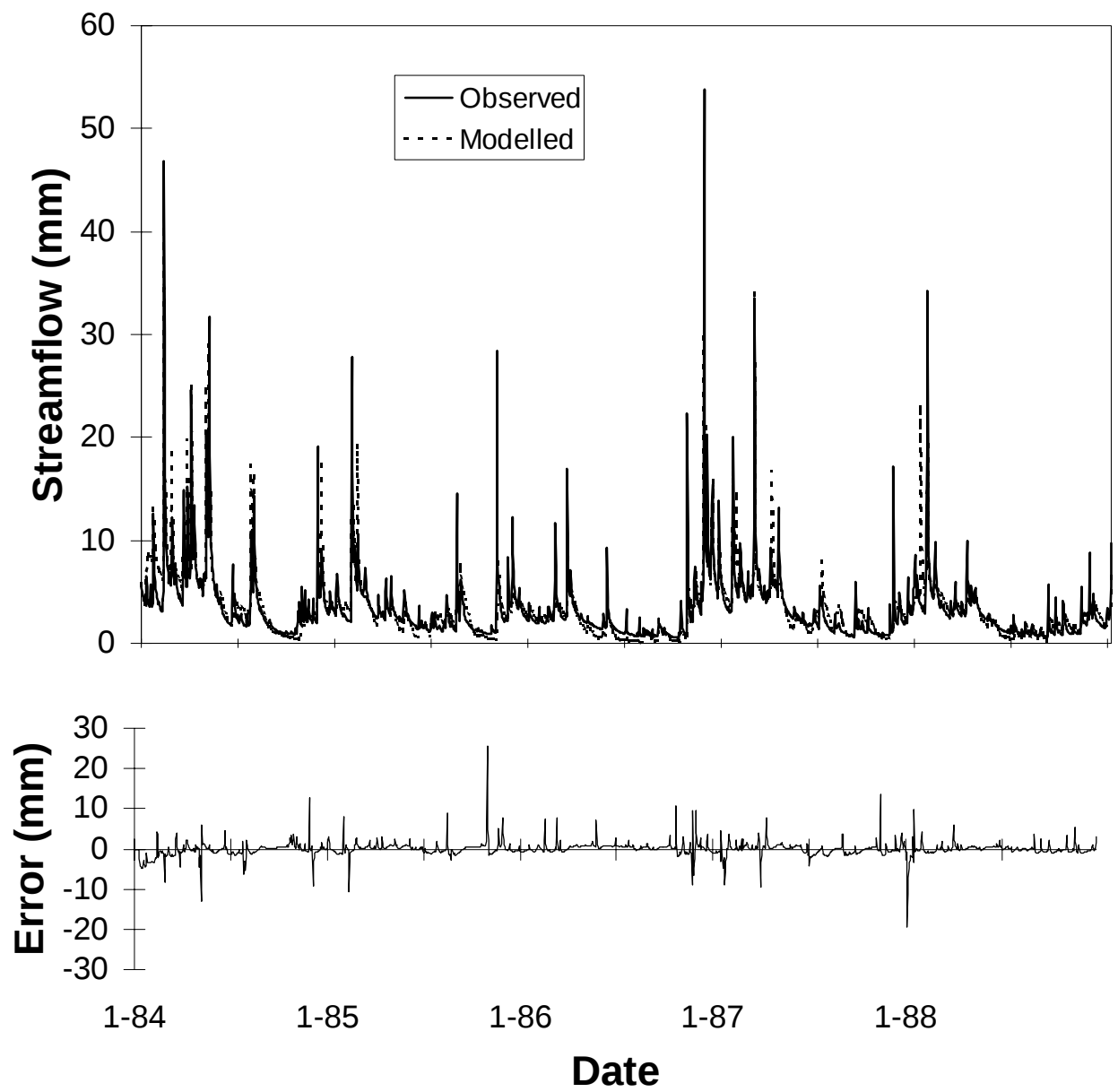


Figure 4.8b Simulation model fit for Coweeta watershed 36

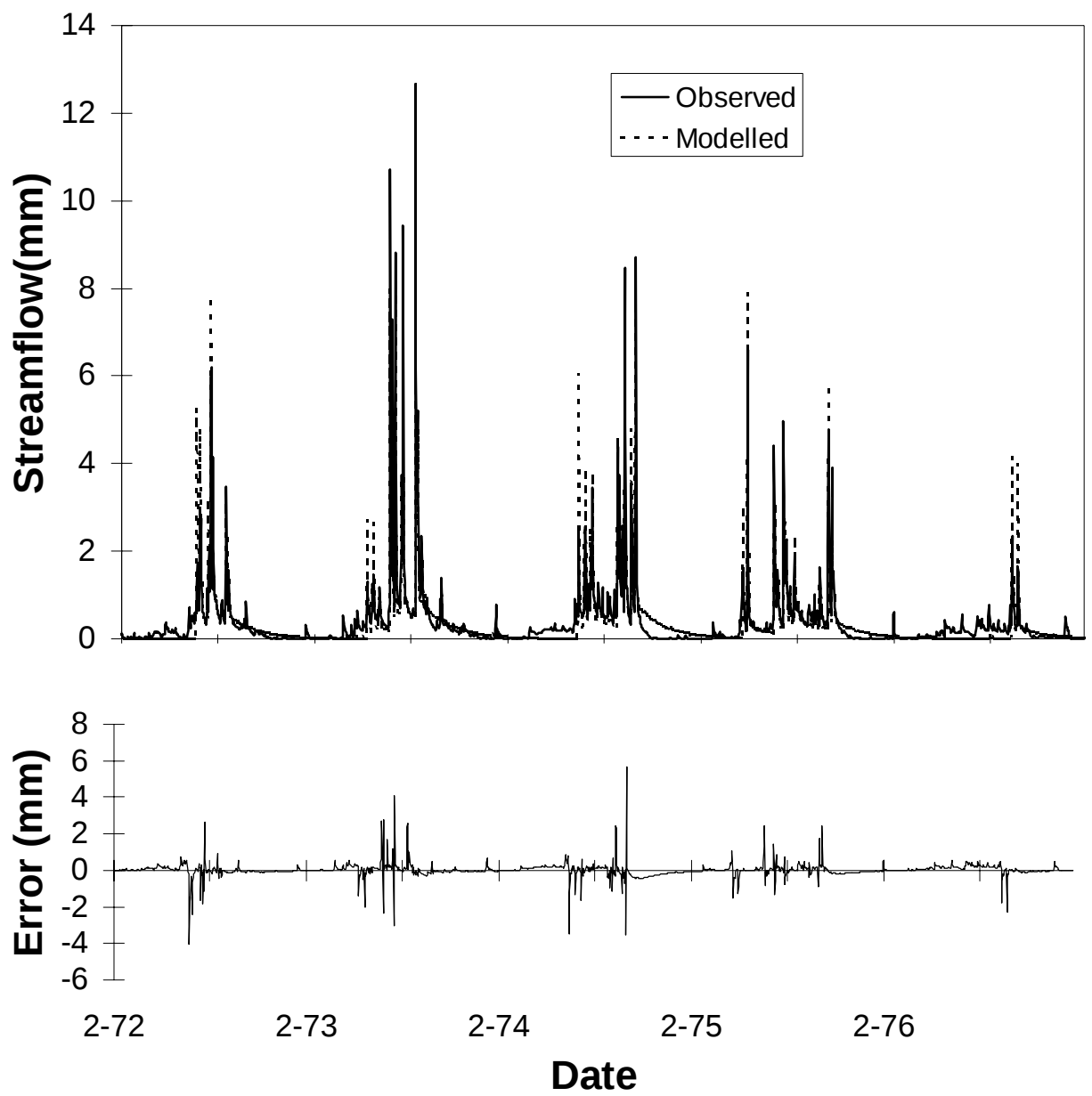


Figure 4.8c Simulation model fit for Scott Creek

in winter. Comparison of Figure 4.9 and Figure 4.8c demonstrates the strong relationship between CMD and streamflow. Figure 4.10 shows the ET predicted for the Scott Creek catchment during the simulation period. Figure 4.11 compares the mean monthly rainfall and ET from 1961 until 1990 for Coweeta Watersheds 34 and 36 as calculated by our model and that calculated at a representative site within the Coweeta Hydrological Laboratory using the Thornthwaite (1948) method (Swift 1998). This provides a broad scale check on the ET predicted by our model.

4.5 Discussion and conclusions

The parameterisations chosen for both ET and discharge are the simplest ones tested, but which have been judged to perform well. Indeed the model tested here performed as well if not better than several other more complex parameterisations investigated, indicating that higher levels of parameterisation are not warranted on these data sets. However, selection of the most appropriate parameterisation in the future will require more extensive testing on longer data sets and across different hydroclimatologies.

Table 4.2 demonstrates the ability of the model to fit the streamflow characteristics of each catchment when calibrated on only two years of data, with R^2 ranging from 0.87 to 0.9. It appears to perform equally well in humid regions as in semi-arid regions. Notably, the bias remains low in all cases. The models for Coweeta Watershed 34 and Scott Creek produce zero bias, and Coweeta Watershed 36 yielded a bias of -0.05 mm/day. These low biases indicate that the total ET losses have been estimated well. Measured ET, preferably on a daily basis, are required though if we are to have confidence in the estimated ET losses per day. The parameter c_4 was the only parameter to demonstrate significant identifiability problems in the calibration procedure. This emphasises the difficulty in establishing a clear value for the CMD at which water ceases to drain to the stream, especially in humid catchments where modelled moisture never ceases to drain to the stream during the calibration period. A longer calibration period may reduce this uncertainty.

Clearly from Table 4.3 the model performs well under simulation conditions, with R^2 ranging from 0.76 to 0.84. As expected the R^2 values are lower than during the calibration period but remain quite high. For both Coweeta catchments the bias has increased significantly but remains small relative to the streamflow. The simulation period bias for the Scott catchment remains zero, indicating again that the ET losses have been estimated well overall. Figure 4.8 demonstrates the simulated model fit for the three catchments. Both Figures 4.8b and 4.8c show consistently good fits while Figure 4.8a shows the model had some trouble fitting the hydrograph recession for Watershed 34.

Figure 4.10 shows an example of the daily ET losses predicted by the model for the simulation period in the Scott catchment. Comparison with Figure 4.11 shows the influence of CMD on the evaporation. The seasonal nature of ET is not as obvious as for CMD. The lowest values occur towards the end of summer when CMD is still high and temperatures are falling, demonstrating the trade off between available soil moisture and available energy, which drives the evaporative process.

Figure 4.12 shows that the rainfall observed for Watersheds 34 and 36 is higher than that for the representative site within the Coweeta Hydrological Laboratory, with Watershed 36 expectedly having the most rainfall since it is at the highest elevation. The representative site rainfall and ET almost coincide in July suggesting that in some years soil moisture may be low enough to affect transpiration of the vegetation. Similarly, Watershed 34 rainfall and ET almost coincide in August again suggesting transpiration of the vegetation may be affected in some years. Watershed 36 maintains a much larger gap between rainfall and ET, so the vegetation is less likely to be stressed in this Watershed. ET estimated by our model follows a very similar pattern to that estimated by the Thornthwaite method on a monthly time step. For watershed 36 there is a much larger gap between rainfall and ET losses which is consistent with its annual yield of 64% being almost 50% higher than the annual yield of watershed 34 (46%).

From equation (4.10), we can see that if parameter c_3 was zero then effective rainfall only occurs when CMD is forced below zero by the rainfall. The non-linear module of

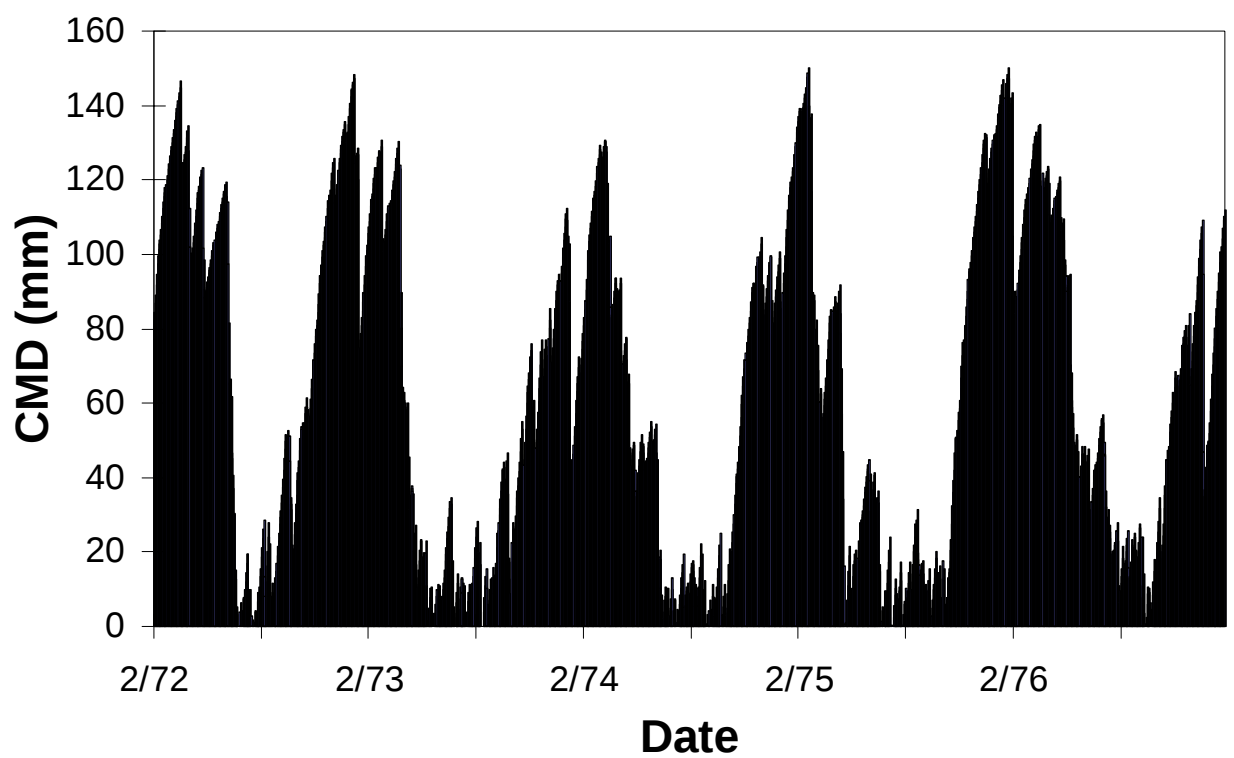


Figure 4.9 Catchment moisture deficit in the Scott Catchment

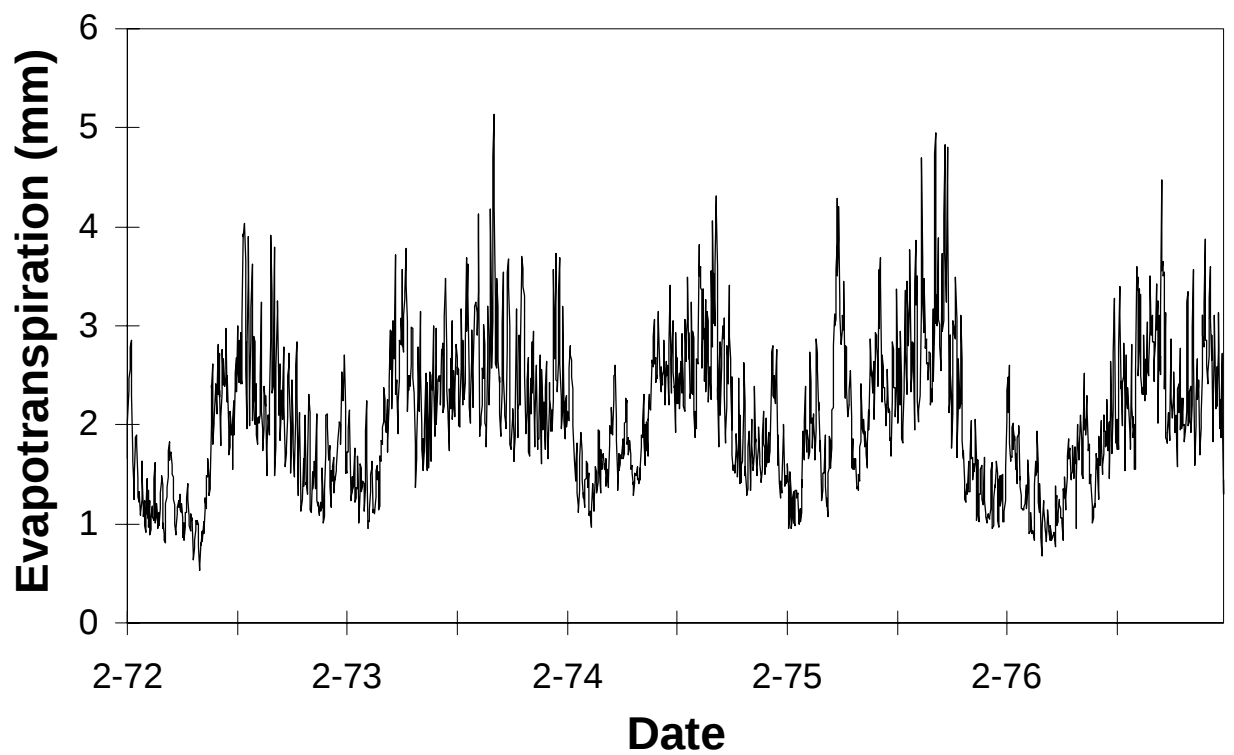


Figure 4.10 Evapotranspiration in the Scott catchment

IHACRES is then acting as a threshold mechanism adjusted by the antecedent moisture conditions. Here the model has collapsed down to five parameters and the non-linear module is acting similarly to the bucket model of (Manabe et al. 1965).

Table 4.2 also shows that the best calibration results for both Coweeta watersheds occur when c_2 is zero. This means that the modelled ET has no dependence on CMD. For the period of calibration this is not entirely surprising as Coweeta has relatively high rainfall, and so the CMD may rarely be large enough to affect the transpiration of the vegetation and therefore ET.

By using this semi-physical approach as described in the accounting scheme in (4.7), the physical interpretation of the parameters will hopefully be made easier than with the previously used statistical approach in IHACRES. It is hoped that, by studying catchments with gauged discharge data and possibly measured ET data, improved relationships can be constructed between the model parameters and landscape attributes, as has been attempted with some success by Post and Jakeman (1996), using the statistical type of non-linear module. These relationships would permit the simulation of streamflow and ET for changes in land use and for ungauged catchments, at least within a similar region. The construction of these relationships is only made possible by the minimal parameterisation of catchment hydrologic response, with the model having at most seven parameters. Three of these parameters are the routing parameters in (4.2) and (4.3).

Only temperature and catchment moisture are investigated as forcing variables for ET here since we have used the modified temperature approach. However, as demonstrated in section 4.2, the availability of other important variables, such as humidity, net radiation and wind speed, would improve the model results.

When refined, this approach to modelling rainfall-ET-runoff can be used to provide the land surface-atmosphere fluxes for a climate model, this is demonstrated in chapter eight. It will be necessary to spatially disaggregate climate forcing variables from the grid scale down to the relevant catchment scale (Brogardly et al. 1993, Hughes and Guttorp 1994, Bates et al. 1998), and to temporally disaggregate daily ET and energy

feedbacks from the land surface to the atmosphere down to a sub-daily time step of the order of an hour. These scale issues are recognised problems (Wood et al. 1990, Beven 1991, Avissar 1995, Kalma and Sivapalan 1995, Raupach and Finnigan 1995a) and are the focus of continuing research. The accounting scheme in (4.7) potentially permits this temporal disaggregation as the ET could be calculated on a finer time-step than the runoff and simply accumulated to the same time-step as the runoff. Provided the ET expression, such as (4.9), is applicable for the sub-daily time step, the model can be calibrated by fitting average daily (or temporally finer) discharge data.

In this approach, parameterisation of all gauged catchments must be undertaken off-line, but once inserted in a GCM or regional climate model the overall computational complexity is little greater than that of a bucket model. These parameterisations are valid at least for historically tested climate conditions, vegetation and land use status. With the construction of relationships between the model parameters and the landscape and vegetation attributes, the model parameterisations would gain an even wider applicability.

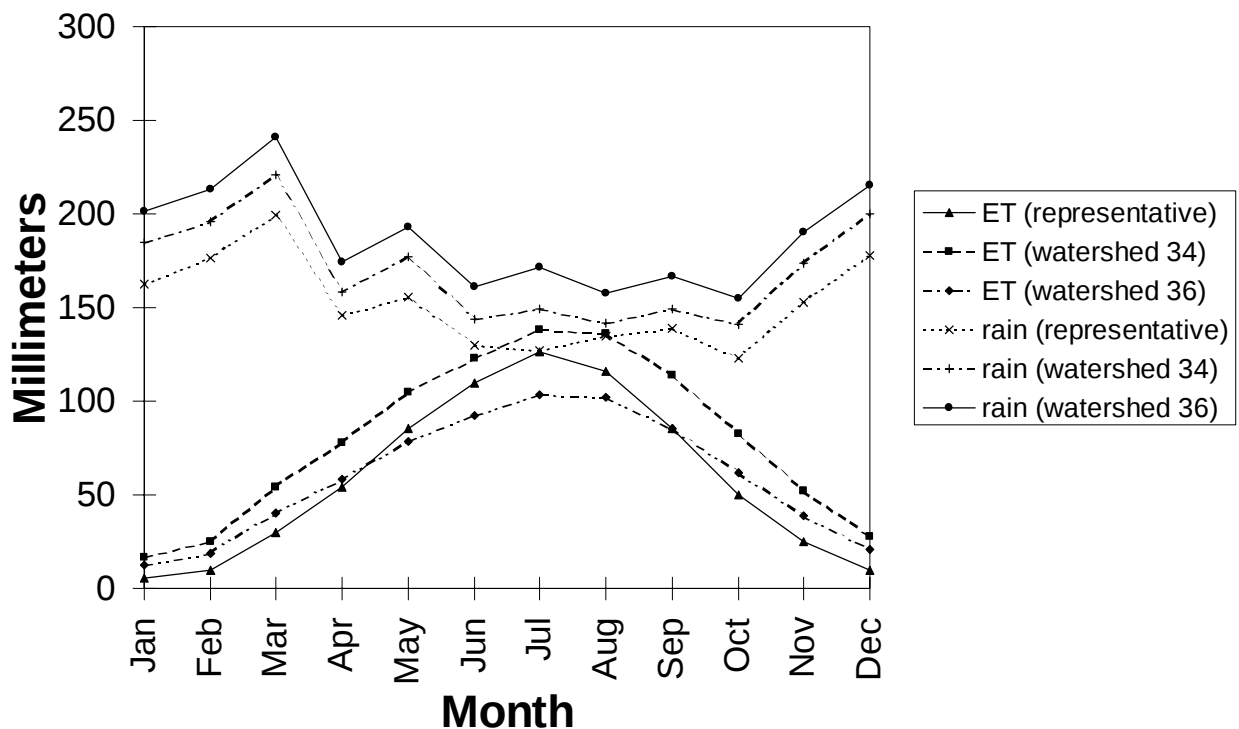


Figure 4.11 Mean monthly precipitation and Evapotranspiration: 1961 – 1990

Part II: Climate Modelling



Before an attempt is made to determine the effect on streamflow predictions of including the rainfall-runoff model CMD-IHACRES in a LAM, it seems worthwhile to first look at the overall performance of the LAM, and in particular to investigate the relative importance of the land surface parameterization (or SVAT) when compared with the rest of the model. To this end, this part will analyze the performance of three different LAM/SVAT combinations over the FIFE region in central Kansas. The three combinations are MM5/BATS, MM5/SHEELS and RegCM2/BATS. This allows comparisons between two identical LAMs with different SVATs, or between two different LAMs with identical SVATs.

Chapter 5

Review of Climate Models

5.1 Introduction

Climate models have developed over the last forty years or so from the early numerical weather prediction models. As such, they began as attempts to perform short term weather prediction, and have since evolved to look at long term climate interactions. In more recent times the impetus behind the development of these models has changed somewhat from an attempt to model the fundamental characteristics of the atmosphere to a desire to provide predictions of the impacts on the climate system of changes such as anthropogenic increases in atmospheric CO₂ or massive deforestation.

The climate system itself is a very complex multicomponent system, with many interconnections and feedbacks between components. Henderson-Sellers and McGuffie (1987) distinguish four basic types of model:

1. Energy balance models are one-dimensional models predicting the variation of surface temperature with latitude. Simplified relationships are used to calculate the terms contributing to the energy balance in each latitude zone. Significant studies using energy balance models were performed by Budyko (1969) and Sellers (1969) along with others (Cess 1976, North et al. 1981).
2. One-dimensional radiative-convective models compute the vertical (usually globally averaged) temperature profiles by explicit modelling of the radiative processes and a 'convective adjustment' which re-establishes a predetermined lapse rate. These models have been used to investigate the vertical profiles of temperature and water vapour in the atmosphere

(Manabe and Wetherald 1967, Ramanathan and Coakley 1979, Hummel and Kuhn 1981) as well as various feedback mechanisms (Wang and Stone 1980, Hansen et al. 1981, Rossow et al. 1982).

3. Two-dimensional statistical dynamical models deal explicitly with surface processes and dynamics in a zonally averaged framework and have a vertically resolved atmosphere. Some applications of these models can be found in Sellers (1976), Ohring and Adler (1978), Saltzman (1978) and Taylor (1980).
4. Global climate models (GCMs) and limited area models (LAMs). The three dimensional nature of the atmosphere and/or ocean is incorporated. In these models an attempt is made to represent most physical processes believed to be important. Descriptions of some GCMs are given in Gordon and Stern (1982), Boer et al. (1984), McGregor et al. (1993a) and Kiehl et al. (1998).

In this thesis the climate influence on (land) surface water, and the feedback from this to the atmosphere is of primary interest. GCMs and LAMs were chosen as the best tool to investigate these dynamics because they contain within them the most plausible representation of the physical processes involved, as well as LAMs being readily applied at a regional scale. They also have the added advantage of requiring global datasets to run them. These datasets are interpolated through areas with sparse data, allowing their use over these areas and hence satisfying a major criteria for use in this thesis. Little attention will be given to the other types of climate models, as well as ocean related components of modern coupled GCMs, since they are not explicitly used in this thesis.

5.2 The climate system

Climate is generally taken to be the statistical state of the atmosphere observed as weather over a number of years. In order to model the climate all the complex interactions between, as well as changes within, all the components of the climate system – the atmosphere, ocean, land, ice and snow, and terrestrial and marine biota need to be modelled. The Global Atmospheric Research Program (GARP) of the

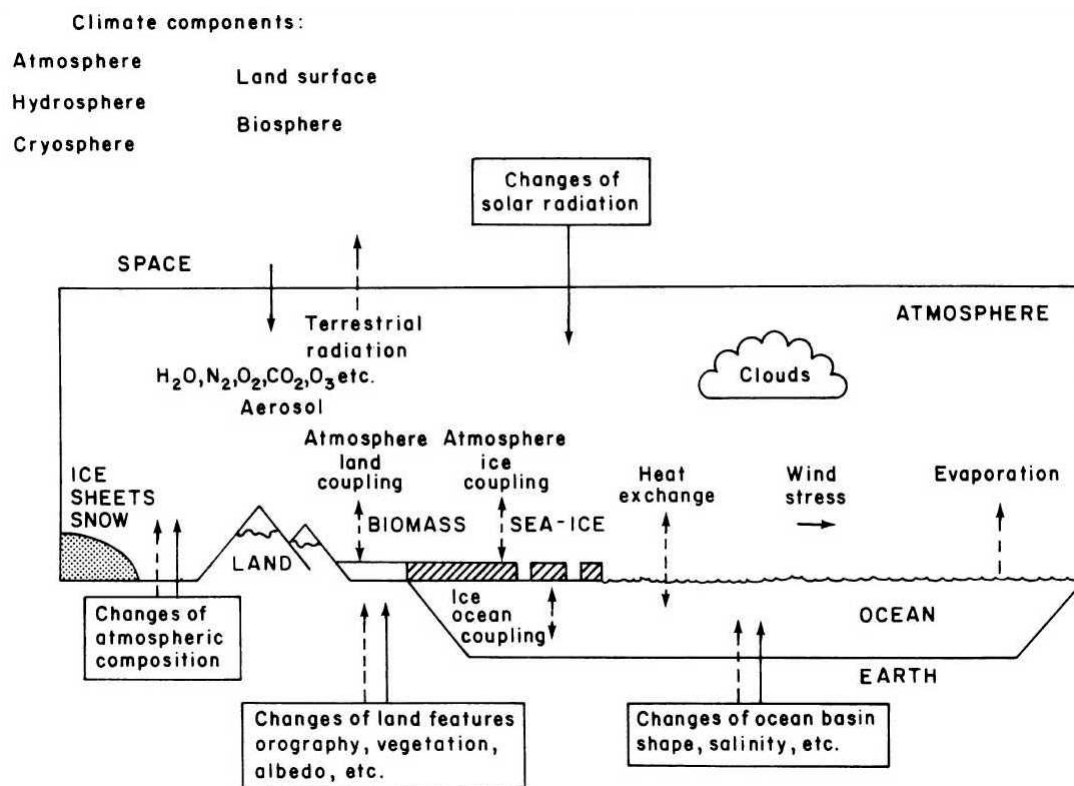


Figure 5.1 Schematic illustration of the components of the climate system. Full arrows are examples of external processes and dashed arrows are examples of internal processes (GARP 1975).

World Meteorological Organization defined the climate system as being composed of the atmosphere, hydrosphere, cryosphere, land surface and biosphere. Figure 5.1 is a representation of the components of the climate system and their interactions. There are several texts which describe the climate system (Trewartha and Horn 1980, Henderson-Sellers and McGuffie 1987, McIlveen 1992, Peixoto and Oort 1992, Trenberth 1992, IPCC 1996).

As Figure 5.1 shows there are many interactions and feedbacks within the climate system. The feedback mechanisms operate in response to some imposed change in the climate. These imposed changes can have external causes such as changes in the Earth's orbit around the sun (the Milankovitch theory), changes in solar activity such as those associated with the sunspot cycle or even collisions with comets or large meteorites. Internal causes of climate change include volcanic eruptions, changes in the chemical composition of the atmosphere such as increasing levels of carbon dioxide or sulfates, and changes to the land surface such as deforestation and desertification.

5.3 Global Climate Models

Due to the history of the evolution of GCMs they are often thought of as having three primary components: (i) dynamics, (ii) physics, (iii) other (land surface, ocean, ice, etc.). All GCMs have as fundamental considerations:

1. Conservation of momentum

$$\frac{D\mathbf{v}}{Dt} = -2\boldsymbol{\Omega} \wedge \mathbf{v} - \delta^{-1}\nabla p + \mathbf{g} + \mathbf{F} \quad (5.1)$$

2. Conservation of mass

$$\frac{D\delta}{Dt} = -\delta\nabla \cdot \mathbf{v} + C - E \quad (5.2)$$

3. Conservation of energy

$$\frac{DI}{Dt} = -p \frac{d\delta^{-1}}{dt} + Q \quad (5.3)$$

4. Ideal gas law

$$p = \delta RT \quad (5.4)$$

where $\mathbf{v}$ = velocity relative to the rotating Earth,

t = time,

$$\frac{D}{Dt} = \text{total time derivative} \left[= \frac{\partial}{\partial t} + \mathbf{v} \cdot \nabla \right],$$

$\boldsymbol{\Omega}$ = angular velocity vector of the Earth,

δ = atmospheric density,

$\mathbf{g}$ = apparent gravitational acceleration,

p = atmospheric pressure,

$\mathbf{F}$ = force per unit mass,

C = rate of creation of atmospheric constituents,

E = rate of destruction of atmospheric constituents,

I = internal energy per unit mass $[= c_v T]$,

Q = heating rate per unit mass,

R = gas constant,

T = temperature,

c_v = specific heat of air at constant volume.

A GCM should also conserve enstrophy (the root mean square of the vorticity).

5.3.1 Dynamics in GCMs

The dynamics of the atmosphere are calculated either on a Cartesian grid or using spectral transform methods. Figure 5.2, taken from Henderson-Sellers and McGuffie (1987), is a schematic showing the differences between these two approaches.

Essentially Cartesian grid GCMs divide the atmosphere into a series of ‘boxes’.

Variables are tracked only at a series of grid points spaced in latitude and longitude, while the vertical spacing can vary depending on the importance of a particular region of the column. Generally there will be more layers near the Earth’s surface and in the

troposphere. Spectral models on the other hand, deal with the atmospheric dynamical fields in the form of waves. The data fields are transferred to and from grid space at every time step via fast Fourier transforms and Gaussian quadrature. Vertical transfers, surface processes and grid-point physics are incorporated after the transformation into grid space.

Whichever coordinate system is used, a series of highly nonlinear partial differential equations (PDEs) must be solved using numerical techniques. An introduction to some of these techniques can be found in Hack (1992). More complete developments of the numerical methods can be found in GARP (1979, Haltiner and Williams (1980) and Staniforth and Cote (1991).

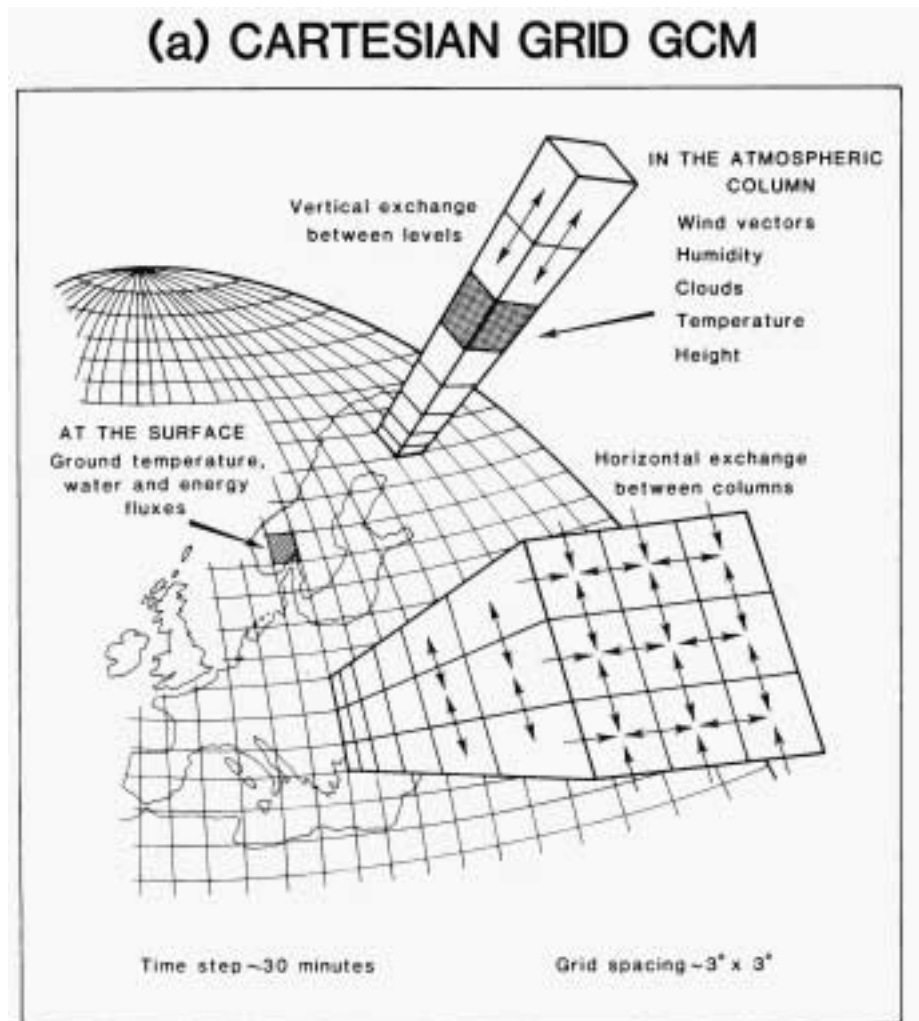


Figure 5.2(a) caption below

Figure 5.2 The construction of (a) a cartesian grid GCM and ; (b) a spectral GCM. In a cartesian grid GCM horizontal and vertical exchanges are handled in a straightforward manner between adjacent columns and layers. In a spectral GCM vertical exchanges are computed in grid-point space, while horizontal flow is computed in spectral space. The method of transfer between spectral and grid-point space can be seen reading around Figure 2.3(b) from point 1 to 4. From Henderson-Sellers and McGuffie (1987).

(b) SPECTRAL GCM

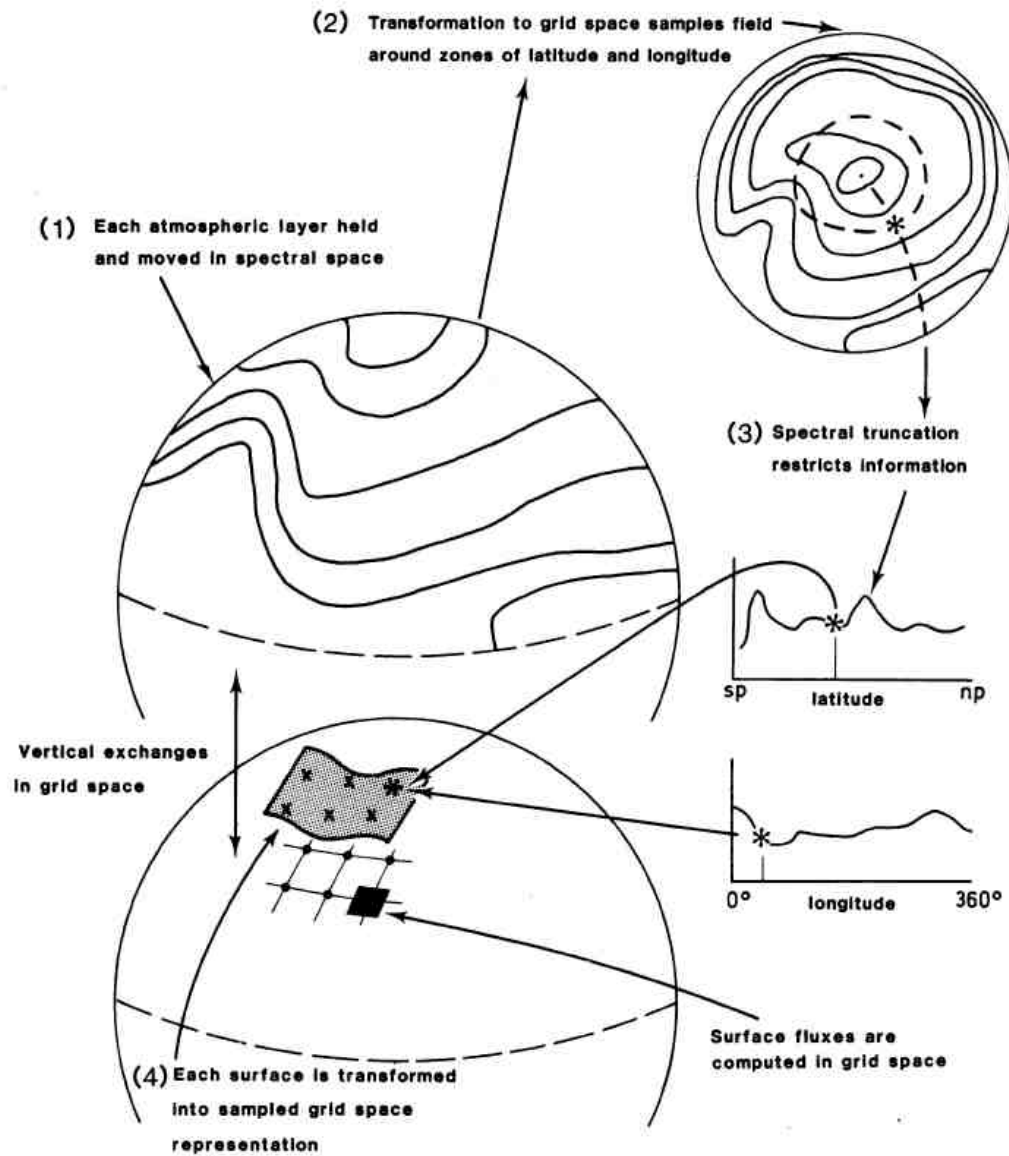


Figure 5.2 (b)

5.3.2 Physics in GCMs

Physical processes accounted for in GCMs typically fall into the following processes:

1. The radiation
2. The boundary layer
3. The convection
4. The large scale precipitation .

Figure 5.3 gives an example of a flowchart of the implementation of a GCMs physics as found in Kiehl (1992). Assuming initial data are available for the prognostic variables (wind (u), temperature (T) and specific humidity (q)), the model calculates initial fluxes for use in the planetary boundary layer (PBL) and surface components of the model. These, along with the thermodynamic and moisture profiles at each gridpoint, are used to test whether the atmospheric column is stable or unstable (**A**). If unstable, a convection parameterization is used to determine the convective heating and moistening terms. Otherwise, if saturated, the stable condensation process is invoked. Based on the type of condensation process, cloud fractions are assigned to model layers. Condensational heating and cloud amounts are stored for further use (**B**). Radiative fluxes and heating rates are then calculated based on the thermal, moisture and cloud profiles in the atmosphere. Mechanical dissipation terms are then determined. At this point all forcing and dissipative terms for the PDEs are available, and a numerical solution technique is applied to obtain new values for the prognostic variables.

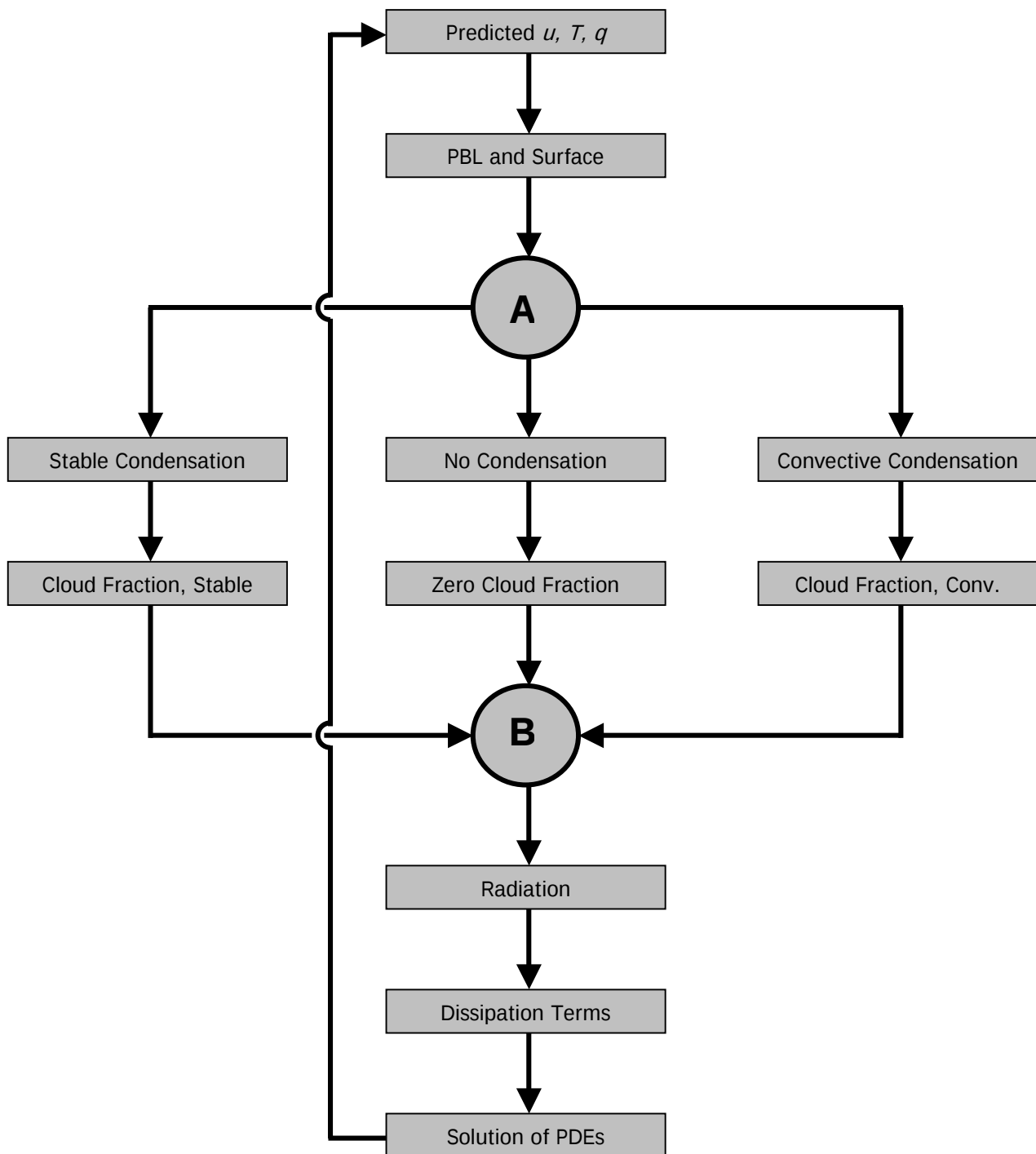


Figure 5.3 Diagram of the procedures employed in the atmospheric component of a GCM. From (Kiehl 1992).

More explicit mathematical descriptions of the physics parameterizations can be found in Chang (1977), Henderson-Sellers and McGuffie (1987) and Kiehl (1992).

5.3.3 Other components of a GCM

5.3.3.1 Ocean circulation models

The ocean was originally treated quite simply in GCMs with either sea surface temperatures or the meridional energy transport of the oceans being prescribed. The most recent versions of GCMs include coupling the atmospheric general circulation to an oceanic general circulation model. Ocean circulation is driven by salinity as well as temperature gradients and has a much longer timescale to atmospheric circulation.

Descriptions of the ocean components of GCMs can be found in Abarbanel and Young (1986), Anderson and Willebrand (1989), Haidvogel and Bryan (1992), IPCC (1996) and Pedlosky (1996) and will not be repeated here.

5.3.3.2 Cryosphere models

Cryosphere models include predictions of snow cover, sea ice and possibly permafrost. The presence (or absence) of snow cover or sea ice substantially affects the surface albedo in radiation calculations as well as the ocean-atmosphere interactions. This thesis focuses on snow free areas and hence no detail on cryosphere modelling will be given here, however it is important when performing GCM model runs. For more detailed discussions see Oerlemans (1982), Flato and Hibler (1990), Lemke et al. (1990), Hibler and Flato (1992) and Veen (1992).

5.3.3.3 Land surface/Biosphere models

These component models have progressed from the very simple surface hydrology ‘bucket’ model of Budyko (1956) and Manabe et al. (1965) to the complex soil-vegetation-atmosphere transfer (SVAT) schemes of Dickinson et al. (1986), Sellers et al. (1986), Avissar and Pielke (1989), Noilhan and Planton (1989) and Bonan (1996). These models are designed to explicitly deal with the transfer of radiation and turbulence as well as the exchange of heat and water vapour at the land-atmosphere interface.

The major problems faced by SVAT schemes involve dealing with the heterogeneity inside a single GCM grid square (typically $\sim 300\text{km} \times 300\text{km}$), and the lack of data with which to initialize and validate the models. These problems are being addressed, with the continuing increase in computing power reducing the GCM grid size and with a series of intensive large-scale field experiments and ongoing monitoring increasing the data available. As presently used, initializing a SVAT model involves at best taking informed guesses while the land surface heterogeneity within each grid square is dealt with in a simple proportional manner.

An overview of modelling the land surface and biospheric processes within a climate modelling frame can be found in Sellers (1992) and IPCC (1996). The Project for Intercomparison of Land-surface Parameterization Schemes (PILPS) (Henderson-Sellers et al. 1993b) is investigating the performance of a large number of SVAT schemes under various conditions. Through a series of publications (Henderson-Sellers et al. 1995, Henderson-Sellers et al. 1996a, Shao and Henderson-Sellers 1996,

Chen et al. 1997, Timbal et al. 1997) they have given insight into the current state of SVAT schemes, their abilities and limitations.

5.3.3.4 Atmospheric chemistry models

Atmospheric chemistry models are concerned with the transport and chemical interactions of the constituents of the atmosphere. They came to the fore with the discovery of the hole in the ozone layer, which was largely attributed to the interaction between O₃ and CFCs. They are also important in keeping track of many greenhouse gases and other molecular species in the atmosphere such as HO_x, NO_x, CO₂, CH_x, N₂O and more recently sulfates. Overviews and further reading in this area can be found in Brasseur and Solomon (1986), Seinfeld (1986), Rood (1987) and Brasseur and Madronich (1992).

5.3.4 Uncertainties in GCM model structure

Many uncertainties are associated with the structure and use of a GCM. Many sensitivity and comparison studies have been done in an attempt to make explicit and understand these uncertainties. An assessment of these uncertainties is presented in IPCC (1996) where they note that

Clouds, the hydrological cycle and the treatment of the land surface remain the largest areas of uncertainty in climate models, and are generally the cause of the largest intermodel differences in both control and sensitivity experiments.

They also draw attention to GCM sensitivity to the spatial resolution and the initial and boundary conditions.

Of particular interest in this thesis is the influence and sensitivity of the land surface, especially the land surface hydrology. Intercomparisons of SVATs such as those produced during PILPS (Shao et al. 1994) indicate major discrepancies in runoff rates between models, which leads to large differences in evapotranspiration. In IPCC (1996) pp.219 it is stated that

The current lack of a physically-based and adequately validated treatment for runoff may be the biggest single obstacle to achieving an SVAT adequate for climate modelling.

Addressing this major uncertainty in climate modelling is a primary aim of this thesis. The uncertainties in surface runoff are affected to a large extent by the uncertainties in cloud formation and precipitation, water vapour transport and evapotranspiration. This compounding of uncertainties can make it difficult to attribute error to a particular component. Improving the surface runoff component, which can be tested against observed data, should facilitate further assessment and improvements in related components of the climate model.

5.4 Limited area models

Since GCM grid spacing is very large, of the order of 100s of km², there is often a need to gain more information about subgrid scale phenomena. The spatial distribution of variables within grid cells is of particular importance when looking at the regional impacts of climate change or weather related phenomena. One way to achieve this downscaling is to use a limited area model (LAM) nested within a GCM.

Also known as regional climate models, LAMs share many of the same features as GCMs in terms of the parameterizations of their dynamics and physics, though they are generally run at much higher spatial and temporal resolution. LAMs differ however in their need to assimilate lateral boundary and initial conditions from GCMs.

GCM simulations attempt to reproduce realistic intensities and frequencies for each type of major synoptic system. If a LAM is nested within a GCM simulation it makes possible the production of a realistic detailed climatology. Previous reviews of LAMs and their simulations can be found in (McGregor et al. 1993c, McGregor 1997, Giorgi and Mearns 1999). The earliest examples of LAM integrations were very short, lasting only a few days. More recently, integrations of several years have allowed investigations into seasonal cycles and annual variability.

In this thesis two LAMs were used. The first is the fifth generation Penn State/NCAR Mesoscale Model (MM5) adapted for climate studies with the addition of either the BATS land surface parameterization or the SHEELS land surface parameterization and is described in section 5.4.1. The second LAM used is NCARs second-generation regional climate model, RegCM2, which is described in section 5.4.2. A summary of the main components of the LAMs can be found in table 5.1. Section 5.4.3 briefly outlines the methodology commonly used in regional climate modelling. Section 5.4.4 describes results from LAM simulations. Section 5.4.5 presents aspects of regional climate modelling where difficulties might be encountered.

Table 5.1 Summary description of the LAMs

LAM	MM5	RegCM2
longwave radiation scheme	broadband emissivity method (Stephens 1984)	band-absorptance technique including contributions of CO ₂ , O ₃ , H ₂ O and clouds (Kiehl and Briegleb 1991)
shortwave radiation scheme	scattering and absorption by clouds, clear air and water vapour (Grell et al. 1994)	δ -Eddington approximation (Joseph et al. 1976)
stable precipitation	(Dudhia 1989)	(Hsie et al. 1984)
convective precipitation	Grell scheme (Grell 1993)	Grell scheme (Grell 1993)
planetary boundary layer	nonlocal-K approach (Hong and Pan 1996)	nonlocal-K approach (Holtslag et al. 1990b)
LAND SURFACE PARAMETERIZATION	BATS	SHEELS
number of layers for temperature	3	3
temperature methodology	Force-restore (Deardorff 1978)	Force-restore (Deardorff 1978)
number of layers for soil moisture	3 nested	11 discrete
soil moisture methodology	Darcy's law	Darcy's law
canopy methodology	Penman/Monteith	Penman/Monteith

5.4.1 Description of the MM5 Limited Area Model

This section provides a brief description of the fifth-generation Penn State/NCAR Mesoscale Model (MM5). A more complete description can be found in Grell et al. (1994). MM5 and its predecessors have been used in many studies to date (Anthes et al. 1985, Lapenta and Seaman 1990, Kuo et al. 1992, Grell 1993, Hines et al. 1995).

MM5 does not come with a SVAT component as standard, instead the standard surface model physics include a high-resolution Blackadar-type planetary boundary layer (Zhang and Anthes 1982) and a surface energy budget. Oncley and Dudhia (1995) evaluated the surface flux estimates of this model. They found that the model flux parameterization was quite sensitive to the roughness length and the moisture availability parameter, M , and that there was no obvious way to select M a priori. They suggest that to overcome these problems a more detailed model of surface exchange is required, i.e. a SVAT. Here we use MM5 with a SVAT incorporated, which we would expect to improve the surface flux estimates.

This implementation of MM5 is a hydrostatic, compressible, primitive equation, terrain following σ vertical coordinate model. Here σ is defined as

$$\sigma = \frac{p - p_{top}}{p_s - p_{top}} \quad (5.5)$$

where p is the pressure, p_{top} is the pressure specified to be the model top and is constant, and p_s is the prognostic surface pressure.

The spatial grid is determined using a Lambert conformal projection. The influence of the lateral boundaries is determined using a ‘sponge boundary’ condition which is given by

$$\left(\frac{\partial \alpha}{\partial t} \right)_n = w(n) \left(\frac{\partial \alpha}{\partial t} \right)_{MC} + (1 - w(n)) \left(\frac{\partial \alpha}{\partial t} \right)_{LS} \quad (5.6)$$

where α represents any variable, MC denotes the model calculated tendency, LS the large-scale tendency which is obtained either from observations or large scale model simulations (one-way nesting), n is the displacement in grid points from the nearest boundary ($n = 1$ on the boundary) up to a maximum of five. The weighting coefficients $w(n)$ take values between zero and one, increasing away from the boundary.

The LAMs presented in this chapter have almost identical dynamical components and hence the following model description will focus on the model physics.

5.4.1.1 MM5 Model Physics

Below is an outline of various physics parameterizations found within MM5, including the atmospheric radiation, precipitation, planetary boundary layer and the land surface scheme.

5.4.1.1.1 Atmospheric Radiation

The atmospheric radiation parameterization in MM5 (Dudhia 1989) interacts with the atmosphere, cloud and precipitation fields, and with the surface using a separate scheme for the longwave and the shortwave radiation.

Longwave radiation scheme

Water vapour is the primary clear air longwave absorber. Since this absorption is strongly spectral in character, MM5 utilizes the broadband emissivity method (Stephens 1984). This involves using a precalculated emissivity function, ε , which represents the frequency-integrated spectrum of water vapour, weighted by a suitable envelope function. The upward and downward emissivity are given as functions of the water vapour path modified by the temperature and pressure. Given these emissivity functions the upward (ε_u) and downward (ε_d) fluxes at any model level are given by

$$F^{\uparrow} = \int_0^{\infty} B(T) d\epsilon_u \quad (5.7)$$

and

$$F^{\downarrow} = \int_0^{\infty} B(T) d\epsilon_d \quad (5.8)$$

The quantity $d\epsilon$ is calculated for each layer using the temperature (T) of the layer and the frequency-integrated Plank function $B = \sigma_{SB} T^4$, where σ_{SB} is the Stefan-Boltzmann constant.

The method for dealing with cloud water can be found in Stephens (1978). Here the cloud water is assumed to have a constant coefficient which is slightly different for upward and downward radiation. The cloud absorption coefficient for upward radiation is $\alpha_{cu} = 0.130 \text{ m}^2 \text{g}^{-1}$ and for downward radiation $\alpha_{cd} = 0.158 \text{ m}^2 \text{g}^{-1}$. To combine these with water vapour absorption, the transmissivities are multiplied since clouds are assumed to be “gray bodies”, i.e. they affect the entire spectrum, not just particular wavelengths. The net emissivity is then

$$\epsilon_{tot} = 1 - T_v T_c \quad (5.9)$$

with

$$T_v = 1 - \epsilon_{vapour} \quad (5.10)$$

and

$$T_c = \exp(-\alpha_c u_c) \quad (5.11)$$

where u_c is the cloud water path (liquid mass per unit area).

Ice cloud is assumed to be composed of hexagonal plate-like crystals that do not reflect longwave radiation and are sufficiently thick to be “black”, it is possible to estimate an absorption coefficient as an integrated cross-sectional area. This absorption coefficient takes the value $\alpha_i = 0.0735 \text{ m}^2 \text{g}^{-1}$, or about half that of cloud water.

For rain and snow, MM5 takes into account the size distribution of the particles. The size spectrum changes with precipitation intensity so the absorption coefficient varies with precipitation amount. The effective absorption coefficient is given by

$$\alpha_p = \frac{1.66}{2000} \left(\frac{\pi N_0}{\rho_r^3} \right)^{1/4} \text{ m}^2 \text{ g}^{-1} \quad (5.12)$$

where ρ_r is the particle density. For constants used in the explicit moisture scheme the absorption coefficients take values of $2.34 \times 10^{-3} \text{ m}^2 \text{ g}^{-1}$ for snow and $0.33 \times 10^{-3} \text{ m}^2 \text{ g}^{-1}$ for rain. The effective water path for a layer of Δz meters thickness is given by

$$u_p = (\rho q_r)^{3/4} \Delta z \times 1000 \text{ gm}^{-2} \quad (5.13)$$

so that the transmissivity is given by

$$T_p = \exp(-\alpha_p u_p) \quad (5.14)$$

This transmissivity is multiplied by the others in equation 5.9 to give ε_{tot} . Note that rain and snow have less effect on the longwave flux by a few orders of magnitude.

Carbon dioxide is less easily treated since it cannot be assumed “gray”. That is its absorption is concentrated in a band of infrared wavelengths. An overlap method, discussed by Stephens (1984), is used to include this effect. In practical terms, the spectrum is divided into a carbon dioxide region and a non-carbon dioxide region. The former requires overlapping of the carbon dioxide transmissivity function while the latter does not. The relative weights of these two regions is slightly temperature dependent, but they add to give the total absorption.

Shortwave radiation scheme

The downward component of shortwave flux is evaluated taking into account the effects of: solar zenith angle, which influences the downward component and path length;

clouds, which have an albedo and absorption; and clear air, where there is scattering and water vapour absorption. Thus,

$$S_d(z) = \mu S_0 - \int_z^{top} (dS_{cs} + dS_{ca} + dS_s + dS_a) \quad (5.15)$$

where μ is the cosine of the zenith angle, S_0 is the solar constant and the subscript c is for clouds, s for scattering and a for absorption.

The cloud albedo and absorption are bilinearly interpolated from tabulated functions of μ and $\ln(w/\mu)$ (w is the vertically integrated cloud water path length) derived from Stephens' (1978) theoretical results.

Clear-air water vapour absorption is calculated as a function of water vapour path allowing for solar zenith angle. The absorption function is from Lacis and Hansen (1974). The method is a similar integration-difference scheme to that described above for cloud. Clear-air scattering is taken to be uniform and proportional to the atmosphere's mass path length, again allowing for the zenith angle, with a constant giving 20 percent scattering in one atmosphere.

The longwave and shortwave fluxes at the surface, calculated from the atmospheric radiative scheme, are used in the energy budget of the land surface.

5.4.1.1.2 Precipitation

Precipitation processes are essentially divided into those processes that can be resolved at the grid resolution, commonly referred to as “stable” precipitation, and those not resolved, referred to as “convective” precipitation. This terminology may be misleading at very fine grid resolutions where convective precipitation may be resolved but it will be used here in order to conform with the wider literature.

Stable precipitation

This scheme is activated whenever grid-scale saturation is reached. Essentially it removes super-saturation as precipitation and adds the latent heat to the thermodynamic equation taking into account additional variables such as cloud and rain water, snow and ice. The ice and snow treatments can be found in Dudhia (1989).

The mixing ratio equations for water vapour, q_v , cloud water (ice), q_c , and rain water (snow), q_r , are given below

$$\begin{aligned} \frac{\partial p^* q_v}{\partial t} = & -m^2 \left[\frac{\partial p^* u q_v / m}{\partial x} + \frac{\partial p^* v q_v / m}{\partial y} \right] - \frac{\partial p^* q_v \dot{\sigma}}{\partial \sigma} \\ & + p^* (-P_{RE} - P_{CON} - P_{II} - P_{ID}) + D_{qv} \end{aligned} \quad (5.16)$$

$$\begin{aligned} \frac{\partial p^* q_c}{\partial t} = & -m^2 \left[\frac{\partial p^* u q_c / m}{\partial x} + \frac{\partial p^* v q_c / m}{\partial y} \right] - \frac{\partial p^* q_c \dot{\sigma}}{\partial \sigma} \\ & + p^* (P_{ID} + P_{II} - P_{RC} - P_{RA} + P_{CON}) + D_{qc} \end{aligned} \quad (5.17)$$

$$\begin{aligned} \frac{\partial p^* q_r}{\partial t} = & -m^2 \left[\frac{\partial p^* u q_r / m}{\partial x} + \frac{\partial p^* v q_r / m}{\partial y} \right] - \frac{\partial p^* q_r \dot{\sigma}}{\partial \sigma} \\ & - \frac{\partial V_f \rho g q_r}{\partial \sigma} + p^* (P_{RE} + P_{RC} + P_{RA}) + D_{qr} \end{aligned} \quad (5.18)$$

where P_{CON} is condensation (and freezing for $T < 0^\circ\text{C}$) of water vapour into cloud (ice) at water saturation, P_{RA} is accretion of cloud by rain (ice by snow), P_{RC} is conversion of cloud to rain (ice to snow), P_{RE} is evaporation (sublimation) of rain (snow), P_{II} is the initiation of ice crystals, and P_{ID} is the sublimation/deposition of cloud ice. The fall speed of rain or snow is V_f .

The diameter of the relevant droplet is given by D which is assumed to follow a Marshall-Palmer size distribution. Droplet fall speeds are taken to be of the form $V(D) = aD^b$. For rain, the Marshall-Palmer intercept parameter is $N_0 = 8 \times 10^6 \text{ m}^{-4}$, $a = 841.99667$ and $b = 0.8$ for V in ms^{-1} and D in meters. For snow we have $N_0 = 2 \times 10^7 \text{ m}^{-4}$, $a = 11.72$ and $b = 0.41$.

For the latent heat we need to know about all changes in state of the water. As snow falls through the 0°C level, it immediately melts to rain, P_{RM} . Advection of ice or snow downwards or of rain or cloud upwards through this level also melts or freezes the particles, P_{MF} . Hence the latent heating is

$$\dot{Q} = L(P_{RE} + P_{ID} + P_{II} + P_{CON}) + L_m(P_{RM} + P_{MF}) \quad (5.19)$$

where $L = L_v$ (latent heat of condensation) for $T > 0^\circ\text{C}$ and $L = L_s$ (latent heat of sublimation) for $T < 0^\circ\text{C}$, and L_m is the latent heat of fusion, i.e. $L_m = L_s - L_v$.

Convective precipitation

In this implementation of MM5 the Grell scheme (Grell 1993) for convective precipitation is used. Figure 5.4 shows the simple conceptual picture of the way this parameterization functions. Clouds are pictured as two steady-state circulations, caused by an updraft and a downdraft. Direct mixing between cloudy air and environmental air occurs only at the top and bottom of the circulations.

With mass flux being constant with height and no entrainment or detrainment occurring along the cloud edges, we have

$$m_u(z) = m_u(z_b) = m_b \quad (5.20)$$

and

$$m_d(z) = m_d(z_0) = m_0 \quad (5.21)$$

where m_u and m_d are the mass flux of the updraft and downdraft respectively. Similarly z_b and z_0 are the originating levels of the updraft and downdraft and hence m_b and m_0 are simply the mass fluxes of the updraft and downdraft at their originating level.

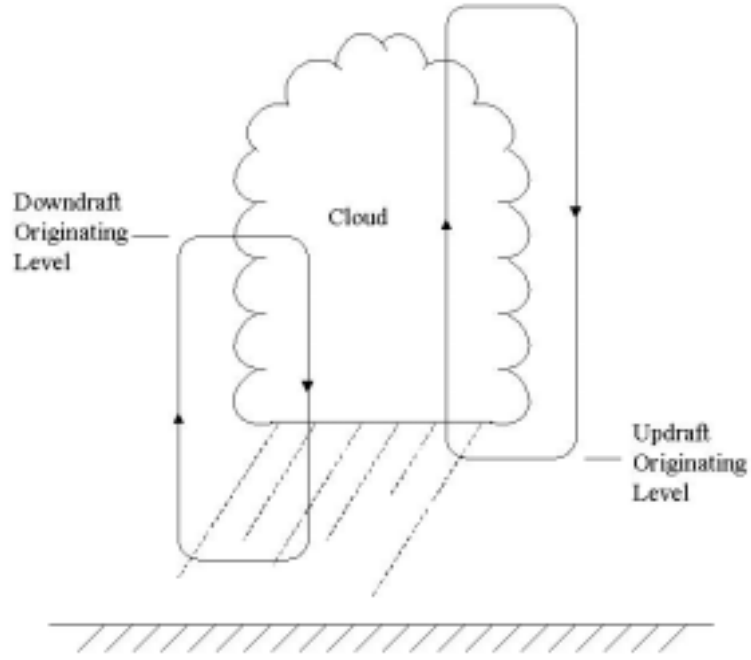


Figure 5.4 Conceptual diagram of convection parameterized in Grell scheme. Adopted from Grell et al. (1994).

Assuming that the conditions at originating levels is given by the environment, for any thermodynamic variable, the budget inside the cloud simply becomes

$$\alpha_u(z) = \tilde{\alpha}(z_b) + S_u(z) \quad (5.22)$$

and

$$\alpha_d(z) = \tilde{\alpha}(z_0) + S_d(z) \quad (5.23)$$

where α is a thermodynamic variable, the tilde denotes an environmental value and S represents sources and sinks. For moist static energy, h ,

$$\tilde{h}(z) = C_p \tilde{T}(z) + gz + L\tilde{q}(z) \quad (5.24)$$

equations 5.22 and 5.23 become

$$h_u(z) = \tilde{h}(z_b) \quad (5.25)$$

and

$$h_d(z) = \tilde{h}(z_0) \quad (5.26)$$

The originating levels of updraft and downdraft are given by the levels of maximum and minimum ambient moist static energy, respectively.

Given boundary conditions, equations 5.20 to 5.26 have two unknowns, m_b and m_0 . In order to leave only one unknown variable, the originating mass flux of the downdraft is made a function of the updraft mass flux and the re-evaporation of convective condensate.

$$m_0 = \frac{\beta I_1 m_b}{I_2} \quad (5.27)$$

where I_1 is the amount of condensation integrated over the whole depth of the updraft normalized by the updraft mass flux, I_2 is the evaporation in the downdraft normalized by the downdraft mass flux, and β is the fraction of updraft condensate that re-evaporates in the downdraft; that is, $1 - \beta$ is the precipitation efficiency. Rainfall is then given by

$$R = I_1 m_b (1 - \beta) \quad (5.28)$$

By compensating mass fluxes and detrainment at cloud top and bottom the heating and moistening feedback into the large scale can be determined. The scheme also includes the cooling effect of moist convective downdrafts.

In order to complete the scheme, a closure assumption is required to relate the mass flux at the bottom of the updraft to the large scale forcings. MM5 assumes that clouds stabilize the environment as fast as the large scale destabilizes it (Arakawa and Shubert 1974). This can be expressed as

$$m_b = \frac{ABE'' - ABE}{\Delta t NA} \quad (5.29)$$

where ABE is the buoyant energy available to a cloud, ABE'' is the production of available buoyant energy by the large scale motions during the time Δt , and NA is the rate of change of available buoyant energy per unit of mass flux.

5.4.1.1.3 Planetary Boundary Layer

The Planetary Boundary layer (PBL), also referred to as the atmospheric boundary layer, parameterization used in these configurations of MM5 was originally developed by Hong and Pan (1996) for inclusion in the National Center for Environmental Prediction Medium-Range Forecast model (NCEP MRF). They based the parameterization on the scheme proposed by Troen and Mahrt (1986) which has been further generalized and reformulated by others (Holtslag et al. 1990a, Giorgi et al. 1993b, Holtslag and Boville 1993). This approach is referred to as the “nonlocal- K ” approach. They start by expressing the turbulent diffusion equations for prognostic variables as

$$\frac{\partial \xi}{\partial t} = \frac{\partial}{\partial z} \left[K_\xi \left(\frac{\partial \xi}{\partial z} - \gamma_\xi \right) \right] \quad (5.30)$$

where ξ represents a prognostic variable, K_ξ is the eddy diffusivity coefficient and γ_ξ is a correction to the local gradient that incorporates the contribution of the large scale eddies to the total flux.

Above the PBL, free atmospheric diffusion is calculated using the “local- K ” approach (Louis 1979). Inside the PBL, Troen and Mahrt (1986) express the momentum diffusivity coefficient as

$$K_{zm} = kw_s z \left(1 - \frac{z}{h}\right)^p \quad (5.31)$$

where p is the profile shape exponent taken to be 2, k is the von Karman constant (= 0.4), z is the height from the surface, and h is the height of the PBL. The mixed layer velocity scale, w_s , is given by

$$w_s = u_* \phi_m^{-1} \quad (5.32)$$

where u_* is the surface frictional velocity scale, and ϕ_m is the wind profile function evaluated at the top of the surface layer. The gradient correction terms for the potential temperature, θ , and specific humidity, q , are given by

$$\gamma_\xi = b \frac{\overline{(w' \xi')}}{w_s} \quad (5.33)$$

where $\overline{(w' \xi')}$ is the corresponding surface flux for θ and q , and b is a coefficient of proportionality. Profile functions identical to those used in surface-layer physics are employed to ensure compatibility between the surface layer top and the bottom of the PBL. For unstable and neutral conditions

$$\phi_m = \left(1 - 12 \frac{0.1h}{L}\right)^{-1/3} \quad (5.34)$$

for u and v , and

$$\phi_t = \left(1 - 16 \frac{0.1h}{L}\right)^{-1/2} \quad (5.35)$$

for θ and q .

For the stable regime

$$\phi_m = \phi_t = \left(1 + 5 \frac{0.1h}{L}\right) \quad (5.36)$$

Here L is the Monin-Obukhov length scale. The top of the surface layer is estimated as $0.1h$. From the derivation of Troen and Mahrt (1986) we find that $b = 7.8$.

The PBL height is found by the iterative solution of

$$h = Rib_{cr} \frac{\theta_{va} |U(h)|}{g(\theta_v(h) - \theta_s)} \quad (5.37)$$

where Rib_{cr} is the critical bulk Richardson number, $U(h)$ is the horizontal wind speed at h , θ_{va} is the virtual potential temperature at the lowest model level, $\theta_v(h)$ is the virtual potential temperature at h , and θ_s is the appropriate temperature near the surface.

The eddy diffusivity for temperature and moisture, K_{zt} , is computed from K_{zm} in equation 5.31 by using the Prandtl number relationship

$$Pr = \left(\frac{\phi_t}{\phi_m} + bk \frac{0.1h}{h} \right) \quad (5.38)$$

where Pr is a constant within the whole mixed boundary layer.

5.4.1.2 Land Surface Parameterization

The MM5 model was run with two variations of the land surface parameterization. One parameterization used is the Biosphere-Atmosphere Transfer Scheme (BATS) version 1e which has been derived from Dickinson et al. (1986) and is described in Dickinson et al. (1993). The second parameterization is the spatially-distributed land surface flux model SHEELS, the Simulator for Hydrology and Energy Exchange at the Land Surface. SHEELS can be applied at local scales by assuming surface homogeneity within a grid cell, or at larger scales in which the sub-grid scale variability of

vegetation, soil and topography properties is represented statistically. SHEELS is based on BATS. SHEELS (Smith et al. 1993, Laymon and Crosson 1995) has retained the physical treatment of vegetation properties and the surface flux parameterizations of BATS, although the sub-surface processes in SHEELS differ significantly. The nested soil layer approach of BATS has been converted to a discrete layer configuration in SHEELS, in which the number and depth of layers are flexible.

5.4.1.2.1 Biosphere-Atmosphere Transfer Scheme (BATS)

The BATS model incorporates a single vegetation layer (or canopy layer), a multiple layer soil scheme, and provision for snow cover on the land surface. BATS requires the specification of 23 parameters at the beginning of an integration, 16 of these parameters are associated with each vegetation type, a further 6 parameters are associated with each soil texture class and the final parameter is related to the albedo (soil colour classes). BATS uses 18 distinct vegetation types, 12 soil texture and 8 soil colour classes.

When coupled to a climate model, the vegetation type, soil texture, and soil colour need to be specified for each grid point, along with the initial soil moisture, and ground and foliage temperatures. From the climate model, BATS requires as input: wind components, air density, temperature, and water vapour mixing ratio at the lowest atmospheric level, surface radiant fluxes at solar and infrared wavelengths, and precipitation. From these and other internally generated quantities, BATS calculates the temperature of the surface soil, deep soil, canopy foliage and canopy air, the soil moisture in three layers, snow cover, and surface fluxes of momentum, heat and moisture. The surface fluxes are then fed into the momentum, thermodynamics and water vapour equations of the climate model as lower boundary conditions.

The sensitivity of BATS to these parameters has been investigated in several papers. Wilson et al. (1987) used BATS in a stand-alone mode and performed many 10-day simulations to test the model sensitivity. Henderson-Sellers (1993) also used BATS in stand-alone mode but in order to overcome some of the limitations present in the Wilson et al. (1987) experiments, she used 2-year integrations and a factorial experimental

design. They concluded that BATS was most sensitive to extreme changes in vegetation roughness length, soil porosity and a factor describing the sensitivity of the stomatal resistance of vegetation to the amount of photosynthetically active solar radiation as well as two factor interactions which include vegetation roughness length.

While stand-alone experiments can be useful in simple land surface experiments, it has been shown that the technique can lead to misleading results. For example, Koster and Eagleson (1990) compared results derived from stand alone experiments, a single column (atmospheric) model and a climate model and showed that the results from the stand alone experiments were incompatible with the other two techniques due to the lack of feedbacks. Pitman (1993) performed sensitivity tests with BATS linked to a single column model. He found BATS to be most sensitive to the roughness length, the fractional vegetation cover and the rooting ratio which is significantly different from the sensitivities found in Henderson-Sellers (1993).

BATS has been used in many studies to date, for example Dickinson and Kennedy (1991), Giorgi (1991), Henderson-Sellers et al. (1993a) and Timbal and Henderson-Sellers (1998). It has generally been found to perform well, especially when compared to other land surface parameterizations. A schematic illustrating the features included in BATS is shown in figure 5.7.

Soil temperature

The soil temperature model is based on a generalization of the force restore method of Deardorff (1978) and is described in Dickinson (1988). The surface soil temperature, T_{g1} , is calculated using the differential equation below

$$C\Delta t \frac{\partial T_{g1}}{\partial t} + 2AT_{g1} = B \quad (5.39)$$

where Δt is the time step in seconds, and A is related to the diurnal frequency, C is related to the thermal inertia of freezing and B includes a term proportional to the net surface heating, h_s ,

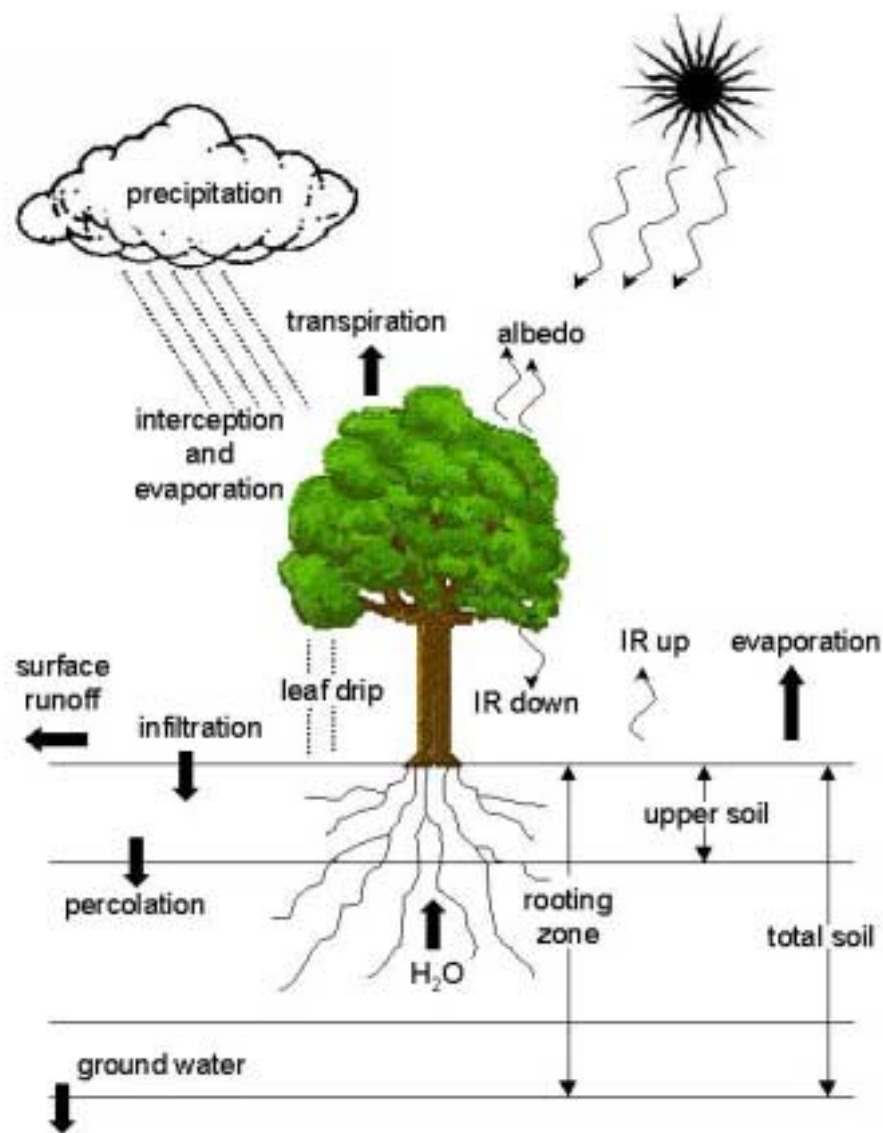


Figure 5.5 Schematic diagram showing the major features of BATS. Adapted from Dickinson et al. (1993).

$$h_s = S_g + F^\downarrow - F^\uparrow - F_s - L_{v,s}F_q - L_f S_m \quad (5.40)$$

where S_g is the solar flux absorbed over bare ground at the earth's surface, $F^\downarrow - F^\uparrow$ is the net IR (long wave) flux from atmosphere to bare ground, F_s is the atmospheric sensible heat flux from the ground to the atmosphere, F_q is the atmospheric moisture flux from the ground to the atmosphere, $L_{v,s}$ is the latent heat of evaporation or sublimation, L_f is the latent heat of fusion and S_m is the rate of snow melt.

These coefficients are then used to advance the surface soil temperature from the N th to $N+1$ th time step following

$$T_{g1}^{N+1} = \frac{B + (C - A - B')T_{g1}^N}{C + A - B'} \quad (5.41)$$

here B' is the derivative of B with respect to temperature.

The subsurface temperature is identified with the annual temperature wave and corresponds to temperature at a depth of roughly 1m. It is described in Dickinson (1988) as

$$(1 + F_{CT2})\Delta t \frac{\partial T_{g2}}{\partial t} + 2A_2 T_{g2} = c_4 \nu_a \Delta t T_{g3} + \frac{D_a}{D_d} \nu_a \Delta t \quad (5.42)$$

where c_4 is a coupling coefficient to soil untouched by annual wave. At present $c_4 = 0$, except under permafrost where we take $c_4 = 1$, $T_{g3} = 271.0$ (deep soil temperature). The seasonal frequency is $\nu_a = \nu_d / 365$. The term A_2 is given by

$$A_2 = \left(c_4 + \frac{D_a}{D_d} \right) 0.5 \nu_a \Delta t \quad (5.43)$$

in the absence of snow

$$D_a = \left(\frac{\nu_d}{\nu_a} \right)^{1/2} D_d \quad (5.44)$$

and in the presence of snow, both D_a and D_d are weighted averages according to the depth of snow.

$$F_{CT2} = \frac{\sqrt{2}L_f (S_{rw} - F_{rr})}{\rho_s c_s \Delta T Z_r} \geq 0 \quad (5.45)$$

where S_{rw} is the rooting zone soil water, Z_r is the depth of the soil rooting layer, and $F_{rr} = 0.15Z_r$ the unfrozen soil water.

Water fluxes

Once precipitation reaches the ground it either infiltrates the surface or is lost to surface runoff. For water, the soil is represented by three layers, all have a top surface at the soil air interface, but with lower surface at increasing depth. Parameters used to represent soil moisture include:

S_{sw} = surface soil water representing water in the upper layer (depth Z_u) of soil

S_{swmax} = maximum upper soil water

S_{rw} = water in the rooting zone depth Z_r of soil

S_{rwmax} = maximum root zone soil water

S_{tw} = total water in the soil to depth Z_t

S_{twmax} = maximum total water

Since all three layers reach the soil surface they all gain the same amount of water from rainfall, P_r , and lose the same amount from evaporation, F_q , and surface runoff, R_s .

Fluxes between soil layers affect the different (but overlapping) reservoirs differently. Their conservation equations are written

$$\frac{\partial S_{sw}}{\partial t} = P_r(1 - \sigma_f) - R_s + \Psi_{w1} - \eta E_{tr} - F_q + S_m + D_w \quad (5.46)$$

$$\frac{\partial S_{rw}}{\partial t} = P_r(1 - \sigma_f) - R_s + \Psi_{w2} - E_{tr} + S_m + D_w \quad (5.47)$$

$$\frac{\partial S_{tw}}{\partial t} = P_r(1 - \sigma_f) - R_s - R_g - E_{tr} - F_q + S_m + D_w \quad (5.48)$$

where σ_f is the fractional foliage cover for each grid point, R_s is the surface runoff, Ψ_w is the rate of transfer of water by diffusion to the upper soil layer from the lower, η is the fraction of transpiration from the top soil layer, E_{tr} is the transpiration, F_q is the moisture flux from the ground to the atmosphere (negative F_q represents dew formation), S_m is the rate of snow melt, D_w is the rate of excess water dripping from leaves per unit land area, and R_g is the ground water runoff.

Infiltration and percolation to ground water depends on various soil properties, in particular the soil texture (e.g. Clapp and Hornberger 1978). Water diffuses through the soil with a diffusivity

$$D_s = K_{w0} \phi_0 B s^{B+2} \quad (5.49)$$

where K_{w0} is the saturated hydraulic conductivity, ϕ_0 is the soil water suction for saturated soil and B is the exponent defined in Clapp and Hornberger (1978). Values for these parameters can be found in Table 3 of Dickinson et al. (1993). Also s is the volume of water divided by volume of water at saturation.

Besides the diffusive movement, there is gravitational drainage which dominates the flow for large enough length scales. This provides the subsoil drainage (groundwater runoff) expression

$$R_g = K_{w0} s^{2B+3} \quad (5.50)$$

In order to calculate the evaporative terms the following parameterization is adopted

$$F_q = \min(F_{qp}, F_{qm}) \quad (5.51)$$

where F_{qp} is the potential evaporation and F_{qm} is the maximum moisture flux through the wet surface that the soil can sustain. F_{qp} is calculated, using equation 5.58 with $f_g = 1$, as the evaporation from a wet surface with the same aerodynamic characteristics as the soil surface. Details on the calculation of F_{qm} can be found in Dickinson et al. (1993).

Movement of water from the rooting zone into the surface soil layer is parameterized by

$$\Psi_{w1} = C_{\beta 1} (s_1 - s_2) \quad (5.52)$$

and from the total column into the rooting zone by

$$\Psi_{w2} = C_{\beta 2} (s_0 - s_1) \quad (5.53)$$

where the coefficients $C_{\beta 1}$ and $C_{\beta 2}$ are defined in Dickinson et al. (1993).

Guided by the criteria that there should be small surface runoff at the soil moisture of field capacity and complete surface runoff at saturated soil, surface runoff is parameterized by

$$R_s = \begin{cases} \left(\frac{\rho_w}{\rho_{wsat}} \right)^4 G & T_{g1} \geq 0^\circ C \\ \left(\frac{\rho_w}{\rho_{wsat}} \right) G & T_{g1} < 0^\circ C \end{cases} \quad (5.54)$$

where ρ_{wsat} is the saturated soil water density and ρ_w is the soil water density weighted toward the top layer, as defined by

$$\rho_w = \rho_{wsat} \frac{s_1 + s_2}{2} \quad (5.55)$$

also

$$G = P_r + S_m - F_q \quad (5.56)$$

The sensible heat fluxes are obtained using

$$F_s = \rho_a C_p C_D U (T_{g1} - T_a) \quad (5.57)$$

where ρ_a is the surface air density, C_p the specific heat for air, C_D is the drag coefficient over land, and U the wind speed. Similarly, the moisture flux (from the surface) to the atmosphere F_q is given by

$$F_q = \rho_a C_D U f_g (q_g - q_a) \quad (5.58)$$

where q_g is the saturated specific humidity at the temperature of the surface, q_a is the specific humidity of the model lowest level, and f_g is a wetness factor, which has the value of 1.0 except for diffusion-limited soil surfaces, where it is defined by the ratio of actual to potential ground evaporation

$$f_g = \frac{F_q}{F_{qp}} \quad (5.59)$$

Energy fluxes

Downward radiation is partially reflected from the surface depending on the albedo.

Values for vegetation albedo are prescribed and can be found in Table 2 of Dickinson et al. (1993). The shortwave albedo for bare soil, A_{LBG} , is taken to be

$$A_{LBG} = A_{LBGO} + 0.01 \left(11 - \frac{40 S_{sw}}{Z_u} \right) \quad (5.60)$$

where A_{LBGO} is the albedo for a saturated soil. The longwave albedo of bare soils is twice the shortwave albedo. Dry and saturated soil albedos can be found in Table 3 of Dickinson et al. (1993).

The treatment of vegetation is an extension of the one-layer “big leaf” approach of Monteith (1981). BATS has separate ground-energy equations, resistances for transfer from above the canopy to air within the canopy and from air within the canopy to the foliage surfaces, as well as allowing for partial wetting of the canopy. At each land grid point a fractional vegetation cover, σ_f , is prescribed.

The surface area of vegetation per unit area of ground consists of transpiring surfaces specified by a leaf area index (L_{AI}) and non-transpiring surfaces (including dead vegetation) specified by a stem area index (S_{AI}). The S_{AI} is constant for each land type, while the L_{AI} has a seasonal variation. The fractional area of leaves and stems covered by water, $\tilde{L}_w$, is required in order to determine the evaporation from them. Following Deardorff (1978)

$$\tilde{L}_w = \left(\frac{W_{dew}}{W_{DMAX}} \right)^{2/3} \quad (5.61)$$

where W_{dew} is the total water intercepted by the canopy and W_{DMAX} is the maximum water the canopy can hold. The fraction of foliage surface free to transpire, L_d , is then defined by

$$L_d = \frac{(1 - \tilde{L}_w)L_{AI}}{L_{AI} + S_{AI}} \quad (5.62)$$

Values of L_{AI} and S_{AI} depend on vegetation type and can be found in Table 2 of Dickinson et al. (1993).

The water on wet foliage evaporates per unit wetted area according to

$$E_f^{WET} = \rho_a r_{la}^{-1} (q_f^{SAT} - q_{af}) \quad (5.63)$$

where q_f^{SAT} is the saturation specific humidity at the temperature of the foliage T_f , q_{af} is the specific humidity ratio of air within the canopy, and r_{la} is the aerodynamic resistance to moisture and heat flux of the foliage molecular boundary layer per unit foliage projected area. Equation 5.62, if negative, gives the rate of accumulation of dew.

The conductance of heat and vapour flux from the foliage is given by

$$r_{la}^{-1} = C_f \times \left(\frac{U_{af}}{D_f} \right)^{1/2} \quad (5.64)$$

where $C_f = 0.01 \text{ms}^{-1/2}$, D_f is the characteristic dimension of the leaves in the direction of wind flow, and U_{af} is the magnitude of the wind velocity within the foliage layer.

Similar to equation 5.63, the heat flux from the foliage is given by

$$H_f = \sigma_f (L_{AI} + S_{AI}) r_{la}^{-1} \rho_a C_p (T_f - T_{af}) \quad (5.65)$$

The evaporative flux from canopy surfaces that are only partly wet now follows from equation 5.63 as

$$E_f = r'' E_f^{WET} \quad (5.66)$$

where

$$r'' = 1 - \delta(E_f^{WET}) \left[1 - \tilde{L}_w - L_d \left(\frac{r_{la}}{r_{la} + r_s} \right) \right] \quad (5.67)$$

where r_s is the ‘stomatal’ resistance, $\tilde{L}_w$ and L_d are defined by equations 5.61 and 5.62 respectively and where δ is a step function that is one for positive argument and zero for zero or negative argument.

Transpiration occurs only from dry leaf surfaces

$$E_{tr} = \delta(E_f^{WET}) L_d \left(\frac{r_{la}}{r_{la} + r_s} \right) E_f^{WET} \quad (5.68)$$

The term stomatal resistance, as used here, refers to the total mechanical resistance encountered by diffusion from inside a leaf to outside. The stomatal resistance factor is taken to be

$$r_s = r_{s \min} \times R_f \times S_f \times M_f \times V_f \quad (5.69)$$

The factors on the right hand side have been discussed by others, e.g. Jarvis (1976) and Hinckley et al. (1978). R_f represents the dependence of the stomatal resistance on solar radiation, S_f represents the dependence on temperature, M_f represents the dependence on soil moisture and V_f represents the dependence on vapour pressure deficit.

All these factors are discussed in more detail in Dickinson et al. (1993). In particular M_f depends on the ability of plant roots to take water from the soil for a given level of root moisture. Essentially M_f is used to guarantee that transpiration does not exceed a maximum value.

The air within the canopy has negligible heat capacity and so heat flux from the foliage, H_f , and from the ground, H_g , must be balanced by heat flux to the atmosphere, H_a ,

$$H_a = H_f + H_g \quad (5.70)$$

where H_f was given in equation 5.65 and flux to the atmosphere is given by

$$H_a = \rho_a \sigma_f C_p C_D U_a (T_{af} - T_a) \quad (5.71)$$

with C_p being the specific heat of air, U_a the magnitude of atmospheric wind above the canopy, and C_D the aerodynamic bulk transfer coefficient between canopy air and atmosphere above, assumed to be the same for heat and moisture as for momentum. The flux from the soil under the canopy is assumed to be

$$H_g = \rho_a C_p \{C_{SOILC} \sigma_f U_{af} \} (T_{g1} - T_{af}) \quad (5.72)$$

where C_{SOILC} assumed to be 0.004 is the transfer coefficient between canopy air and underlying soil.

Now solving equation 5.70 for T_{af} we get

$$T_{af} = \frac{c_A T_a + c_F T_f + c_G T_{g1}}{c_A + c_F + c_G} \quad (5.73)$$

where

$$c_A = \sigma_f C_D U_a \quad (5.74)$$

$$c_F = \sigma_f (L_{AI} + S_{AI}) r_{la}^{-1} \quad (5.75)$$

$$c_G = C_{SOILC} \sigma_f U_{af} \quad (5.76)$$

are conductances for heat flux, to the atmosphere above the canopy, and from the foliage, and the ground, respectively. Similarly the canopy air is assumed to have zero capacity for water vapour storage so that the flux of water from the canopy air, E_a , balances the flux from the foliage, E_f , and the flux from the ground, E_g ,

$$E_a = E_f + E_g \quad (5.77)$$

where E_f was defined in equation 5.66, and

$$E_a = \rho_a c_A (q_{af} - q_a) \quad (5.78)$$

$$E_g = \rho_a c_G f_g (q_{g,s} - q_{af}) \quad (5.79)$$

The quantity $q_{g,s}$ is the saturated soil water vapour concentration and f_g is a wetness factor, defined as the ratio of actual to potential ground evaporation as obtained from the soil moisture parameterization. Solving equations 5.77 – 5.79 for q_{af} gives

$$q_{af} = \frac{c_A q_a + c_V q_f^{\text{SAT}} + c_G f_g q_{g,s}}{c_A + c_V + f_g c_G} \quad (5.80)$$

where $c_V = r'' c_F$ is the average conductance of foliage to water vapour flux.

5.4.1.2.2 Simulator for Hydrology and Energy Exchange at the Land Surface (SHEELS)

As mentioned earlier, the physics of SHEELS are based on those present in BATS. The main difference between them occurring in the sub-surface hydrologic processes. Instead of the nested three layer approach of BATS, SHEELS uses a discrete layer approach with five 2cm thick layers in the top 10cm of soil, a root zone containing three 30cm thick layers and a lower zone extending to 10m depth and divided into three layers.

By considering the contributions of infiltration, evaporation, transpiration, diffusion and gravitational drainage, SHEELS determines the change in soil moisture content in each of the soil layers. The Green-Ampt equation is used to calculate the infiltration, I , based on the amount of precipitation reaching the soil surface. Surface runoff is based on the local slope angle (ϕ) and infiltration excess:

$$R_u = (P - I) \bullet \sin \phi \quad (5.81)$$

The change in depth of water in soil layer i (d_i) due to water exchange with the atmosphere can be expressed as

$$\frac{\partial d_i}{\partial t} = I_i - E_i - T_i \quad (5.82)$$

where I_i , E_i and T_i are the amounts of infiltration, evaporation and transpiration attributed to layer i . These terms are proportions of the total quantities I , E and T , and are determined by applying weighting functions for each variable.

The vertical fluxes of moisture within the soil are formulated using Darcy's law

$$q_\theta = -K \frac{\partial \psi}{\partial z} - K \quad (5.83)$$

where q_θ is the vertical water flux, θ is the volumetric water content, $K=K(\theta)$ = hydraulic conductivity and $\psi = \psi(\theta)$ = hydraulic matric potential.

Applying mass continuity and expressing terms as functions of θ yields the diagnostic equation

$$\frac{\partial \theta}{\partial t} = \frac{\partial q_\theta}{\partial z} = \frac{\partial}{\partial z} \left[D(\theta) \frac{\partial \theta}{\partial z} \right] + G(\theta) \frac{\partial \theta}{\partial z} \quad (5.84)$$

where $D(\theta) = K \frac{\partial \psi}{\partial \theta}$ = diffusion coefficient, and

$$G(\theta) = \frac{\partial K}{\partial \theta} = \text{gravitational coefficient.}$$

The diffusion term is solved using the Crank-Nicholson numerical scheme, and the functions $\psi(\theta)$ and $K(\theta)$ are given by the empirical parameterizations of Clapp and Hornberger (1978).

5.4.2 Description of the RegCM2 Limited Area Model

Included here is a brief description of the second-generation regional climate model (RegCM2) developed at NCAR. A more complete description can be found in Giorgi et al. (1993b), Giorgi et al. (1993c) and Shields et al. (1994). RegCM2 has been used in many studies to date (Giorgi (1991), Bates et al. 1993, Giorgi et al. 1994, Bates et al. 1995, Small et al. 1999b).

The original NCAR regional climate modelling system, RegCM, was derived from MM5's predecessor, MM4. Many of MM4's physics parameterizations were modified to adapt its use to long term climate simulation. In a sense RegCM2, MM5/BATS and MM5/SHEELS are parallel developments of MM4 intended to improve the climate simulation. As such they share many common features, and only those features which are different will be expanded upon below.

As with MM5, this implementation of RegCM2 is a hydrostatic, compressible, primitive equation, terrain following σ vertical coordinate model. Here σ is defined as

$$\sigma = \frac{p - p_{top}}{p_s - p_{top}} \quad (5.85)$$

where p is the pressure, p_{top} is the pressure specified to be the model top and is constant, and p_s is the prognostic surface pressure.

The spatial grid uses a Lambert conformal projection similar to MM5. The lateral boundary conditions used differ from those used in MM5, and are based on the relaxation procedure of Davies and Turner (1977). This procedure uses Newtonian and diffusion terms to gradually drive the model solution toward specified large scale values inside a buffer area. For the variable α this can be written as

$$\left(\frac{\partial \alpha_{MC}}{\partial t} \right)_n = F(n)F_1(\alpha_{LS} - \alpha_{MC}) - F(n)F_2 \nabla^2(\alpha_{LS} - \alpha_{MC}) \quad (5.86)$$

where F_1 and F_2 are given by

$$F_1 = \frac{0.1}{\Delta t} \quad (5.87)$$

$$F_2 = \frac{\Delta s^2}{50\Delta t} \quad (5.88)$$

Here Δt is the model time step and Δs is the model gridpoint spacing. $F(n)$ is given the exponential functional form

$$F(n) = \exp \left[-\frac{(n-2)}{N_I} \right] \quad (5.89)$$

where N_I is a constant. Here n has a maximum of 8 which means that the eight outermost grid points are directly effected by the lateral boundary conditions while for MM5 only the 5 outermost gridpoints are effected.

5.4.2.1 RegCM2 Model Physics

5.4.2.1.1 Atmospheric Radiation

The RegCM2 radiation parameterization is the same as that found in the NCAR Community Climate Model version 2 (CCM2) (Hack et al. 1993). This scheme interacts with the atmosphere and clouds as well as explicitly accounting for absorption/emission by O_3 , H_2O , CO_2 and O_2 .

Longwave radiation

Longwave fluxes are calculated in both up and down directions for each model level. Absorptivities and emissivities are employed to solve the transfer equations. Thus, the clear sky fluxes at a half level k are

$$F_{clr}^{\downarrow}(p_k) = B(0)\varepsilon(0, p_k) + \int_0^{p_k} \alpha(p', p_k) \frac{dB}{dp'}(p') dp' \quad (5.90)$$

and

$$F_{clr}^{\uparrow}(p_k) = \sigma_{SB} T_S^4 - \int_{p_k}^{p_s} \alpha(p', p_k) \frac{dB}{dp'}(p') dp' \quad (5.91)$$

where $B(p) = \sigma_{SB} T^4(p)$ is Stefan-Boltzmann's law, and the *absorptivity* is defined as

$$\alpha(p, p') = \frac{1}{\frac{dB}{dT}(p')} \int A_{\tilde{\nu}}(p', p) \frac{dB_{\tilde{\nu}}}{dT}(p') d\tilde{\nu} \quad (5.92)$$

and the *emissivity* is

$$\varepsilon(o, p) = \frac{1}{B(0)} \int A_{\tilde{\nu}}(o, p) B_{\tilde{\nu}}(0) d\tilde{\nu} \quad (5.93)$$

where $A_{\tilde{\nu}}$ is the absorptivity due to a given gas, $B_{\tilde{\nu}}(p')$ is Planck's function, and $\tilde{\nu}$ is the wavenumber.

For CO_2 and O_3 , the band-absorptance technique is used to evaluate α and ε . This method uses the fact that gas absorption is limited to a finite spectral width. Planck functions are evaluated at the center of the bands, and integration over $\tilde{\nu}$ is carried out for $A_{\tilde{\nu}}$. Thus,

$$\alpha_{\text{CO}_2}(p, p') = \frac{1}{4\sigma_{SB} T^3(p')} \frac{dB_{\text{CO}_2}}{dT'}(p') A_{\text{CO}_2}(p', p) \quad (5.94)$$

where B_{CO_2} is evaluated for $\tilde{\nu} = 667 \text{ cm}^{-1}$ and $A_{\text{CO}_2}(p', p)$ is the broad-band absorptance from Kiehl and Briegleb (1991). Similarly,

$$\varepsilon_{CO_2}(0, p) = \frac{1}{\sigma_{SB} T^4(0)} B_{CO_2}(0) A_{CO_2}(0, p) \quad (5.95)$$

For ozone,

$$\alpha_{O_3}(p, p') = \frac{1}{4\sigma_{SB} T^3(p')} \frac{dB_{O_3}}{dT'}(p') A_{O_3}(p', p) \quad (5.96)$$

and

$$\varepsilon_{O_3}(0, p) = \frac{1}{\sigma_{SB} T^4(0)} B_{O_3}(0) A_{O_3}(0, p) \quad (5.97)$$

where A_{O_3} is the ozone broad-band absorptance from Ramanathan and Dickinson (1979). Voigt line profile effects for CO_2 and O_3 are included using the method described in Kiehl and Briegleb (1991). The method of Ramanathan and Downey (1986) is used for water vapor absorptivities and emissivities as well as the overlap treatment between water vapour and other gases. The total absorptivity is given by

$$\alpha(p, p') = \alpha_{CO_2}(p, p') + \alpha_{O_3}(p, p') + \alpha_{H_2O}(p, p') \quad (5.98)$$

and the total emissivity is

$$\varepsilon(0, p) = \varepsilon_{CO_2}(0, p) + \varepsilon_{O_3}(0, p) + \varepsilon_{H_2O}(0, p) \quad (5.99)$$

Clear-sky fluxes are then obtained by integrating equations 5.92 and 5.93. The downward longwave clear-sky flux at the surface is then

$$F_{clr}^{\downarrow}(p_s) = B(0)\varepsilon(0, p_s) + \int_0^{p_s} \alpha(p', p_s) \frac{dB}{dp'}(p') dp' \quad (5.100)$$

while the upward flux at the surface is just

$$F^{\uparrow}(p_s) = \sigma_{SB} T_s^4 \quad (5.101)$$

The downward cloudy-sky flux at the surface is

$$F^{\downarrow}(p_s) = F_{clr}^{\downarrow}(p_s) f_{clear} + \sigma_{SB} T^4(p_{clbk}) A'(p_{clbk}) + \sum_{k=3}^{KMAX} \left\{ \sigma_{SB} T^4(p_{clbk}) + \int_{p_{clbk}}^{p_s} \alpha(p', p_s) \frac{dB}{dp'}(p') dp' \right\} f_{cld}(k) \quad (5.102)$$

where p_{clbk} is the pressure level of the cloud base at k .

The cloud emissivity is accounted for by defining an effective cloud amount for each model layer, where the broadband emissivity, ε , is defined as

$$\varepsilon(p_k) = 1 - \exp(-0.1LWP(k)) \quad (5.103)$$

where the cloud liquid water path (u_c) is calculated using a prescribed, meridionally and height varying, but time independent, cloud water density (ρ_l):

$$u_c(k) = \rho_l^0 h_l \left[e^{-z_{k+1}/h_l} - e^{-z_k/h_l} \right] \quad (5.104)$$

where z_k is the height of the k th layer interface, h_l is a meridionally varying, empirically derived local liquid water scale, and ρ_l^0 is equal to 0.18gm^{-3} .

Shortwave radiation

The shortwave radiation is dealt with using the δ -Eddington approximation of Joseph et al. (1976) and also Coakley et al. (1983) and is described in Briegleb (1992). It has been shown that this approximation simulates the effects of multiple scattering quite well.

The solar spectrum is divided into 18 discrete spectral intervals (7 for O_3 , 1 for the visible, 7 for H_2O , and 3 for CO_2). The RegCM2 model atmosphere is composed of a discrete vertical set of horizontally homogenous layers within which radiative heating rates are to be specified. Within each of these layers several radiatively active constituents are present in a homogenous mix. Solar irradiance, the cosine of the solar

zenith angle, and surface reflectivity for direct and diffuse radiation in each spectral interval, are specified.

The δ -Eddington method for RegCM2 involves evaluating the δ -Eddington solution for the reflectivity and transmissivity for each layer in the vertical. The layers are then combined together, accounting for multiple scattering between the layers, which allows evaluation of upward and downward spectral fluxes at each interface boundary between layers. This procedure is repeated for all spectral intervals to accumulate broadband fluxes, from which the heating rate can be evaluated from flux differences across each layer.

The δ -Eddington approximation allows for gaseous absorption by O₃, CO₂, O₂ and H₂O. Molecular scattering and cloud water droplet scattering/absorption are included. A summary of the spectral intervals and the absorption/scattering data used in the formulation is given in Briegleb (1992).

For cloud scattering and absorption, the radiative parameterization of Slingo (1989) for liquid water droplet clouds is employed. In this parameterization, the optical properties of the cloud droplets are represented in terms of the prescribed liquid water path (u_c , see equation 5.104) and effective radius,

$$r_e = \int r^3 n(r) dr / \int r^2 n(r) dr \quad (5.105)$$

where $n(r)$ is the cloud drop size distribution over radius r

The basic physical dependencies of the optical properties on the effect radius, r_e , can be expressed in the parametric equations

$$\tau_c = u_c \left(a + \frac{b}{r_e} \right) \quad (5.106)$$

$$\omega = 1 - c - dr_e \quad (5.107)$$

$$g = e + fr_e \quad (5.108)$$

The quantity τ_c is the cloud extinction optical depth (0 to ∞), ω is the particle single scattering albedo (0 to 1), g is the asymmetry parameter (-1 to +1), and $a...f$ are positive constant coefficients for 4 spectral ranges: 0.25-0.69 μm , 0.69-1.19 μm , 1.19-2.38 μm , and 2.34-4.00 μm . For use within RegCM2 though, the cloud droplet effective radius is fixed at 10 μm .

Partial cloudiness and cloud overlap radiative effects are represented in the following manner. A parameterization that gives results approximately equal to the random overlap assumption, *without* the computational cost of calculating the spectrum of cloud cases and which gives the proper limits of zero cloud cover and complete cloud cover in a single layer, is utilized. The cloud extinction optical depth (τ_c) for each layer is modified as: $\tau'_c = \tau_c A_c^{3/2}$ and A_c is the fractional cloud cover in the layer; the power $3/2$ was found necessary to give results approximately the same as the random overlap assumption.

The δ -Eddington scheme is implemented so that the solar radiation is evaluated once every model hour over the sunlit portions of the model domain. The surface albedo is specified in two wavebands (0.2-0.7 μm , and 0.7-5.0 μm) and distinguishes albedos for direct and diffuse incident radiation. Albedos for ocean surfaces, geographically varying land surfaces and sea ice surfaces are distinguished. Ozone is prescribed, and CO_2 is assumed to be uniformly mixed with constant mass mixing ratio. Diagnostic cloud amount (A_c) is evaluated every model hour just prior to the solar radiation calculation.

5.4.2.1.2 Precipitation

Stable precipitation

Stable precipitation occurs in two ways: first, all supersaturated water vapour at a given grid point instantaneously precipitates; secondly, the more comprehensive explicit

approach of Hsie et al. (1984) is used. This approach is similar to the scheme in MM5 as described in section 4.1.1.2. Hsie extended the dry model of Keyser and Anthes (1982) to account for moist conditions by adding prognostic equations for water vapour, cloud water and rainwater. These equations are directly comparable to equations 5.16 to 5.18.

$$\begin{aligned} \frac{\partial p^* q_v}{\partial t} = & -\frac{\partial p^* q_v u}{\partial x} - \frac{\partial p^* q_v \dot{\sigma}}{\partial \sigma} - p^* v \frac{\partial q_v}{\partial y} \bigg|_p + p^* (P_{RE} - P_{CON}) \\ & + f_{\rho_s} (u_{gs} + c) \frac{\partial \sigma v q_v}{\partial \sigma} + p^* F_H q_v + p^* F_V q_v \end{aligned} \quad (5.109)$$

$$\begin{aligned} \frac{\partial p^* q_c}{\partial t} = & -\frac{\partial p^* q_c u}{\partial x} - \frac{\partial p^* q_c \dot{\sigma}}{\partial \sigma} - p^* (P_{RA} + P_{RC} - P_{CON}) \\ & + f_{\rho_s} (u_{gs} + c) \frac{\partial \sigma v q_v}{\partial \sigma} + p^* F_H q_c + p^* F_V q_c \end{aligned} \quad (5.110)$$

$$\begin{aligned} \frac{\partial p^* q_r}{\partial t} = & -\frac{\partial p^* q_r u}{\partial x} - \frac{\partial p^* q_r \dot{\sigma}}{\partial \sigma} + p^* (P_{RA} + P_{RC} - P_{RE}) \\ & - g \frac{\partial \rho q_r v_t}{\partial \sigma} + f_{\rho_s} (u_{gs} + c) \frac{\partial \sigma v q_v}{\partial \sigma} + p^* F_H q_r \end{aligned} \quad (5.111)$$

where u and v are the cross front and along front components of velocity, u_g is the geostrophic wind in the x direction, F_H is the horizontal diffusion and F_V represents vertical turbulent mixing effects. The coordinate system moves towards the east with a constant phase speed c . Other terms are described in section 4.1.1.2. Similarly to the scheme in MM5, a Marshall-Palmer distribution is assumed for raindrop size.

Convective precipitation

The convective precipitation scheme implemented in RegCM2 is the Grell scheme (Grell 1993). This is the same scheme used in MM5 and described in some detail in section 4.1.1.2. The stable and convective precipitation schemes are coupled as described by Zhang et al. (1988).

5.4.2.1.3 Planetary boundary layer

The parameterization of the PBL present in RegCM2 is based on Holtslag et al. (1990b). It is a nonlocal- K approach based on Troen and Mahrt (1986) and is quite similar to the approach taken in MM5 and described in section 4.1.1.3. The two schemes though differ in their representation of the wind profile function. The wind profile functions used in MM5 (equations 4.30 to 4.32) are based on the profile functions of Businger et al. (1971), while here the velocity scale has been rewritten in terms of the profile functions of Dyer (1974). For stable conditions we have

$$\phi_t = \begin{cases} 1 + 5\frac{z}{L} & 0 \leq \frac{z}{L} \leq 1 \\ 5 + \frac{z}{L} & \frac{z}{L} > 1 \end{cases} \quad (5.112)$$

which is comparable to equation 5.36 but with a change under very stable conditions ($z/L > 1$). Under unstable conditions we have

$$\phi_m = \left(1 - 15\frac{z}{L}\right)^{-1/3} \quad (5.113)$$

$$\phi_t = \left(1 - 15\frac{z}{L}\right)^{-1/2} \quad (5.114)$$

These equations are equivalent to equations 5.34 and 5.35 in MM5's PBL scheme. The PBL height and eddy diffusivity for temperature and moisture are then found using equations 5.37 and 5.38, as outlined in section 5.4.1.1.3.

5.4.2.2 Land Surface Parameterization

RegCM2 uses the BATS version 1e (Dickinson et al. 1993) land surface scheme as its land surface parameterization. A description of this scheme can be found in section 5.4.1.2.1. It is the same as the BATS implementation used with MM5.

5.4.3 Basic methodology

5.4.3.1 Types of simulation

McGregor (1997) presents a table containing the resolution, duration and region covered by LAM simulations performed throughout the world. Model simulations have been performed with resolutions ranging from 20km up to 0.5°. LAMs have been used in many regions around the world though Europe, the USA and Australia have received the most attention probably due to the fact that the modelling groups exist predominantly in these locations.

With regard to duration, LAM simulations have been performed using a variety of options. One option is to use a perpetual month approach, as used in McGregor and Walsh (1993) which consisted of a 300-day perpetual January run. That is, the LAM was forced using GCM data for a particular January, repeated 10 times. Another is to run a sequence of individual simulations for a given month driven by different years of an analysis or GCM. More recently multi-year seasonally varying runs have become more common. These allow soil moisture and temperature to evolve realistically over longer time scales.

5.4.3.2 Nesting procedures

LAMs may be nested in either analysis or GCM output. Here analysis refers to the use of a GCM to interpolate recent past climate using whatever observations are available. Typically the output is available two or three times per day, and these are interpolated in time and space to the lateral boundary points as required during the simulation. One-way nesting involves the LAM being driven at the boundaries by the analysis or GCM, without providing any feedback to the analysis or GCM itself. This type of nesting is the most common since analysis or GCM data can be archived and used to drive a LAM simulation at a later date once a region of interest has been established. Two-way nesting involves feedbacks from the LAM to the GCM, this nesting is considerably more computationally expensive since the GCM and LAM are necessarily run simultaneously.

The GCM (or analysis) communicates to the LAM through the lateral boundary. There are many options for the provision of lateral boundary conditions. In Anthes et al. (1987) five different schemes which have been implemented with their mesoscale model are listed. These are: fixed; time-dependent; time-dependent and inflow/outflow-dependent; sponge (Perkey and Kreitzberg 1976); and relaxation (Davies and Turner 1977). Other variations also exist such as that found in Giorgi et al. (1993a) where they modified the lateral boundary condition for water vapour to include a zero gradient condition for outflow.

Some LAM based studies have also used the double-nesting method (McGregor and Walsh 1994). This involves nesting a LAM within a GCM using a relatively coarse grid, and nesting another LAM within this LAM using a much finer grid.

5.4.3.3 Initialization

As well as needing lateral boundary conditions for the entire duration of the simulation, LAMs also need to be initialized throughout the entire domain. This includes initializing surface, soil and vegetation characteristics that gain importance with the finer horizontal resolution. The presence of landscape boundaries, for example the land/sea boundary or boundaries between different vegetation or soil types, leads to discontinuities in variables such as surface temperature and albedo, as well as soil temperature and moisture. Special interpolation methods need to be employed to initialize these variables near such boundaries.

Another consideration during initialization is the vertical interpolation. Due to differences in the vertical levels in the atmosphere, often the result of differences in the representation of orography between a LAM and a GCM, vertical interpolation of the GCM atmospheric fields (especially temperature) to the pressure levels of the LAM are required. These pressure levels themselves must be adjusted to the height of the LAM orography. This vertical compensation is also required near the boundary rows whilst nesting, if GCM and LAM orography differ.

5.4.3.4 Choice of domain size

Domain size refers to the total spatial extent of the LAM simulation. Experiments have been performed to study the effects of domain size on the LAM simulation (Jones et al. 1995, Podzun et al. 1995). They essentially found that the LAM domain should be sufficiently small that the synoptic circulation does not depart far from that of the driving GCM, but it should also be sufficiently large to allow development in the LAM of features which occur on a scale too small to be resolved by the GCM.

5.4.4 Limited area model simulations

A review of regional climate model simulations having a duration of at least a month and nested within a GCM or analyses is given in McGregor (1997). A general conclusion from the LAM investigations is that LAMs can improve the patterns for temperature and precipitation within the region but they have trouble overcoming any bias in the GCM driving data. Much of the improvement in the surface fields of prognostic variables is attributed to the enhanced representation of orography.

In 1999 the Journal of Geophysical Research published a special section dedicated to the regional climate modelling system developed at NCAR (RegCM). It contains papers that deal with RegCM testing and development, applications to climate studies and coupling with other components of the climate system. Giorgi and Shields (1999) use three year simulations over the continental United States to analyze the performance of the simplified explicit moisture scheme and three cumulus convection schemes. They found that the Grell scheme gave the best performance at monthly to seasonal timescales. Giorgi et al. (1999) focused their analysis on the radiative transfer processes, in particular on cloud processes. The model was run over an East Asia domain. They suggested a number of adjustments, which improve the model performance, but also emphasized that the representation of cloud radiative processes is an area in which the model is in need of improvement. Kato et al. (1999) analyzed the high-resolution performance of the model over East Asia. They found that the improved representation

of topographical features leads to an improvement in the simulation of precipitation patterns.

Several papers have demonstrated RegCMs utility when applied to regional climate studies. All have found that the model performs well in simulating interannual variability at the regional scale. This performance though depends on the season, geographical setting and the location of the region in relation to the air mass circulations, which determine precipitation. Regions covered include eastern Africa (Sun et al. 1999a, b), central Asia (Small et al. 1999a) and continental United States (Giorgi and Shields 1999).

There are also several papers addressing the coupling of regional climate models with other components of the climate system. These include coupling to sea ice and mixed layer ocean models for the arctic region (Lynch et al. 1999, Pinto et al. 1999), coupling to a full aerosol tracer model (Qian and Giorgi 1999), coupling to crop models (Mearns et al. 1999) and coupling to a lake model (Small et al. 1999b).

5.4.5 Issues in regional climate modelling

In this section three of the major issues involved in regional climate modelling. These issues have been part of an ongoing debate over the last several years and as such comprehensive discussions can be found in several papers including McGregor (1997) and Giorgi and Mearns (1999). Here only a brief account of each issue is presented.

Spin-up issue

Spin-up refers to the process by which initial (estimated) values for various variables will relax into values more representative of the region and time of interest. In a LAM it can be thought of as the time taken by the lateral boundary information to pervade the model domain and generate a dynamical equilibrium between the lateral boundary conditions and the internal model physics and dynamics. Spin-up time for the atmospheric component of a LAM is typically on the order of a few days. Spin-up time

for the land surface parameterization though may take anything from a few weeks up to several years for the most complicated schemes (Robock et al. 1998).

“Garbage in, garbage out” issue

Large-scale circulations simulated by regional climate models do not deviate far from those of the driving fields supplied as lateral boundary conditions. This necessarily implies that if a GCM incorrectly places a major storm track or other large-scale circulation then this misplacement will be reflected in the LAM. This is what is meant by the “garbage in, garbage out” problem. Fortunately, with continued improvements in GCMs this problem is diminishing.

Lateral boundary condition issue

Warner et al. (1997) reviewed the problem of lateral boundary conditions in regional numerical weather prediction models, this review is also relevant to LAMs. The methods used to treat lateral boundary conditions in current LAMs are mathematically ill-posed problems. The abrupt change of grid size by several times at the lateral boundaries can distort wave propagation and reflection properties. Several techniques have been developed to deal with this problem (see section 5.4.3.2) which have proven to be adequate for most regional model applications. The main problem that has been encountered with these techniques is the occurrence of spurious precipitation near the downwind domain boundaries. The problem is usually local in nature and the thermodynamic effects are absorbed by the lateral boundary condition terms, although the surface energy and water budgets near the boundary may be affected.

5.5 Climate Model Validation and Intercomparison

All models of real phenomena need to be validated against observations in order to assess their predictive skill. Validation of climate models is a difficult task to perform given that data are generally collected at a point while the climate models supply areally

averaged estimates. The models also provide estimates of many variables at each point which should be validated but frequently there is only a limited variety of data available for validation purposes. These problems have led to the existence of two types of validation experiment for climate models as well as the growth in the number of intercomparison experiments being performed.

The first type of validation experiment consists of using data collected as routine synoptic weather observations, often by a nation's meteorological service. This allows the use of a large number of observation stations (or satellite based observations) covering a significant spatial expanse. Unfortunately relatively few variables are collected routinely, these would include near-surface temperature and humidity, precipitation and possibly streamflow.

The second type of validation experiment entails the use of data collected as part of an intensive field experiment. While observations relating to many more variables than are routinely measured are made, they tend to cover a relatively small spatial extent and can be thought of as point values. Typically these field experiments collect data relating to the surface energy and water balance.

Obvious shortcomings of the two validation approaches available to model developers has led to the rise in intercomparison studies. These studies aim to improve our understanding of the implications of various physical parameterizations by analyzing the range of responses due to differences between models. They may also use observational data to validate the results using one of the methods above. Several intercomparison projects are currently actively performing studies using large numbers of participating models while intercomparison studies of a small number of models (two or three) generally involve the comparison of an old version of a model to a newly updated version.

Section 5.5.1 briefly reviews climate model validation studies of both types outlined above. Section 5.5.2 reviews intercomparison studies, including brief descriptions of each of the major intercomparison projects. The focus of this review is the land surface though other components of a climate model will be discussed briefly.

5.5.1 Validation

5.5.1.1 Routine synoptic validation

Routine synoptic validation simply refers to using routinely collected data to validate the model. This type of data is collected at thousands of meteorological stations throughout the world. Modelers wishing to validate a model may take these data, interpolate them in some fashion to create a spatially averaged field, and use them as observations against which to test the model. Another option available to modelers is to use spatially averaged data sets which have been produced by others. Production of these data sets requires the use of an interpolation model which may involve anything from relatively simple averaging techniques to observational analysis performed with a GCM. More recently, large scale satellite-derived observational data sets have also become available for validation.

Collecting data from many stations and calculating spatial averages can be a time consuming task though it does have the advantage of giving the modeler a better idea of the expected error range in the observational dataset. Below are some examples which are relevant to the development of RegCM2. Giorgi (1991) used data from 1436 meteorological stations to validate the performance of different physics parameterizations on summertime precipitation as simulated by the Pennsylvania State University – NCAR Mesoscale Model version 4 (MM4). Some 390 stations in western United States were used by Giorgi et al. (1993a) to provide precipitation and temperature observations for the validation of RegCM1. Giorgi et al. (1993b) used precipitation and surface temperature data from 308 European stations to validate the inclusion of a new planetary boundary-layer and radiative transfer parameterizations in RegCM2. Data compiled by the Japan Meteorological Agency were used by Kato et al. (1999) to investigate the effects of increasing resolution as well as simply validating the RegCM2 model over East Asia. Many other climate modelers have used data, largely precipitation and temperature, in a similar fashion to validate their models.

One of the few easily obtainable datasets which is a spatial integrator by nature is river runoff. Being indicative of an area, rather than a point measurement, means runoff data lends itself well to verifying the overall performance of the surface branch of the hydrological cycle. Some examples of these studies include Russell and Miller (1990) and Kuhl and Miller (1992) who used data from 33 rivers around the world to verify, unsuccessfully, the runoff estimated by the GCM developed at NASA/Goddard Institute for Space Studies. On a smaller scale, Dumenil and Todini (1992) used runoff data from the Arno river in northern Italy to validate the use of a rainfall-runoff scheme in the Hamburg climate model. Data from the Mississippi basin were used by Liston et al. (1994) to evaluate GCM land surface hydrology parameterizations. One of the difficult tasks when using runoff data for validation is the separation of errors in forcing (the precipitation field) from the errors in the surface model (Miller et al. 1994).

There currently exists a number of large scale data sets that can be used to validate results from GCM runs some of which are presented in Table 5.2. This form of validation has been used since the earliest attempts to validate climate models such as that by Holloway and Manabe (1971) who used the temperature dataset of Newell et al. (1969) in an attempt to validate an early version of the GCM developed at the Geophysical Fluid Dynamics Laboratory (GFDL). Some examples which demonstrate the continued use of these datasets for climate model validation include Garratt et al. (1993) who used the dataset of Henning (1989), Kiehl et al. (1998) who used several datasets including Warren et al. (1988) and Xie and Arkin (1996) and the study of Small et al. (1999b) which utilized the dataset of Legates and Willmott (1990). In this type of validation it may not be obvious whether errors in the variables can be linked to deficiencies in the model being validated or to the procedure used to create the dataset itself.

The use of satellite derived datasets for climate model validation is also increasing, for example, Kiehl et al. (1998) utilized the dataset of Hurrell and Campbell (1992) while Small et al. (1999b) used multi channel sea surface temperature data derived from the NOAA advanced very high resolution radiometer satellite platform. Another often used climate model validation dataset is that derived from an observational analysis run of a GCM, here a GCM is assumed to be the best tool for interpolating between point

observations. For example Giorgi and Marinucci (1996) used an observational analysis performed with the ECMWF GCM to validate the performance of the RegCM2 LAM over Europe.

5.5.1.2 Intensive field experiment validation

Since the mid-eighties there have been a number of field experiments which have taken measurements suitable for use in the validation or evaluation of climate model parameterizations. A summary of these experiments is presented in Table 5.2 and some examples of the data sets being used for validation are given here.

Data collected at an experiment in Manaus, Brazil have been used in validating climate model parameterizations by Abramopoulos et al. (1988) and Garratt et al. (1993). Jacquemin and Noilhan (1990) tested a land parameterization using data from HAPEX-MOBILHY, these data were also used by Bougeault et al. (1991) in the validation of the land surface parameterization implemented in the French Weather

Table 5.2 Large-scale data sets that can be used to validate surface-related GCM results

<i>Dataset</i>	<i>Variable</i>	<i>Area</i>
Korzoun et al. (1977)	Runoff	Global
Jaeger (1983)	Rainfall	Global
Warren et al. (1988)	Cloud cover	global
Henning (1989)	Energy balance components	Global
Legates and Willmott (1990)	Rainfall	Global
Vinnikov and Yesserkepova (1991)	Soil moisture	Russia
Wallis et al. (1991)	Rainfall, runoff, temperature	USA
Mintz and Serafini (1992)	Water balance components	Global
GPCP (Rudolf 1993)	Rainfall	Global
GRDC (Dumenil and Bengtsson 1993)	Runoff	50 major rivers
Hollinger and Isard (1994)	Soil moisture	Illinois, USA

Xie and Arkin (1996)	Rainfall	global
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Table 5.3 Recent field experiments that collected data suitable for use in the validation of climate model parameterizations

<i>Experiment</i>	<i>Reference</i>	<i>Location</i>	<i>Data</i>
ARME, 1983-85	Shuttleworth et al. (1984b, a)	Amazon basin, Brazil	surface energy and water balance, soil wetness
HAPEX- MOBILHY, 1986	Andre et al. (1986)	Southwest France	as above, info on mesoscale variability
Cabauw, 1987	Beljaars and Viterbo (1994)	Netherlands	surface energy and water balance
FIFE, 1987, 1989	Sellers et al. (1992)	Kansa, USA	as above
KUREX, 1987-88		basin of river Seym, Russia	as above
LOTREX, 1988		Germany	as above
Niger, 1988		Northern Sahel, Niger	as above
Manaus, 1988		Amazon basin, Brazil	as above
SEBEX, 1989-90	Wallace et al. (1991)	Sahel	as above, soil wetness
HEIFE, 1990		Tibetan plateau, Gobi desert	surface energy and water balance
EFEDA, 1991-95		Central Spain	as above
ABRACOS, 1991-95	Shuttleworth et al. (1991)	Amazon basin, Brazil	as above
BOREAS, 1991-95	Sellers et al. (1997)	Canada	as above, snow measurements
HAPEX- SAHEL, (1997)	Goutorbe et al. (1997)	Sahel	surface energy and water balance

1992	(1997)		balance
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Service (Meteo-France) LAM. Kuchment et al. (1993) tested a SVAT using data collected during KUREX, while Ducoudre et al. (1993) used ARME data to validate their own SVAT. Betts et al. (1993) used the 1987 FIFE data against which to validate the ECMWF operational model and then later, to validate the ECMWF reanalysis model (Betts et al. 1998). The FIFE data was also used by Bosilovich and Sun (1995) to validate the land surface parameterization in the Purdue Mesoscale Model. Oncley and Dudhia (1995) used data collected during STORM-FEST in the evaluation of surface fluxes from the MM5 model. Data collected during the BOREAS experiment was used by Bonan et al. (1997) to validate the land surface model (LSM) developed at NCAR for use within the CCM3 GCM. For a review of field experiments suitable for use in the validation of climate model parameterizations see Shuttleworth (1991). There are also several examples of field experiment data being used within major intercomparison projects as outlined in the next section.

5.5.2 Intercomparison

Intercomparison studies were originally performed by modeling groups when comparing a new version with an old version of their model. With the advent of the IPCC process broader cooperation between modeling groups has been encouraged and has led, in part, to the inception of an abundance of intercomparison projects. A review of some of the most relevant intercomparison projects is presented below. This is followed by a review of some intercomparison studies that have occurred outside the auspices of the major projects.

5.5.2.1 Project to Intercompare Regional Climate Simulations (PIRCS)

The aim of PIRCS is to provide a common framework for evaluating the strengths and weaknesses of regional climate models and their component procedures through systematic, comparative simulations. PIRCS experiment 1 involved eight LAMs run on a domain covering the continental United States for a period of two months covering May 15 to July 15, 1988. Some results from this experiment can be found in Takle et al. (1999). They found that the LAMs were able to reproduce bulk temporal and spatial characteristics of meteorological fields, in particular the 500 hPa height field was well simulated by participating models. They found that large-scale precipitation was simulated well in terms of time and location though amounts often varied from observations, while convective precipitation is represented only in a stochastic sense with less agreement in temporal and spatial patterns. Simulated surface energy budget was also compared to FIFE observations. While the simulated results show broad agreement with the FIFE observations, significant scatter among results meant that no strong conclusions could be drawn.

Future PIRCS experiments include work to examine model ability to predict the Great Plains low level jet, organized mesoscale convective systems, and the relationship between these two phenomena. Also planned are studies to examine the effect of uncertainties in initial and boundary conditions especially in relation to soil moisture. Exploration of model implementation details such as placement of lateral boundaries and methods for assimilating large-scale data into limited area models will also be addressed. Finally, models will be evaluated for their ability to assist climate change impact assessments in areas such as agriculture, human health and water resources.

5.5.2.2 Atmospheric Model Intercomparison Project (AMIP)

A series of intercomparison projects aimed at the improvement of components of climate models have been ongoing for the last decade or so. AMIP focuses on the atmospheric circulation component as described by Gates (1992). AMIP is an international effort to undertake the systematic validation, diagnosis and intercomparison of the performance of atmospheric general circulation models under

realistic conditions. The first phase of AMIP calls for the simulation of the climate of the decade 1979-1988 using the observed monthly-averaged distributions of sea-surface temperature and sea ice as boundary conditions along with a common prescribed atmospheric CO₂ concentration and solar constant. AMIP is one of the older intercomparison projects and has spawned more than two dozen diagnostic subprojects and many publications including Anderson (1996), Lau et al. (1996), Sperber and Palmer (1996), Ferranti et al. (1997) and Zwiers and Kharin (1998). Thirty one modeling groups have been involved in AMIP, a summary of AMIP results to date is presented below.

Gates et al. (1998) found that although there are model outliers in each simulated variable examined, validation of the AMIP models ensemble mean shows that the average large-scale seasonal distributions of pressure, temperature and circulation are reasonably close to what are believed to be the best observational estimates available. Large intermodel differences in the tropics occur in the large-scale structure of the ensemble mean precipitation and ocean surface heat flux. The total cloudiness is rather poorly simulated globally, particularly in the Southern Hemisphere. The models simulation of the seasonal cycle (as represented by the amplitude and phase of the first annual harmonic of sea-level pressure) closely resembles the observed variation in almost all regions. The ensemble's simulation of the interannual variability of sea-level pressure in the tropical Pacific is reasonably close to that observed, though the amplitude of major El Ninos is underestimated. The interannual variability in mid-latitudes, on the other hand, is not simulated as well. When analyzed in terms of the variability of the evolution of their combined space-time patterns in comparison to observations, the AMIP models are seen to exhibit a wide range of accuracy, with no single model performing best in all respects.

In one of the AMIP subprojects evaluation of the soil moisture simulations was undertaken (Robock et al. 1998). These simulations were compared to the soil moisture observational dataset of Vinnikov and Yeserkepova (1991) which contains data from 130 observation station inside Russia, and also the dataset of Hollinger and Isard (1994) which contains data from 17 stations within Illinois in the United States. They found that the model simulated results differ substantially from the observations and from each

other. Winter soil moisture variations in high latitudes were simulated poorly by all models, with soil moisture being kept almost constant while observations show that soil moisture varies almost as much in winter as in other seasons. In general, interannual variations were not captured by the models. They found that, generally speaking, a few months were sufficient for spinning up the soil moisture however the most complex parameterizations, such as those based on the Simple Biosphere model (SiB) could take several years to reach equilibrium with the model climatology.

In order to understand better the nature of these errors and to accelerate the rate of model improvement, an expanded and continuing project (AMIP II) is being undertaken in which analysis and intercomparison will address a wider range of variables and processes, using an improved diagnostic and experimental infrastructure.

5.5.2.3 Project for Intercomparison of Land Surface Parameterization Schemes (PILPS)

According to Henderson-Sellers et al. (1993b) PILPS is a project designed to improve the parameterization of the continental surface, especially hydrological, energy, momentum and carbon exchanges with the atmosphere. The PILPS science plan incorporates enhanced documentation, comparison, and validation of continental surface parameterization schemes by community participation. PILPS is an important exercise because existing intercomparisons, although piecemeal, demonstrate that there are significant differences in the formulation of individual processes in the available land surface schemes. These differences are comparable to other recognized differences among current global climate models such as cloud and convection parameterizations. It is also clear that too few sensitivity studies have been undertaken with the result that there is not yet enough information to indicate which simplifications or omissions are important for the near-surface continental climate, hydrology and biogeochemistry. PILPS emphasizes sensitivity studies with and intercomparisons of existing land surface parameterizations and the development of areally extensive datasets for their testing and validation.

A summary of the status of PILPS up until 1995 can be found in Henderson-Sellers et al. (1996b). Presented here is a review of PILPS related publications to date. The project

has been broken down into four phases. PILPS Phase 1 concentrated on intercomparison of simple off-line integrations using model-derived atmospheric forcing. In Phase 2, off-line integration results from different land-surface schemes were compared with observed fluxes. In Phase 3, which was conducted jointly with AMIP as subproject 12, the land-surface schemes were incorporated as components of their host GCM, allowing for the comparison of the global continental climate simulated by each GCM. PILPS is currently completing Phase 4, in which the performance of a single host atmospheric model, coupled to surface schemes, will be evaluated.

PILPS phase 1

In Phase 1 of PILPS a series of land surface models were forced “off-line” with data generated from a GCM and results from the final year of a multi-year equilibrium simulation were reported and analyzed. Early results demonstrated a large degree of disagreement among the models about the partitioning of energy between sensible and latent heat. Pitman et al. (1993) reported that the simulations for a tropical forest grid point ranged by 90 Wm^{-2} for the sensible heat flux and 80 Wm^{-2} for the latent heat flux. For a grassland grid point simulations ranged by 44 Wm^{-2} for sensible heat and 33 Wm^{-2} for the latent heat flux. These results lead to a supplementary series of experiments in phase 1, the results of which are reported in Pitman et al. (1999). In these experiments values for each land surface parameter were provided, results were quality controlled and analyzed, focusing on the scatter simulated amongst the models. These extra controls succeeded in marginally reducing the range reported for sensible heat over tropical forest to 79 Wm^{-2} , while the range of latent heat predictions remained unchanged. Over grassland the ranges fell to 34 Wm^{-2} and 27 Wm^{-2} for sensible and latent heat fluxes respectively. They also found large differences in the simulated runoff and soil moisture and at the monthly timescale. They suggest that this casts doubt on the reliability of land surface schemes and further claim that ‘It is a priority to resolve the disparity in the simulations, understand the reasons behind the scatter and to determine whether this lack of agreement in decoupled tests is reproduced in coupled experiments.’

The data from PILPS phase 1 were later used by Koster and Milly (1997) to investigate the interplay between transpiration and runoff formulations. They constructed a monthly

water balance model that reproduced the annual and seasonal water balances of the different PILPS schemes. Analysis of this model led them to identify two quantities that characterize a land surface parameterization's formulation of soil water balance dynamics: 1) the efficiency of the soils evaporation sink integrated over the active soil moisture range, and 2) the fraction of this range over which runoff is generated. Regardless of the surface parameterization's complexity, the combination of these two derived parameters with rates of interception loss, potential evaporation, and precipitation provides a reasonable estimate for the parameterization's simulated annual water balance. They emphasize the importance of the schemes runoff formulation which often has not received much attention in the past. This is shown through simple sensitivity experiments that 'demonstrate that even a "perfect" description of canopy structure and stomatal behavior, toward which many land surface models strive, does not ensure realistic evaporation rates if the runoff formulation remains relatively crude or incompatible with the evaporation formulation.'

PILPS phase 2

In phase 2 of PILPS the intercomparison is "offline" as in phase 1, but point-based observations are used as the atmospheric forcing and observed flux data are available for evaluation of performance of the schemes. Three different datasets were utilized during phase 2, the Cabauw dataset (Beljaars and Viterbo 1994) was used during phase 2(a), while phase 2(b) used data from HAPEX-MOBILHY (Andre et al. 1986) and phase 2(c) used data from the Red-Arkansas river basin which was compiled by the National Climatic Data Center (NCDC) (Abdulla 1995). Phase 2(d) was recently completed using data from Valdai, Russia, the results of which are yet to be published. Results from phase 2(a) suggest that a good parameterization of runoff is particularly important due to the linkage between energy budget and water budget, and the feedback between these components. They also found that schemes with similar structure or underlying philosophies often exhibited very different behaviors.

Phase 2(b) of PILPS was focused on quantifying the differences in soil moisture predictions among participating land surface parameterizations. The results from this phase can be found in Shao et al. (1994) and Shao and Henderson-Sellers (1996). One of the main conclusions was that no scheme did well both during the main growing

period of the crop and over the full annual cycle. During the intensive observation period the total evapotranspiration predicted by the models varied from 81 – 167 mm with the measured value being 124 mm (Mahfouf et al. 1996). Most of the models overestimate bare soil evaporation and large differences remain between model predictions with the range in cumulative evaporation during the first four months being 135 mm. This is due to very large differences in the parameter relating moisture stress to the reduction of evaporation from the potential rate (Desborough et al. 1996). Drainage displays a considerable range among models, from 40 to 300 mm, with a mean of around 200mm while observations suggest 290 mm of water leaves the soil via subsurface processes.

The major goal of PILPS phase 2(c) is to evaluate the ability of current land-surface schemes to reproduce measures energy and water fluxes over multiple seasonal cycles across a climatically diverse, continental-scale basin. A supplementary objective was to test the ability of the schemes to calibrate their parameters using data from small catchments and to transfer this information from the calibration basins to other small catchments and to the computational grid boxes. The results from this phase are discussed in Wood et al. (1998), Liang et al. (1998) and Lohmann et al. (1998). Results from this phase were consistent with earlier PILPS experiments in terms of differences among models in predicting energy and water fluxes. The mean annual Bowen ratio varied from 0.52 to 1.73 as compared to the data estimated value of 0.92. The run-off ratios varied from a low of 0.02 to a high of 0.41, which compares to an observed value of 0.15. They also found that in general, those schemes that did not calibrate performed worse, not only on the validation catchments, but also at the scale of the entire modeling domain.

Results from PILPS phase 2 suggest that individual land-surface schemes capture specific aspects of the complex system with reasonable accuracy but no one scheme captures the whole system satisfactorily and consistently.

PILPS phase 3

PILPS phase 3 is the joint PILPS-AMIP project on land-surface processes: AMIP diagnostic subproject 12. Analysis of the surface energy budget for these coupled

simulations can be found in Qu and Henderson-Sellers (1998). They compared observations and offline results from Cabauw (phase 2(a)) and Valdai (phase 2(d)) to the coupled simulation results. They found that the normalized measures of the scatter among land-surface simulations show that the range from the coupled simulations is larger than that from the offline simulations, on both annual and seasonal time scales. It was hoped that feedback from the host atmospheric model would reduce the disagreement among PILPS land-surface schemes but this was not the case.

PILPS phase 4

Phase 4(a) evaluates the operation of land-surface parameterization schemes fully coupled to a particular GCM, while phase 4(b) has the land-surface schemes coupled to a LAM. The GCM chosen for this study is CCM3 and the LAM chosen is the Limited Area Prediction System (LAPS) developed by the Bureau of Meteorology and Research Centre (BMRC), Melbourne, Australia. This phase is currently ongoing.

Some results from phase 4(b) are presented in Timbal and Henderson-Sellers (1998). In this study they coupled four land-surface schemes to the LAPS LAM. They found that the BUCKET scheme had several shortcomings compared to more detailed schemes, however very noticeable scatter still remains among the more complex land-surface parameterizations. They also found significant differences in calculated precipitation due to the surface feedbacks of the different land-surface schemes. Of particular concern were the differences among participating schemes in the extreme cases, such as initialization of soil moisture to field capacity or wilting point. Further investigations are planned using higher resolutions and a more elaborate initialization technique.

5.5.2.4 Other intercomparison studies

As well as the above, more formal, projects there have been several other studies which have performed intercomparisons of various climate model parameterizations. Rowntree and Lean (1994) compared the standard hydrological parameterization from the UK Meteorological Office model to the ARNO scheme. They found that the ARNO scheme was an improvement over the standard model but that it was unable to reproduce

summer runoff. Polcher et al. (1996) compared three land surface schemes used in GCMs. They found large differences in simulated evaporation, runoff and soil moisture. They claim to have shown that ‘the surface resistance and field capacity were essential parameters in determining the annual cycle of evaporation and that a representation of sub-grid scale variability of soil moisture had an important impact on runoff.’

Chen et al. (1996) compared the simulation of land surface evaporation for four land surface parameterizations over FIFE. The simplest of these models was the simple bucket model with two parameters (Manabe et al. 1965), and the most complex model is the simplified Simple Biosphere (SSiB) model of Xue et al. (1991) with 22 parameters. They conclude that some complexity in the canopy resistance scheme is important in reducing both the overestimation of evaporation during wet periods and underestimation during dry periods with the two most complex models performing the best. They also demonstrated that simply increasing complexity of the model does not necessarily improve performance with the most complex model (22 parameters) performing similarly to the second most complex model (15 parameters).

Desborough (1999) compared various versions of the CHAmeleon Surface Model (CHASM). He demonstrates that a combination of surface energy balance complexity and aerodynamic parameterization can be used to explain the gross simulation differences obtained in PILPS. His results also suggest that explicit canopy resistance, canopy interception and bare ground evaporation are of limited value with respect to overall land-surface model precision in simulating the annual, monthly and seasonally averaged diurnal cycles of evaporation but that they may be important for simulating evaporation at the daily time scale.

Leung et al. (1999) intercompared three LAMs which were used to simulate an extreme flood event over eastern Asia. They found that each model simulated the gross flood conditions reasonably well, though significant differences were found in the simulated energy and hydrological cycles, especially over cloudy areas. The reasons for this include the simulation of the amount and vertical distribution of clouds, the treatment of cloud radiative feedbacks, and the representation of land surface processes. They also

note that 'One specially important criterion is the radiation balance which has serious implications for long term climate simulations.'

5.5.2.5 Conclusions

A common theme of all the intercomparison studies is that no one model performs better overall than the other models. Each model performs well in the simulation of some variables, while performing poorly in the simulation of others. In general the models are much better at reproducing large scale atmospheric phenomena than phenomena at the land surface. The simulation of precipitation displays reasonable agreement between models, especially in terms of stable precipitation, while convective precipitation exhibits a more stochastic nature and hence larger disagreement between models.

At the land surface significant scatter is consistently found in the energy and water fluxes. The largest scatter is possibly associated with the simulated runoff, where differences in runoff ratio of an order of magnitude have been found between models. This scatter in energy and water fluxes is exacerbated in the presence of clouds. The feedback between the land surface parameterization and the atmospheric circulation component has been found to increase the scatter between models, emphasizing the need to get these feedbacks correct.

Finally the studies have shown that having a more complex model, with more parameters, does not lead to better performance. Indeed, it often makes it difficult to determine the reasons for model simulation inaccuracies. Finally it is noted that calibration of the land surface scheme produced significant improvement in model performance. This suggests that the land cover and soil parameters, established by limited experiments but applied universally, are a source of significant errors. The process of calibrating the model can be thought of as simply improving the estimates of these parameters for the particular conditions present in the area of interest.

Chapter 6

Regional Climate Model Performance Analysis and Comparison: the Energy Budget

6.1 Introduction

This chapter investigates the performance of the LAMs in simulating the surface energy budget. These terms are of particular importance when looking at climate change scenarios since near surface air temperature is the most widely quoted prognostic variable of climate change, and the one in which the most confidence is placed. Emphasis must also be placed on the connections between the energy and hydrological budgets which occur intimately through the evaporative process as well as in the soil moisture influence over surface temperatures.

The surface energy budget is given in equation 5.41 and can be rewritten for a non-snow affected areas simply as

$$h_s = R_{net} - F_s - L_v F_q \quad (6.1)$$

where R_{net} is the net incident radiation at the surface.

Evidence has shown that over the course of a year the net surface heating (or ground heat flux) h_s , is negligibly small compared to the other terms in equation 6.1. When taking a long term view in a stable climate the net surface heating should be essentially zero. On the other hand, in a changing climate we would expect a long term drift of the net surface heating. Hence the net surface heating is an important indicator of long term climate change. Unfortunately it is difficult to measure and can be a source of considerable variation between climate models themselves.

Confidence in any predictions of climate change then requires confidence in the model's representation of the energy budget. The representation of the energy budget in three LAMs are intercompared and tested against observations in the remainder of this chapter.

6.2 Experiment Design

6.2.1 LAM setup

Domain and Period

The simulation domain for the experiment is centered over the FIFE region of the Konza Prairie, Kansas and covers an area of $1200 \times 1260 \text{ km}^2$ (longitude $\times$ latitude) in the central United States. The experiment was run for a total of two years covering 1987 and 1988. The LAMs (MM5/BATS, MM5/SHEELS and RegCM2) all used the same resolution with 17 levels in the vertical, horizontal grid point spacing of 20km and a time step of 1 minute.

The period chosen is long enough to investigate the LAMs' representation of the seasonal cycle, and ends in 1988 which was considered an extreme drought year in the central United States (Trenberth and Guillemot 1996) and provides a good test of the LAMs' ability to reproduce an extreme period. The domain allows comparison with the intensive observations collected during FIFE as well as data collected within the Konza Long Term Ecological Reserve (LTER) site situated within FIFE.

Initial and Boundary Conditions

Atmospheric initial and boundary conditions were extracted from the analysis produced by the European Centre for Medium-Range Weather Forecasts (ECMWF). The analysis is treated as output from a “perfect” model of the atmosphere for the periods simulated and therefore assumes that differences between LAM output and observations represent simulation errors due to construction and assimilation of lateral boundary conditions and internal shortcomings of the LAMs. This is not strictly true and it must be remembered that some of the errors encountered will be due to errors in the analysis.

The aim of the current study is not to optimise the LAM parameterisations for a particular location, but rather to compare the performance of the schemes. As such, the LAM parameters have the values that are used as the default for this location. That is, the parameter values have been predefined using global scale datasets.

6.2.2 Observations

The FIFE experimental domain consists of 15×15 km tallgrass prairie on an undulating topography. The experiment itself is described in detail in Sellers et al. (1992). The FIFE dataset consists of nearly continuous observations of surface parameters from the Portable Automated Mesonet stations (PAM) and four intensive field campaigns (IFCs) with surface flux observations, all conducted in 1987 (from 26 May to 7 June, from 25 June to 11 July, from 6 to 21 August and from 5 to 16 October). The PAM data (10 stations) consists of wind at 5.4m, temperature and humidity at 2m, together with radiometric measure of the ground surface temperature, downward shortwave radiation, net radiation, downward longwave radiation and rainfall. During the IFCs, fluxes of sensible, latent and ground heat fluxes were measured at 13 stations with Bowen ratio and eddy correlation methods. The FIFE data have been quality controlled, edited and averaged by Betts and Ball (1998), resulting in a single time series that represents the average over an area of 15×15 km square.

The LAM results used in this study were given by the gridpoint centered within the FIFE site. Each gridpoint is representative of an area 20×20 km square, which is somewhat larger than the FIFE site itself. There are several reasons why the comparison is meaningful. Conditions over the FIFE grassland site were relatively homogenous, so that simple averaging of the data gave a representative mean. The Konza prairie itself covers over 50,000 km², and the diurnal cycle over land integrates over considerable advection distances (up to 100-200 km) (Betts et al. 1998).

6.3 Results

6.3.1 Net Incident Radiation

6.3.1.1 Clear Sky Radiation

To begin the analysis of the performance of the LAMs surface energy budget an investigation into the simulated net incident radiation was conducted. This investigation is initially limited to rain-free summer days, that is, days for which all three LAMs and the observations show no precipitation. This simplifies the analysis by removing the influence of clouds on the radiation. This criteria leaves six days in 1987: 9th and 10th of May, and the 4th, 5th, 6th and 7th of June. Due to fewer observations only three days were available in 1988: 13th and 16th of May, as well as the 13th of June. The diurnal variation in the net longwave, net shortwave and overall net incident radiation are examined below.

Net longwave radiation

The net longwave radiation for all nine rain-free summer days is shown in Figure 6.1. Note that the local solar noon occurs at 1820 UTC (Coordinated Universal Time – formerly GMT). Significant differences between the observed and simulated net longwave radiation at the surface can be seen in Figure 6.1. These differences can be explained by differences in the downward longwave radiation at the surface, differences in the surface albedo (which is related to the upper layer soil moisture, see equation 5.60) and differences in the longwave radiation emitted by the surface (which is directly related to surface temperature via Stefan-Boltzmann's law).

Explaining the difference between MM5/BATS and MM5/SHEELS is slightly simplified by the same treatment of longwave radiation in the atmosphere providing the same incident longwave radiation at the surface. Hence these differences can be explained totally in terms of the albedo and the longwave radiation emitted at the surface. In 1987 MM5/SHEELS generally simulates lower surface temperatures than MM5/BATS, this leads directly to lower longwave emissions from the surface and

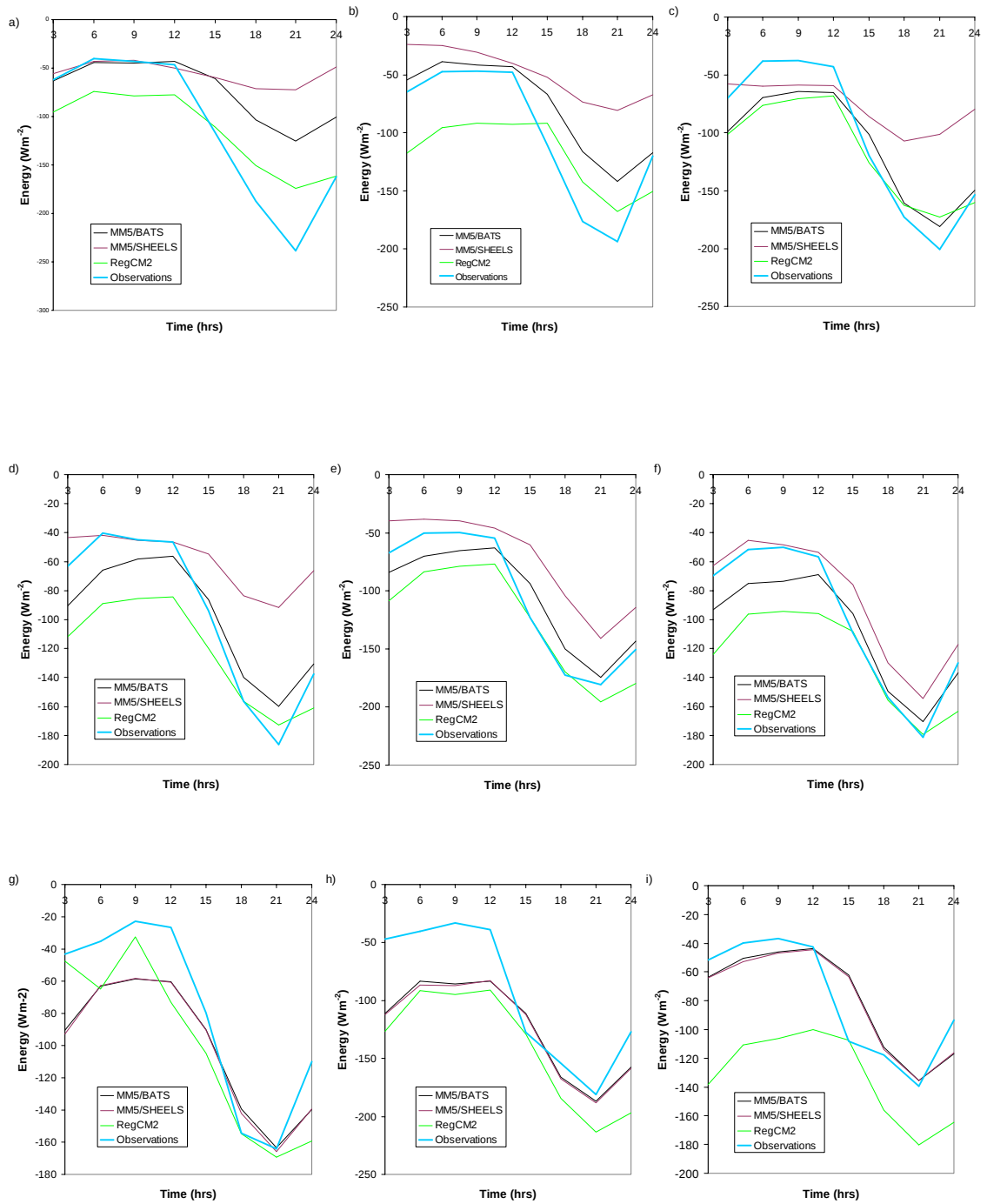


Figure 6.1 Three hour average net longwave radiation at the surface on rain-free summer days. **a)** 9/5/87; **b)** 10/5/87; **c)** 4/6/87; **d)** 5/6/87; **e)** 6/6/87; **f)** 7/6/87; **g)** 13/5/88; **h)** 16/5/88 and **i)** 13/6/88.

hence a higher net longwave radiation. During 1987 MM5/SHEELS demonstrates the least amount of diurnal variation in the net longwave radiation. This difference is related to the diurnal temperature variation where MM5/BATS produces generally higher temperatures and a larger range such that the largest differences between MM5/SHEELS and MM5/BATS tend to occur with the highest temperatures. This leads, due Stefan-Boltzmann's law, to the significantly larger longwave emissions simulated by MM5/BATS during the day, when compared to MM5/SHEELS. This can be seen most clearly in Figure 6.1 a, b, c and d.

The albedo of the vegetated fraction (close to 0.8) remains the same in each model while the albedo of the bare ground fraction is dependant upon the top layer soil moisture (equation 5.60). This dependence is such that the drier the top layer of soil is, the higher the albedo. This effect is relatively minor compared to the surface temperature effect since it only applies to around 0.2 of the grid cell surface. Here, MM5/SHEELS has a generally higher top layer soil moisture in 1987, which would imply a generally higher albedo and hence lower net longwave radiation. This effect is reasonably accounted for by the surface temperature difference throughout 1987, except perhaps for the night of the 9th of May 1987 where the effects simply cancel each other out.

In 1988 a very different relationship is evident. In this drier year the MM5 based LAMs demonstrate much stronger similarities with both the surface temperature and the upper layer soil moisture being very similar. This is reflected in the extremely similar net longwave radiation simulated by the two LAMs and can be seen in Figure 6.1 g, h and i.

Examining the difference between RegCM2 and MM5/BATS is a more complicated since the different longwave radiation treatments in MM5 and RegCM2 lead to different downward longwave radiation at the surface. Again surface temperature has a major effect on the longwave radiation emitted from the surface while the albedo has a secondary effect. RegCM2 simulates the lowest temperatures of the LAMs and, if all else was equal, this would lead to a higher net longwave radiation at the surface. However this is not what can be seen in Figure 6.1. Here RegCM2 consistently simulates the lowest net longwave radiation. This apparent contradiction can be

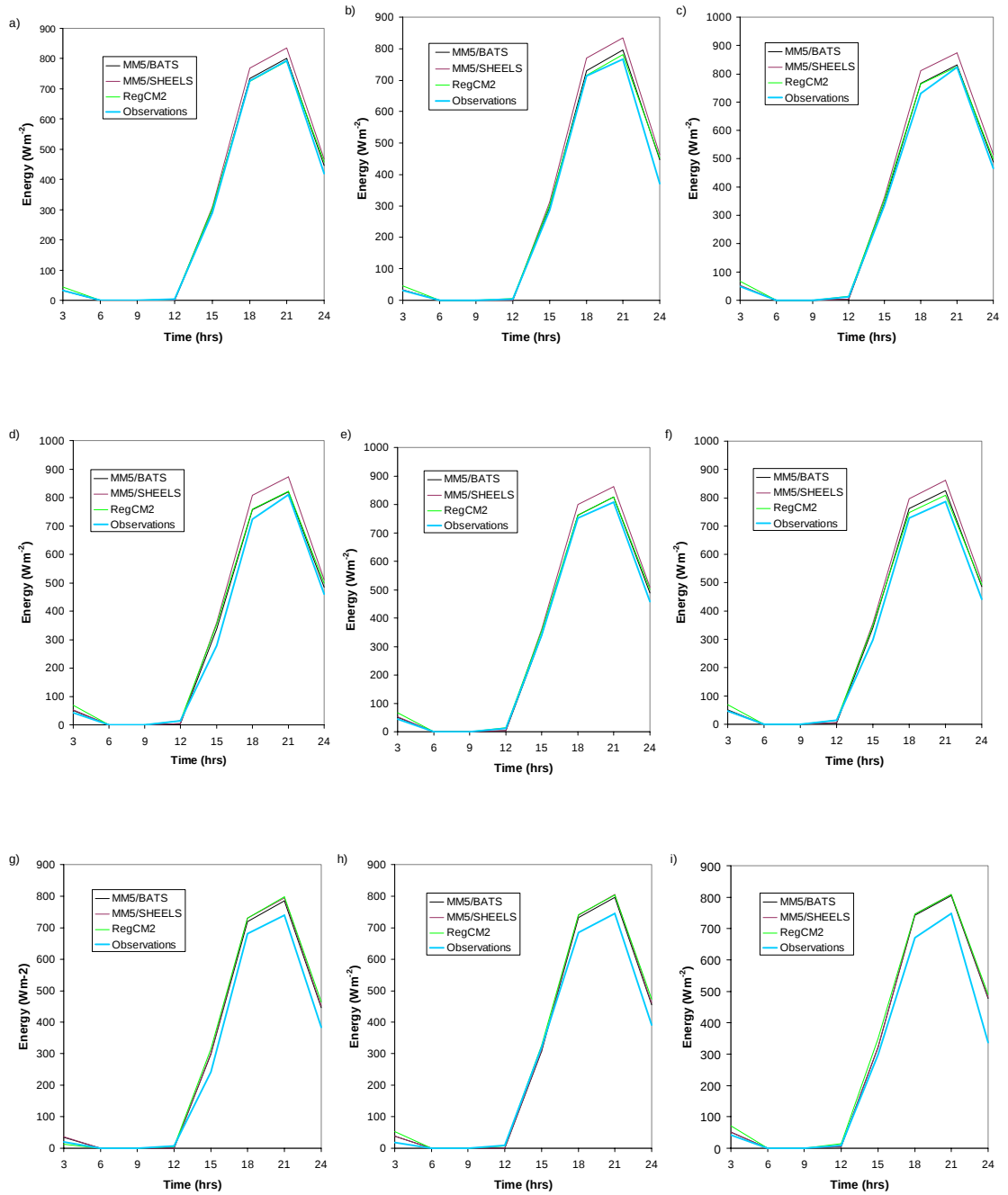


Figure 6.2 Three hour average net shortwave radiation at the surface on rain-free summer days. **a)** 9/5/87; **b)** 10/5/87; **c)** 4/6/87; **d)** 5/6/87; **e)** 6/6/87; **f)** 7/6/87; **g)** 13/5/88; **h)** 16/5/88 and **i)** 13/6/88.

explained by RegCM2 producing the smallest downward longwave radiation at the surface.

Under clear sky conditions then, more longwave radiation is able to reach the surface under the relatively simple scheme present in the MM5 LAMs than under the scheme present in RegCM2. The MM5 scheme depends critically on the pre-calculated broadband emissivity function, while the RegCM2 scheme (equations 5.90-5.97) explicitly calculates the effects of the dominant gases, CO₂ and O₃. The observed downward radiation tends to lie between the RegCM2 and MM5 LAM results, with RegCM2 performing marginally better at reproducing the diurnal cycle of the downward longwave radiation flux.

In terms of albedo we note that MM5/BATS has a higher soil moisture, and therefore a lower albedo, for most of the rain-free days. RegCM2, however, has higher soil moisture and therefore lower albedo on the fourth of June 1987 and on the 13th of May 1988. The effects of this can be seen in Figure 6.1 where in both c) and g) the RegCM2 simulated net longwave radiation is, at least for part of the time, higher than the net longwave radiation simulated by the MM5 LAMs.

When comparing against observations in 1987 no LAM captures the diurnal variation of the net longwave radiation very well. Overnight (up to 12 UTC) MM5/SHEELS performs best while during the day RegCM2 seems to perform best. During 1988, the drier year, the MM5 based LAMs produce a marginally better performance than RegCM2, due primarily to the difference in downward longwave radiation. Overall, no LAM clearly outperforms any of the others and the longwave radiation budget at the surface remains a source of error. Fortunately the longwave radiation budget is a minor part of the total incident radiation budget and as such is not as important as the shortwave radiation budget.

Net shortwave radiation

The net shortwave radiation for these rain-free summer days can be found in Figure 6.2. Clearly there is much better agreement between the LAMs and the observations for the shortwave radiation than was seen for the longwave radiation. The downward shortwave

radiation is almost identical for all three LAMs, despite RegCM2 using the more complex and complete shortwave radiation scheme. While the LAMs may be in close agreement with each other, they all overestimate compared to the observed downward shortwave radiation.

Any differences between the LAMs can be explained in terms of upward shortwave radiation which is determined by the downward shortwave radiation and the surface albedo. The surface albedo is a combination of the albedo of the vegetation, which is the same for all three LAMs, and the albedo of the bare ground fraction, which is dependant on upper layer soil moisture. Upper layer soil moisture is then a controlling factor on the upward shortwave radiation. In 1987, MM5/SHEELS has the highest upper layer soil moisture, followed by MM5/BATS, with RegCM2 slightly drier again. As a consequence MM5/SHEELS has the smallest albedo, followed by MM5/BATS and then RegCM2. As can be seen in Figure 6.2 a to f, this leads to MM5/SHEELS having the highest net shortwave radiation, with MM5/BATS and RegCM2 being both lower than MM5/SHEELS and closer to the observations. In 1988, the dry year, the upper layer soil moisture simulated by the LAMs are much closer together. This results in the LAMs all having similar albedos and hence similar net shortwave radiation budgets. All three LAMs overestimate the net shortwave radiation under dry conditions (Figure 6.2 g to i).

Net incident radiation

The net incident radiation is obtained by combining the net longwave radiation with the net shortwave radiation. This is shown for the rain-free summer days in Figure 6.3. The shape of the net incident radiation graphs shown in Figure 6.3 are dominated by the shape of the shortwave radiation since the magnitude of the longwave radiation is less than a quarter that of the shortwave radiation. Up until around 1200 UTC the net incident radiation is dominated by the longwave radiation budget, while after this the shortwave radiation budget dominates.

During 1987 the MM5 based LAMs overestimate the daytime longwave radiation budget which exacerbates the overestimation of the shortwave radiation. This causes the MM5 based LAMs, particularly MM5/SHEELS, to significantly overestimate the net incident daytime radiation (Figure 6.3 a to f). RegCM2 tends to have lower estimates for

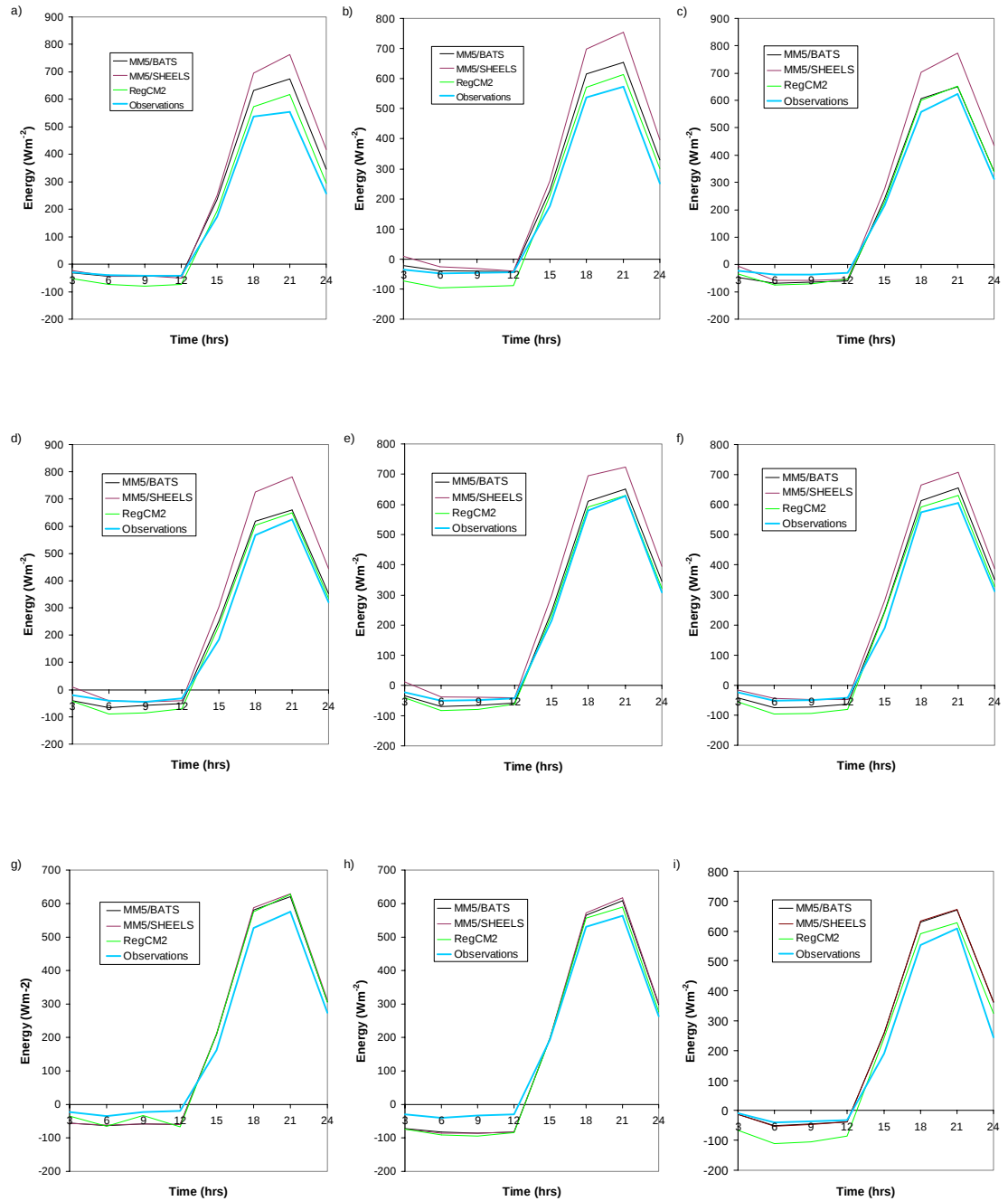


Figure 6.3 Three hour average net incident radiation at the surface on rain-free summer days. **a)** 9/5/87; **b)** 10/5/87; **c)** 4/6/87; **d)** 5/6/87; **e)** 6/6/87; **f)** 7/6/87; **g)** 13/5/88; **h)** 16/5/88 and **i)** 13/6/88.

the daytime longwave radiation and hence its net incident radiation remains closer to the observed than the MM5 based LAMs. During 1988 the MM5 based LAMs perform much better at reproducing the observed net longwave radiation than they did in 1987. As a result the range of the daytime net incident radiation (Figure 6.3 g to i) remains quite close to that seen in the daytime net shortwave radiation (Figure 6.2 g to i).

6.3.1.2 Total Net Incident Radiation

The daily net incident radiation simulated by the LAMs is compared to observations in Figure 6.4. RegCM2 performs best compared to observations available. All three LAMs simulate similar net incident radiation during winter but can differ by as much as 100 Wm^{-2} during summer. It is during this summer period that the MM5 based models significantly overestimate the energy available due to radiation, while RegCM2 compares favourably with the observations for the same period.

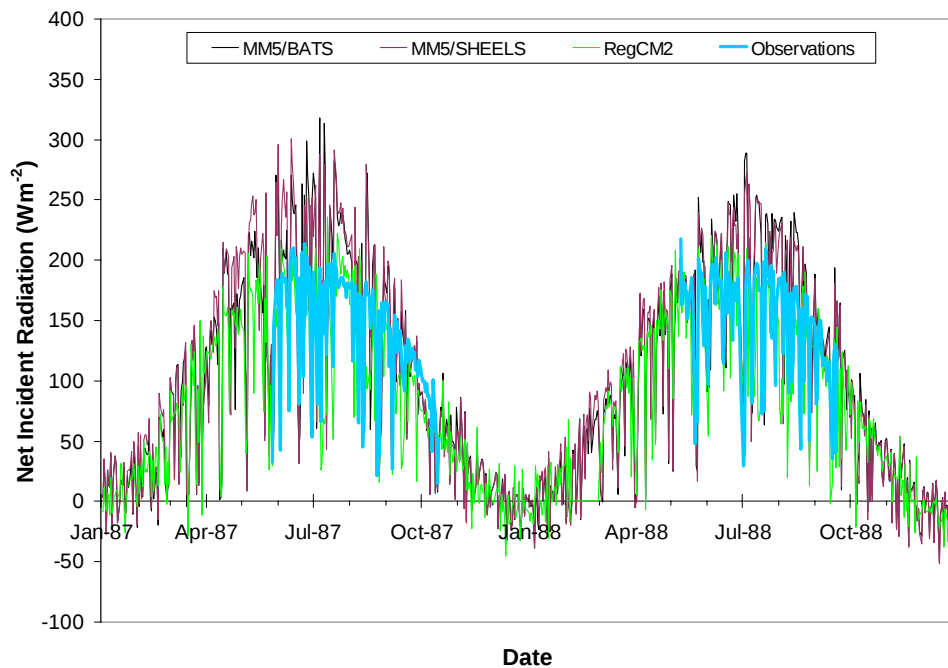


Figure 6.4 Daily net incident radiation for all three LAMs

Means and standard deviations of the daily net incident radiation during the common observation period in each year are presented in Table 6.1. RegCM2 appears to best reproduce the observations in both years, as discussed in relation to Figure 6.4, compared to the other LAMs. It can be seen in Table 6.1 that both MM5 based LAMs over estimate the mean daily incident radiation by 23 – 30 Wm^{-2} and RegCM2 underestimates the mean by 12 – 19 Wm^{-2} . That is, RegCM2 does come closest to reproducing the observed mean daily incident radiation but this reproduction is not as good as it appears in Figure 6.4. All the LAMs overestimate the standard deviation of the daily net incident radiation though again RegCM2 performs the best, while the MM5 based LAMs significantly over estimate the variation inherent in the time series. Note also that there is little difference between 1987 and 1988 in the observations and LAM simulations despite 1988 being a drought year. The small decrease in the variance seen in the dry year is to be expected due to the reduced occurrence of clouds.

The cumulative net incident radiation during observation periods in both 1987 and 1988 are presented in Figure 6.5. The statistics found in Table 6.1 are borne out here, with the MM5 based LAMs overestimating and RegCM2 underestimating. It can be seen in Figure 6.5 a) that MM5/SHEELS simulates more net radiation than MM5/BATS in 1987. This trend is reversed in the drier year, 1988 (Figure 6.5 b).

Table 6.1 Means and standard deviations of daily daily net incident radiation over the common observation period; 27 May to 19 September.

	1987 (Wm^{-2})		1988 (Wm^{-2})	
	Mean	St. Dev.	Mean	St. Dev.
Observations	150.8	45.4	148.3	43.8
MM5/BATS	173.2	73.9	178.0	57.9
MM5/SHEELS	180.5	68.6	174.5	57.5
RegCM2	131.6	54.4	136.8	48.3

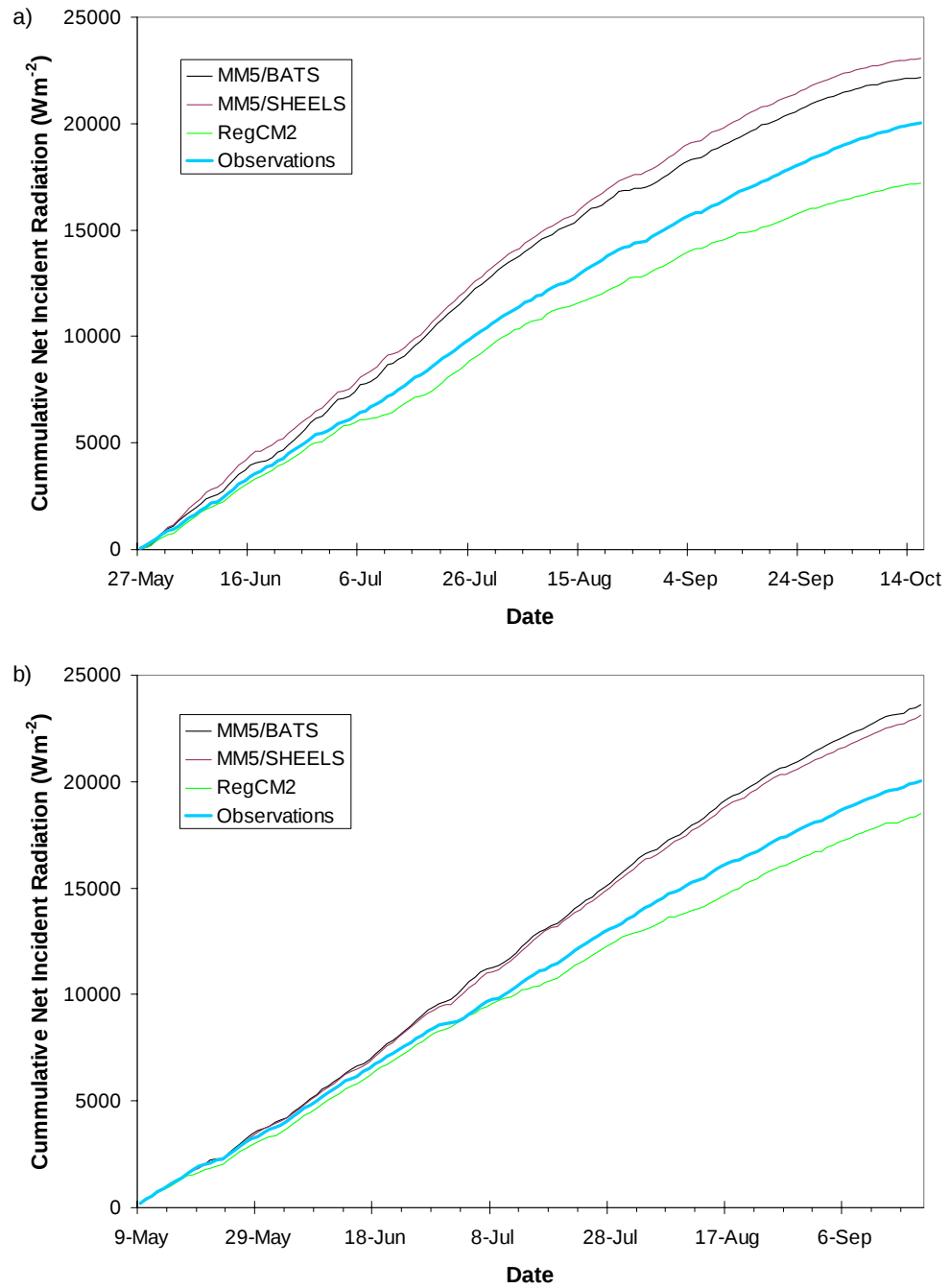


Figure 6.5 Cumulative net incident radiation for all three LAMs during the observation period in **a)** 1987 and **b)** 1988

6.3.2 Latent Heat

The latent heat (evapotranspiration) provides a vital link between the energy budget and the water budget at the land surface. Here the focus is on the latent heat (expressed as energy) as it pertains to the energy budget. Figure 6.6 presents the daily latent heat simulated by the three LAMs, as well as observations taken in both years. Along with section 7.3.2, Figure 6.6 shows the LAMs performing reasonably well at reproducing the observed latent heat flux in 1987 but all three LAMs underestimate the latent heat in 1988.

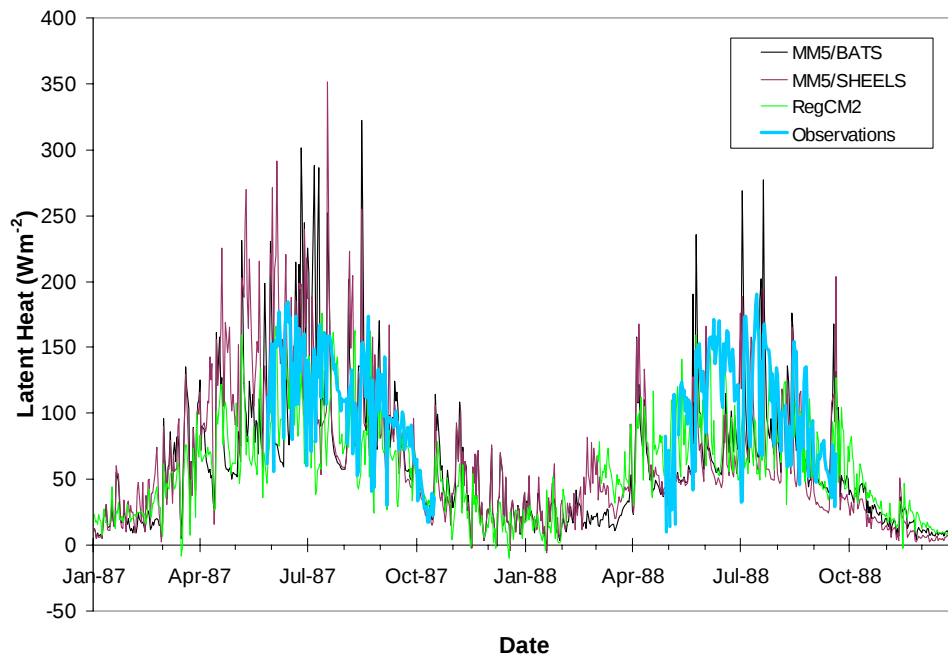


Figure 6.6 Daily latent heat simulated by all three LAMs.

In order to further investigate the relationship between net incident radiation and latent heat flux, the daily evaporative fraction (EF) is presented in Figure 6.7. Here evaporative fraction is defined as

$$EF = \frac{L}{R_{net}} \quad (6.1)$$

where L is the latent heat flux and R_{net} is the net incident radiation.

The three LAMs simulate values for EF which are close to the observed during the 1987 observation period (Figure 6.7 a). Of the three LAMs RegCM2 performs the best with the MM5 based LAMs occasionally, but significantly, overestimating the EF . These overestimations are caused by the simulation of an unusually low amount of net incident radiation while the latent heat flux does not change very much. This means, for instance, that on the 8th of October 1987 the MM5 based LAMs simulate the production of latent heat flux to be between four or five times larger than the energy made available by the net incident radiation. This extra energy is supplied through the ground heat flux, the wind and most of all, the conversion of sensible heat flux.

In 1988 (Figure 6.7 b) it can be seen that the LAMs consistently underestimate the EF , while still producing the occasional significant overestimation. Again these overestimations occur when the net incident radiation is unusually low and they gain much of their energy from the conversion of sensible heat. During 1988, no unusually high values for EF are observed though during 1987 the observed EF almost reaches two on several occasions. These occasions are also associated with low net incident radiation and the conversion of sensible heat, as for the high EF values simulated by the LAMs. The high EF values simulated by the LAMs are significantly larger than the highest observed values. This suggests that the LAMs convert energy from the sensible to the latent heat flux much more readily than is observed in the field.

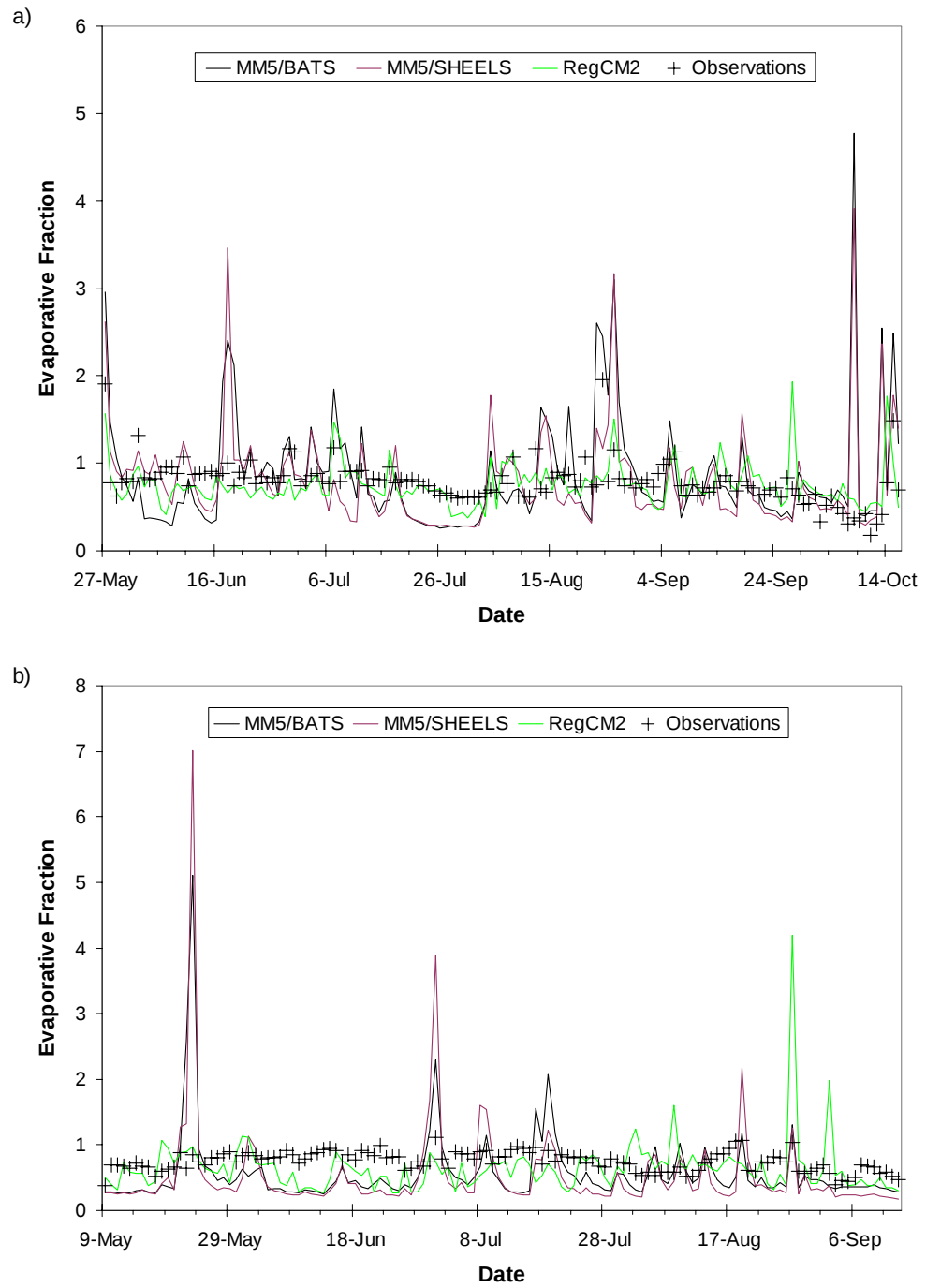


Figure 6.7 Evaporative fraction simulated by all three LAMs during the observation period in **a)** 1987 and **b)** 1988.

6.3.3 Sensible Heat

Sensible heat is defined as the transfer of energy from the surface via thermal conduction. The daily sensible heat flux simulated by the three LAMs can be found in Figure 6.8. The mean and standard deviation statistics for the common observation period in each year can be found in Table 6.2. It is clear from both the figure and the table that the LAMs generally overestimate the sensible heat, with RegCM2 coming closest to the observations. In particular RegCM2 is much better at reproducing the observed variance of the time series, which the MM5 based LAMs significantly overestimate. Some of the difference between the LAMs can be explained in terms of the differences in the simulated wind speed (see Figure 6.13) and its relationship to sensible heat given in equations 5.65 and 5.71.

The change from 1987 to 1988 is also of interest. The observed mean daily sensible heat increases by approximately 11 Wm^{-2} and the standard deviation increases by almost 4 Wm^{-2} when comparing the wet year (1987) to the dry year (1988). All three LAMs simulate an increase in the mean daily sensible heat in the drier year. However these increases are much larger than that observed with MM5/BATS, MM5/SHEELS and RegCM2 increasing by 37, 47 and 18 Wm^{-2} respectively. All three LAMs also simulate a decrease in the standard deviation of the daily sensible heat time series in the drier year. This trend is the opposite of what is observed.

Table 6.2 Means and standard deviations of daily sensible heat over the common observation period; 27 May to 19 September.

	1987 (Wm^{-2})		1988 (Wm^{-2})	
	Mean	St. Dev.	Mean	St. Dev.
Observations	20.95	18.29	32.23	22.00
MM5/BATS	52.25	52.54	89.01	27.56
MM5/SHEELS	55.15	49.03	102.46	29.01

RegCM2	39.92	24.95	57.68	21.12
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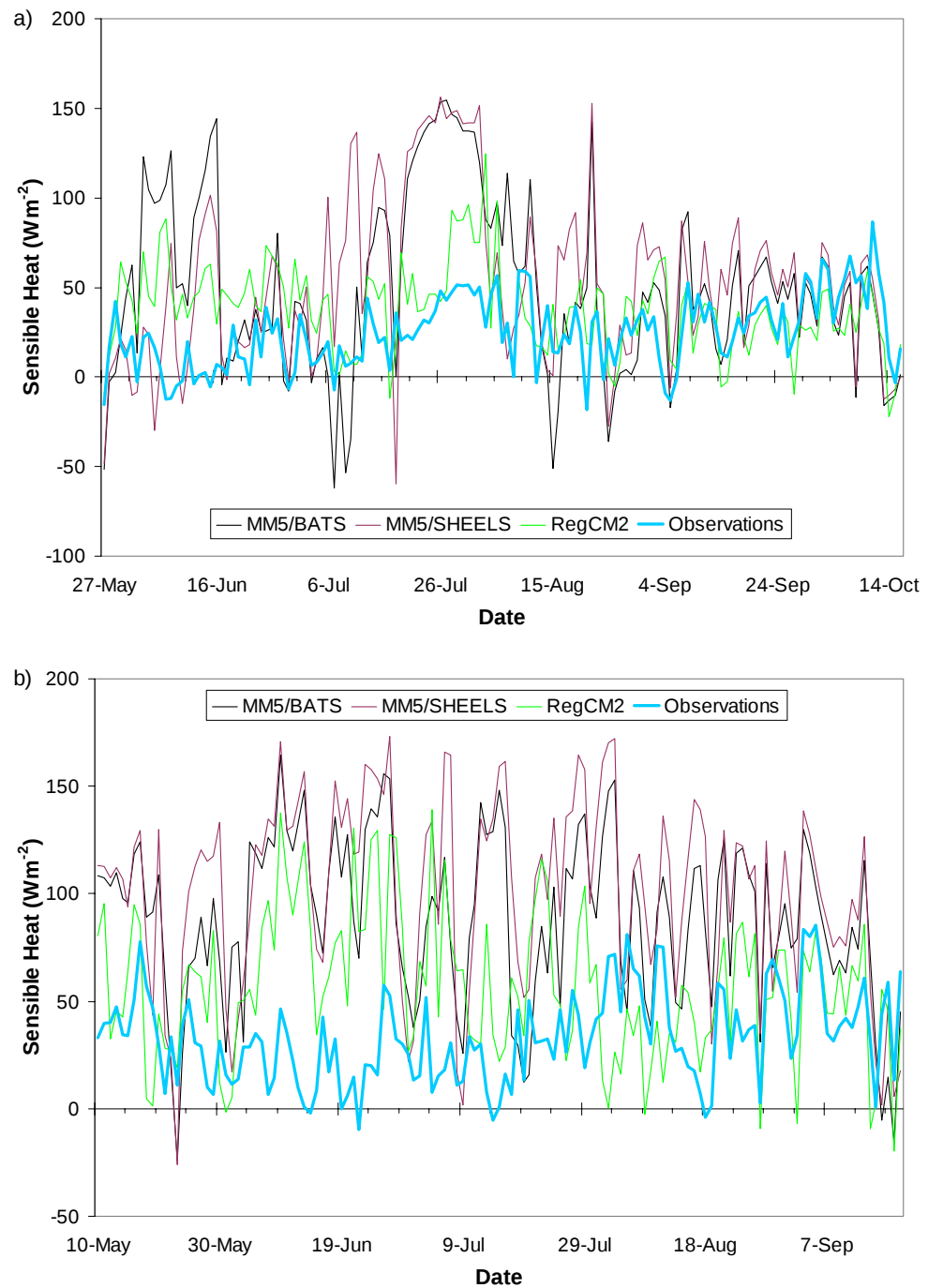


Figure 6.8 Daily sensible heat flux simulated by all three LAMs during the observation period in **a)** 1987 and **b)** 1988

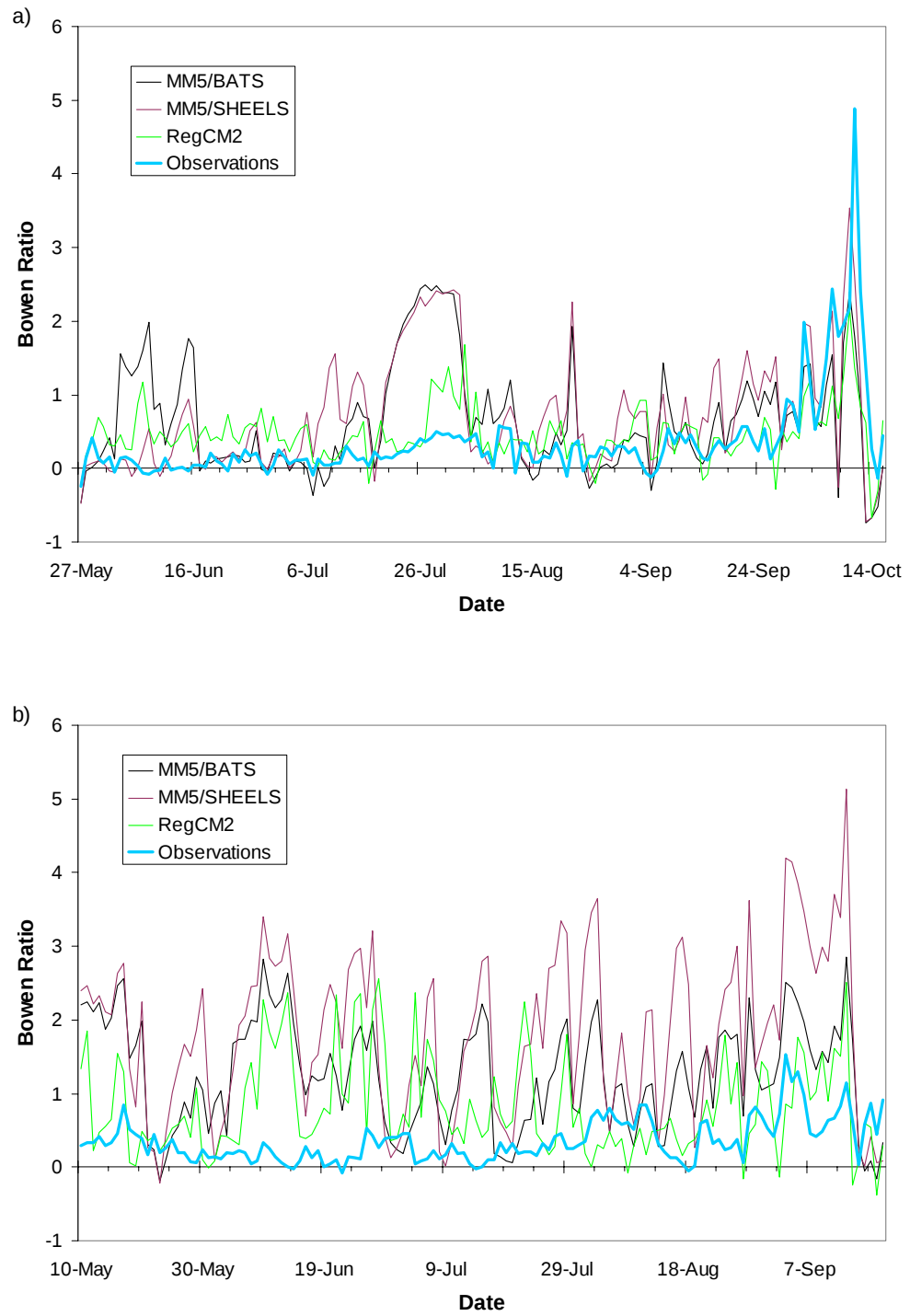


Figure 6.9 Daily Bowen ratio for the three LAMs during the observation period in **a) 1987** and **b) 1988**.

The Bowen ratio provides a measure of the relative distribution of the net radiation into sensible and latent heat and is given by equation 3.5. The daily Bowen ratio simulated by the three LAMs during the observation periods in 1987 and 1988 are presented in Figure 6.9. It is clear that generally the LAMs significantly and consistently overestimate the Bowen ratio. This emphasises previous results which showed the latent heat to be underestimated and the sensible heat overestimated.

Again the MM5 based LAMs produce the largest variance in the daily Bowen ratio time series. All three LAMs perform best in 1987, the wetter year, while performing quite badly in 1988. The observed Bowen ratio increases in 1988 compared to 1987, especially early in the observation period, but remains at around 0.5 for the most part. The three LAMs also have higher Bowen ratios in 1988 than 1987. However early in the 1988 period the LAMs often have a Bowen ratio close to 2 compared to the observed 0.5. Hence, we observe twice as much latent heat flux as sensible heat flux but the LAMs simulate twice as much sensible heat flux as latent heat flux.

7.3.4 Net Surface Heating

The net surface heating is defined by equation 5.41. A comparison of the daily net surface heating simulated by the LAMs with observations taken in each year is presented in Figure 6.10. Again the MM5 based LAMs simulate too much variance in the daily time series, while RegCM2 is much closer to the observed (Table 6.3). This is exemplified by the extreme values produced by the MM5 based LAMs, which are many times larger than the extreme observations.

The presence of soil moisture increases the heat conductance of the soil and therefore, in years which are unusually dry, heat conductance through the soil would be retarded compared to a normal year. This would be reflected in a decrease in the variance of the surface heating as is shown in Figure 6.10, and recorded as the standard deviation in Table 6.3, where the observed variance of the surface heating decreases

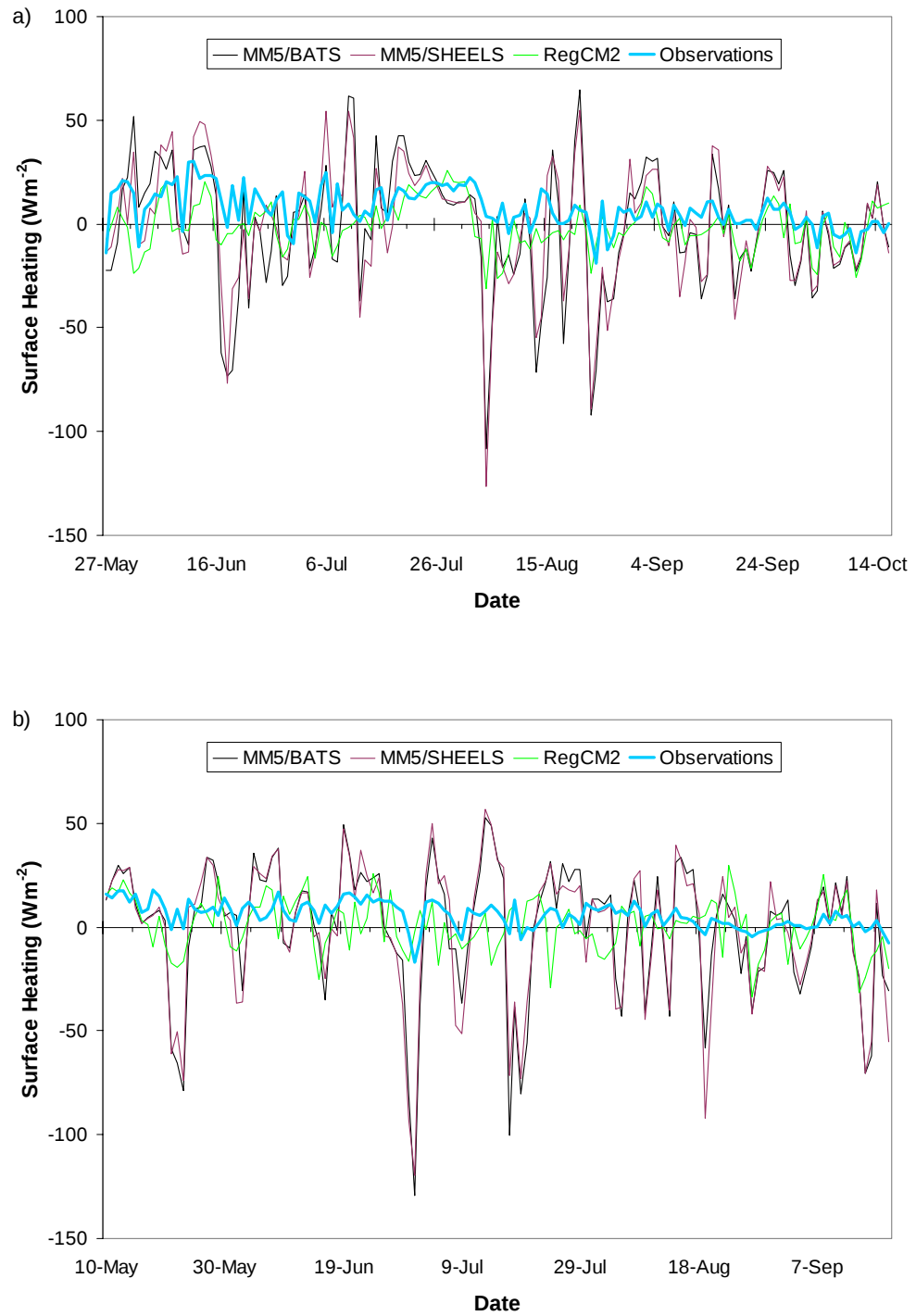


Figure 6.10 Daily net surface heating simulated by all three LAMs during observation periods in **a)** 1987 and **b)** 1988.

from 9.52 Wm^{-2} in 1987 to 5.96 Wm^{-2} in 1988. This change in the variance of the surface heating is not reproduced by any of the LAMs, which demonstrate very little change in the variance between the wet and dry year.

Figure 6.11 presents the cumulative surface heating for the LAMs and observations during the observation periods in a)1987 and b)1988. Clearly the LAMs do not come close to reproducing the cumulative observed surface heating through summer in both 1987 and 1988. The observations show a continual increase in the cumulative surface heating through these summer periods, whereas in 1987 the LAMs all simulate the surface heating to be negative by the end of August. In 1988 the LAMs simulate an increasing cumulative surface heating until around mid-June, after which the cumulative surface heating remains steady until the end of the observation period. The observations on the other hand, show the cumulative surface heating continuing to increase until the end of the period.

Table 6.3 Standard deviations of daily net surface heating over the common observation period; 27 May to 19 September.

	Observations	MM5/BATS	MM5/SHEELS	RegCM2
1987	9.52	32.01	30.47	11.16
1988	5.96	31.58	32.17	12.41

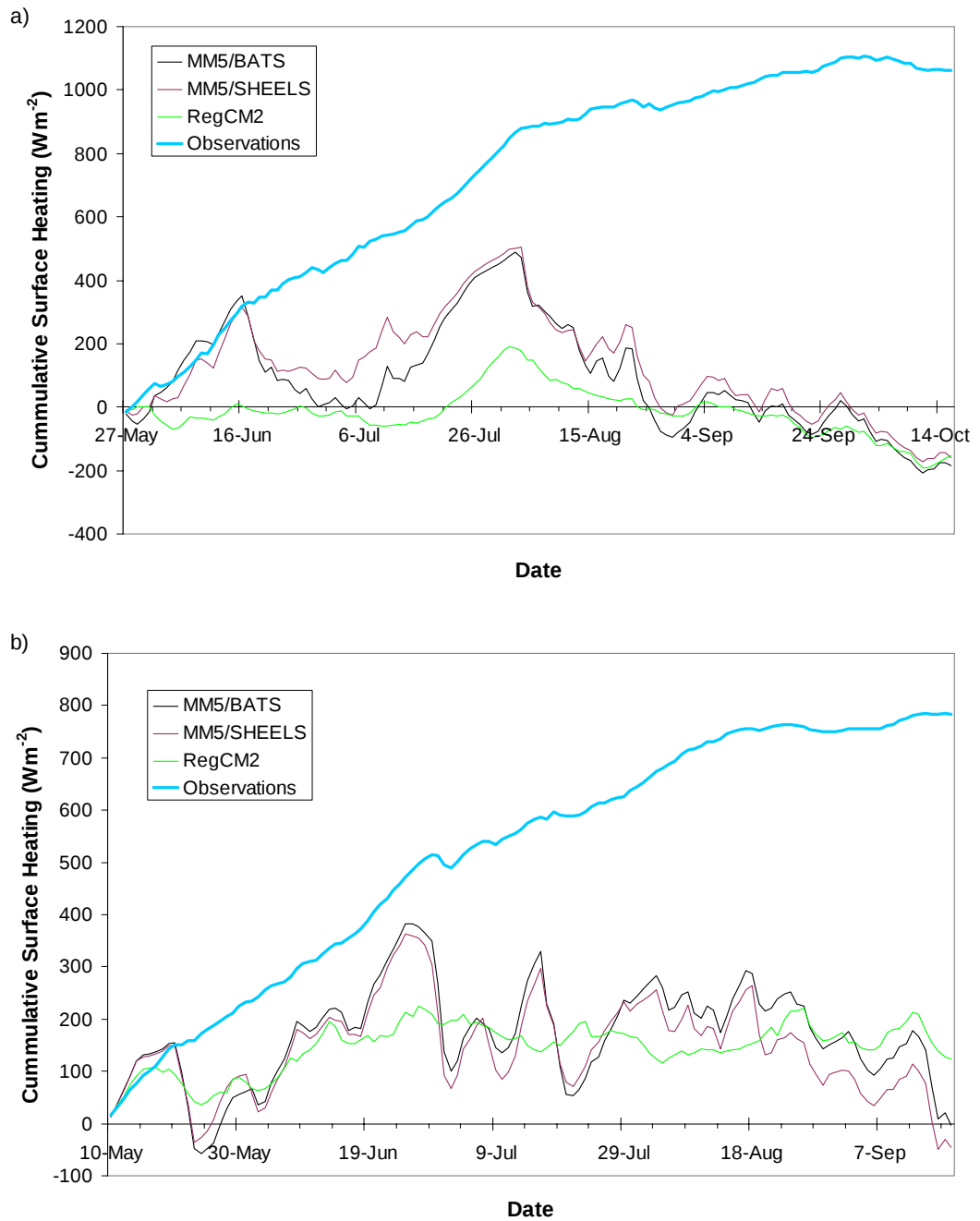


Figure 6.11 Cumulative surface heating during the observation period in
a) 1987 and b) 1988.

7.3.5 Near Surface Temperature and Wind

One of the most widely quoted prognostic variables of climate change is the near surface air temperature, climatological means of which are considered to be a robust indicator of climate change. This says little about the performance of the LAMs in terms of reproducing the observed near surface air temperature on a daily basis. Figure 6.12 presents the daily mean near surface air temperature for the three LAMs and the observations. The LAMs perform well at reproducing these observations, with disagreement occurring during summer in both years. The MM5 based LAMs tend to overestimate the temperature during summer while RegCM2 tends to underestimate it. This difference may be partly explained by the difference in wind speed between the models and observations.

The mean monthly near surface wind speed for all three LAMs and observations is presented in Figure 6.13. Clearly the MM5 based LAMs consistently overestimate the wind speed, while RegCM2 consistently underestimates it. This difference in simulated wind speed is transferred to the temperature through equations 5.73 to 5.76.

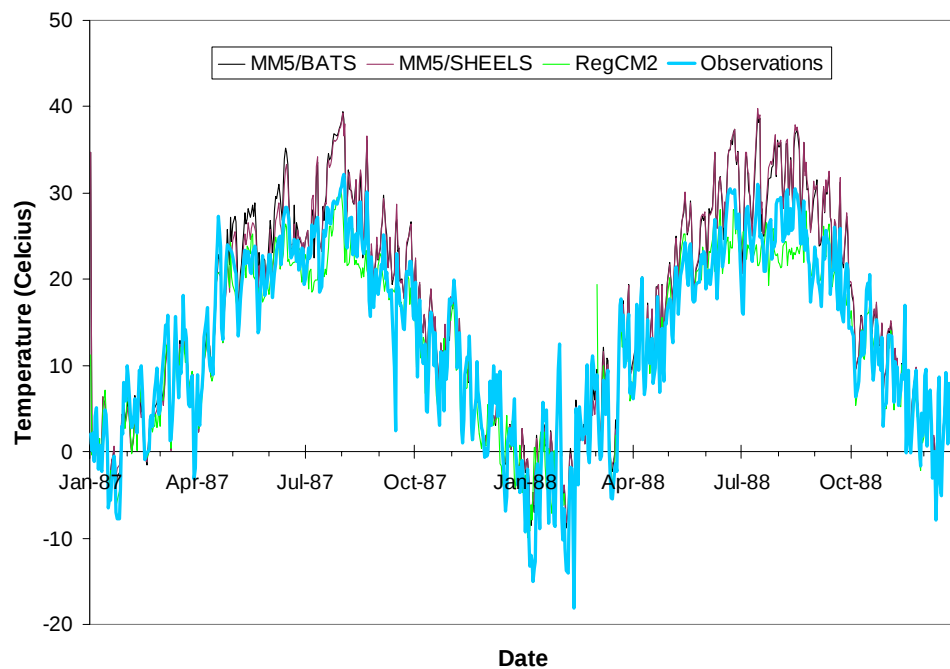


Figure 6.12 Daily near surface air temperature.

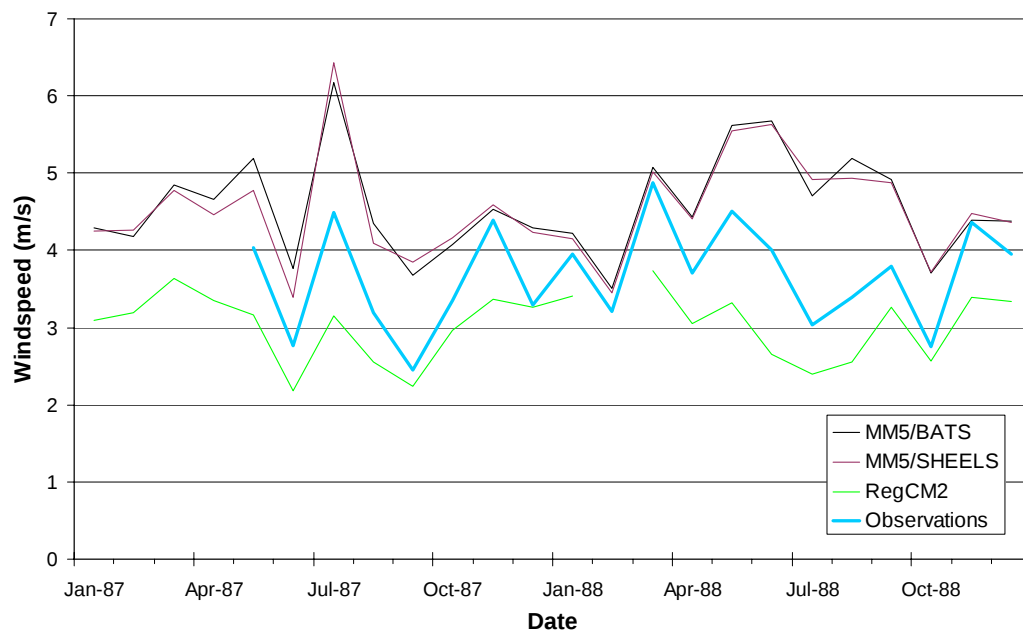


Figure 6.13 Mean monthly near surface wind speed.

6.3.6 Surface Energy Balance

The cumulative energy balance over FIFE for each LAM during the observation periods in 1987 and 1988 is presented in Figure 6.14 and 6.15 respectively. The observed energy balance for 1987 is shown in Figure 6.14 a). It can be seen that the majority of the incident energy is used in the production of latent heat, around 77%. Sensible heat consumes the next largest portion of the incident energy with almost 18% being used. The surface heating consumes the least amount of energy though it is still significant at just over 5%.

In 1987 the cumulative energy balance of all three LAMs is quite similar, as can be seen in Figure 6.14 b) to d). All the LAMs simulate just over two thirds of the incident energy being consumed as latent heat The remainder being consumed as sensible heat,

while the surface heating has negligible effect. Hence, in the 1987 observation period the LAMs underestimate both the latent and surface heating while simultaneously

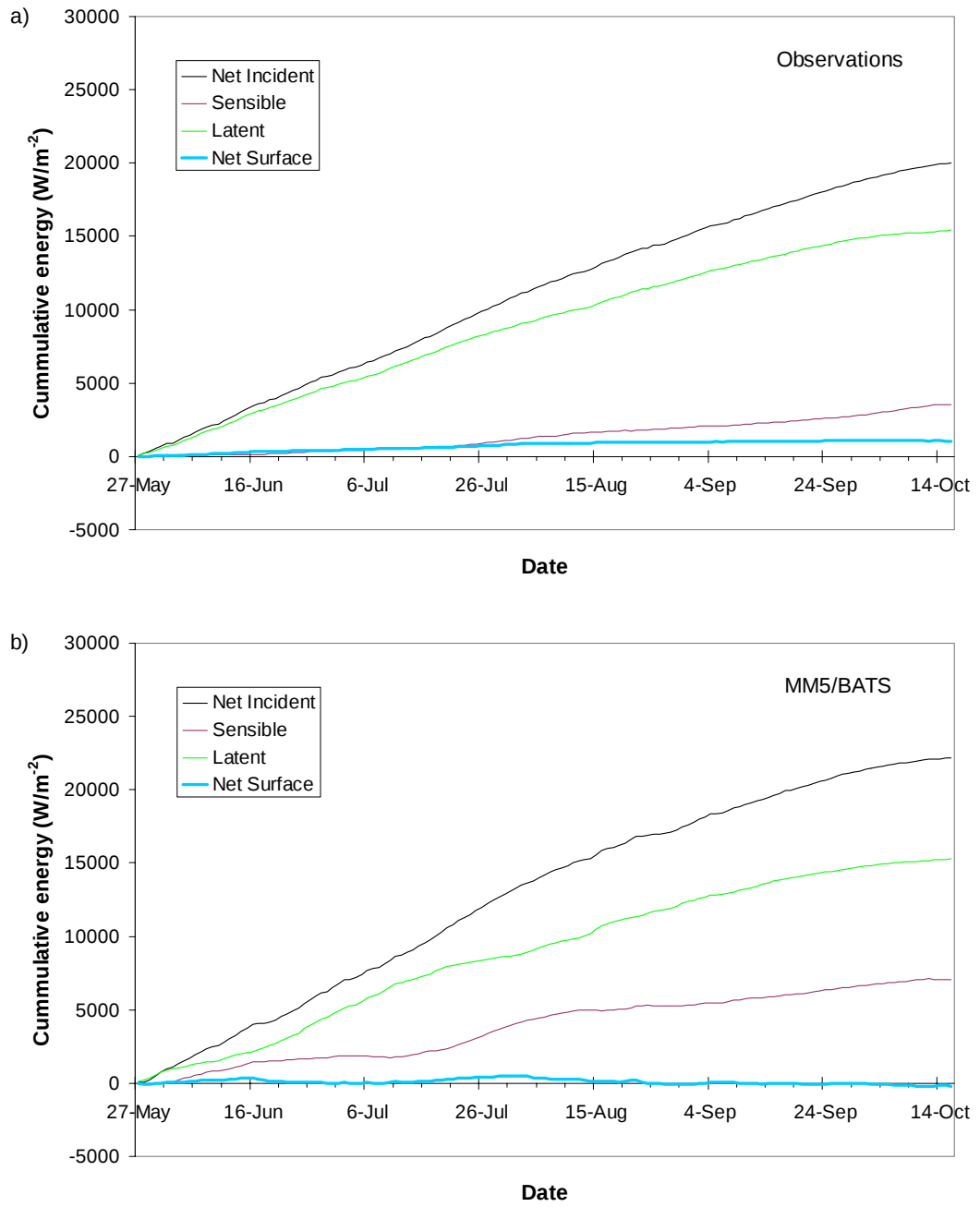


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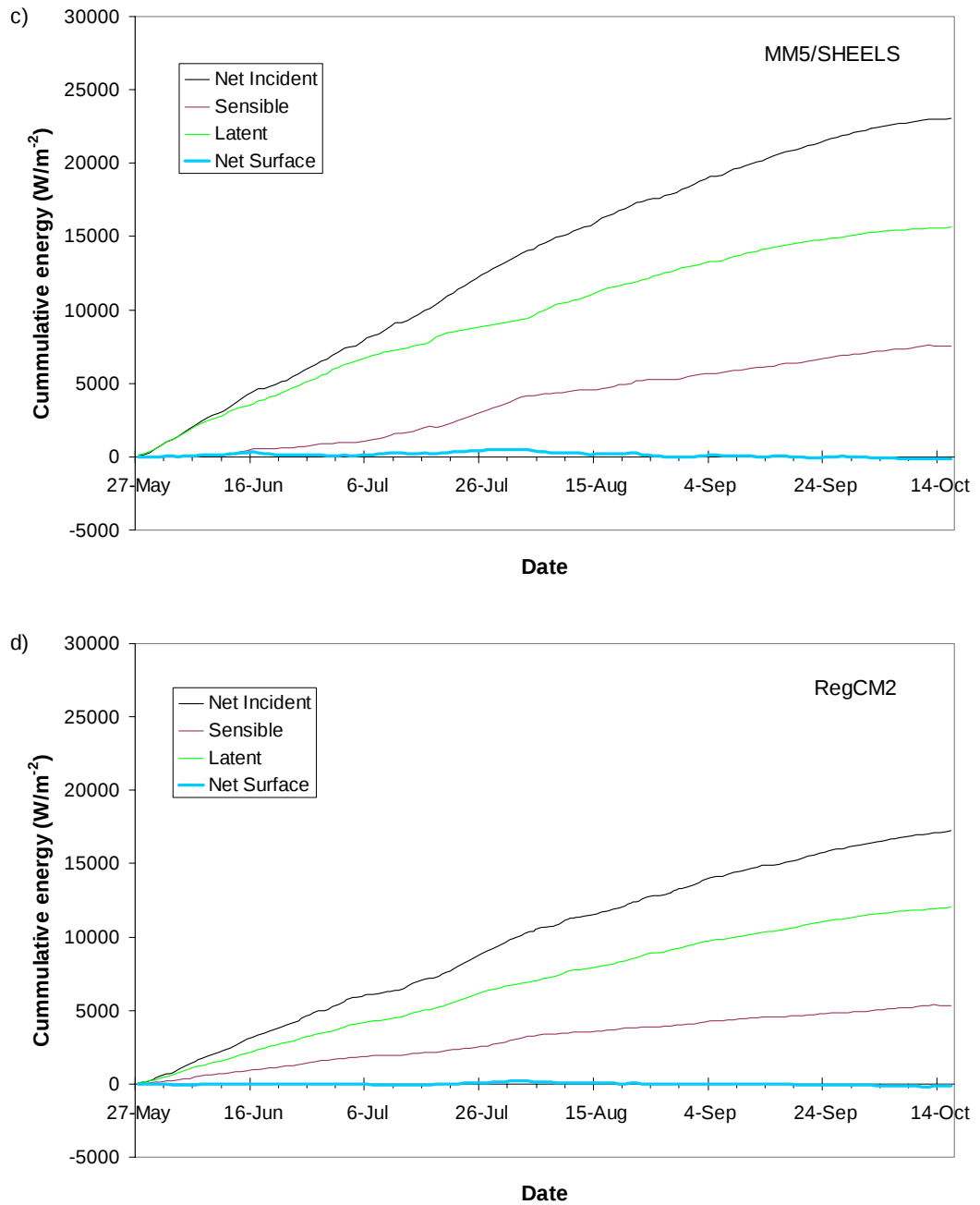


Figure 6.14 Cumulative energy balance during the 1987 observation period:
a) Observations; **b)** MM5/BATS; **c)** MM5/SHEELS and **d)** RegCM2.

overestimating the sensible heating by almost double the observed proportion. It is also noteworthy that the MM5 based LAMs overestimate the incident energy while RegCM2 underestimates it. This means that the MM5 based LAMs actually perform reasonably well at reproducing the absolute quantity of latent heat, while RegCM2 performs better at reproducing the sensible heat.

The observed cumulative energy balance during the 1988 observation period is quite similar proportionally to that seen in 1987. During this drier year the observations show that the latent heat proportion falls by around 4%, while the sensible heat proportion rises by about the same amount. The surface heating still accounts for around 5% of the incident energy. Under the drier conditions in 1988 the LAM simulations of the energy balance diverge. MM5/BATS splits the incident energy evenly between the latent and sensible heating. MM5/SHEELS simulates a larger proportion being consumed by sensible heat than by latent heat (opposite to what is observed), and RegCM2 comes closest to the observations with the latent heat accounting for the majority of the incident energy. However, the proportion of the incident energy that RegCM2 consumes as latent heat is still some 15% lower than that observed.

In general then, the energy balance simulated by the LAMs is quite poor. Under normal to wet conditions (1987) the LAMs at least show considerable consistency among themselves. However, under dry conditions (1988) not only do the LAMs perform poorly at reproducing the observed energy balance, they also demonstrate considerable disagreement among themselves.

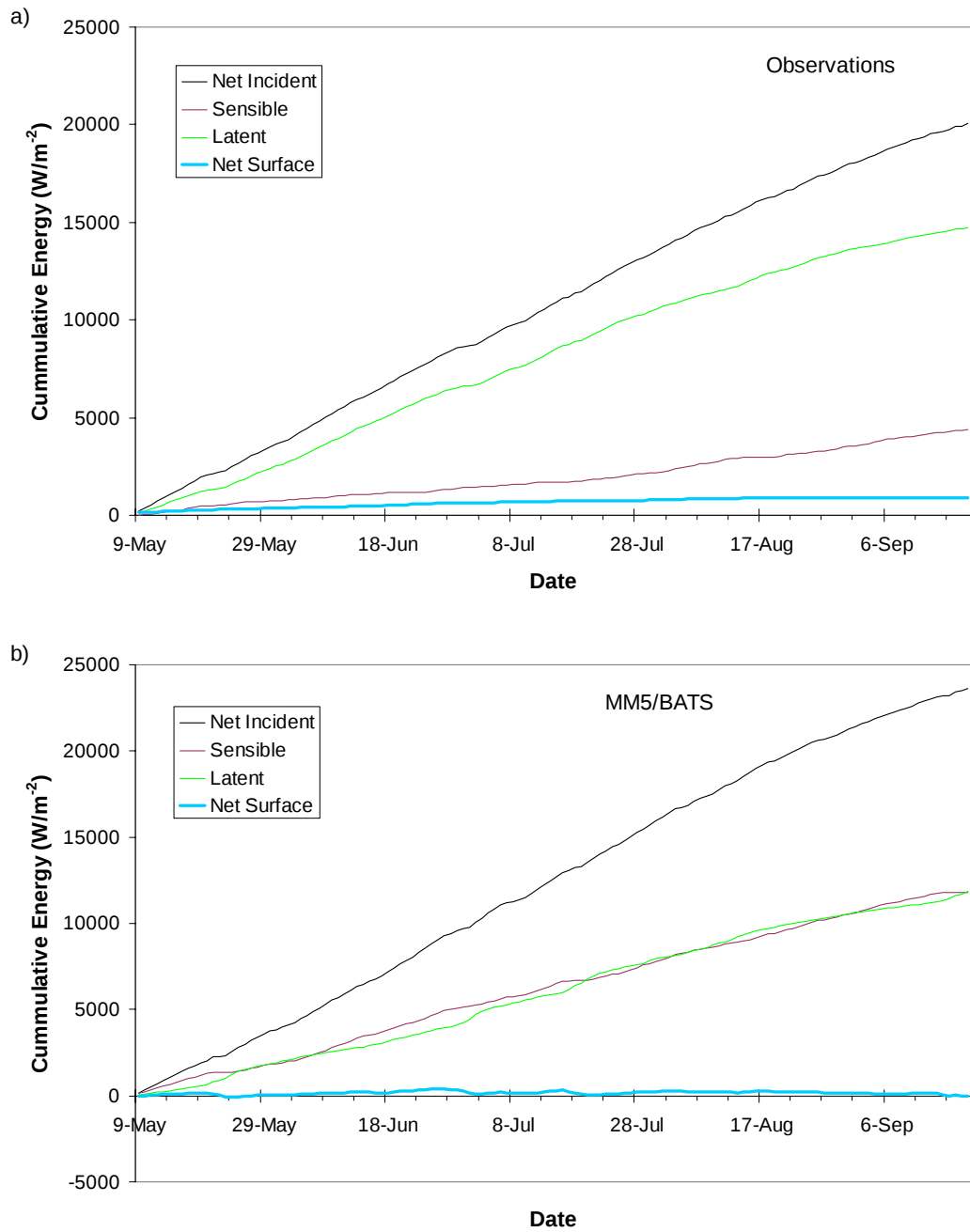


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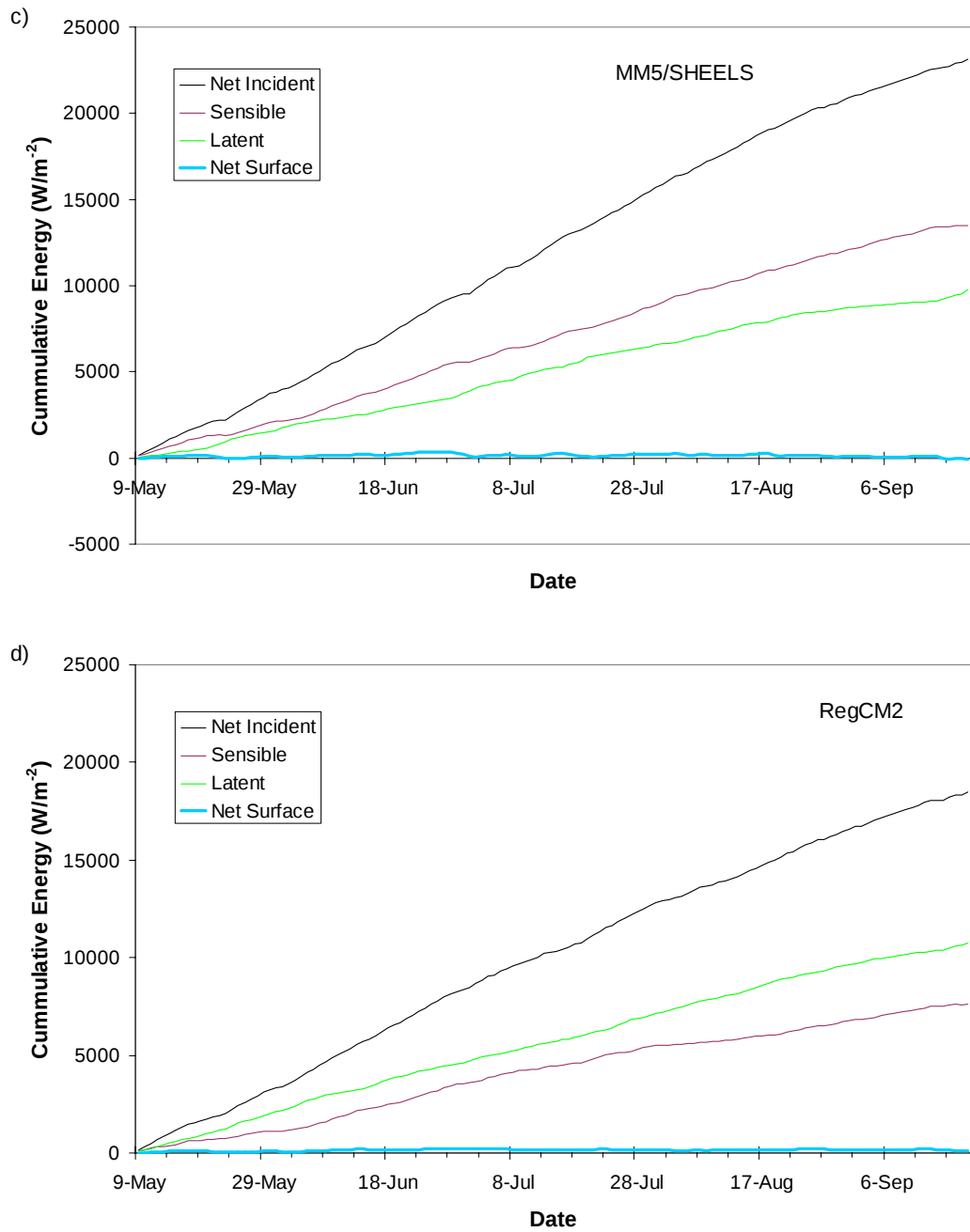


Figure 6.15 Cumulative energy balance during the 1988 observation period:
a) Observations; **b)** MM5/BATS; **c)** MM5/SHEELS and **d)** RegCM2.

7.4 Conclusions

Using the LAM simulations performed over the FIFE region, along with observations taken during several intensive field campaigns, intercomparisons and analysis of the temporal distribution of energy budget terms were conducted. These simulations had a two year duration, starting in 1987 and included the drought year 1988. Analysis of the net incident radiation reveals reasonable agreement with observation, with errors of around 10% to 15% in the cumulative total. The MM5 based LAMs overestimate compared to the observations, while RegCM2 underestimates. Much of the difference between the LAMs occurs in the net longwave radiation where RegCM2's more complicated atmospheric longwave radiation scheme simulates significantly less downward longwave radiation. Overall, no LAM performs best at simulating the diurnal cycle of the net incident radiation though RegCM2 does perform best on a daily basis.

The three LAMs produce reasonable simulations of the daily latent heat in 1987 but significantly underestimate in 1988, the dry year. In terms of the evaporative fraction, RegCM2 comes closest to reproducing the observed. In particular, the high EF values simulated by the MM5 based LAMs are significantly larger than the highest observed values. This suggests that the LAMs convert energy from the ground and sensible heat to the latent heat flux much more readily than observed in the field.

The simulation of the sensible heat flux by all three LAMs demonstrates fairly poor reproduction of the observed values on a daily basis, particularly in early summer, when they consistently overestimate. Of the three LAMs, RegCM2 comes closest to reproducing the observed sensible heat. Comparison of the daily Bowen ratio reveals overestimation by all three LAMs. This overestimation is larger in early summer each year. Again, the LAMs perform worse in 1988, the dry year. Of the three LAMs, RegCM2 simulates a Bowen ratio closest to that observed.

RegCM2 performs the best in terms of reproducing the observed daily net surface heating with the MM5 based LAMs demonstrating considerably more variance than is observed. In a cumulative sense though all the LAMs perform poorly. The observations

indicate a sustained summer warming of the surface which is not reproduced by any of the LAMs. It is usually assumed that over a full annual cycle the net surface heating would be negligible. This is indeed what the three LAMs simulate though it would be useful to compare a full annual cycle of observed net surface heating. The observations available indicate a strong seasonality in the net surface heating which the LAMs do not capture. This then introduces seasonal errors into related variables such as the latent and sensible heat fluxes.

In terms of the overall energy balance there was general agreement amongst the LAMs and observations in 1987, the wet to normal year, while during the dry year, 1988, there is considerable disagreement. This indicates that comparing LAMs under extreme conditions, such as those present in the drought year 1988, emphasises the differences between them. Comparing LAMs under normal conditions then may not reveal model differences which become apparent under extreme conditions. This suggests that while it is important to continue to assess and compare the performance of LAMs over various land use types, much of the potential information from such studies may not be obtained unless extreme climatic conditions are used.

In almost all the surface energy related variables investigated here, the MM5 based LAMs simulate considerably more variance than RegCM2, which itself tends to simulate larger variance than the observations. This larger variance of the MM5 based LAMs compared to RegCM2 is related to the larger net incident radiation available to drive the MM5 land surface parameterisations. On the other hand, RegCM2 has less net incident radiation than is observed, yet still simulated a larger variance. This suggests that BATS, and probably SHEELS, inherently simulate more variation than is observed. That is, the ground heat and moisture stores are able to change significantly faster than what is observed in the field, at least for this particular land type.

In the overall simulation of the energy balance RegCM2 performs better than the MM5 based LAMs due in part to more consistent simulations of wet and dry years. MM5/SHEELS tends to simulate the largest differences between the wet and dry years with the energy balance in the dry year being particularly poor.

Chapter 7

Regional Climate Model Performance Analysis and Comparison: the Water Budget

7.1 Introduction

The previous chapter analyzed the performance of the limited area models (LAMs) in terms of their prediction of energy variables. This chapter focuses on their performance in terms of water fluxes. Currently there exists considerable disagreement among climate models in terms of various water fluxes in their hydrological cycle representations (e.g. Lohmann et al. (1998)). Under global warming conditions potential changes in various aspects of the hydrological cycle, such as runoff and soil moisture, are of major interest due to their impact on ecosystem function. This is particularly true in arid and semi-arid regions where successful modelling of the water balance is particularly difficult (Ye 1996).

A general form of the surface water budget is given in equation 3.1. Similarly to the ground heat flux, the change in soil moisture in a stable climate should be negligible over the long term. Unlike the ground heat flux though, soil moisture can change considerably over the short to medium term. Even on an interannual basis the soil moisture can vary greatly from flood years to drought years. Given long term climate change it seems reasonable to assume there would be an associated change in the soil moisture regime. Isolating this change from the natural variability is very difficult, especially given the large range of soil moisture estimates of climate models (e.g. Shao et al. (1994)).

Confidence in any predictions of climate change then requires confidence in the associated models' representation of the water budget. In the remainder of this chapter the representation of the water budget in three LAMs (MM5/BATS, MM5/SHEELS and RegCM2) are intercompared and tested against observations.

7.2 Experiment Design

Results presented in this chapter were obtained from the same model runs discussed in section 6.2.1. Briefly the three LAMs were run for two years on a domain covering an area of $1200 \times 1260 \text{ km}^2$, centered over the FIFE region. The LAMs were run with 17 levels in the vertical, horizontal grid point spacing of 20km and a time step of 1 minute.

The model results are compared to observation taken during FIFE (section 6.2.2) as well as data collected as part of the Konza LTER site, which includes Kings Creek catchment in the north west corner of the FIFE domain.

7.3 Results and Discussion

7.3.1 *Precipitation*

When studying the water balance at the land surface, inputs to the system come essentially from precipitation. The analysis begins with comparison of the simulated precipitation and its components.

7.3.1.1 **Convective precipitation**

Convective precipitation tends to occur over a limited spatial extent. This means that while it may be possible to estimate the convective precipitation on a regional basis, it is much more difficult to locate each convective event accurately. This implies that point comparisons of precipitation simulated by different convective schemes may not be very informative. This section presents results obtained using LAMs containing the same convective scheme (the Grell scheme) but differing in other physical parameterisations. Thus differences in convective rainfall are indicative of the influence of other LAM components on the convective parameterisation.

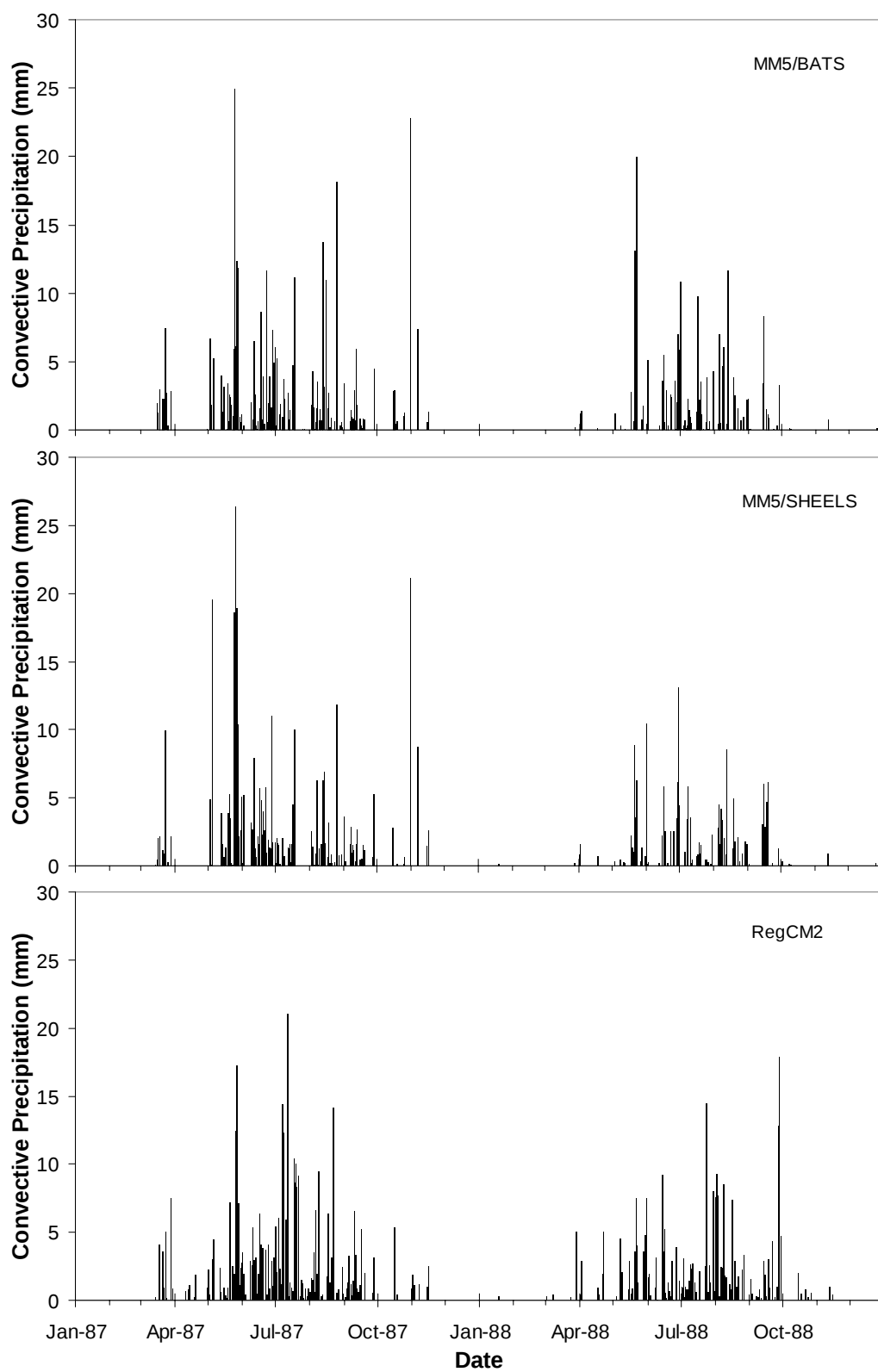


Figure 7.1 Daily convective precipitation simulated by the three LAMs

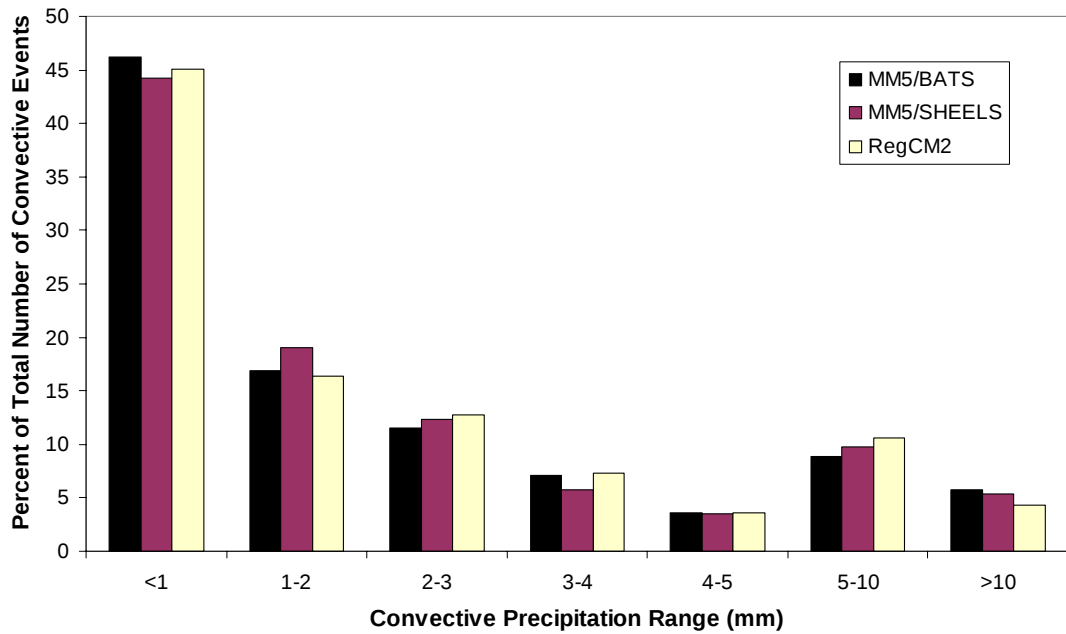


Figure 7.2 Comparative histogram of convective precipitation events

The simulated daily convective precipitation for all three models is shown in Figure 7.1. Significant variation is evident among the models, with RegCM2 displaying significantly different behaviour to the MM5 based LAMs. The two MM5 based LAMs exhibit similar event timing though major differences may exist in the quantity simulated. Over the two year integration MM5/BATS simulates one convective event less than MM5/SHEELS, while RegCM2 simulates an extra 50 events.

The distribution of the magnitude of the convective precipitation events is shown in Figure 7.2. In percentage terms the Grell scheme produces similar distributions for all three models. This implies that once convective precipitation has occurred in the LAMs, similar ranges of mass flux (or buoyant energy), via equations 5.28 and 5.29, are produced. However, RegCM2 is more likely to encounter unstable atmospheric conditions which induce convective precipitation. The cause of this may be related to differences in the PBL parameterization, the coupling between the convective and stable precipitation parameterizations and/or differences in the predicted land surface evaporation.

A similar seasonal cycle for each model is shown in Figure 7.3, with convective precipitation occurring almost exclusively in the summer months (May to September). Much more convective precipitation was produced in 1987 than 1988 for all models. Much less variation among the models is seen in 1988, which was the drier year of the two. At least some of the variability between models, seen mainly in 1987, can be explained by the spatial mismatch between models for the locations of these local convective storms.

RegCM2 produces a significantly larger total amount of convective precipitation over the two years, producing around 120 mm more than either of the MM5 based LAMs (Figure 7.4). This is almost certainly related to atmospheric stability rather than local production of evaporation since RegCM2 produces the least amount of evaporation overall. MM5/SHEELS produces more convective precipitation in 1987 than MM5/BATS but less in 1988. This is almost certainly related to the simulation of evaporation shown in Figure 7.12 where MM5/SHEELS produces more evaporation than MM5/BATS in 1987, but less in 1988.

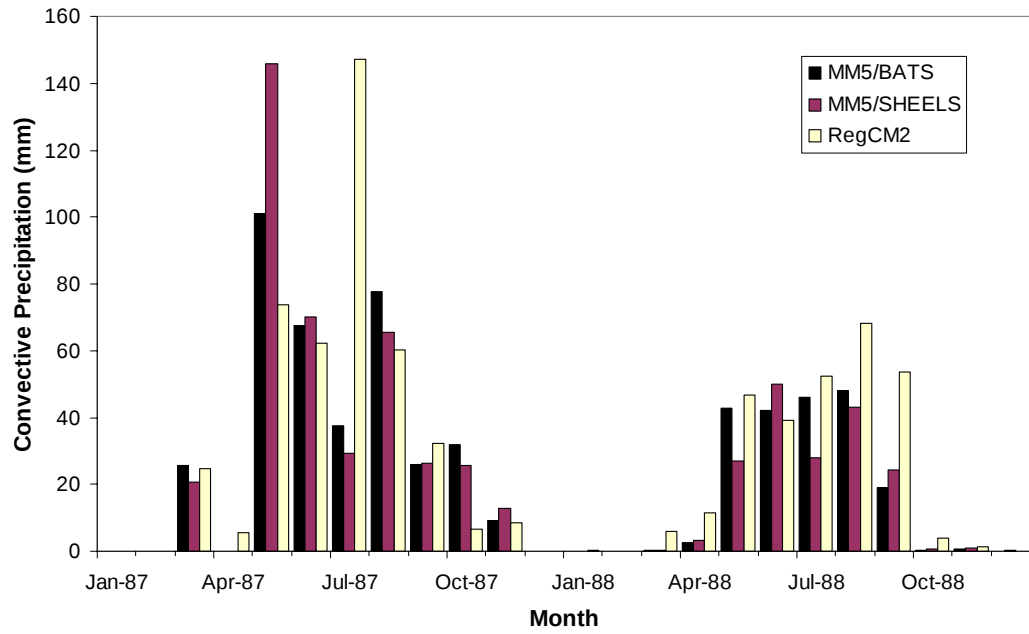


Figure 7.3 Monthly convective precipitation totals

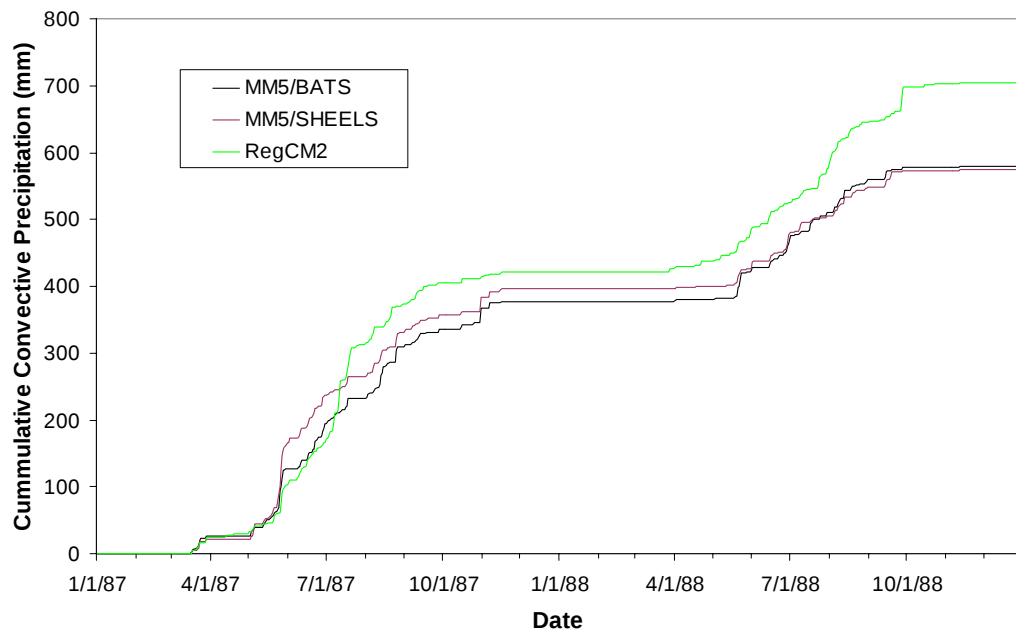
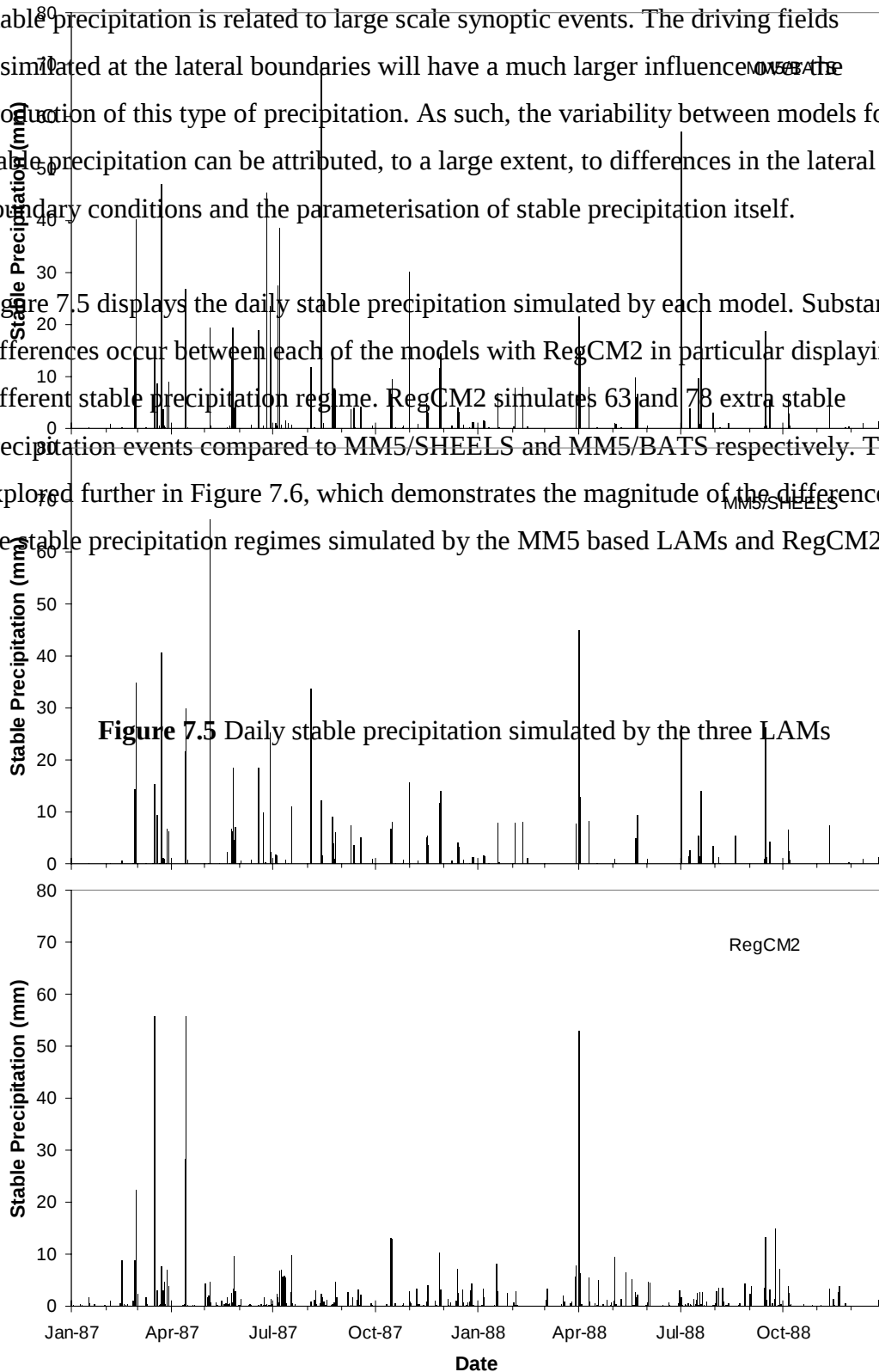


Figure 7.4 Cumulative convective precipitation for all three LAMs

7.3.1.2 Stable precipitation

Stable precipitation is related to large scale synoptic events. The driving fields assimilated at the lateral boundaries will have a much larger influence over the production of this type of precipitation. As such, the variability between models for stable precipitation can be attributed, to a large extent, to differences in the lateral boundary conditions and the parameterisation of stable precipitation itself.

Figure 7.5 displays the daily stable precipitation simulated by each model. Substantial differences occur between each of the models with RegCM2 in particular displaying a different stable precipitation regime. RegCM2 simulates 63 and 78 extra stable precipitation events compared to MM5/SHEELS and MM5/BATS respectively. This is explored further in Figure 7.6, which demonstrates the magnitude of the difference in the stable precipitation regimes simulated by the MM5 based LAMs and RegCM2.



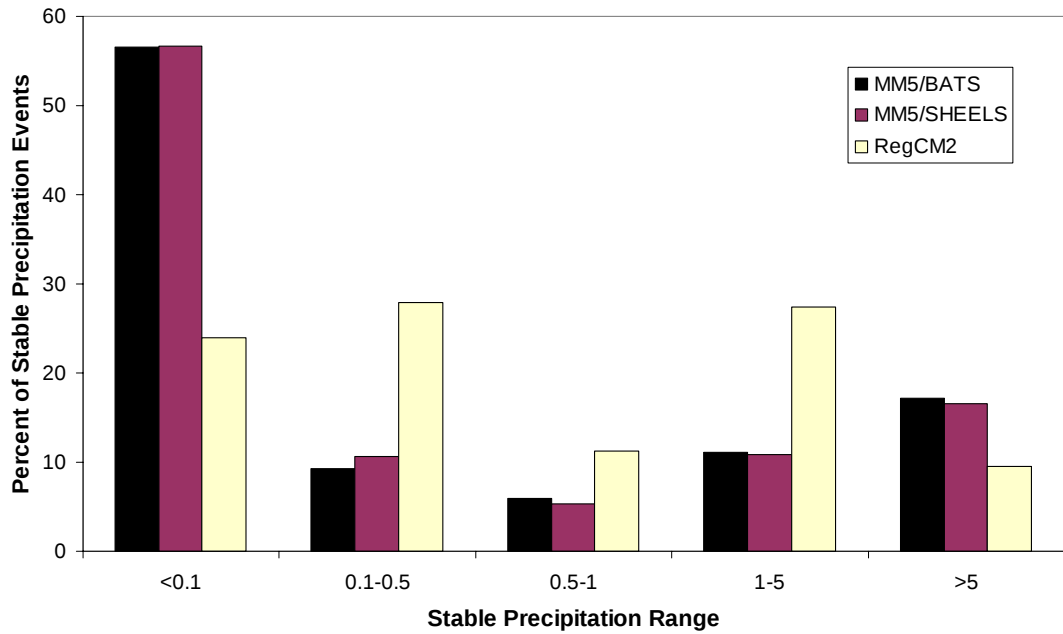


Figure 7.6 Comparative histogram of stable precipitation events

Over half of the stable precipitation events simulated by the MM5 based LAMs produced less than 0.1 mm of precipitation compared to around 25% of the events simulated by RegCM2. The MM5 based LAMs also simulated events greater than 5 mm almost 20% of the time while RegCM2 did the same only around 10% of the time. That is, approximately 75% of the stable precipitation events simulated by the MM5 based models produce extreme quantities of precipitation. RegCM2, on the other hand, simulated 65% of events to lie within the range 0.1 mm to 5 mm. Thus it seems that the inclusion of a more physical lateral boundary condition (equations 5.86-5.89) as well as horizontal diffusion and vertical turbulent mixing in the stable precipitation parameterisation (equations 5.109-5.111) in RegCM2 moderates the stable precipitation quantities simulated, while also causing the simulation of a substantially larger number of events.

Investigation of monthly total stable precipitation, as shown in Figure 7.7, reveals no clear seasonal variation unlike the convective precipitation. On a monthly basis all the LAMs demonstrate a similar amount of variation in their stable precipitation totals

which implies that the extreme values seen in the daily totals for the MM5 based models are scattered throughout the year.

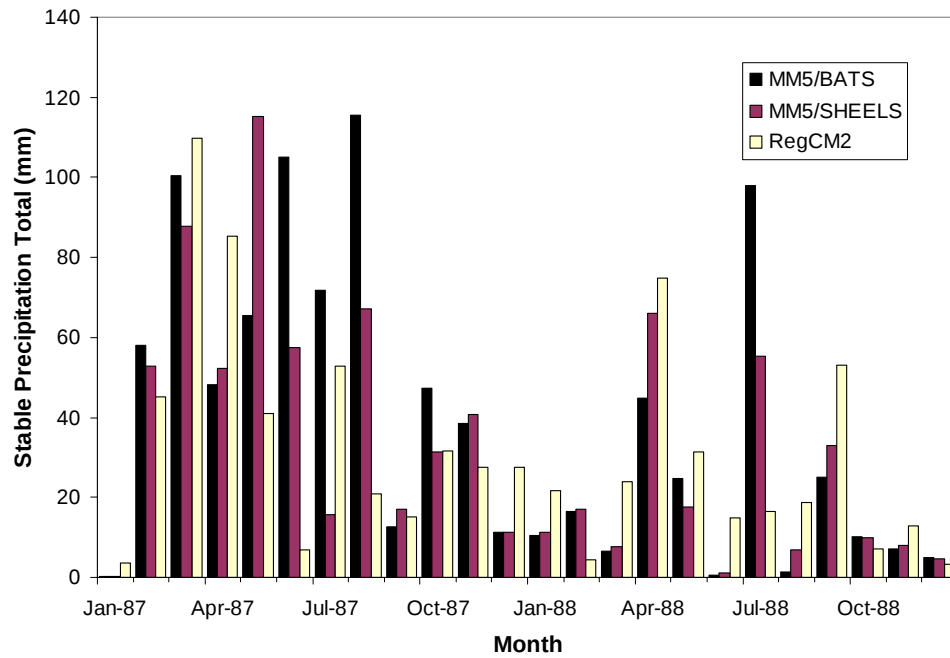


Figure 7.7 Monthly stable precipitation totals

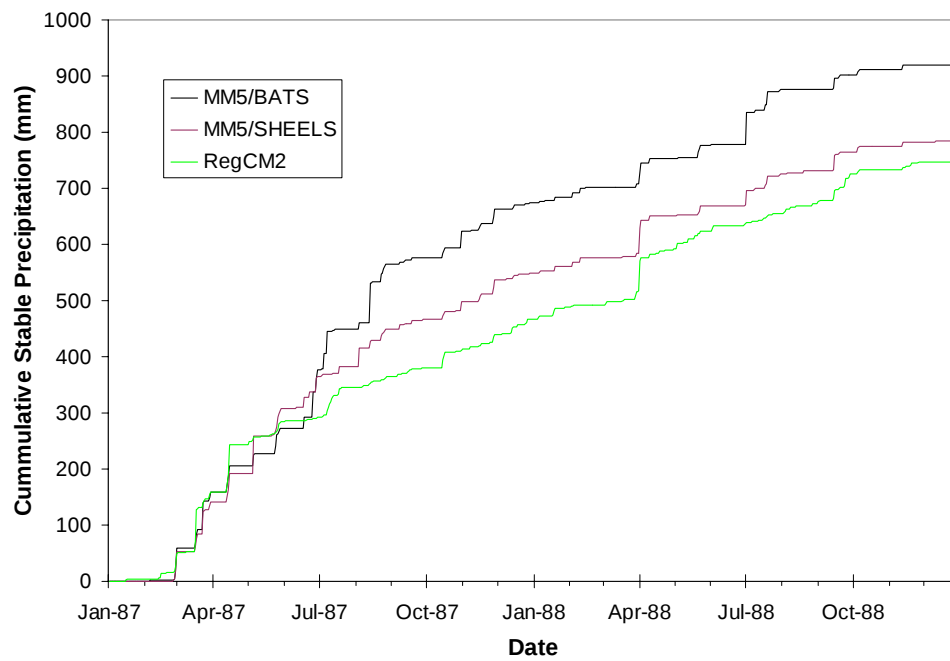


Figure 7.8 Cumulative stable precipitation for all three LAMs

Over the two years RegCM2 produces the least amount of stable precipitation in total. This is despite simulating many more stable precipitation events (Figure 7.8). The larger proportion of extreme events simulated by the MM5 based LAMs is enough to guarantee that they produce larger overall totals. It can be seen in Figure 7.8 that a difference of almost 150 mm exists between the total stable precipitation of the two MM5 based LAMs. This implies that differences in land surface treatment can have a significant impact on the simulation of stable precipitation as well as convective precipitation.

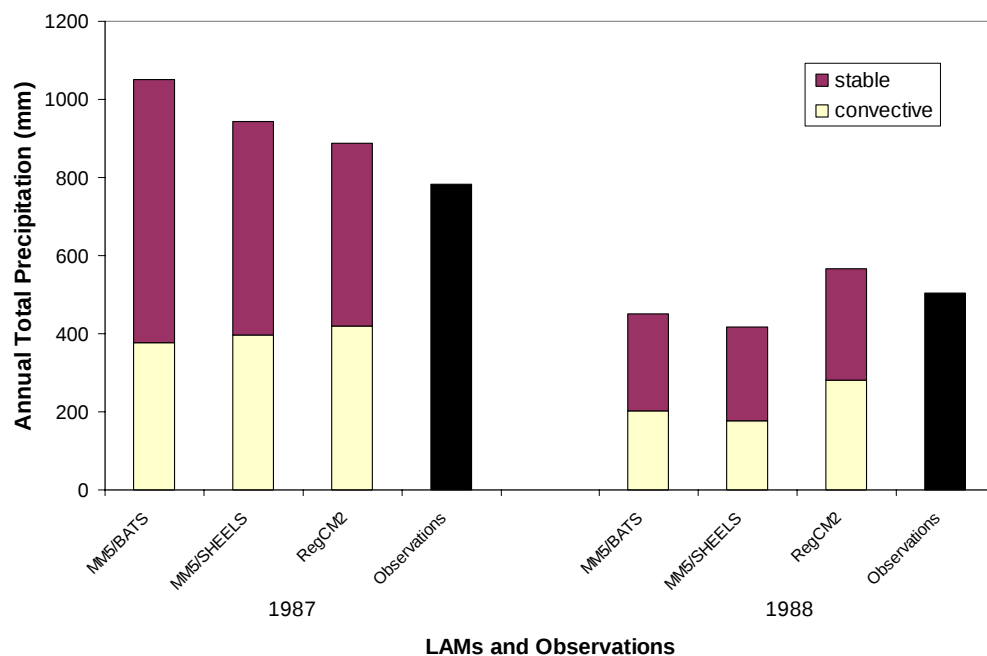


Figure 7.9 Contribution of convective and stable precipitation to the annual total

7.3.1.3 Total precipitation

Observations of precipitation do not discriminate between stable and convective precipitation. Here comparisons are made between total precipitation simulated by the three LAMs and observations taken over the two years. Figure 7.9 shows the annual totals along with the contributions made by convective and stable precipitation for each of the LAMs. In both years RegCM2 simulates an almost even split between convective and stable precipitation. MM5/SHEELS simulates 57% to 58% stable precipitation in both years. MM5/BATS on the other hand, simulates about 64% of the precipitation as stable precipitation in 1987, with this dropping to around 55% of the precipitation in 1988. MM5/BATS simulates total precipitation furthest from the observations in 1987, with an extra 270 mm of rain falling. In 1988 MM5/SHEELS performs the worst, underpredicting the precipitation by almost 90mm. MM5/SHEELS exhibits the largest change in precipitation between 1987 and 1988 indicating a greater sensitivity to the drier conditions experienced in 1988 through the soil moisture parameterisation. The simulations of precipitation in all three LAMs appear to be more sensitive to the wider conditions (1987 being wet and 1988 being dry) than the observations, with the RegCM2 sensitivity being closest to that observed. Overall RegCM2 appears to reproduce annual precipitation totals better than the other LAMs though it seems to consistently overestimate.

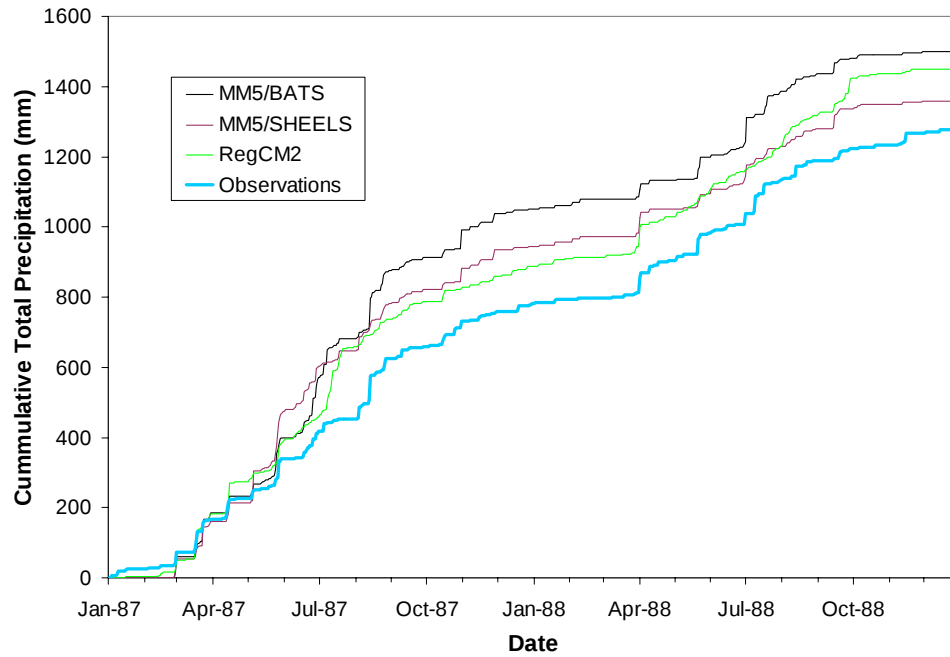


Figure 7.10 Cumulative total precipitation for all three LAMs

While RegCM2 simulates the annual totals better overall, MM5/SHEELS simulates the cumulative total better due to its overestimation of precipitation in the wetter year (1987) and a compensating underestimation of precipitation in the dry year (1988). This can be seen in Figure 7.10. All three LAMs overestimate the total precipitation over the two year period with the overestimates ranging from 77 mm for MM5/SHEELS up to 218 mm for MM5/BATS. As a proportion of the observed precipitation the simulations range from 6% to 17% overestimation.

7.3.2 Evapotranspiration

In the previous chapter the ET (latent heat) was expressed in terms of an energy flux, here it is expressed in terms of a mass flux. Observations of ET were only taken during intensive observation periods and as such only exist for a few months each year. Presented here is an analysis of the performance of the LAMs in terms of the simulation of ET. Particular attention is paid to the periods for which observations exist. Many components of a LAM influence the simulation of ET at the land surface. These include

the energy input to the system (radiation parameterisation), the moisture input to the system (precipitation parameterisation) and most importantly the allocation of a proportion of each input to ET (land-surface parameterisation). The relationship between ET and the energy input to the system was examined in section 6.3.2. Here the focus is on the relationship with other terms in the water budget.

An examination of monthly total ET, shown in Figure 7.11, gives an indication of the seasonal cycle simulated by the LAMs. Note that RegCM2 encountered numerical stability problems in February 1988 and hence this month was excluded from the RegCM2 simulated response. In general, all the LAMs follow the same seasonal cycle with the maximum occurring during summer and the minimum during winter, which mirrors the seasonal cycle of the incident radiation. In all the LAMs this cycle is less pronounced in 1988 (dry year). The LAMs all agree that less ET occurred during 1988 with RegCM2 simulating a relatively small decrease, MM5/SHEELS simulating a very large decrease and MM5/BATS somewhere in between the two. The observations, on the other hand, indicate that only a marginal decrease occurred. In terms of the difference between the two years (a wet and a dry year), RegCM2 therefore performs the best but suffers from a consistent underestimation bias compared to the observations.

MM5/SHEELS however, displays a much greater sensitivity to the available moisture conditions than is seen in the observations. During 1987 MM5/SHEELS tends to simulate the most ET of the three LAMs, even slightly overestimating compared to the observations. During 1988 MM5/SHEELS tends to simulate the least amount of ET of the three LAMs, which is a significant underestimation when compared to the observations. In 1987 MM5/BATS performs the best of the three LAMs, though it significantly underestimates the ET in 1988 along with the other two LAMs. The ranking of the three LAMs from the most sensitive to available moisture conditions, to

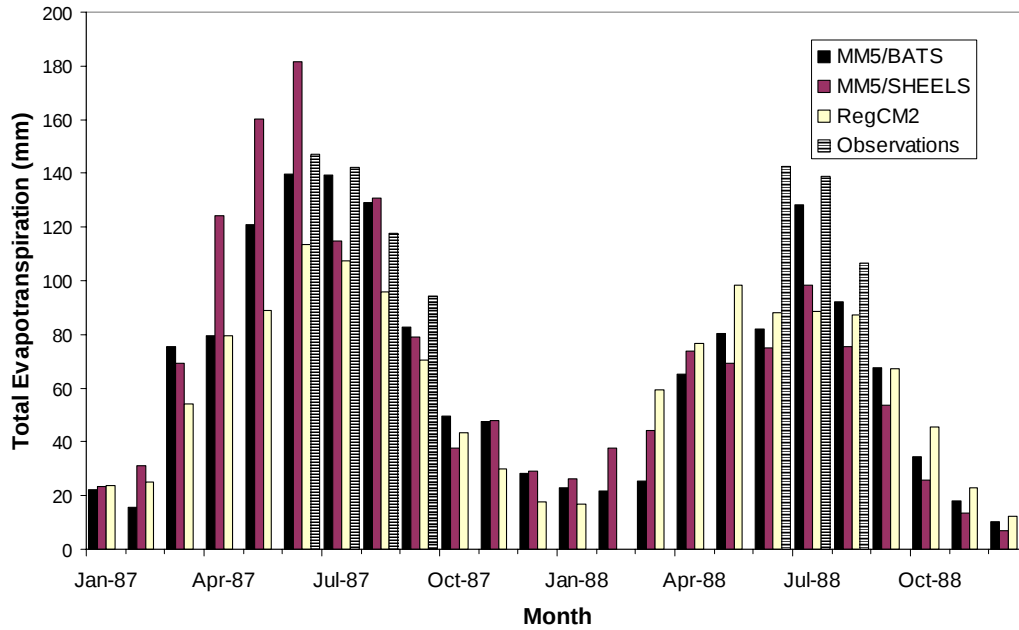


Figure 7.11 Monthly total ET simulated by all three LAMs

the least is MM5/SHEELS, MM5/BATS and RegCM2. Thus it appears that the SHEELS land-surface parameterisation exhibits a greater range of variability, on a monthly basis, than BATS. This variability is further exacerbated by the greater variability present in the precipitation simulated by the MM5 based LAMs.

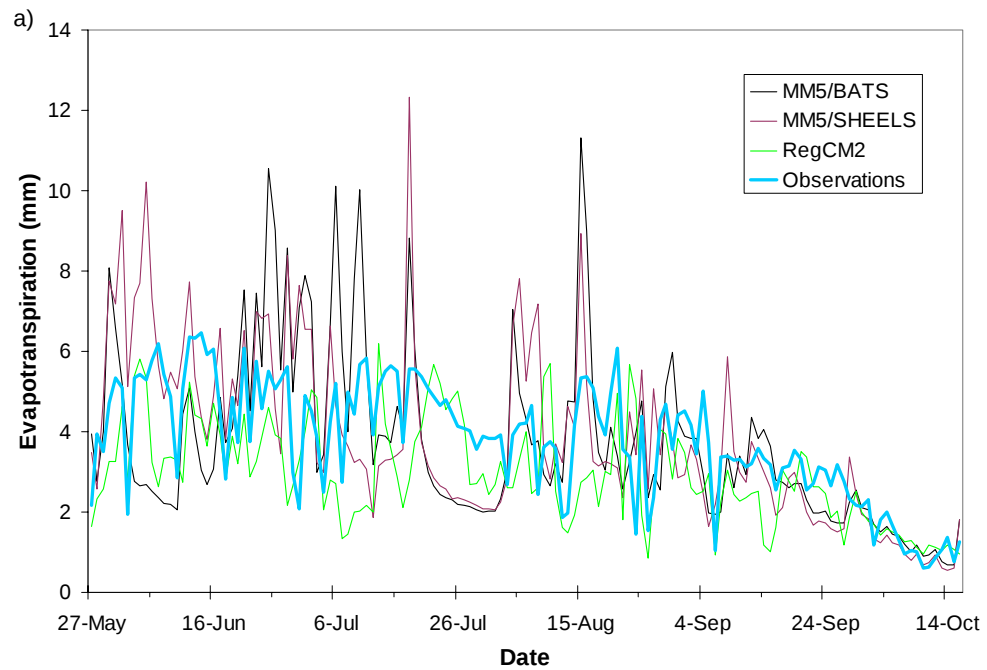
The daily ET estimates simulated by the LAMs is presented in Figure 7.12. The daily variation simulated by the MM5 based LAMs is significantly greater than that simulated by RegCM2 and is present in the observations during 1987 (Figure 7.12 a). The MM5 based LAMs daily ET estimates also, on average, appear to be closer to the observed than RegCM2. In 1988 (Figure 7.12 b) all the LAMs appear to underestimate the ET on average, while the variance appears much closer to the observed variance than for 1987.

Quantitative measures for the variance and mean are given in Table 7.1 for the common observation period in each year. It can be seen that the MM5 based LAMs reproduce the mean daily ET well in 1987, but massively overestimate the variance present (given by the standard deviation), while RegCM2 significantly underestimates the daily ET in 1987 but reproduces the observed variance well. During 1988 though, all the LAMs

significantly underestimate the mean daily ET, with the MM5 based LAMs doing well in reproducing the observed variance and RegCM2 underestimating this

Table 7.1 Means and standard deviations over the common observation period;
27 May to 19 September.

	1987		1988	
	Mean (mm)	St. Dev. (mm)	Mean (mm)	St. Dev. (mm)
Observations	4.230	1.194	3.887	1.412
MM5/BATS	4.250	2.107	3.115	1.490
MM5/SHEELS	4.396	2.031	2.556	1.340
RegCM2	3.228	1.197	2.757	0.983



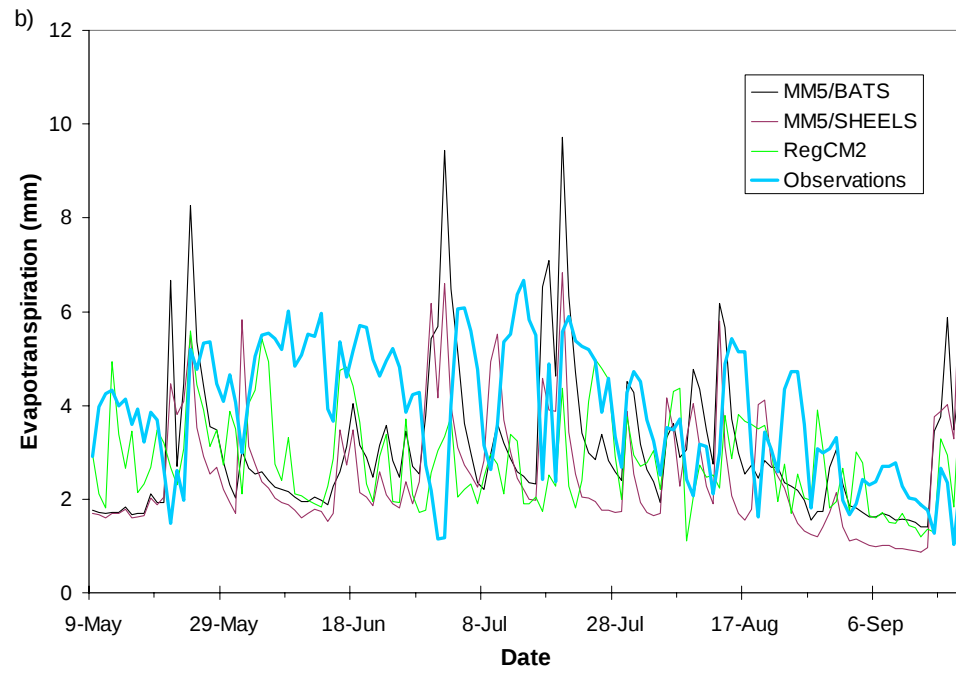


Figure 7.12 Daily ET simulated by the three LAMs **a)** during the 1987 observation period and **b)** during the 1988 observation period.

variance. It is worth noting that moving from a wet to a dry year a decrease in the mean daily ET accompanied by an increase in the variance can be observed, while all three LAMs expect the variance to decrease in the drier year. MM5/SHEELS again demonstrates large sensitivity to the available moisture conditions, having by far the largest change in mean daily ET from 1987 to 1988.

Figure 7.13 presents the cumulative ET simulated by each model through the observation period each year. This serves to emphasise some of the findings above. In 1987 (Figure 7.13 a) the two MM5 based LAMs simulate the total ET quite well, while RegCM2 underestimates this total by over 100 mm. The MM5 based LAMs simulate similar ET throughout the 1987 observation period except during June when MM5/BATS initially simulates less ET than MM5/SHEELS, but later in June MM5/BATS simulates more ET. This can be explained by the extra precipitation simulated by MM5/SHEELS late in May, as can be seen in Figure 7.10. This extra moisture available to MM5/SHEELS allows it to produce ET at the potential rate throughout early June. By the end of June the total precipitation simulated by MM5/BATS has reached the level simulated by MM5/SHEELS and the difference between the total ET simulated by each model is reduced.

Figure 7.13 b) shows all three LAMs significantly underestimating the total ET during the 1988 observation period. MM5/SHEELS simulates the least amount of ET during 1988 while during 1987 it simulated the most. During this observation period MM5/BATS consistently simulated more ET than MM5/SHEELS. This is reflected in Figure 7.10 where it can be seen that MM5/BATS consistently simulates more total precipitation than MM5/SHEELS throughout the 1988 observation period.

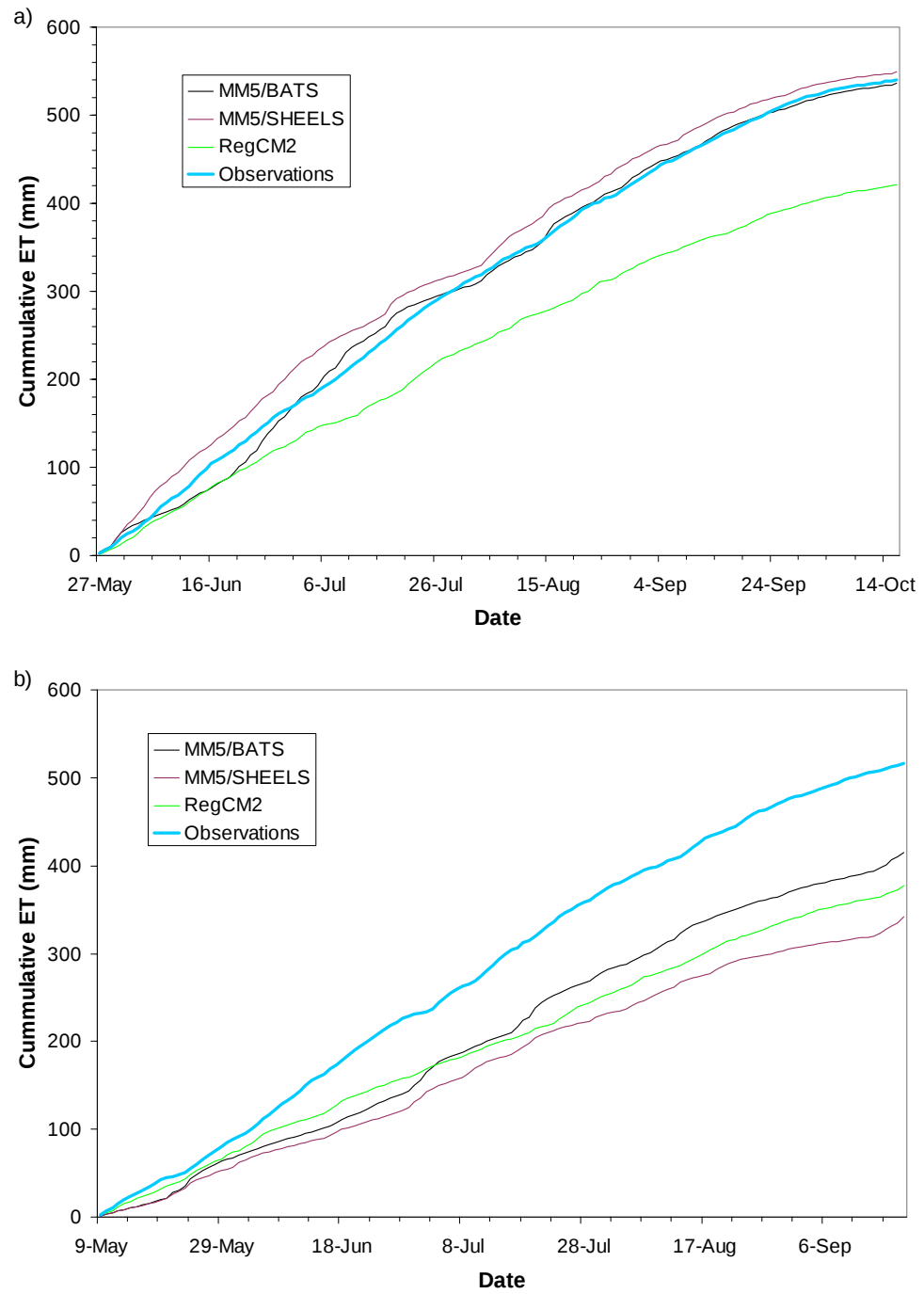


Figure 7.13 Cumulative ET for all three LAMs during the observation period in **a) 1987** and **b) 1988**.

7.3.3 Runoff

In this section the runoff simulated by each of the LAMs is compared with observations taken from Kings Creek Catchment in the North-West quadrant of the FIFE site. The runoff simulated by the LAMs is representative of runoff from a 20 km × 20 km grid square. This runoff is then scaled down to be representative of an area of approximately 11 km² (the area covered by Kings Creek catchment) using an area proportionality approach. This can be justified in the sense that since the LAMs represent the grid square using a single land cover type, assuming homogeneity within the grid square for the land-surface parameters.

The monthly runoff totals are presented in Figure 7.14. All three LAMs agree with the observations in that much less runoff is predicted to occur in 1988 than 1987. That, however, is as far as the similarities go. The observations display strong seasonality with very little runoff occurring before March and after June in both years, with the peak occurring in March or April. This spring domination of runoff is not simulated by the LAMs, which tend to simulate runoff throughout the year. During the wet year (1987) MM5/BATS simulates significant runoff to occur from March through to

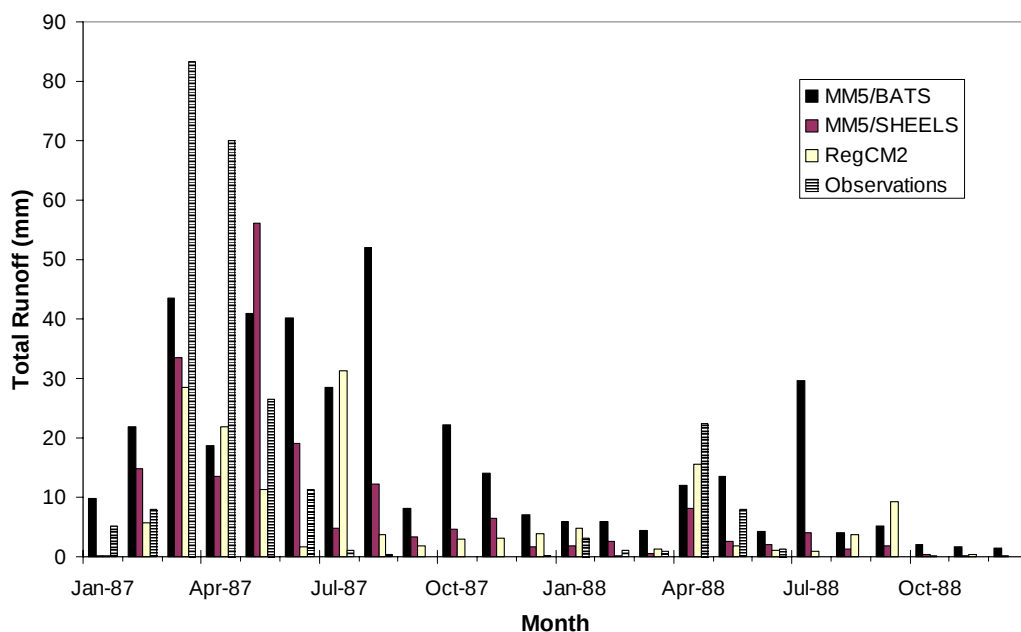


Figure 7.14 Monthly total runoff simulated by all three LAMs

August with no real peak within this span. MM5/SHEELS performs best in terms of reproducing the observed seasonal cycle with the majority of runoff occurring at the correct time of year though the peak occurs in May, some two months after the observed peak. RegCM2 simulates a runoff peak in May, in agreement with observations, but it also simulates another runoff peak in July, which is not observed at all.

During 1988 runoff is observed to occur in April and May with almost no runoff occurring in any other month. All three LAMs simulate runoff from April to September. MM5/BATS simulates most runoff to occur between April and July, with July having the largest single total even though no runoff is observed then. MM5/SHEELS again performs best at reproducing the seasonal cycle with the runoff peak occurring in April, the same month as the observed peak., though it does simulate runoff to continue to occur much later in the year than is observed. RegCM2 also reproduces well the peak in April though, as in 1987, a secondary peak is simulated to occur later in the year, this time it occurs in September.

Daily runoff simulated by the LAMs can be seen in Figure 7.15. Clearly, none of the LAMs is able to reproduce a hydrograph such as is seen in natural streamflow. All the LAMs produce extremely spiky response to the driving precipitation, which is reminiscent of a bucket model style overflow mechanism. For RegCM2 and MM5/BATS the surface runoff is given by equation 5.54, which simulates runoff (in the absence of snow) as a proportion (depending on soil moisture) of the precipitation minus ET. This formulation then allows runoff to occur only on days when precipitation is greater than ET. Unlike observed runoff then, BATS cannot simulate surface runoff on days that do not have any precipitation predicted. MM5/SHEELS has a similar problem in the simulation of runoff though here the runoff is related to the precipitation minus the infiltration (equation 5.81).

The cumulative runoff for all three LAMs is given in Figure 7.16. The massive range of runoffs simulated by the LAMs is clear here. MM5/BATS simulates almost twice as

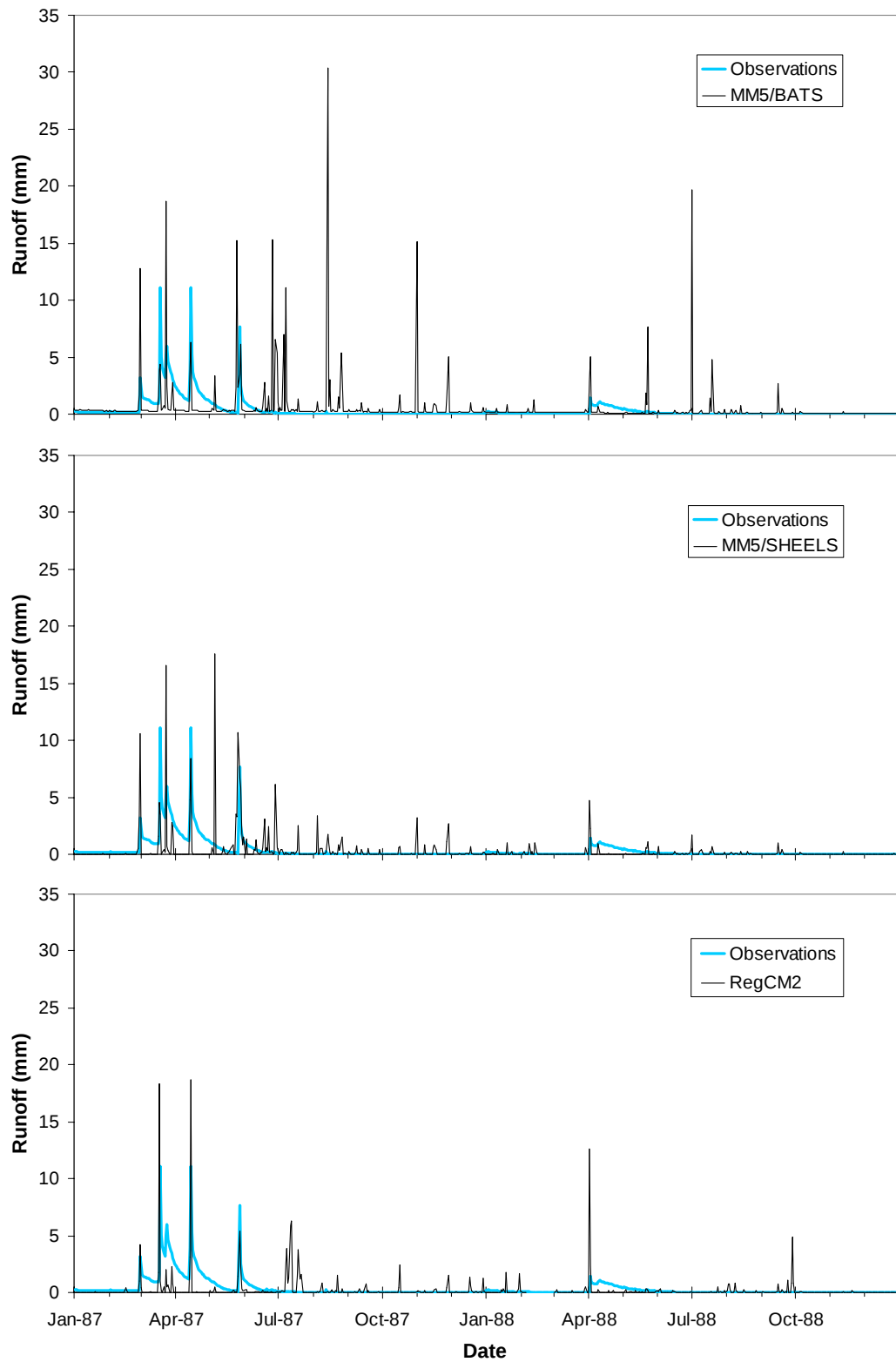


Figure 7.15 Daily runoff simulated by all three LAMs

much runoff as is observed over the two year period, overestimating the runoff by 178 mm. MM5/SHEELS underestimates the total runoff but is closest to the observed of all the LAMs, underestimating by only 24 mm, and RegCM2 underestimates the total runoff by 64 mm. The difference in runoff simulated by MM5/BATS and RegCM2 can be explained in part by the higher overall precipitation simulated by MM5/BATS and perhaps more importantly by the higher proportion of large precipitation events (greater than 5 mm) simulated by the MM5 based LAMs and shown in Figure 7.6. This can also be seen in Figure 7.15 where MM5/BATS simulates many more runoff events larger than 10 mm than either of the other two LAMs.

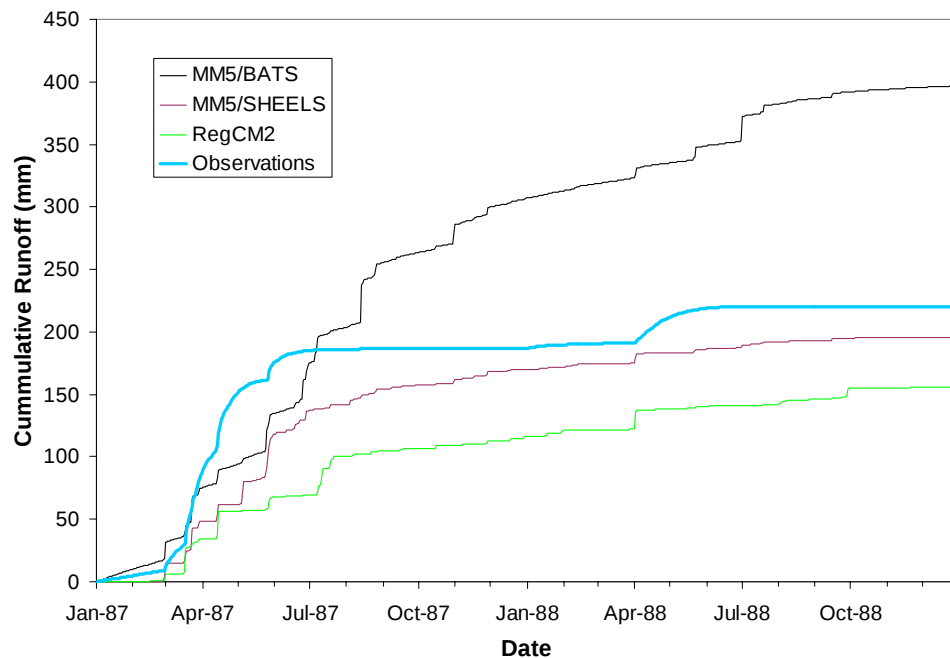


Figure 7.16 Cumulative runoff simulated by all three LAMs

7.3.4 Soil Moisture

Intercomparison of soil moisture simulated by the three LAMs is presented in this section. Two layers in the soil profile are considered, the surface layer which is

considered to be the top 10 cm, and the root zone taken to be the top 1 m of soil. Observations of soil moisture in the surface layer were taken using a gravimetric water

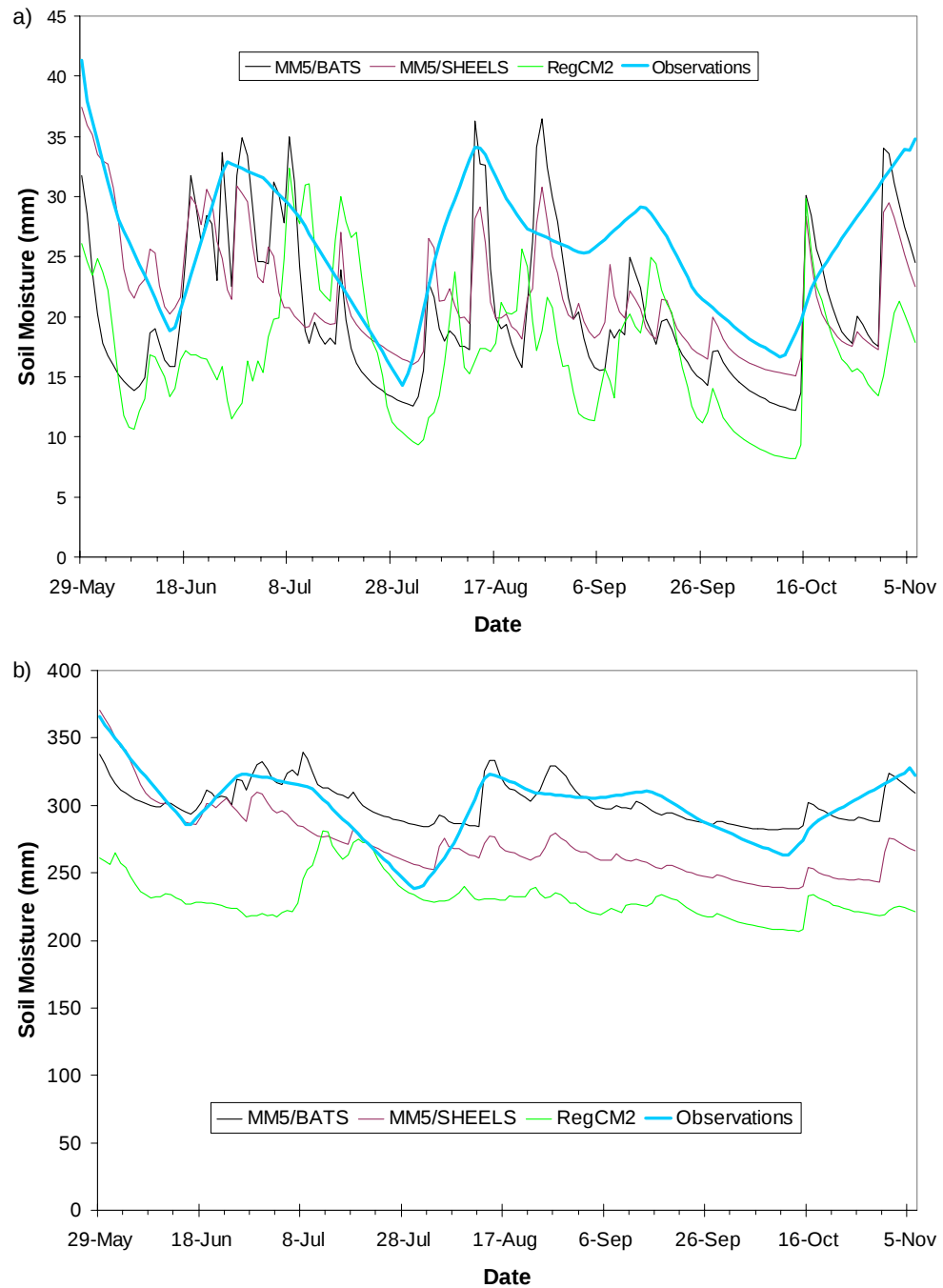


Figure 7.17 Soil moisture simulated in 1987 by all three LAMs in the
a) surface layer and **b)** root zone.

content method (weight of water in the sample/weight of dried sample), while deeper observations of soil moisture were taken using a neutron probe, down to a depth of two metres. Here only the top metre is considered. Measurements were made at as many as 30 sites, then an areal average for each soil layer was compiled by Betts and Ball (1998). In 1987 the observations began on the 29th of May and continued until the sixth of November, while in 1988 the observations began on the 11th of April and finished on the 29th of September.

Surface layer soil moisture is important due to its influence over the evaporation from bare soil (equation 5.51) and the surface runoff (equations 5.54 and 5.81). The observed soil moisture in the surface layer, shown in Figure 7.17 a), displays a much smoother response than is simulated by any of the LAMs. The MM5 based LAMs perform best at reproducing the observed surface layer soil moisture, with RegCM2 consistently simulating the least amount of soil moisture. The range of soil moisture values observed in the surface layer is reproduced by the LAMs reasonably well. Over this observation period a decrease in surface soil moisture of around 7 mm is observed. This decrease is simulated accurately by both MM5/BATS and RegCM2. MM5/SHEELS however, simulates a decrease more than double that observed.

The root zone soil moisture exerts a strong influence on the transpiration (equation 5.68 and 5.69) and sub-surface temperature (equation 5.42-5.45) and a secondary influence on the surface runoff. The observed soil moisture in the root zone can be seen in Figure 7.17 b). Again the observations produce a smoother time series than is simulated by any of the LAMs. RegCM2 simulates less root zone soil moisture than either of the other two LAMs or the observations, throughout the period under consideration. Overall, MM5/BATS performs best at reproducing the observed moisture in the root zone. Over this observation period a decrease of 43 mm is observed in the root zone soil moisture. RegCM2 simulates a decrease of 40 mm, MM5/BATS a decrease of 29 mm and MM5/SHEELS a decrease of around 104 mm.

Similar graphs for the 1988 observation period can be seen in Figure 7.18. Again it is clear that RegCM2 consistently simulates the least amount of soil moisture throughout the soil profile. In the surface layer all the LAMs begin the period with surface moisture

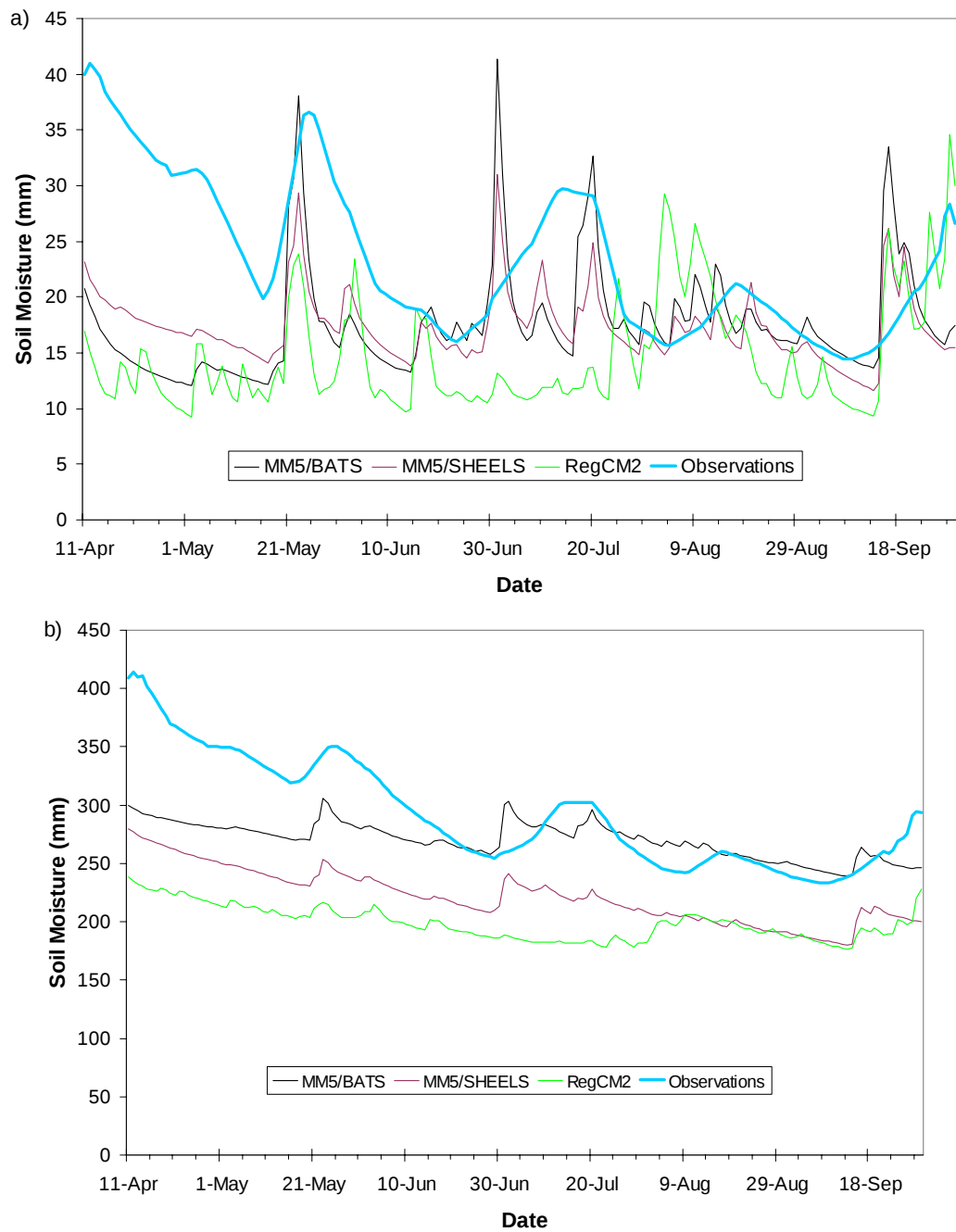


Figure 7.17 Soil moisture simulated in 1988 by all three LAMs in the
a) surface layer and **b)** root zone.

estimated to be around half of what was observed. By the end of the period though, this massive underestimation has been reduced considerably since the observations display a decrease through the period, while the LAMs simulate the surface moisture to remain fairly constant (MM5/BATS and MM5/SHEELS) or even increase (RegCM2). The observations also show a considerable drying out of the root zone with an overall decrease of 115 mm, which would be expected during a year that is considered a drought year. While all three LAMs simulate a decrease in the root zone moisture during this period, the decrease simulated is much less than that observed. Of the LAMs, MM5/SHEELS simulated the largest decrease in root zone soil moisture (79 mm).

Clearly the SHEELS land surface parameterisation simulates a greater range of soil moisture values than BATS. This can be attributed to the use of many thin soil layers in SHEELS.

7.3.5 Surface Water Balance

The cumulative water balance over FIFE for each LAM during the observation periods in 1987 and 1988 are presented in Figures 7.18 and 7.19 respectively. These figures include the change in soil moisture down to a depth of 1 m. The soil extends further than this though, with the median soil depth being 1.5 m (Duan et al. 1996). Also included in these figures is the residual of the water balance, $\mathfrak{R}$, defined by

$$\mathfrak{R} = P - ET - R - \Delta SM \quad (7.1)$$

This residual is a measure of the water imbalance at the surface. Figure 7.18a) presents the observed water balance in 1987. Over this summer period, the observations have a residual of –41 mm. This agrees well with the water imbalance found by Betts and Ball (1998), who used soil moisture data down to two metres depth. In earlier studies Viterbo and Beljaars (1995) found a larger water imbalance of 100 mm, while Duan et al. (1996) found a smaller water imbalance of only 20 mm. They each used precipitation

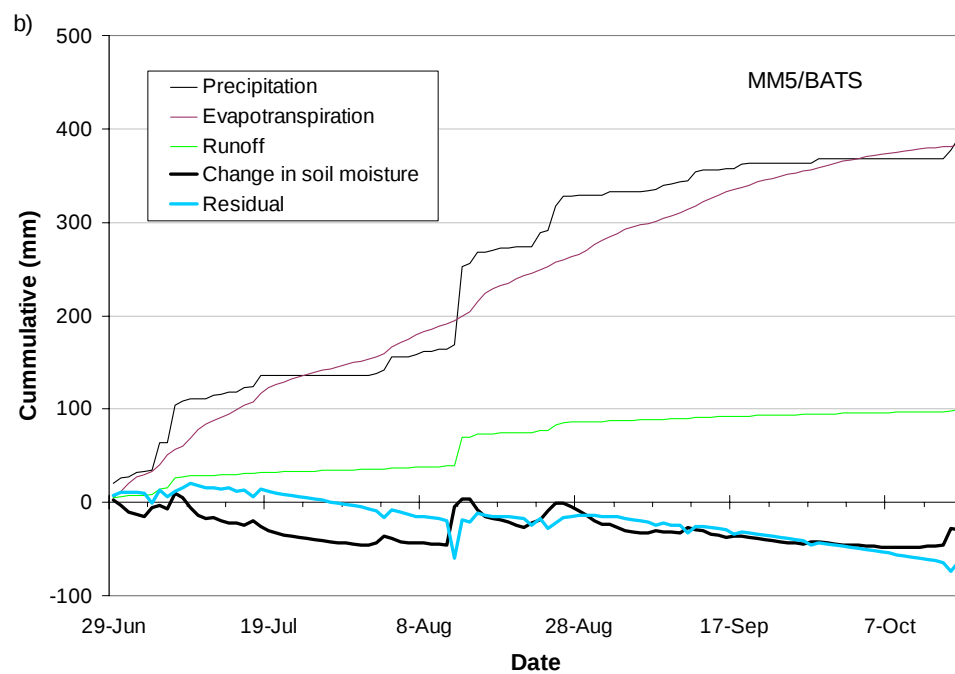
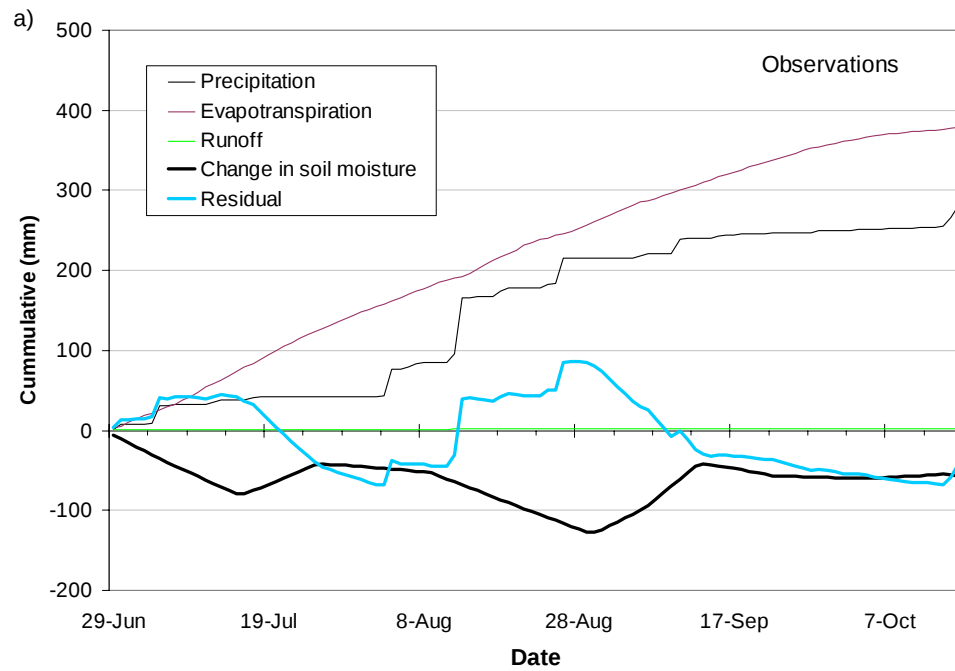


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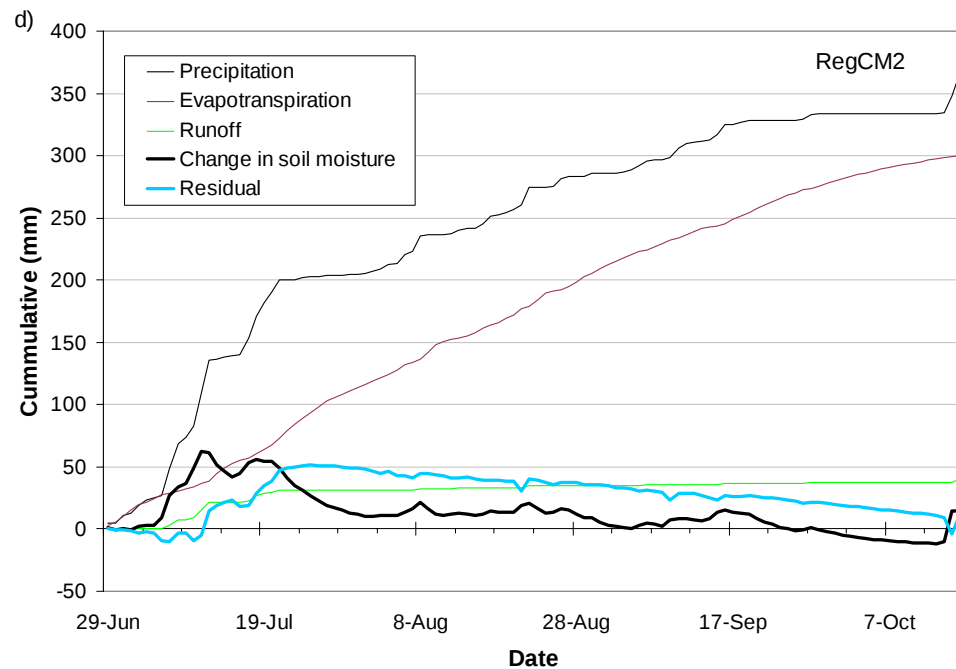
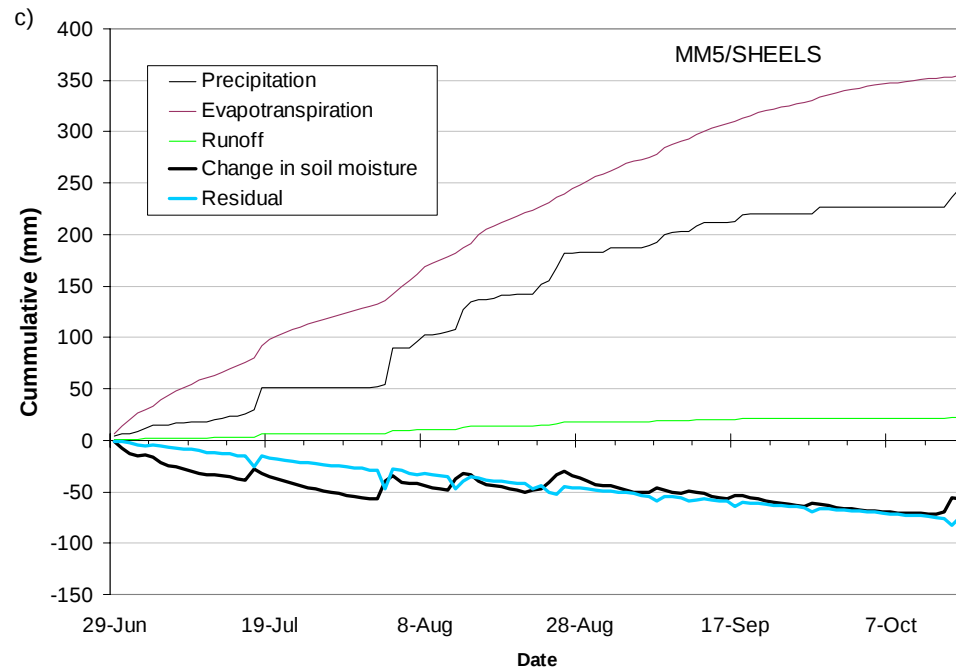


Figure 7.18 Cumulative water balance during the 1987 observation period:
a) Observations; **b)** MM5/BATS; **c)** MM5/SHEELS and **d)** RegCM2.

time series determined by averaging over different sets of rain gauges. This discussion is included to emphasise how difficult it is to model the long term hydrological balance at the surface satisfactorily (Viterbo and Beljaars 1995).

The observed water balance for 1988 is presented in Figure 7.19 a). This time the residual is considerably larger than in 1987, being around -150 mm. While some of this water imbalance may be accounted for by changes in soil moisture below 1 m depth (Betts and Ball 1998), much of it remains unexplained. Difficulties involved in the accurate measurement of basic variables such as precipitation, evaporation and soil moisture may be contributing factors.

The water balance simulated by MM5/BATS is presented in Figure 7.18b). This balance is very different to that observed with the precipitation being slightly larger than ET due to the over-estimation of precipitation by around 100 mm. This is counteracted by the over-estimation of runoff by almost the same amount, while the soil moisture is underestimated. This means that the residual is larger than that observed at around -60 mm.

The water balance simulated by MM5/SHEELS resembles the observed water balance much more than MM5/BATS did. However the combination of under-estimating precipitation and over-estimating runoff leads to a residual of -75 mm. This is the largest residual found in 1987. RegCM2, on the other hand, under-estimates ET while over-estimating precipitation, runoff and the change in soil moisture, leading to a residual of only 10 mm. Despite displaying a soil moisture regime very different to the other LAMs and the observations, RegCM2 does the best job at maintaining its own water balance.

The surface water balance found in 1988 by the observations and simulated by the LAMs (Figure 7.19) show more similarities than in 1987. Here ET is always larger than precipitation, and the residual tends to be larger than the change in soil moisture. The observations display a massive imbalance of -150 mm, closely followed by MM5/BATS with a residual of almost -110 mm. MM5/SHEELS and RegCM2 come

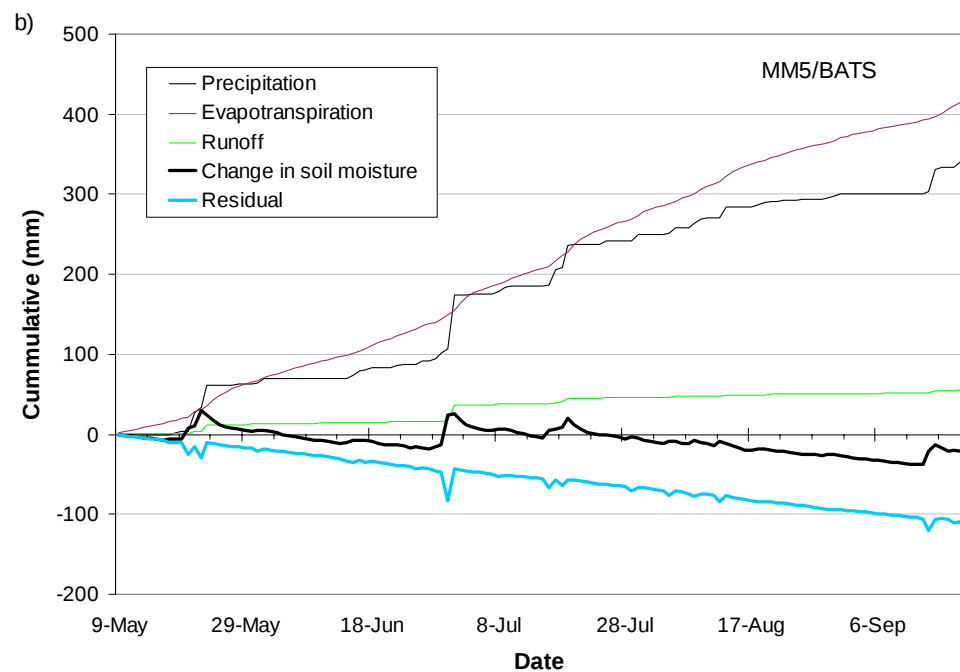
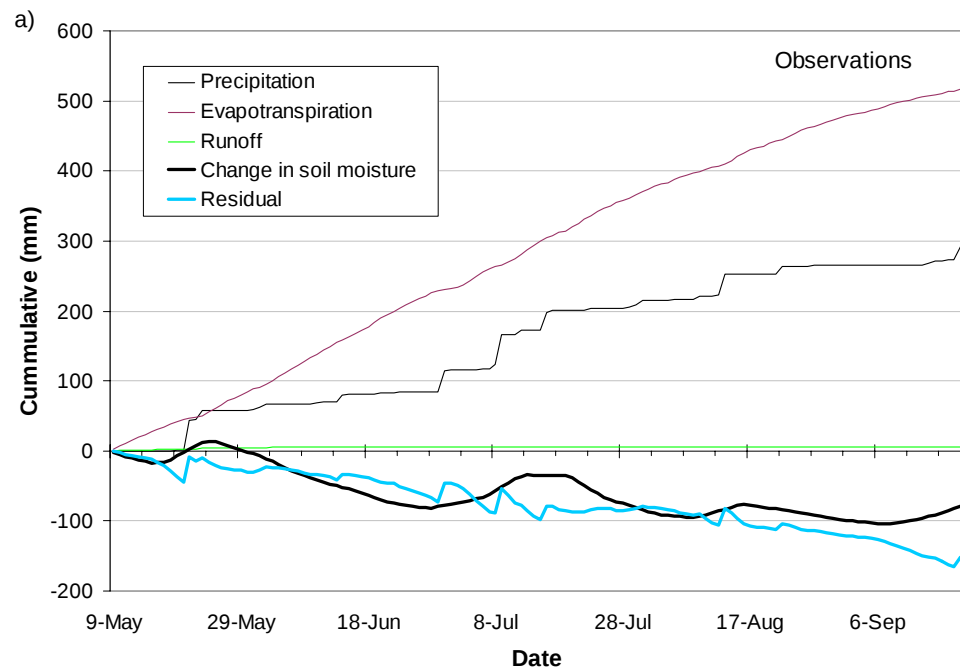


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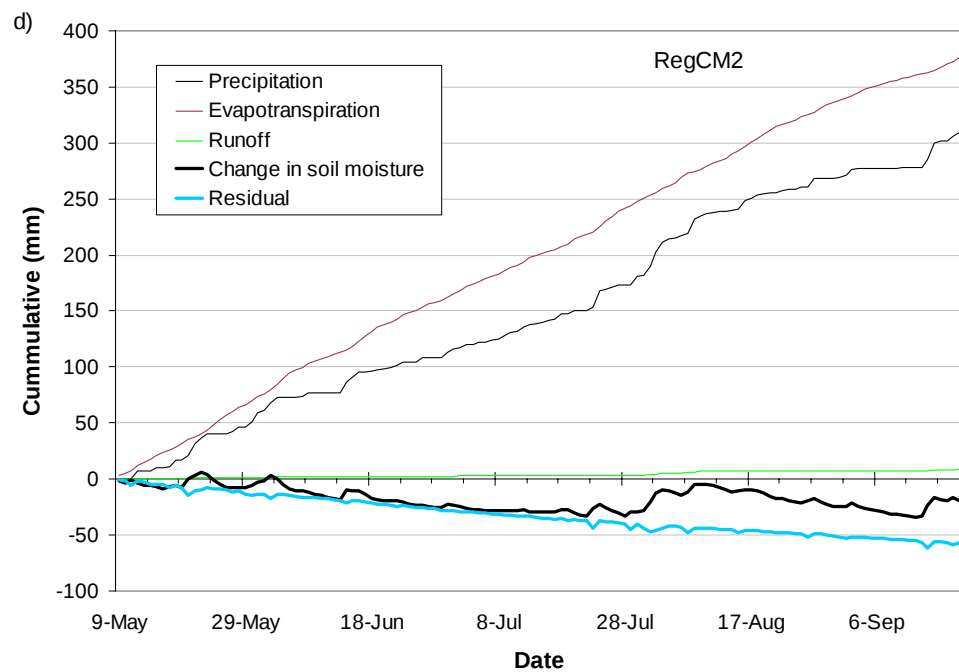
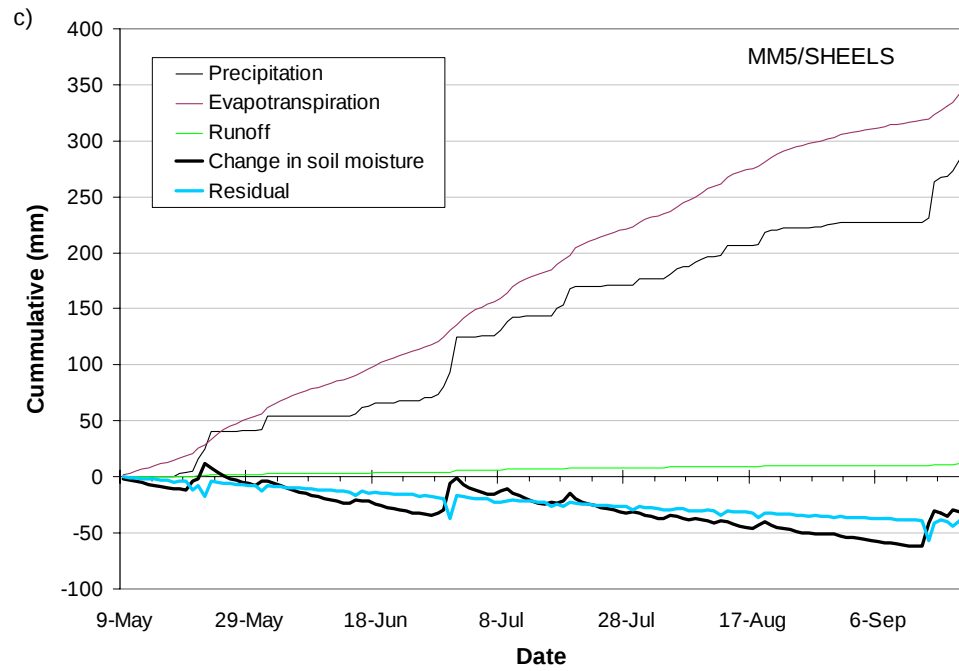


Figure 7.19 Cumulative water balance during the 1988 observation period:

a) Observations; **b)** MM5/BATS; **c)** MM5/SHEELS and **d)** RegCM2.

much closer to maintaining the water balance, with residuals of -40 and -55 mm respectively.

In general then, the water imbalance has increased during the drier year. This is not entirely surprising as soil moisture has been found to increase in spatial heterogeneity under drier conditions (Western et al. 1999). This in turn increases the spatial heterogeneity in runoff generation and ET production, making areally averaged estimates more difficult to obtain with any accuracy. MM5/SHEELS opposes this trend though, simulating the water balance with a smaller residual during the drier year.

7.4 Conclusions

Using the LAM simulations performed over the FIFE region, along with observations taken during several intensive field campaigns, intercomparisons and analysis of the temporal distribution of water budget terms were conducted. These simulations had a two year duration, starting in 1987, and included the drought year 1988. Analysis of the total precipitation reveals RegCM2 to be best able to reproduce the observed annual totals, with a consistent small overestimation. The MM5 based LAMs on the other hand, show a significant overestimation in the total precipitation in 1987, and then underestimate it in 1988. Concentrating on the convective precipitation, MM5/BATS simulates less than MM5/SHEELS in 1987 while this is reversed in 1988. This may be explained by the local production of ET, which follows the same pattern. RegCM2 on the other hand, simulates more convective precipitation than MM5/BATS in both years, despite simulating less surface ET. This is related to differences in the PBL parameterisation which make RegCM2 more likely to encounter unstable atmospheric conditions. Overall differences in the PBL parameterisation seem to have a larger effect on the production of convective precipitation than differences in the local production of ET, especially during dry years.

All three LAMs are able to reproduce the overall seasonal cycle of ET well though the totals and variances can differ substantially. Under normal to wet conditions the MM5 based LAMs simulate the mean daily ET well but substantially overestimate the variance, while RegCM2 significantly underestimates the mean but reproduces the variance well. Under dry conditions all three LAMs significantly underestimate the ET, while the MM5 based LAMs reproduce the variance well and RegCM2 underestimates this variance. The MM5 based LAMs simulated a much larger change in the ET regime from the wet to the dry year compared to RegCM2, which simulated a change quite similar to that observed.

The runoff simulated by the LAMs is in broad seasonal agreement with the observations, including the simulation of significantly less runoff in 1988 compared to 1987. That, however, is as far as the similarities go with runoff generally demonstrating the largest disagreement among the LAMs. Over the two year period the range of total runoff simulated by the three LAMs is as large as the total observed runoff over the same period. This is a significantly larger error than is found in any other variable investigated and suggests that improvements in the runoff formulation are urgently required.

The magnitude of the range of soil moisture values simulated by the LAMs compares favourably with the range found in the observations, though the actual soil moisture values themselves may be biased on the low side (especially for RegCM2). Having many thin layers near the surface allows MM5/SHEELS to generally maintain more soil moisture in the surface layer (top 10cm) during the wet year, despite simulating less precipitation. This trend is somewhat reversed in the dry year. In the reproduction of the observed change in the surface layer soil moisture through the observation periods, RegCM2 is the best performer, being quite close to the observed changes.

In terms of the overall water balance there was general agreement amongst the LAMs and observations in 1988, the dry year, while during the wet year, 1987, there is considerable disagreement. This trend is the opposite of what was found in the overall energy balance. This may be a reflection of the fact that when less water is available to the surface the surface parameterisations are more restricted in how that water is

allocated to the various fluxes, thus encouraging agreement between them. Whereas when water is plentiful at the surface, the surface parameterisation may allocate this extra water in a number of ways, thus increasing disagreement among the models. This implies that comparing LAMs under extreme wet conditions would emphasise the differences among them in terms of the water balance. In combination with the results in Chapter 6, it appears that under dry conditions energy balance related parameterisations are responsible for the majority of the differences between the LAMs, while under wet conditions water balance related parameterisations gain greater importance.

None of the three LAMs performs better in the simulation of all water fluxes though MM5/SHEELS performs marginally better in terms of the overall water balance. This suggests that there may be an advantage in using many thin layers to model the movement of soil moisture.

Part III: Climate - Surface Hydrology Interactions



This last part brings the first two parts together. It contains a study of the improvement in the modelling of surface hydrology by regional climate models, which is gained through the inclusion of a rainfall-runoff model (CMD-IHACRES). The use of CMD-IHACRES in a study of the hydrological impacts of climate change is subsequently demonstrated in a case study centered on a region outside Perth, Western Australia.

Chapter 8

Improving Streamflow Prediction in Regional Climate Models

8.1 Introduction

This chapter focuses on the simulation of runoff in climate models. It presents an investigation into the current performance, in terms of runoff, of the BATS and SHEELS land surface parameterisations as implemented inside the MM5 and RegCM2 LAMs. The effects of including CMD-IHACRES as the runoff parameterisation in these LAMs is explored in an effort to improve the runoff simulations of the LAMs while minimising any additional computational burden.

While only a small fraction of the world's water is present in rivers at any given time, they remain a major component of the earth's hydrological cycle. In particular, they provide a critical pathway for returning water from continents to the oceans. The role of rivers in the long-term global water budget has been discussed by others (Baumgartner and Reichel 1975, Milliman and Meade 1983, Russell and Miller 1990). This freshwater flux from the continents to the oceans influences both the thermohaline circulation in the ocean and the formation of sea-ice via its influence on salinity. In a fully coupled climate model with a closed hydrological cycle, this river sourced freshwater feedback to the oceans can possess a climacteric nature. For example, (Mysak et al. 1990) suggest that salinity anomalies in the North Atlantic Ocean, which could affect the thermohaline circulation, are related to ice transport through the Fram Strait, which in turn may be related to anomalous river flow into the Arctic Ocean.

An ideal land-surface parameterisation in a climate model should be capable of producing realistic time series of water and energy outputs based upon climatic inputs and spatially varying physical descriptors of the land surface (including terrain, soil and vegetation characteristics). Unfortunately, data available to validate climate model hydrologic descriptions of precipitation, ET and soil moisture storage are lacking due to several reasons. Most importantly, they are generally point measurements compared to the areal average values simulated by the climate model. Further discussion can be found in Chapter 2. River runoff on the other hand, is an important spatial integrator of the hydrologic cycle, it is measured more accurately than other components of the hydrologic cycle and river runoff data are readily available. The importance of river

runoff data in validating climate models has been discussed by others, for example Liston et al. (1994), Miller et al. (1994) and Arnell (1995).

Most current land-surface schemes, regardless of whether they are relatively simple (eg. bucket model) or complex (eg. a SVAT scheme), contain highly simplified treatments of runoff. This is particularly true in comparison to the treatment of other hydrological components such as ET. This dichotomy in the treatment of various parts of the hydrological cycle is a cause for concern. Viterbo and Illari (1994) highlight the importance of runoff and soil moisture formulations. While Koster and Milly (1997) note that “even a “perfect” description of canopy structure and stomatal behaviour, toward which many land-surface models strive, does not ensure realistic evaporation rates if the runoff formulation remains relatively crude or incompatible.”

The growing concern over global warming has led to regional climate models being used to conduct climate change impact studies. Further discussion on this can be found in section 9.1. These studies are often aimed at addressing questions relating to fresh water supply for urban areas, changes to flooding regimes or the impact on agricultural and ecological systems, all of which require accurate simulation of runoff. For this reason, and the reasons cited above, improving the runoff simulation in climate models is a pressing concern.

8.1.1 Climate model simulated river runoff studies

Several studies have compared climate model-simulated runoff to observed river runoff in order to validate the model or to investigate the impact of global warming. Russell and Miller (1990) compared annual river runoff generated by the NASA/Goddard Institute for Space Studies (GISS) GCM (Hansen et al. 1983) with observed runoff reported in Milliman and Meade (1983) for 33 of the world’s major rivers. They found that under present climate conditions annual river runoff totals were within 20% of the observed. Miller and Russell (1992) investigated the effect of global warming and found for a doubled CO₂ climate, runoff increased for 25 of the 33 rivers. Kuhl and

Miller (1992) extended this work to examine the seasonal variation of the model-generated river runoff. They found considerable disagreement between model-generated and observed seasonal river runoff, especially in dry river basins. Total river runoff was calculated simply as a sum of runoff generated in each grid cell within the river basin without any attempt to route the flow between grid boxes.

Further studies on large river basins have introduced river runoff routing models, such as that found in Naden (1993), to realistically route the flow between grid cells. Miller et al. (1994) demonstrate the use of a simple routing model to examine the NASA/GISS GCM-generated river runoff for 39 rivers around the world. Liston et al. (1994) use a more complicated routing model applied in the Mississippi basin. Both of these studies found a significant improvement in the seasonal timing of GCM-simulated river runoff resulting from the use of a routing model.

More recently, Lohmann et al. (1998) used a routing model when conducting a spatial and temporal analysis of water fluxes produced by 16 different land-surface parameterisations as part of PILPS phase 2, while Hagemann and Dumenil (1996) introduced their own routing model. In both cases, as well as routing the river runoff between grid cells as was done in the studies above, the runoff produced within each grid cell was ‘routed’ to the outlet of this cell before being further routed through other grid cells to the river basin outlet. This within-cell routing is considered important since most land-surface parameterisations assume that any runoff generated instantly reaches the grid cell boundary. Clearly this is too much of a simplification. Other studies have attempted to combine hydrological and climate models, thus removing the need for this within-cell routing.

8.1.2 Past attempts to combine hydrological and climate models

Possibly the first hydrologic model to be incorporated into a climate model is the Nanjing model (Zhao 1977), variations of which have also been termed the Arno model by Franchini and Pacciani (1991) and the variable infiltration capacity model, or VIC,

(Wood et al. 1992). In this model the fixed capacity of a bucket scheme is replaced by a sub-grid variable field capacity which is predefined by a “storage capacity distribution curve” and the average soil moisture storage capacity is computed through integration over the grid area. Wood et al. (1992) calibrated the parameters for their VIC model over the French Broad River basin from daily stream discharge and pan evaporation, and 6-hourly precipitation and temperature data. Then using outputs from the Geophysical Fluid Dynamics Laboratory (GFDL) GCM to drive the land-surface scheme they found considerable improvement over the bucket model. Dumenil and Todini (1992) incorporated the ARNO model into the Hamburg GCM. They related the rainfall-runoff model parameters to the topography such that the model could be applied to all grids of the GCM. They found this new scheme improved the GCM simulation, especially in terms of runoff.

Variations of this model have also been used for GCM land-surface parameterisations by Stamm et al. (1994) in the GFDL model, by Laval et al. (1994) in the LMD model, by Rowntree and Lean (1994) in the UK Meteorological Office model, and by Habets et al. (1999a, b) in the ISBA land-surface scheme. In all cases the model performance is highly subject to the validity of the storage capacity distribution curve chosen. Moreover they focus on the subgrid variability which is beyond the capacity of validation due to insufficient data, and thus may be difficult to use in practice.

8.2 Experiment design

LAM model simulation results presented in this chapter were obtained from the same model runs discussed in section 6.2.1. Briefly the three LAMs were run for two years on a domain covering an area of $1200 \times 1260 \text{ km}^2$, centered over the FIFE region. The LAMs were run with 17 levels in the vertical, horizontal grid point spacing of 20km and a time step of 1 minute. The rainfall-runoff model developed in Chapter 3, CMD-IHACRES, was applied to Kings Creek catchment.

The model results are compared to observations taken during FIFE (sections 4.2.1 and 6.2.2) as well as data collected as part of the Konza LTER site, which includes Kings Creek catchment in the north west corner of the FIFE domain. Runoff simulated by the LAMs was compared to streamflow from the Kings Creek catchment after it was scaled according to area. Even though a LAM grid point is representative of an area somewhat larger than the FIFE site itself, there are several reasons why the comparison is meaningful. Summer conditions over the FIFE grassland site were relatively homogenous, so that simple averaging of the data gave a representative mean. The Konza prairie itself covers over 50,000 km², and the diurnal cycle over land integrates over considerable advection distances (up to 100-200 km²) (Betts et al. 1998).

The results presented below begin with the calibration and validation of the hydrological model, CMD-IHACRES. This is followed by the comparison of various characteristics of the ‘stand alone’ runoff simulations produced by the models and the observations. As well as daily and cumulative runoff comparisons, differences in the shapes of the flow duration curves and double mass plots are investigated. Stand alone simulations of ET are also compared.

The hydrological model is then run ‘offline’, driven by each of the LAMs. First, the hydrological model is driven by the precipitation and temperature simulated by the LAMs. Here the hydrological model calculates ET using the modified temperature approach embodied in equation 4.9. Secondly, the hydrological model is driven by the precipitation and ET simulated by the LAMs. Thus the simplistic modified temperature approach used in CMD-IHACRES is effectively replaced by the much more complicated SVAT approaches found in BATS and SHEELS.

8.3 Results

8.3.1 CMD-IHACRES calibration

The hydrological model, CMD-IHACRES, was calibrated for the Kings Creek catchment which is found inside the FIFE site. The model was calibrated on daily data, over the two years 1987 and 1988 when the most reliable data was collected. As stated earlier, this data was collected simultaneously at many locations within the FIFE site and combined into an areal average by Betts and Ball (1998). The calibration results are presented in Figure 8.1, with the relevant statistics shown in Table 8.1.

It can be seen in Figure 8.1 that CMD-IHACRES performs well in terms of runoff during the calibration period, and this is reflected in Table 8.1 where the calibration displays a high efficiency and a low bias and %ARPE. While the modelled runoff largely misses the final major event for 1987 it does capture the small runoff in 1988 particularly well. This is indicative of successful soil moisture accounting during the months when the stream is dry.

It can be seen in Table 8.1 that parameter c_2 is zero. It follows from equation 4.9 that CMD-IHACRES simulates the ET to be simply proportional to the temperature. The runoff is split almost evenly between the quick and slow stores ($V_s=0.509$). It takes 2.18 days for runoff to travel through the quick store (τ_q), while it takes almost 46 days for runoff to travel through the slow store (τ_s).

The model runoff performance was then validated on the two preceding years, 1985 and 1986. This data was put together as a historical data series for the FIFE site. It was collected by Kansas State University using a single precipitation gauge, almost 20km from the center of the FIFE site. This precipitation series is therefore less representative of the precipitation within the FIFE site than the areally averaged data used in 1987 and 1988. This is reflected in the simulation results shown in Figure 8.2 and Table 8.1 where it can be seen that the magnitude of the bias has increased marginally while the efficiency has decreased from 0.82 to 0.63.

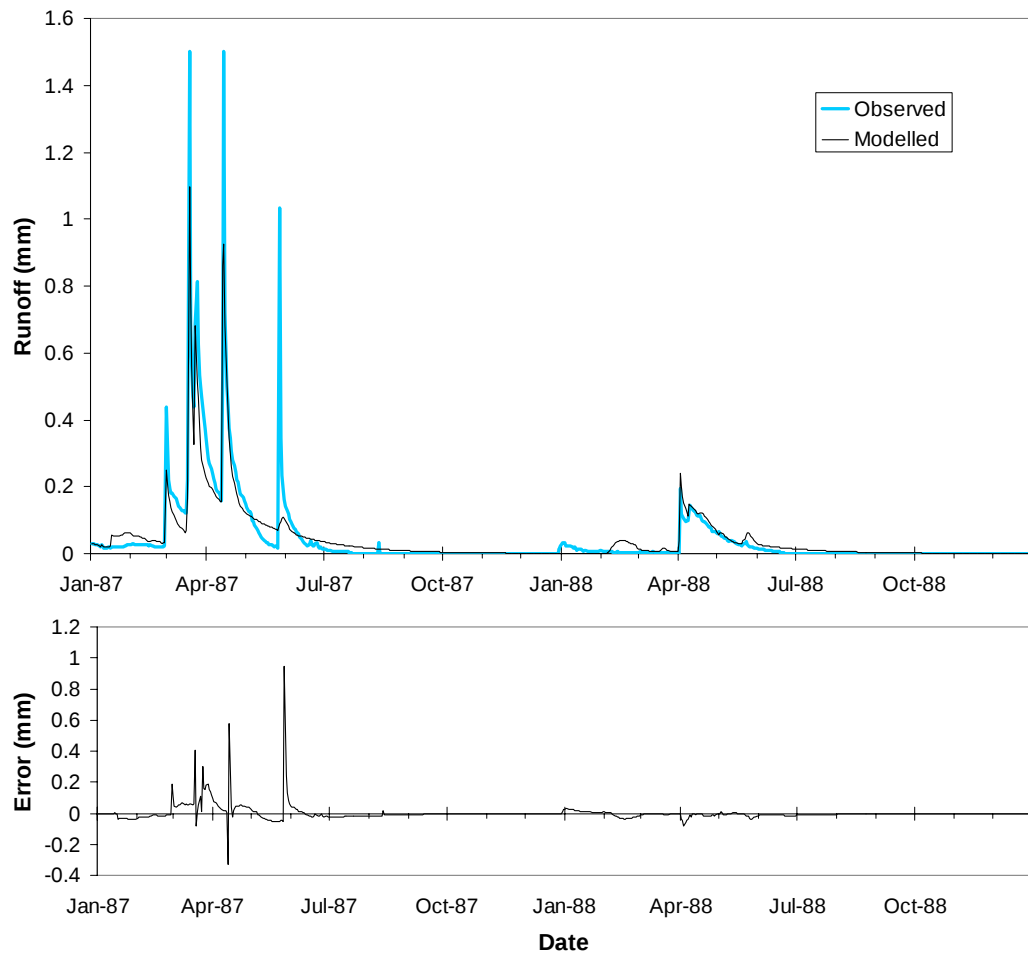


Figure 8.1 CMD-IHACRES calibration results

Table 8.1 CMD-IHACRES calibration and validation results

Model run and period	Model Parameters						Model Performance			
	c_1	c_2	c_3	c_4	τ_q (days)	τ_s (days)	V_s	%ARPE	B (mm/d)	E
Calibration 1/1/87 – 30/12/88	0.14	0	1	35	2.18	45.97	0.509	0.1	0.0	0.82
Validation 1/1/85 – 30/12/86	0.14	0	1	35	2.18	45.97	0.509		-0.05	0.63

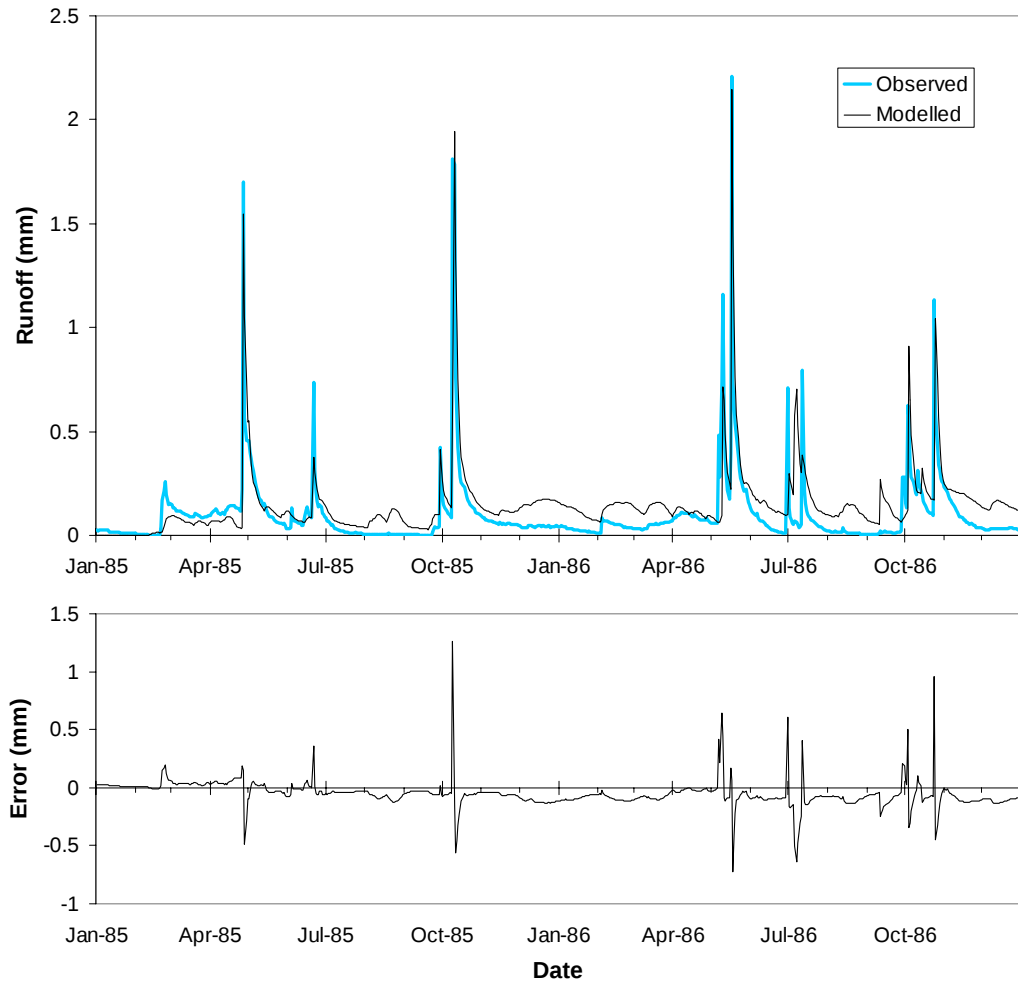


Figure 8.2 CMD-IHACRES validation results

Figure 8.2 shows the calibrated model capturing, in an independent period, the major runoff peaks well except for the peaks found in early July 1986. The observed runoff has two runoff peaks in early July, 12 days apart. The simulation on the other hand, has a single major runoff peak between these two. This is indicative of the mismatch between the runoff observed in Kings Creek and hence the precipitation that falls over the catchment, and the precipitation falling over a single gauge some 20km away.

Similarly, there are many increases in the simulated runoff while the observed runoff displays no response. This can be seen many times during observed low flow periods. This may be related to small scale convective precipitation events which differ in timing and amount between the Kings Creek catchment and the gauge. Given these problems, the validation results suggest that the hydrological model performs adequately in simulating surface runoff for Kings Creek, with this performance improving if the precipitation time series is more representative of the precipitation in the area of interest.

8.3.2 Stand-alone model results

In this section the stand alone model simulations of runoff are compared. Figure 8.3 shows the daily runoff simulated by each of the LAMs, CMD-IHACRES and the observations. The simulation which best reproduces the observations is clearly CMD-IHACRES, it is the only model which produces a reasonable hydrograph. That is, the LAMs are unable to reproduce the streamflow recession after a major runoff event. Instead, runoff simulated by the LAMs consists of a series of (mostly) discrete peaks reminiscent of the overflow runoff response of a simple bucket model. The LAMs also significantly overestimate the magnitude of the largest runoff events, especially MM5/BATS.

Characteristics of the simulated runoff are compared further in Figures 8.4 and 8.5, which display the simulated flow duration curves (fdc) and double mass plots respectively. The period in these figures is composed of two years of daily data. While this may not be a long enough time series for the figures to be truly representative of Kings Creek, it does allow intercomparison of the runoff characteristics simulated in these two years. It can be seen in Figure 8.4 that the fdc simulated by CMD-IHACRES is closest to reproducing the observed fdc. The fdcs simulated by the LAMs differ significantly from that observed, though they are all quite similar to each other. They all significantly underestimate the persistence of low flows compared to the observations.

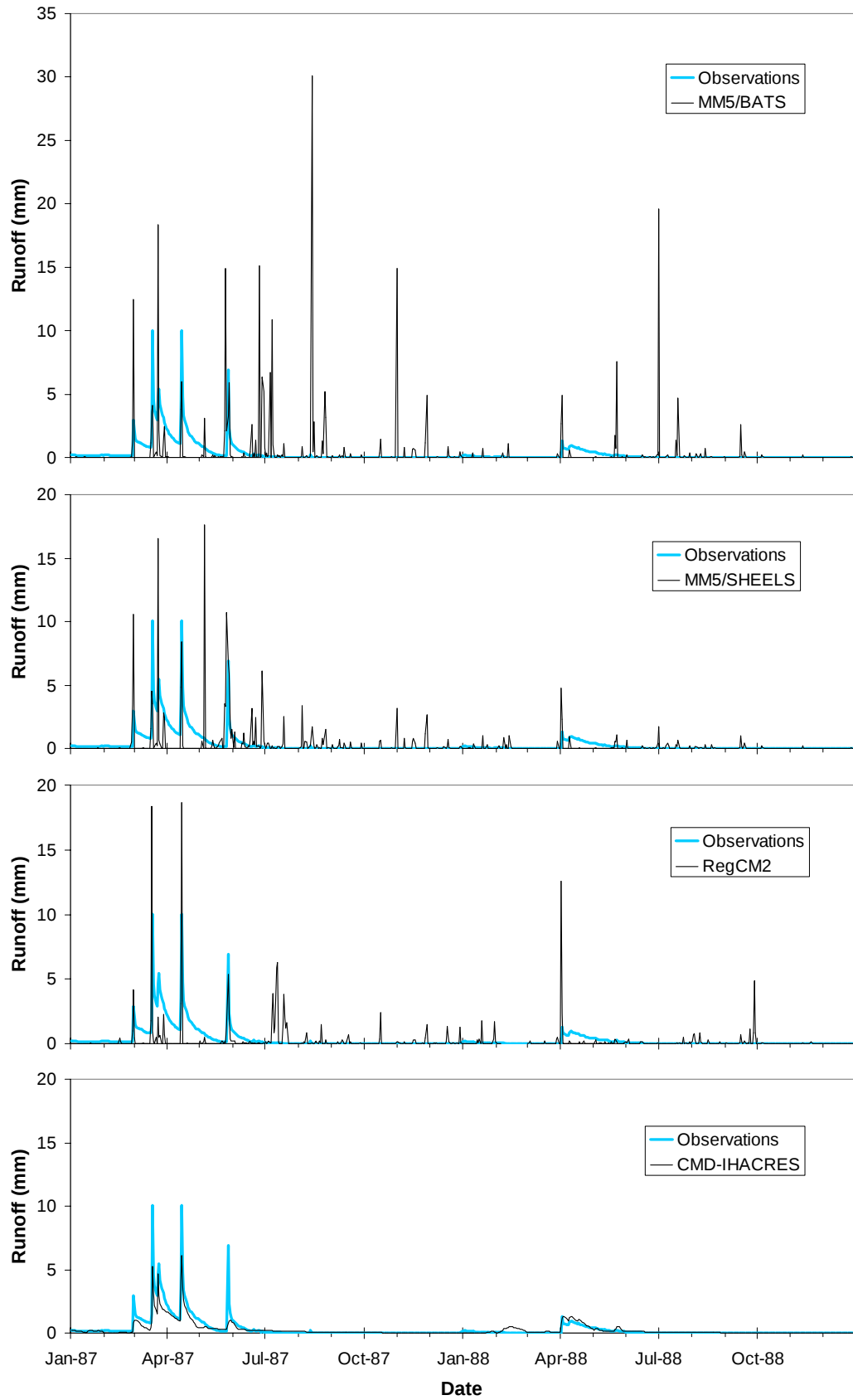


Figure 8.3 Daily runoff simulated by the models

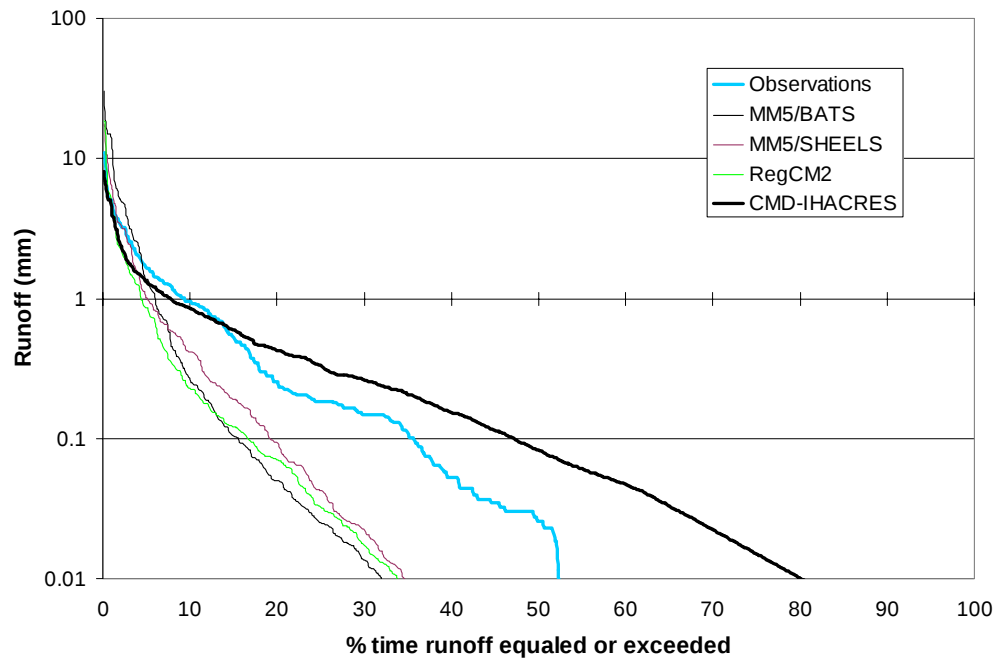


Figure 8.4 Flow duration curves simulated by the models

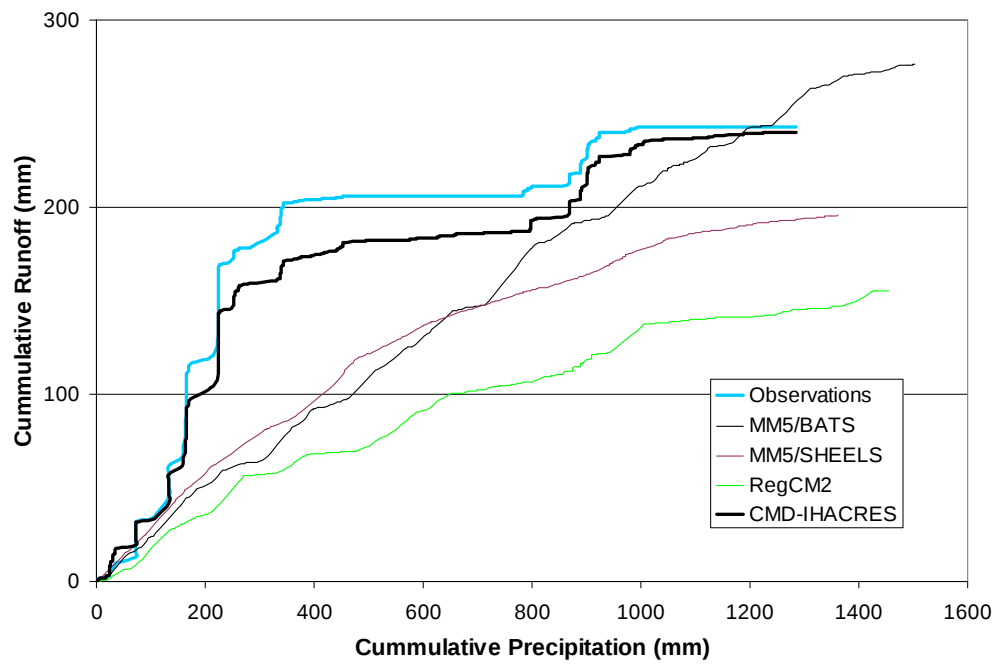


Figure 8.5 Double mass plots simulated by the models.

Figure 8.5 presents the double mass plots simulated by the models. These plots present the relationship between runoff and precipitation in terms of accumulated daily values. Clearly the observed runoff displays a significantly non-linear relationship between runoff and precipitation. CMD-IHACRES also simulates a similar non-linear relationship to that which is observed. While the LAMs all simulate different double mass plots they are all much closer to linear than the observations. This near linearity of the relationship between runoff and precipitation simulated by the LAMs demonstrates clearly the potential for the inclusion of a CMD-IHACRES style runoff parameterisation.

CMD-IHACRES also simulates ET from the catchment. This is compared to the ET simulated by the LAMs in Figures 8.6 and 8.7. Figure 8.6 presents the daily ET simulated by all the models. CMD-IHACRES uses by far the simplest treatment for ET,

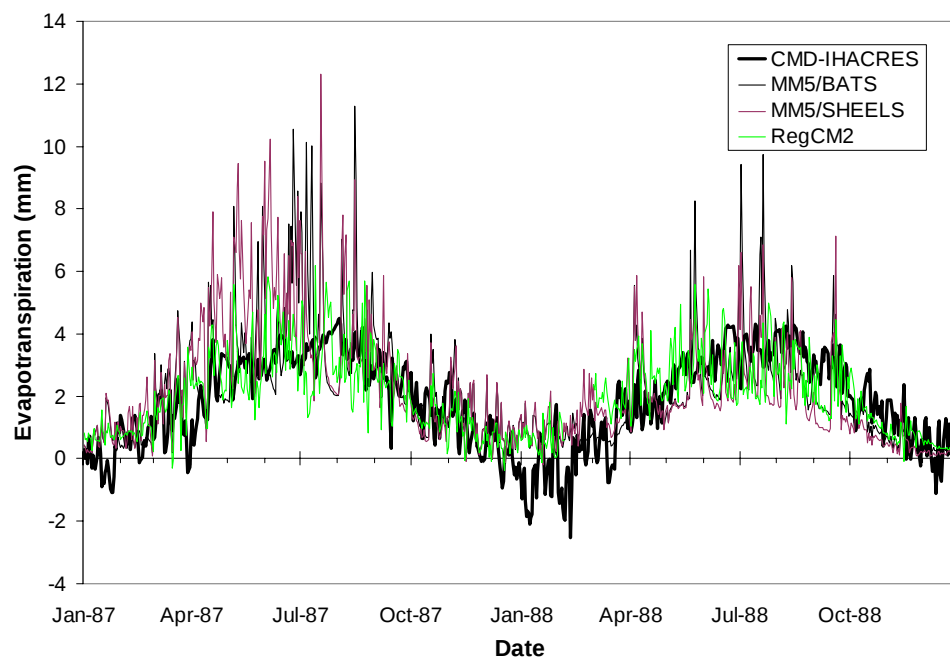


Figure 8.6 Daily ET simulated by the models.

as it is simply represented as being proportional to temperature. The ET simulated by all the models follows a similar overall trend. The LAM-simulated ET was compared in more detail in section 7.3.2. The LAMs simulated greater variance in the daily ET in summer compared to winter while CMD-IHACRES simulated ET with a similar variance throughout the two years. This results in the CMD-IHACRES simulated ET displaying greater variance than the LAMs during winter but less variance during summer. CMD-IHACRES also simulates greater dewfall during winter (ET is negative).

Figure 8.7 compares the daily ET simulated by the models with measurements taken during the observation period in both years. The means and standard deviations for these time series (within the common observation period) can be found in Table 8.2. Clearly CMD-IHACRES simulates significantly less variance than is observed during both observation periods. The mean daily ET during 1987 is significantly underestimated by CMD-IHACRES with most of this underestimation occurring during the first half of the observation period (see Figure 8.7a). CMD-IHACRES comes closest to reproducing the mean daily ET during the 1988 observation period. Figure 8.7b however, reveals that CMD-IHACRES actually underestimates the daily ET early in the observation period, while overestimating the ET later in the period.

Table 8.2 Means and standard deviations of ET over the common observation period; 27 May to 19 September.

	1987		1988	
	Mean (mm)	St. Dev. (mm)	Mean (mm)	St. Dev. (mm)
Observations	4.230	1.194	3.887	1.412
MM5/BATS	4.250	2.107	3.115	1.490
MM5/SHEELS	4.396	2.031	2.556	1.340
RegCM2	3.228	1.197	2.757	0.983
CMD-IHACRES	3.259	0.627	3.424	0.521

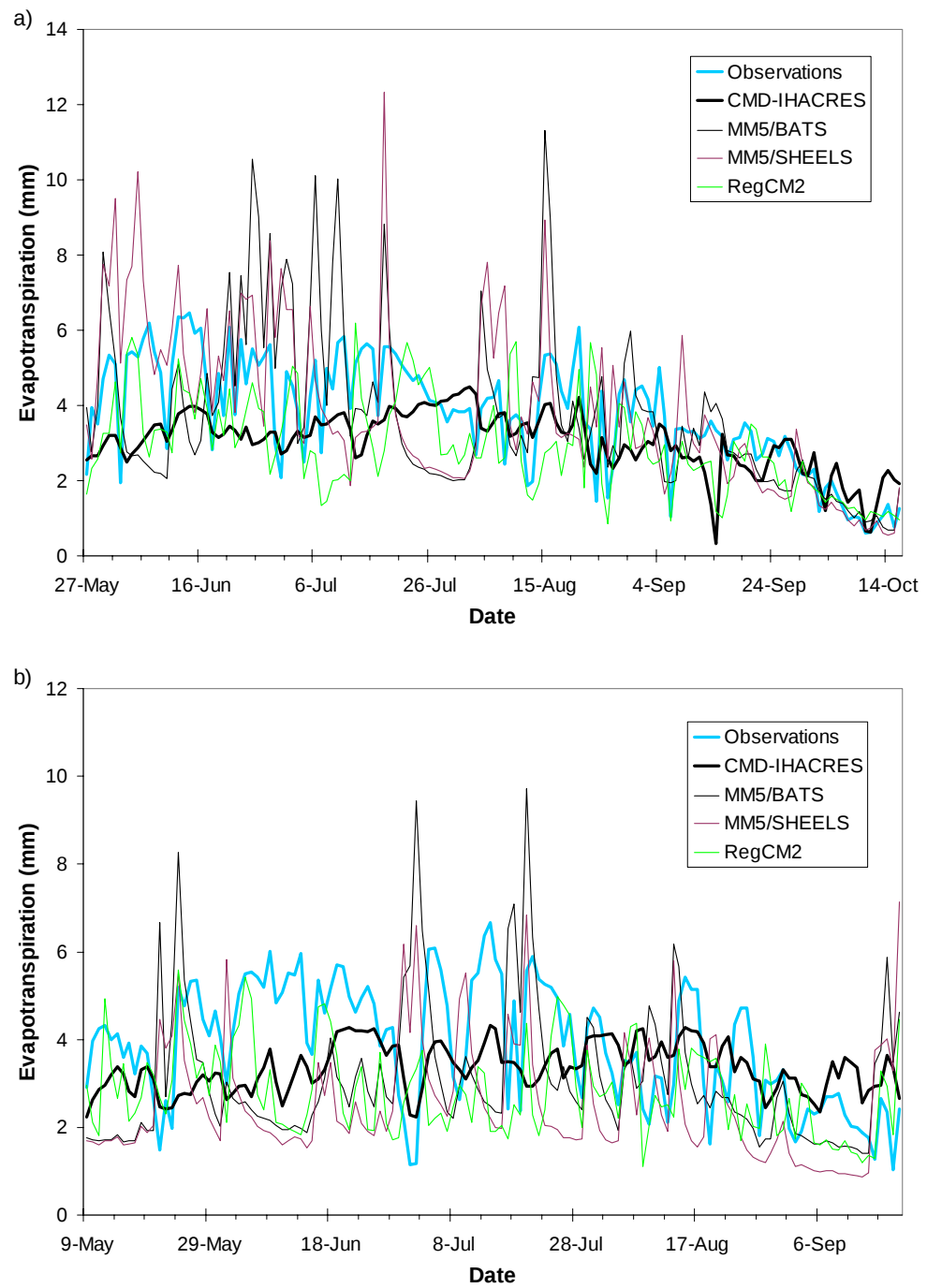


Figure 8.7 Daily ET simulated by the models during the observation period in **a)** 1987 and **b)** 1988

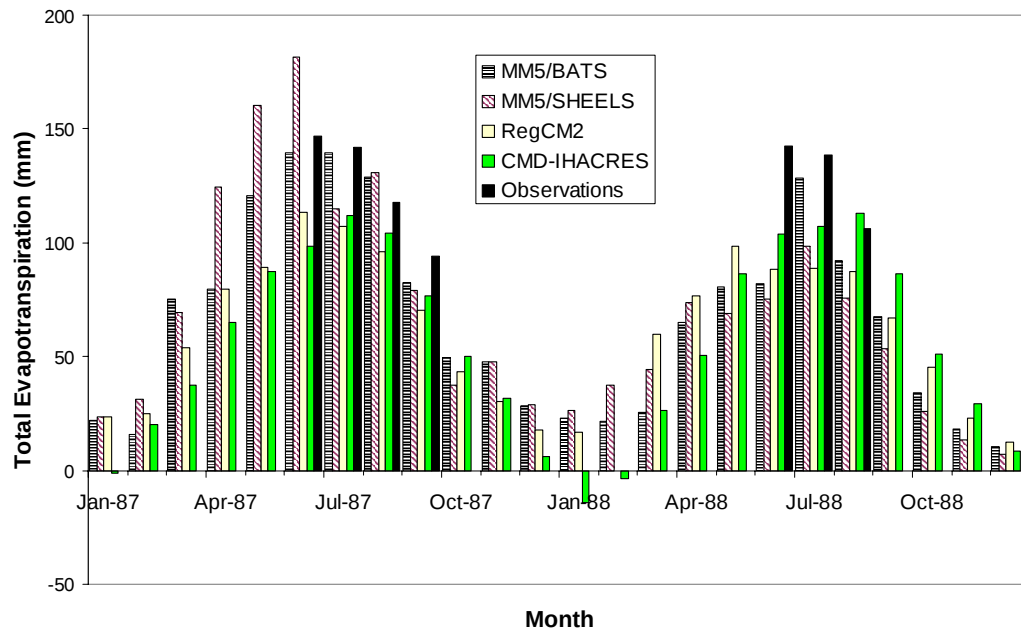


Figure 8.8 Total monthly ET simulated by the models.

The monthly cycle of ET simulated by the LAMs is presented in Figure 8.8, which is similar to Figure 7.11. Here, particular attention is paid to the ET simulated by CMD-IHACRES. The use of the modified temperature approach has moved the ET annual cycle back a month or two compared to the LAMs and observations. That is, the peak monthly ET simulated by CMD-IHACRES in 1987 occurs in July while the LAMs and observations suggest this peak should occur in June. In 1988 CMD-IHACRES simulates the peak ET to occur in August while the LAMs and observations again have this peak in June. Compared to the LAMs, CMD-IHACRES simulates the least ET over the two year period, with significant underestimation occurring early in the year followed by some overestimation late in the year.

8.3.3 Offline simulation results

CMD-IHACRES was then used to simulate runoff from the LAMs by being run offline. That is, CMD-IHACRES was fed the precipitation and temperature time series simulated by the LAMs (run with the BATS or SHEELS land surface parameterisation) as input and simulated the ET and runoff using the calibration parameter set shown in section 8.3.1. These simulated ET and runoff time series were not fed back into the LAMs.

The daily runoff estimated by the offline simulation of CMD-IHACRES with each of the LAMs can be found in Figure 8.9. This is a clear improvement on the streamflow simulated by the LAMs alone. In particular, the scattering of runoff peaks during the end of 1987 and through 1988 have been removed. The MM5/BATS runoff does well compared to the observations except for a few large runoff peaks which occur in July 1987. These runoff events are related to precipitation events simulated by the LAM that were not observed. RegCM2 also simulates extra runoff events in July 1987, again related to precipitation events which were not observed. MM5/SHEELS simulates only a small runoff event in July, unlike the other LAMs, however it does simulate a major runoff event which is not observed during May 1987.

The flow duration curves simulated by the LAMs in stand alone mode and in offline mode are shown in Figure 8.10 to 8.12. Significant changes are found for each of the LAMs. The fdcc for the MM5/BATS offline simulation is marginally closer to reproducing the observed fdcc than the stand alone simulation. The difference between the offline simulation and the observed is a combination of the extra runoff events simulated in July 1987 and a general increased occurrence of low flows. The fdcc for the MM5/SHEELS offline simulation is significantly closer to the observed fdcc due largely to the increased occurrence of low flows. RegCM2 is the only LAM that does not show the offline simulated fdcc to be closer to the observed than the stand alone simulation. Here the combination of an increased occurrence of low flows and the extra runoff simulated during July 1987 has led to a general overestimation of the flows.

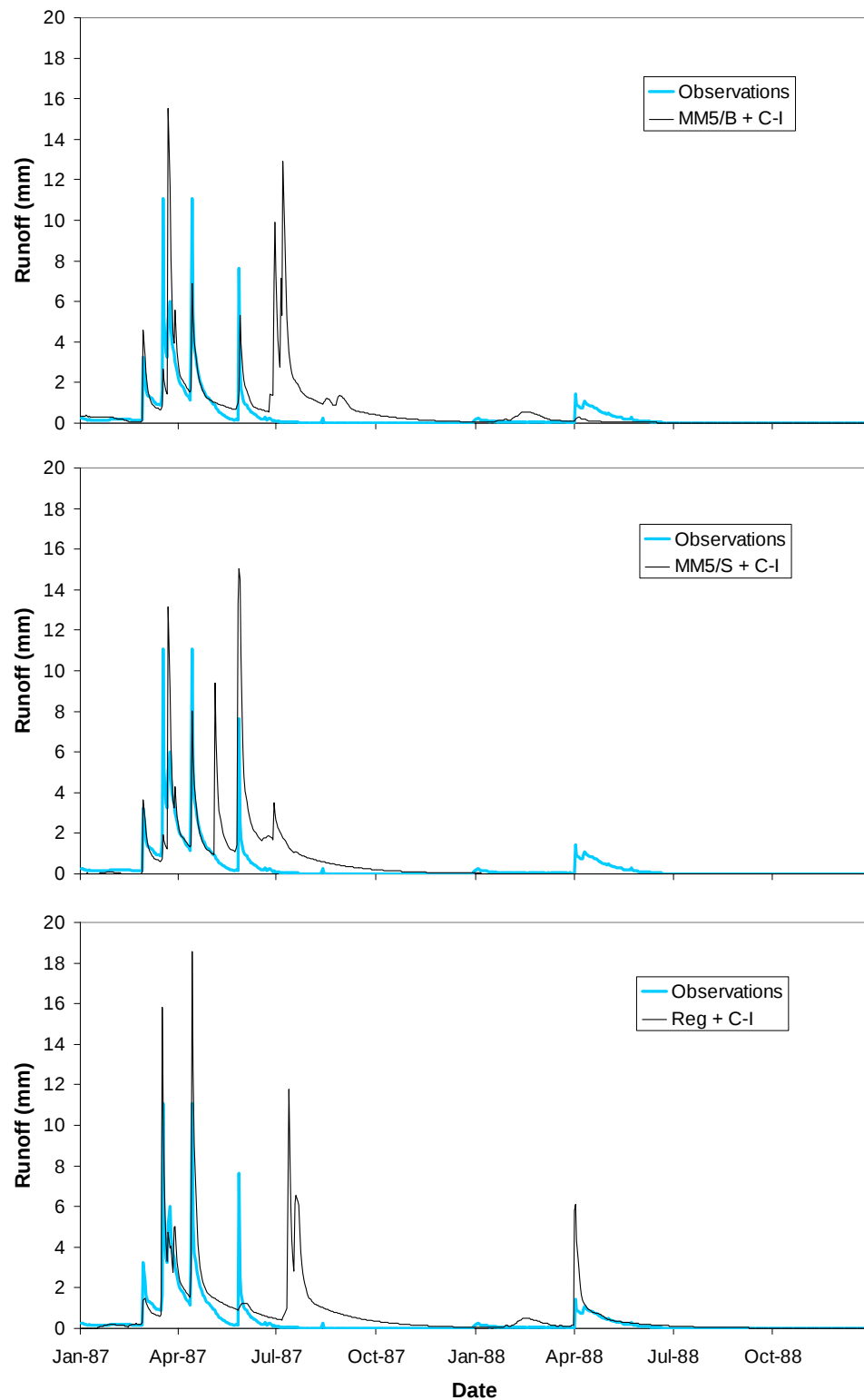


Figure 8.9 Runoff simulated by CMD-IHACRES when run offline with output from each of the three LAMs. MM5/B, MM5/S, Reg and C-I refer to MM5/BATS, MM5/SHEELS, RegCM2 and CMD-IHACRES respectively.

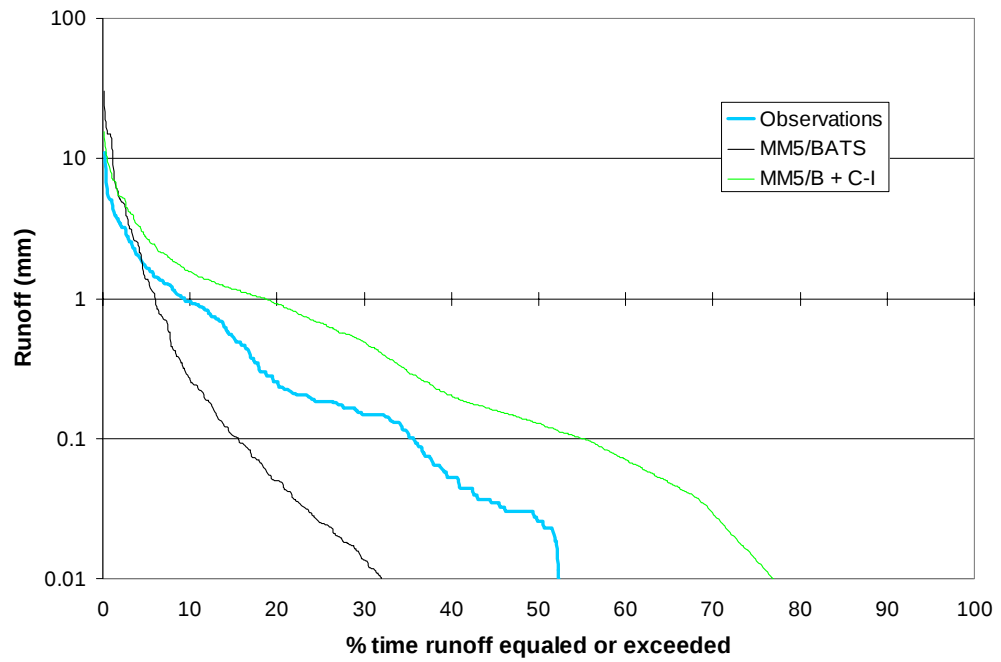


Figure 8.10 Flow duration curves simulated by MM5/BATS alone and by CMD-IHACRES run offline with MM5/BATS.

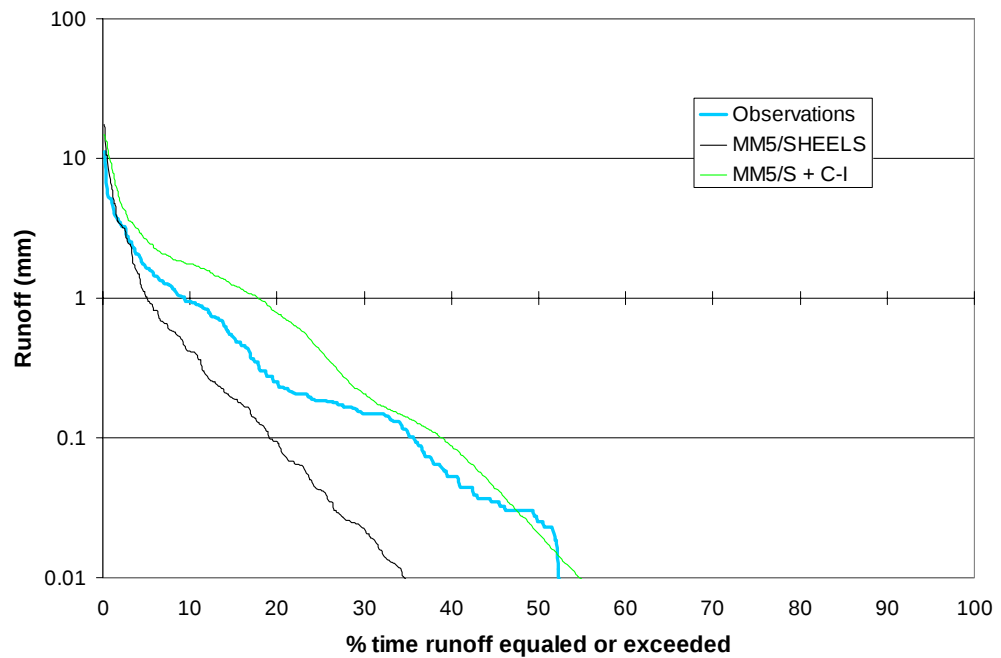


Figure 8.11 Flow duration curves simulated by MM5/SHEELS alone and by CMD-IHACRES run offline with MM5/SHEELS.

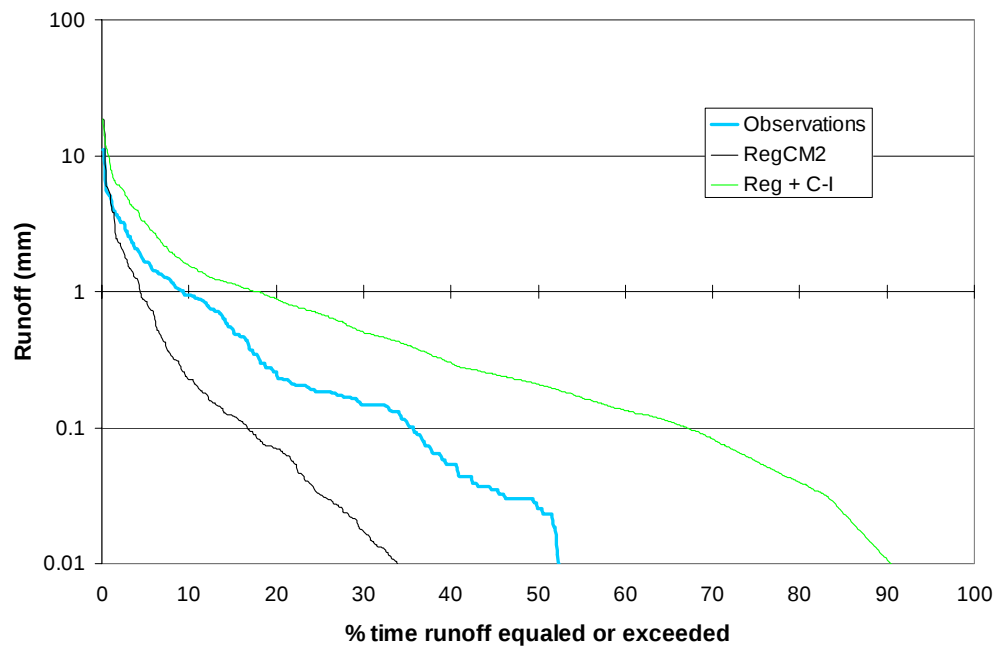


Figure 8.12 Flow duration curves simulated by RegCM2 alone and by CMD-IHACRES run offline with RegCM2.

The average daily runoff for each month simulated by the models in stand alone mode can be seen in Figure 8.13. The observations and CMD-IHACRES show the peak runoff to occur in March/April 1987, with a lesser peak in April 1988. The LAMs on the other hand, do not display such a clear runoff peak in either year. Instead the runoff is spread more evenly through the year. In particular the observed runoff peak in March/April 1987 is not reproduced well by the LAMs. Figure 8.14 is similar to Figure 8.13 but here the simulated runoff comes from the offline simulation of CMD-IHACRES driven by the LAMs. Comparing Figure 8.14 with Figure 8.13 reveals significant improvement in capturing the March/April 1987 runoff peak by all the LAMs. The offline simulated runoff subsequently departs from the observed due to the extra runoff events simulated by the models. These occur in May for MM5/SHEELS and in July for both MM5/BATS and RegCM2.

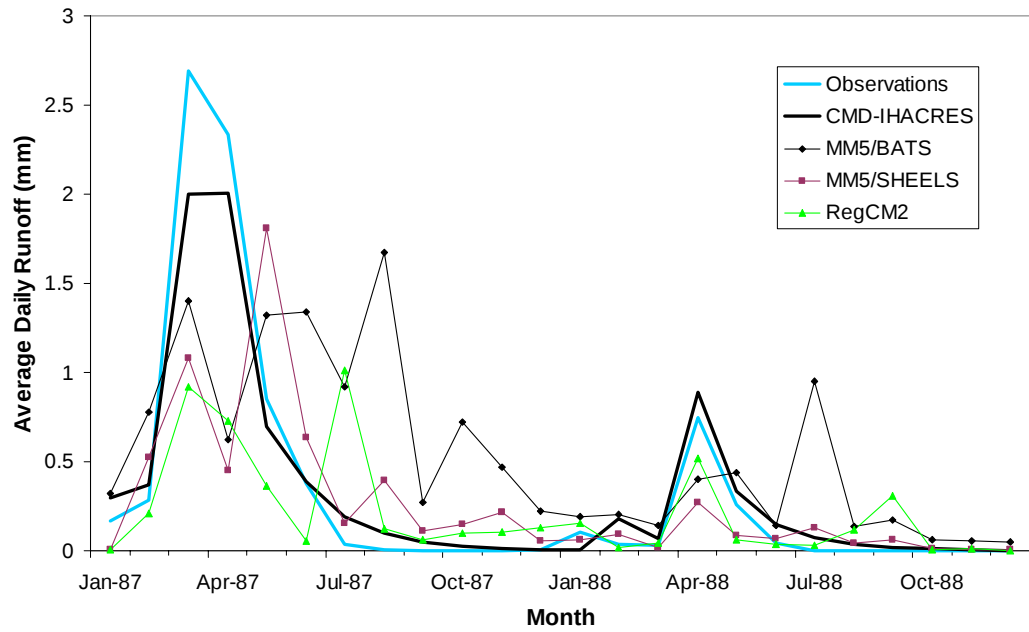


Figure 8.13 Average daily runoff each month simulated by each of the models in stand alone mode.

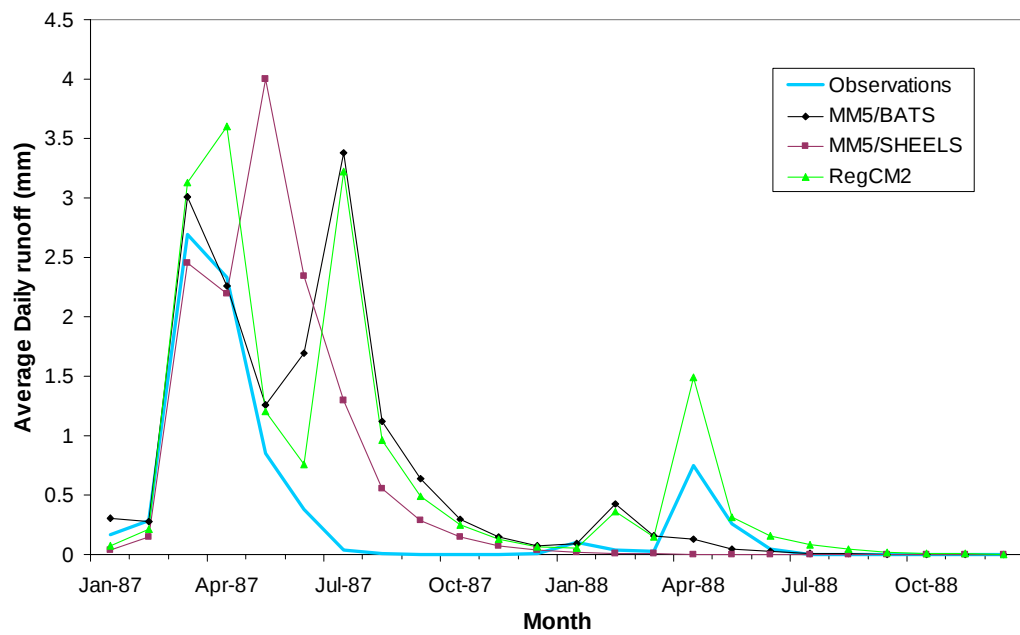


Figure 8.14 Average daily runoff each month simulated by each of the LAMs in offline mode.

The offline simulation results presented above require the precipitation and temperature time series simulated by the LAMs to drive CMD-IHACRES. In this case, CMD-IHACRES then calculates the ET to be directly proportional to the temperature. While this may produce a reasonable first order approximation of the ET, it is clear from chapter three that a more complicated formulation is required to more accurately simulate ET. Section 4.2 demonstrates the improvement in ET simulation gained by changing the modified temperature approach used in CMD-IHACRES and given by equation 4.13, to the Priestly-Taylor (equation 4.14) or Penman-Monteith (equation 4.15) formulations.

BATS and SHEELS both calculate the ET from the surface using a canopy layer and a ground layer. They explicitly account for the interception and evaporation of water by the vegetation canopy, transpiration of the vegetation and evaporation from bare ground. The potential evaporation is calculated using the Penman-Monteith big-leaf approach, which is then amended by the available soil moisture to obtain the actual ET. This approach is conceptually more comprehensive than the modified temperature approach used in CMD-IHACRES.

In order to assess the effect of using such a simplified ET formulation in CMD-IHACRES on the simulated runoff a further set of offline experiments were conducted. This time, instead of providing precipitation and temperature time series, the LAMs provide precipitation and ET time series, thus removing the need to use equation 4.9 at all. The parameters calibrated in section 8.3.1 are used here, however parameters c_1 and c_2 are no longer required as they only appear in equation 4.9.

Figure 8.15 displays the daily runoff simulated in the offline experiments. In general the P&ET offline experiments simulate less runoff than the P&T experiments, this is largely due to the increased ET losses in the P&ET experiments. MM5/BATS simulates the least amount of difference between the two offline experiments. MM5/SHEELS simulated runoff changes much more between the two offline experiments than

MM5/BATS. The unobserved runoff peak occurring in May 1987 in the P&T experiment has been removed in the P&ET experiment, however the P&ET experiment

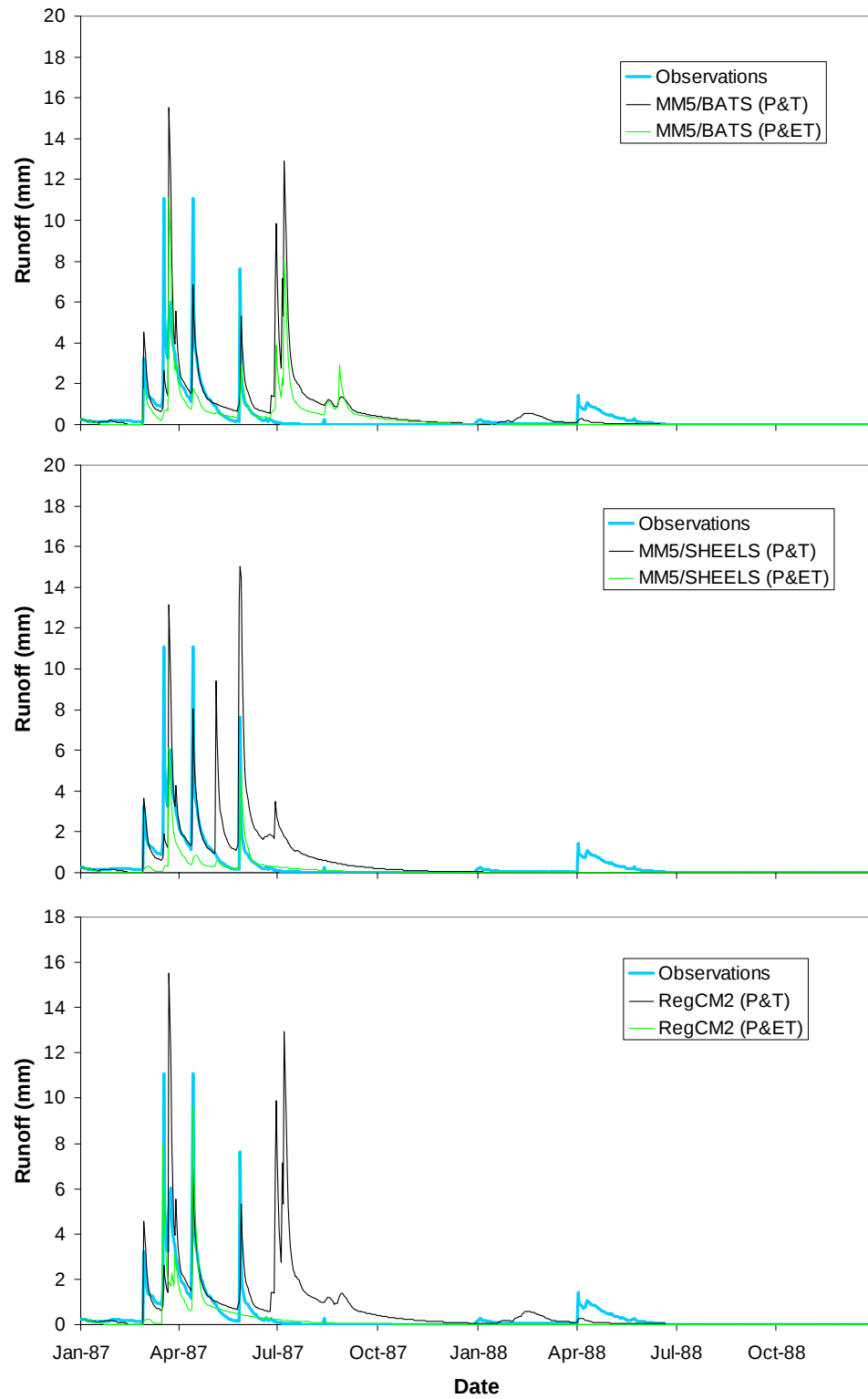


Figure 8.15 Daily streamflow simulated in offline simulations where CMD-IHACRES is driven by precipitation and temperature (P&T) time series or by precipitation and ET (P&ET) time series simulated by the LAMs

is unable to reproduce the runoff peak observed in April 1987. Similarly for RegCM2, the unobserved runoff peak in July 1987 in the P&T experiment has been removed in the P&ET experiment, however the P&ET experiment is unable to reproduce the runoff peak observed in June 1987.

8.3.4 Interplay between ET and runoff formulations

It has been recognised in previous land-surface and climate model intercomparison studies that due to the complexity of interactions among the components of the model, isolating and quantifying a given components contribution to the overall error is very difficult (Timbal and Henderson-Sellers 1998, Wood et al. 1998, Leung et al. 1999). Insight into the intercomparison conducted here may be gained from the definition and comparison of a few bulk quantities which characterise the models water balance dynamics. Koster and Milly (1997) derived two such quantities in terms of their relatively simple monthly water balance model (MWBM).

Analysis of this model led to two quantities that characterise the formulation of the soil water balance dynamics: 1) the efficiency of the soil's evaporation sink integrated over the active soil moisture range $\langle \beta \rangle$, and 2) the fraction of this range over which runoff is generated f_R . Koster and Milly (1997) calibrated the parameters of the MWBM against ET and runoff simulated by various land-surface models and claim that 'Regardless of the land-surface model's complexity, the combination of these two derived parameters with rates of interception loss, potential evaporation, and precipitation provides a reasonable estimate for the land-surface model's simulated annual water balance.'

Here the two quantities defined by Koster and Milly (1997) above are derived, without recourse to the MWBM, for all three LAMs (MM5/BATS, MM5/SHEELS and RegCM2), CMD-IHACRES and the observations. The major differences in the derivations include: Koster and Milly (1997) remove the interception loss from the ET time series while it is included here; they perform a land-surface model run with a prescribed saturated surface to determine the potential evaporation while the method of

Thornthwaite (1948) is used here; they define the “active soil moisture range” in terms of parameters in the MWBM (the minimum value of soil moisture is that for which ET goes to zero, and the maximum value is that for which all precipitation is converted to runoff) while here it is defined simply as the range encountered during the two year simulation. Finally, they require the calibration of eight parameters in the MWBM before they can subsequently derive the two quantities of interest while these quantities are derived comparatively directly.

The derived quantities are independent of the actual soil moisture values, allowing direct comparison between models even though they may simulate substantially different magnitudes for soil moisture. That is, the two derived parameters, $\langle \beta \rangle$ and f_R , in fact describe the *relative* positions of the ET efficiency and runoff functions. Koster and Milly (1997) claim “the absolute positions are, in a sense, irrelevant in terms of the land-surface models response to atmospheric forcing.” While this may be true when focusing on the runoff and ET simulated by a LAM it cannot be considered true when assessing the LAM as a whole. For example, the surface albedo depends directly on the magnitude of the soil moisture as can be seen in equation 5.60.

In order to derive the first of the quantities above, an ET efficiency, β , is defined as

$$E = \beta \cdot E_p \quad (8.1)$$

where E_p refers to the potential evapotranspiration (PET), and here it is calculated using the method of Thornthwaite (1948) as described in section 3.2.6. This calculation was performed using monthly average ET and PET. $\langle \beta \rangle$ is then the average of β across the active soil moisture range which is defined here as the soil moisture range encountered during this two year simulation.

In order to derive the second quantity a linear regression between monthly runoff and soil moisture was first performed. From this regression an estimate of the soil moisture for which runoff is zero was obtained, SM_0 . The fraction of the active soil moisture range over which runoff occurs is then given by

$$f_R = \frac{SM_{\max} - SM_0}{SM_{\max} - SM_{\min}} \quad (8.2)$$

where $SM_{\max}$ and $SM_{\min}$ are the maximum and minimum soil moisture simulated during this two year period respectively. The analogous equation in terms of catchment moisture deficit is given by

$$f_R = \frac{CMD_0}{CMD_{\max}} \quad (8.3)$$

where CMD_0 is the CMD at which runoff reaches zero, and $CMD_{\max}$ is the maximum CMD reached during the two year period. The minimum value for CMD is zero.

Values for these two quantities were established for all the LAMs in both stand alone and offline mode, for CMD-IHACRES as well as for the observations and are presented in Table 8.3. In the case of the observations only data from the observation

Table 8.3 Derived values of $\langle\beta\rangle$ (dimensionless) and f_R (dimensionless) for the observations, CMD-IHACRES and the three LAMs.

Model	$\langle\beta\rangle$	f_R
Observations	0.912	0.74
CMD-IHACRES	0.911	0.67
MM5/BATS	0.843	0.96
MM5/SHEELS	0.815	0.89
RegCM2	0.901	0.83
MM5/B + C-I	0.881	0.72
MM5/S + C-I	0.883	0.66
RegCM2 + C-I	0.911	0.74

periods in each year were used, that is the observational quantities were based on a significantly shorter period than the other quantities and as such would contain greater uncertainty. Nevertheless it does serve to create targets which the models should aim to reproduce.

Of the stand alone experiments CMD-IHACRES is best able to reproduce the observed evaporation efficiency and runoff fraction. RegCM2 also performs reasonably well, while the MM5 based LAMs perform relatively poorly. In the offline experiments the performance of all of the LAMs is improved with RegCM2 performing particularly well in reproducing the observed quantities.

8.3.5 Runoff or Effective Rainfall?

The runoff formulations in LAMs were developed in order to allow the maintenance of a reasonable long term water balance at a grid point. As such, little effort went into obtaining a formulation which embodied the correct runoff timing and instead runoff was considered to be instantly lost from the water balance of a grid cell, after which it was no longer of interest. They thus ignored the movement of water within a grid cell.

This definition of runoff may be more akin to the effective rainfall used within CMD-IHACRES (see Chapter 4) than to the streamflow itself. This would imply that to improve the runoff simulated by the LAMs it may be enough to treat the LAM simulated runoff as effective rainfall which can then be run through the linear routing component of CMD-IHACRES in order to obtain a true runoff time series.

Figure 8.16 presents the runoff simulated by the linear routing component of CMD-IHACRES, calibrated in section 8.3.1, driven by the LAM simulated runoff treated as effective rainfall. While the streamflow simulated here is not particularly close to the observations, it does display behaviour much more reminiscent of the observed streamflow than the stand alone LAM simulated runoff (see Figure 8.3).

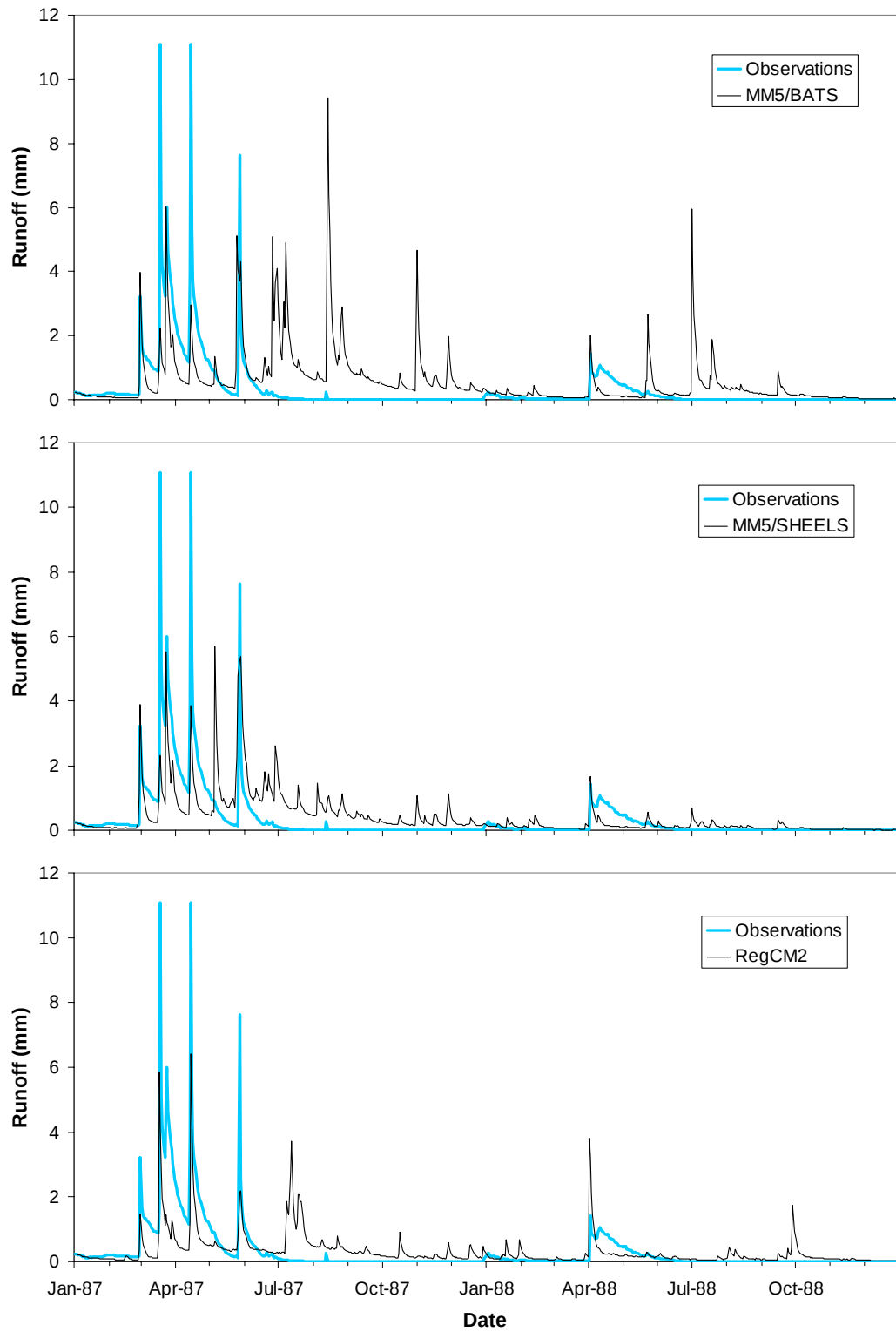


Figure 8.16 Daily runoff simulated using the runoff simulated by each LAM as the effective rainfall to drive the linear component of CMD-IHACRES

8.4 Conclusions

In this chapter the effect of including CMD-IHACRES as the runoff component in the LAMs was investigated. First, CMD-IHACRES was successfully calibrated and validated on the Kings Creek catchment in section 8.3.1. Comparison of the stand alone simulated daily runoff reveals CMD-IHACRES performs significantly better than the LAMs. In fact the LAM simulated runoff is a series of discrete peaks similar to what would be expected from the overflow mechanism of a simple bucket model. This is despite the fact that the LAMs use somewhat more complicated runoff formulations than a bucket model.

In BATS the surface runoff is given by equation 5.54. Thus it depends on both the relative soil moisture and G where, in the absence of snow melt, G is given by

$$G = P - ET \quad (8.4)$$

If G is negative then the surface runoff (R_s) is zero. That is, surface runoff can only occur when the precipitation is greater than the ET. Clearly this is not a realistic assumption for runoff since we can observe runoff even on days with no precipitation.

In SHEELS the surface runoff is given by equation 5.81. Thus it depends on both the local slope angle and the infiltration excess. The local slope angle is simply a constant at any particular location giving

$$R_s \propto P - I \quad (8.5)$$

If the infiltration through the surface layer is greater than the precipitation then no surface runoff occurs. While this approach is quite different to the approach taken in BATS, it nevertheless has similar implications for days with no precipitation. That is, if there is no precipitation there can be no runoff, which is not what is observed.

This inability to simulate runoff without simultaneous precipitation is the major reason for the near linearity found in the double mass plots (Figure 8.5) simulated by the LAMs. It is also largely responsible for the lack of low flows seen in the flow duration curves (Figure 8.4).

Using CMD-IHACRES as the runoff formulation significantly improves the runoff simulated by all three LAMs (Figure 8.9). In particular, runoff is simulated even when no simultaneous precipitation occurs, generally increasing the occurrence of low flows. This is evident in the flow duration curves shown in Figures 8.10 – 8.12. It also significantly improves the seasonal cycle of runoff as can be seen in Figures 8.13 and 8.14.

The improvement in the runoff simulation obtained using a more complicated ET formulation than the modified temperature approach of CMD-IHACRES, as seen in Figure 8.15, suggests that this may be a worthwhile improvement to CMD-IHACRES. If it were possible to calibrate CMD-IHACRES with a complex ET formulation, one would expect an improvement in the runoff simulation over that seen in Figure 8.15 where the model is run using parameter values calibrated to the modified temperature ET formulation. As discussed in section 4.2 however, the data necessary to drive these complex ET formulations are rarely available for calibration purposes. Thus the CMD-IHACRES parameters remain to be determined by calibration with the modified temperature approach as was done here, or in the long term the parameters may be related to landscape attributes and thus determined without calibration (Post 1996, Post and Jakeman 1996).

Examination of the non-dimensional quantities defined in section 8.3.4 reveals the MM5 based LAMs to have low ET efficiency, $\langle \beta \rangle$. That is, they tend to underestimate the proportion of the potential ET which is converted to actual ET when compared to the observed for this two year period. All three LAMs overestimate the fraction of the active soil moisture range over which runoff occurs, f_R . This may be largely related to the fact that the runoff formulation in both BATS and SHEELS is only secondarily

related to soil moisture, while it is primarily a function of precipitation and ET or infiltration. Including CMD-IHACRES as the runoff formulation in the LAMs significantly improves both their ET efficiency and runoff fraction over this two year period. In particular, the combination of RegCM2 and CMD-IHACRES almost precisely reproduces the observed values of these parameters.

Possibly the most promising avenue for further investigation is touched upon in section 8.3.5. Here the LAM simulated runoff is treated as effective rainfall to drive the linear component of CMD-IHACRES. Clearly this is enough to produce a reasonable hydrograph. The parameter values used in the linear component of IHACRES have been calibrated against the effective rainfall time series produced using the catchment moisture deficit accounting scheme. This effective rainfall time series is somewhat different in character to the runoff time series produced by the stand alone LAM simulations, as such the parameter values used may not be appropriate.

In order to obtain appropriate parameter values it may be necessary to change the non-linear component of CMD-IHACRES and/or the runoff formulation in the LAMs such that the effective rainfall time series simulated by CMD-IHACRES and the LAMs demonstrates similar (if not the same) characteristics. This would then allow the calibration of CMD-IHACRES on observed data to produce parameter values appropriate for use with LAM generated effective rainfall time series.

Alternatively, it may be possible to estimate the linear CMD-IHACRES parameter values by investigating the properties of the observed streamflow directly and/or the physical properties of the catchment of interest. A technique for the estimation of the linear CMD-IHACRES parameters based on analysis of the observed streamflow time series is currently under development (Croke in prep.). Various other hydrograph estimation techniques have also been developed by others (Chapman 1996, Da Ros and Borga 1997, Hansen et al. 1997, Moussa 1997).

In summary then, the inclusion of CMD-IHACRES run offline with the LAMs significantly improves the runoff simulation. In particular the combination of RegCM2 and CMD-IHACRES appears to provide the best runoff simulation compared to the

MM5 based LAMs. This suggests that the combination of CMD-IHACRES and a regional climate model may well prove to be of practical use in investigating climate change effects on streamflows in data sparse areas. While these results suggest that further experiments with CMD-IHACRES run on-line with a LAM are warranted, the best way to incorporate CMD-IHACRES into a LAM is not clear. Further work investigating this “online” potential is currently under way.

Chapter 9

Hydrological Impacts of Climate Change on Inflows to Perth, Australia

9.1 Introduction

This chapter presents a climate change impact study focussing on surface hydrology, in particular on flooding and the changes in average recurrence interval (ARI) of flow events up to 1:1000 years. It demonstrates the potential use of CMD-IHACRES to investigate the extent of this impact. While it is acknowledged that considerable uncertainties exist in the development of climate change scenarios and even more in climate change impact assessments, it is important to attempt to quantify change as soon as possible so that identified problems can be addressed in advance by adaptive management. In the context of water resources extreme conditions are of particular concern. Coping with increased flood magnitudes or longer droughts may require long-term planning to ameliorate their impacts.

It is likely that the industrially induced increase of greenhouse gases in the atmosphere could lead to changes of the Earth's climate. For example IPCC (1996) estimates that by 2100 the global mean surface air temperature will increase by between 1°C and 3.5°C. One of the most important aspects of global change science is an estimation of possible impacts of these climatic changes to the world water availability. This is particularly important for any work aimed at supporting the sustainable management and long-term planning of water resources. It is especially important in Australia where the supply of water resources is a major constraint for further development of large urban areas as well as being vital for the environmental protection of aquatic systems.

Most hydrologically-focused climate change impact studies assess changes in water availability as well as extreme events, particularly floods (eg. Schreider et al. (2000b)). These studies are generally performed in order to answer questions about the future water supply to large urban or intensive agricultural areas and, hence, are focused on major water supply catchments. This study undertakes a comprehensive analysis of climate change impacts on the hydrological regime of streams entering the Perth City area, Western Australia. Streamflow entering the Perth urban area from the studied catchments represents some 90% of the total flow of the Swan River, which

subsequently flows through the heart of Perth. As such this is a fairly comprehensive study of the likely impacts on the urban flooding regime for Perth. The analysis is focused on the flooding regime and its implications for the city's urban storm water system, and therefore does not look explicitly at the water supply for urban consumption.

The problem of streamflow response to climate variation is stated in Nemec and Schaake (1982). An overview of climate models employed for climate impacts assessment can be found in Pittock (1988), Tucker (1988, 1990, 1992), Pittock (1993), Leavesley (1994) and Whetton et al. (1994).

Climate impacts assessment for Australian catchments have also been implemented in some previous research works. Close (1988) modelled nine rivers of the Murray-Darling Drainage Division and estimated the possible climate impact on its water resources using the Murray-Darling Basin Commission empirical model. Nathan et al. (1988) applied a deterministic, conceptual rainfall-runoff model, HYDROLOG, to study the climate impact on runoff in the Myponga Weir catchment in Southern Australia and Moggerah Dam in Queensland. Whetton et al. (1993) investigated implications of climate change on floods and droughts in Australia. Chiew et al. (1995) followed the above approach using the Modified HYDROLOG model in order to model 28 benchmark catchments in Australia and estimate the climate impact on their streamflow. Ye (1996) used climate change scenarios developed with a GCM and a stochastic weather generator to investigate the climate change impact on streamflow predicted by three conceptual rainfall-runoff models. He found greater variation between the rainfall-runoff model predictions in arid or semi-arid regions when compared to humid regions. Schreider et al. (1997) analysed the climate impacts on the water availability and extreme events such as floods and droughts in four Basins of the Murray-Darling Drainage Division (Goulburn, Ovens, Kiewa and Upper Murray). Climate change impacts on urban flooding in three Australian urban catchments were analysed in Schreider et al. (2000a).

We begin this chapter with a short review of various climate change impact analysis methods followed by a justification of our model selection. Section 9.2 provides a

demonstration of the climate invariance of CMD-IHACRES parameters. A description of the study site is provided in Section 9.3. Section 9.4 outlines the modelling methodology used, the results of which are presented in Section 9.5. A brief discussion of the implications of these results is given in Section 9.6.

9.1.1 Climate change scenarios

Where it may be generally accepted that the increasing concentrations of greenhouse gases in the atmosphere are causing climate change, there still exists considerable uncertainty in the magnitude, timing and spatial distribution of this change. Results from climate change impact studies, such as this one, depend critically on the climate change scenarios used. Several methods for creating climate scenarios exist, which cover a considerable range of complexity, cost and time demand.

One of the simplest approaches is to use scenarios based on changes to long term seasonal mean values of climate descriptors (as determined for example by GCM runs) to transform present-day climatic time series. This approach has been used in previous studies (Schneider et al. 1997, Feddema 1999, Mehrotra 1999), but has several limitations. Primarily, this approach takes no account of any changes in the variability of the relevant climate descriptors.

The direct use of GCMs as the basis for creating climate scenarios has the advantage of estimating changes in climate due to increased greenhouse gases for a large number of climate variables in a physically consistent manner. These climate variables are consistent with each other within a region and around the world. A major disadvantage of using GCMs is that, although they accurately represent global climate, they are often inaccurate when simulating current regional climate. To overcome this regional inaccuracy the use of regional (limited area) climate models (McGregor and Walsh 1993, Giorgi et al. 1994) or stochastic downscaling of relevant meteorological variables (Hughes 1993, Hughes and Guttorp 1994, Bates et al. 1998) has been promoted in

several studies. Both of these methods share the disadvantage of being costly and time demanding. The use of regional climate models depends intimately on the GCM-supplied boundary conditions, and so may not correct for errors, while the stochastic downscaling technique assumes that synoptic scale relationships are constant over time.

Another approach, used here, utilises a stochastic weather generator to create extended time series for various climate descriptors (Bates et al. 1993, Charles et al. 1993, Semenov and Barrow 1997). These stochastic weather models, fitted to current climatic time series, can be adapted to the generation of synthetic series for future climate using the method presented by Wilks (1992). Adjustments to the model parameters are made in a manner consistent with the changes in monthly statistics derived from comparisons of GCM runs for control and doubled CO₂ conditions.

In this study the output statistics of the CSIRO9 GCM (McGregor et al. 1993b) have been used to look at a 2×CO₂ future climate scenario. When compared with other GCM results, this model is among those predicting the most severe changes. Hence a scenario has been included that embodies about half of the 2×CO₂ predicted change, which we will refer to as the 1.5×CO₂ scenario. These scenarios are consistent with the broad range of global warming projections based on increased atmospheric concentrations of greenhouse gases. They are physically plausible and estimate daily precipitation and mean temperature, which are then used to drive the hydrological model. By using a 1000 year long daily time series it is expected that the potential range of climate variability under each of the CO₂ scenarios would be captured.

Other implicit assumptions involve the role of vegetation in the climate system. Here we have assumed that the overall vegetation response for a given precipitation and temperature input will remain similar over the next century while greenhouse gas concentrations increase. However, the vegetation cover and/or its evapotranspiration response may change with future changes in climatic patterns of temperature, precipitation, solar radiation, as well as fertilisation and stomatal resistance effects related to increases in carbon dioxide. It is worth noting that most GCMs currently use a static model of the land surface where surface characteristics such as roughness length,

albedo, soil and vegetation parameters are specified at the start of a run and do not change regardless of any prolonged change in the predicted climate.

Several studies have investigated the impact of climate on vegetation structure (Busby 1988, Monserud et al. 1993). They imply some movement of vegetation types, in particular the shrinking of boreal zones and the increasing elevation of the tree line in alpine regions. Since our study area doesn't contain any boreal type vegetation, and the period over which the above increases in CO₂ are expected to occur (about 70 years) is too short for forests to grow over considerable areas, we would not expect any significant natural migration of vegetation types. Of course, vegetation structure in a region can change quite rapidly and radically due to direct human intervention. Here we have assumed that any change in land management practices will be minor enough to have no significant impact.

Studies have investigated the biological response of some plants to increased CO₂ in the atmosphere, an introductory overview of which can be found in Kristiansen (1993). While this response varies between species a few general points can be made. There are changes in stomatal conductance associated with higher CO₂ levels, which lead to reduced water exchange per unit leaf area. The higher levels of CO₂ also lead to enhanced leaf growth. These two effects can, depending on species, offset each other in terms of total evapotranspiration response (Lins et al. 1997). Until further knowledge becomes available it seems reasonable to assume that the net effect of these changes on a vegetation stand, which contains many species, will be within the measurement and model error. Thus the assumption that the overall vegetation response will remain similar to that over recent history for the time covered by this study seems quite plausible.

9.1.2 Model selection

The spectral GCM used in this study (CSIRO9) was developed by the CSIRO Division of Atmospheric Research (McGregor et al. 1993a). The model operates with nine

vertical levels in the atmosphere and a horizontal resolution of about $300 \text{ km} \times 600 \text{ km}$. A time-step of 30 minutes was used. The simulated climate data derive from 30 year equilibrium (constant CO_2 concentration) runs for present day (control) and future (doubled CO_2) climates.

The stochastic weather generator used in this study is based on the WGEN generator described by Richardson and Wright (1984). WGEN simulates daily precipitation occurrence and amount, maximum and minimum temperature, and solar radiation. Precipitation occurrence is described by a two state (wet or dry), first-order Markov chain wherein the transition probabilities for a given location are allowed to vary through an annual cycle. Precipitation amounts on wet days (rainfall greater than 0.3mm) have their variation characterised using a gamma distribution. Standardised temperature and solar radiation components are represented as a first-order, trivariate autoregressive process conditioned upon whether the day is wet or dry. WGEN produces stochastic realisations of the variables above maintaining the statistical relationships established from observation.

As discussed in Chapter 2, Wheater et al. (1993) described three types of rainfall-runoff models which could be used for predicting the stream discharge effects of climatic variations: metric, physically-based and conceptual. Metric models are based primarily on observations and seek to characterise system response from these data. Physically-based models use a more classical mathematical-physics formulation of component processes, based on continuum mechanics, and numerical solution techniques to solve the relevant equations. Conceptual type models vary considerably in complexity but are always based on a representation of internal storages and the fluxes between them which are associated with particular hydrological components and processes.

Metric models contain too little process description to be used to make predictions on independent periods not used for model calibration, hence have little applicability for simulating future climate impacts. Physically based models require large computation and data resources. They also have the disadvantage of containing large numbers of parameters, which introduces serious ambiguity in the identification of the parameter values. The catchments chosen for consideration here have been instrumented for a

relatively long period, meaning that streamflow, precipitation and temperature data are available for decades in the majority of gauging sites in these regions. Thus, conceptual lumped rainfall-runoff models seem to be the most suitable type of model for the streamflow analysis required in the region selected and for the particular purposes of this study. The model CMD-IHACRES falls within this class of models. The number of parameters (five or seven) to be fitted is small compared with other conceptual models, yet its performance has been impressive across a range of hydroclimatologies (Jakeman and Hornberger 1993, Post and Jakeman 1996, Schreider et al. 1997).

In the present study the CMD-IHACRES rainfall-runoff model (Evans and Jakeman 1998), was employed for predicting the future climate impacts on streamflow. It is a hybrid metric-conceptual model based on the Instantaneous Unit Hydrograph (IUH) technique. The method represents total streamflow response as a linear convolution of the IUH with rainfall excess or effective rainfall, which is in turn a non-linear function of measured rainfall and temperature. The evaporative losses from the catchment are dealt with through a Catchment Moisture Deficit (CMD) accounting scheme. Chapter four describes the model and defines its parameters in detail.

One advantage of the CMD-IHACRES model is that its parameters reflect the average, lumped properties of the catchment considered. This provides the ability to predict spatially averaged streamflow response. Therefore, the CMD-IHACRES application is suitable even in circumstances which lack spatially distributed catchment input. Being structurally simpler than physically-based models, the conceptual models can easily utilise records of hydrological data for calibration.

Perhaps the most significant argument for the use of conceptual hydrological model CMD-IHACRES in the present work is that the model parameters attempt to reflect the geomorphological and vegetation characteristics of the catchment considered and are little affected by regional climate conditions (Jakeman et al. 1993, Post and Jakeman 1996). A demonstration of this parameter climate invariance is given in section 9.2. CMD-IHACRES parameter values can therefore be established under present climatic conditions and then used without reference to the observed streamflow data. This type of model can be used for streamflow prediction for estimated future climatic conditions,

assuming that the catchment properties considered (landscape, vegetation, building and road structure for the urban catchments) will not change considerably.

9.2 Climate invariance of CMD-IHACRES parameters

An important assumption in the use of a rainfall-runoff model to analyse the effects of changes in climate on streamflows is that the models parameters are not climate dependant. Jakeman et al. (1993) indicates the climate invariance of the parameters in the IHACRES rainfall-runoff model. In order to examine the variability of the non-linear loss module parameters in the CMD-IHACRES model with respect to calibration period and the climate therein, one catchment is selected here for analysis. The Jamieson River (stream gauging station 405218), in the Goulburn River Basin, was selected because it is an example of a catchment of the scale of all but one of the other catchments being examined, is subject to winter dominated rainfall and possesses no significant land use changes over its period of record. The catchment is dominated by state-protected forest, where logging is negligible because this area is mostly covered by unproductive mixed tree species (DNRE 1998). The mean annual rainfall in the lower part of the catchment at the Jamieson Post Office is 1250 mm. However, the climatology of this catchment is characterised by a considerable difference in precipitation levels between its drier lower part, and its headwaters located in an alpine region experiencing around 1500-1800 mm of annual precipitation.

Table 9.1 shows the calibration results for the Jamieson River catchment for years 1972, 1978 and 1985: the values of the non-linear loss module parameters and performance statistics in terms of the Nash-Sutcliffe Efficiency (NSE) and the bias (see Chapter 4) are shown. The NSE is of primary importance in flood related studies as it provides a measure of how well the peaks were modelled. These calibration results were obtained using precipitation data recorded for the meteorological station at Jamieson PO (83017) and temperature from the Lake Eildon meteorological station (88023). Fits to the

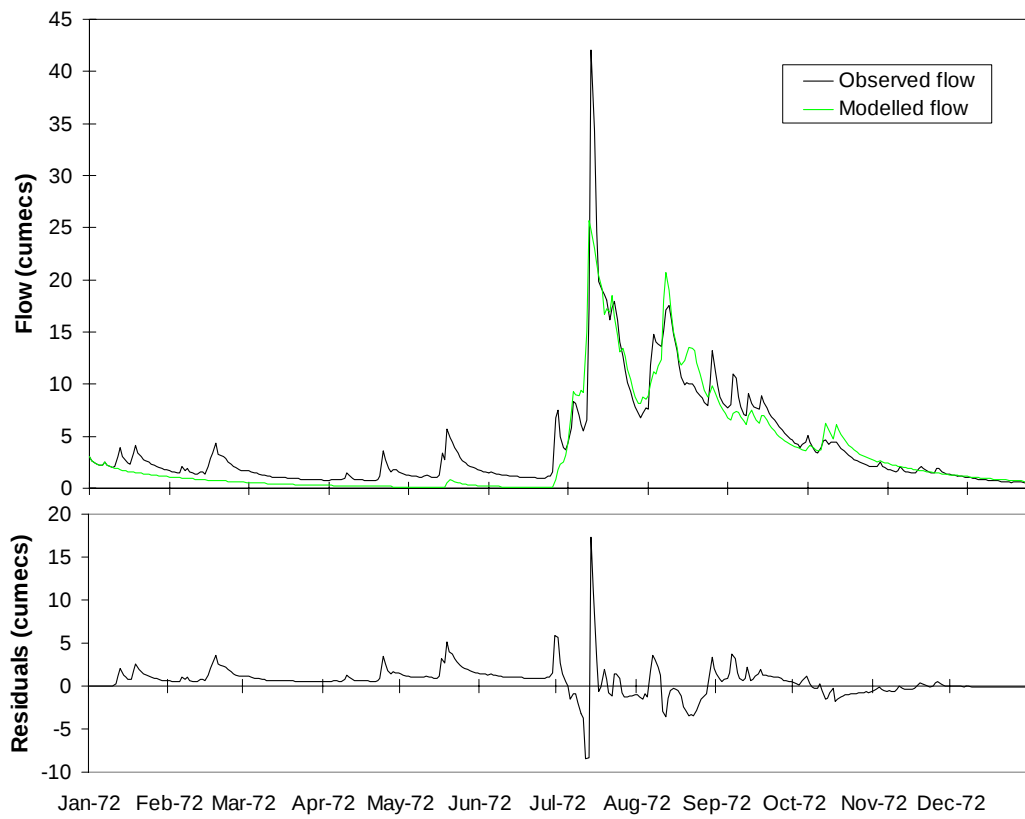
observed streamflow are shown in Figure 9.1.

Simulation runs over the whole period of observation (~20 years) provide a similarly high Nash-Sutcliffe efficiency: 0.753, 0.791 and 0.744 for these three models respectively. The parameters of the non-linear module, as shown in Table 9.1, are very similar from one calibration period to another despite the differences in precipitation conditions during these periods. Annual precipitation varies by $\pm 20\%$. The modelling exercise in the Jamieson catchment demonstrates that the CMD-IHACRES loss module parameters have little climatic dependence, and hence adds confidence to our assessment of the applicability of this model for streamflow assessment under future climatic change impacts.

Table 9.1 Calibration results for the Jamieson River catchment

Model number	Starting date	Loss module parameters				Model efficiency (NSE)	Model bias (m^3/s)	Mean precip (mm)	Mean temp ($^{\circ}\text{C}$)
		c_1	c_2	c_3	c_4				
1	1/1/1972	0.28	0.01	3	35	0.856	0.57	920	13.4
2	1/1/1978	0.28	0.01	3	35	0.906	0.30	1330	13.6
3	1/1/1985	0.28	0.01	4	35	0.829	-0.28	1190	13.3

Model 1



Model 2

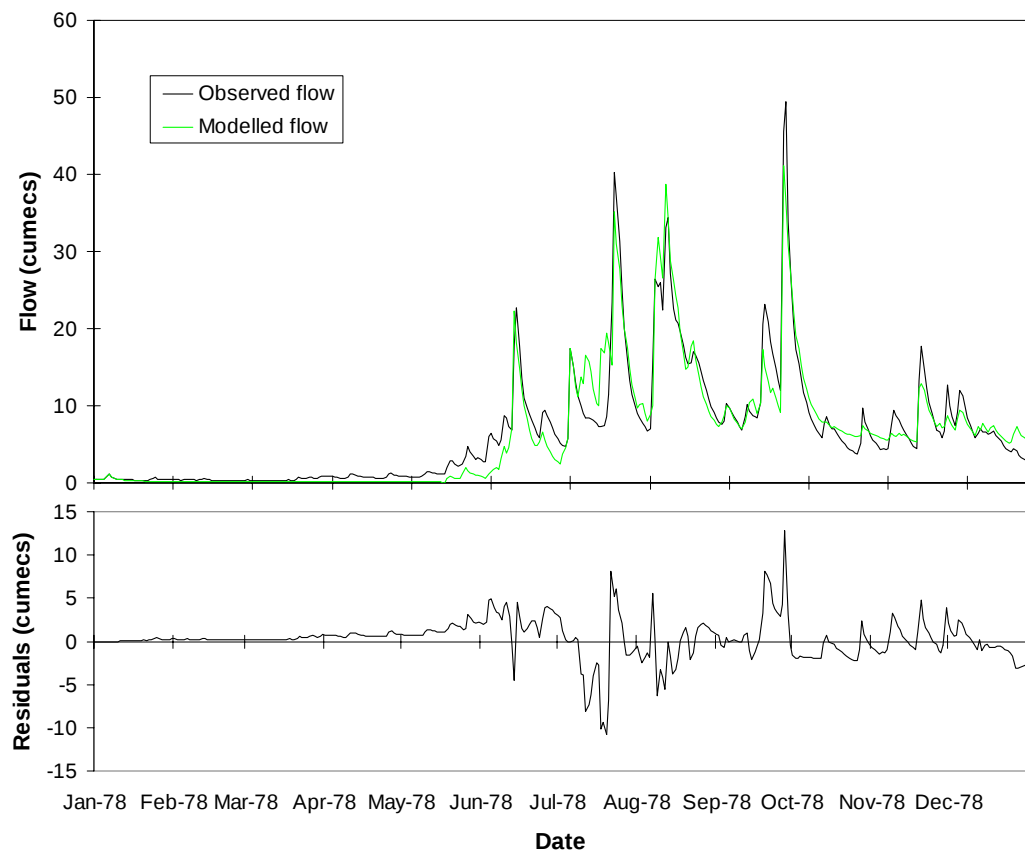


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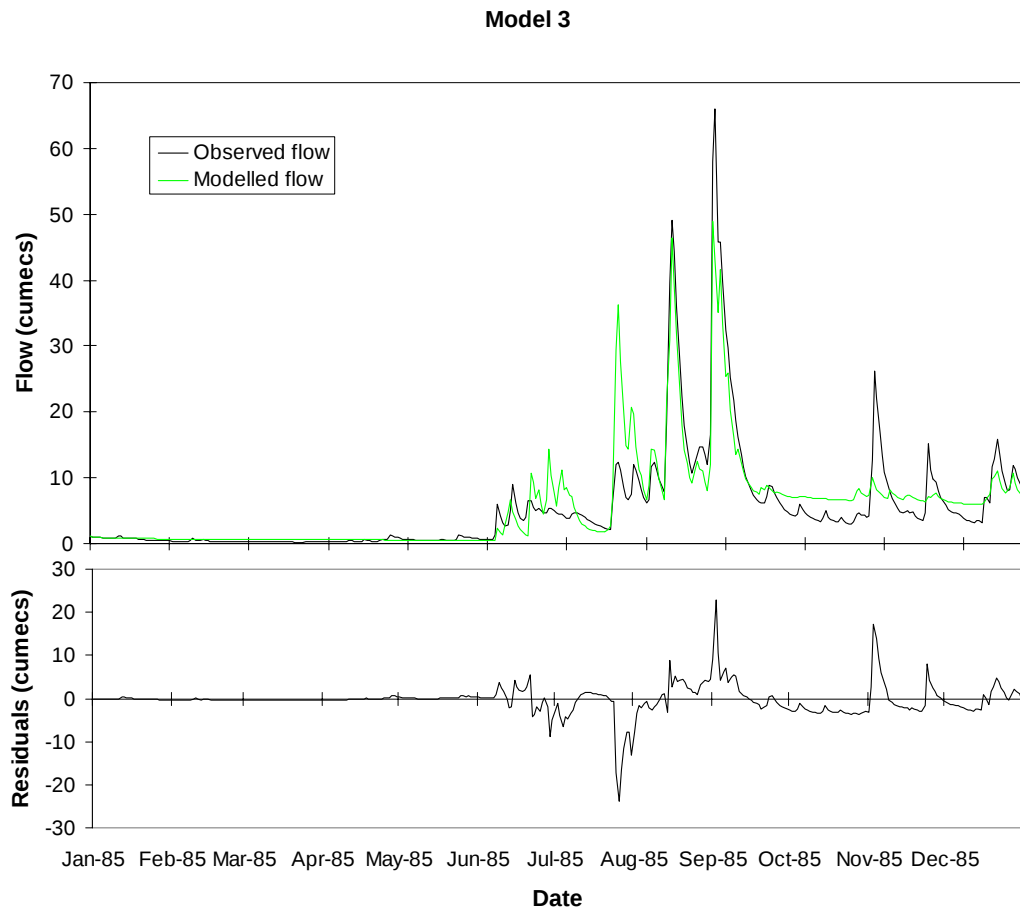


Figure 9.1 Calibration results for the Jamieson River catchment

9.3 Site description of the Western Australian catchments

A schematic of the study area, along with catchment boundaries and gauging sites, is given in Figure 9.2. All the catchments studied drain into the Swan River, which subsequently flows through the city of Perth itself. The area is dominated by winter rainfall, and the streams are ephemeral in nature. Some descriptors of the studied catchments are given in Table 9.2. Clearly the Avon River catchment covers an extremely large area and extends much further from the coast than the other catchments. The majority of the inland parts of this catchment do not contribute flow to the Avon

River itself except during large flow events. During 'normal' conditions a series of inland lakes store the streamflow from the eastern part of the catchment.

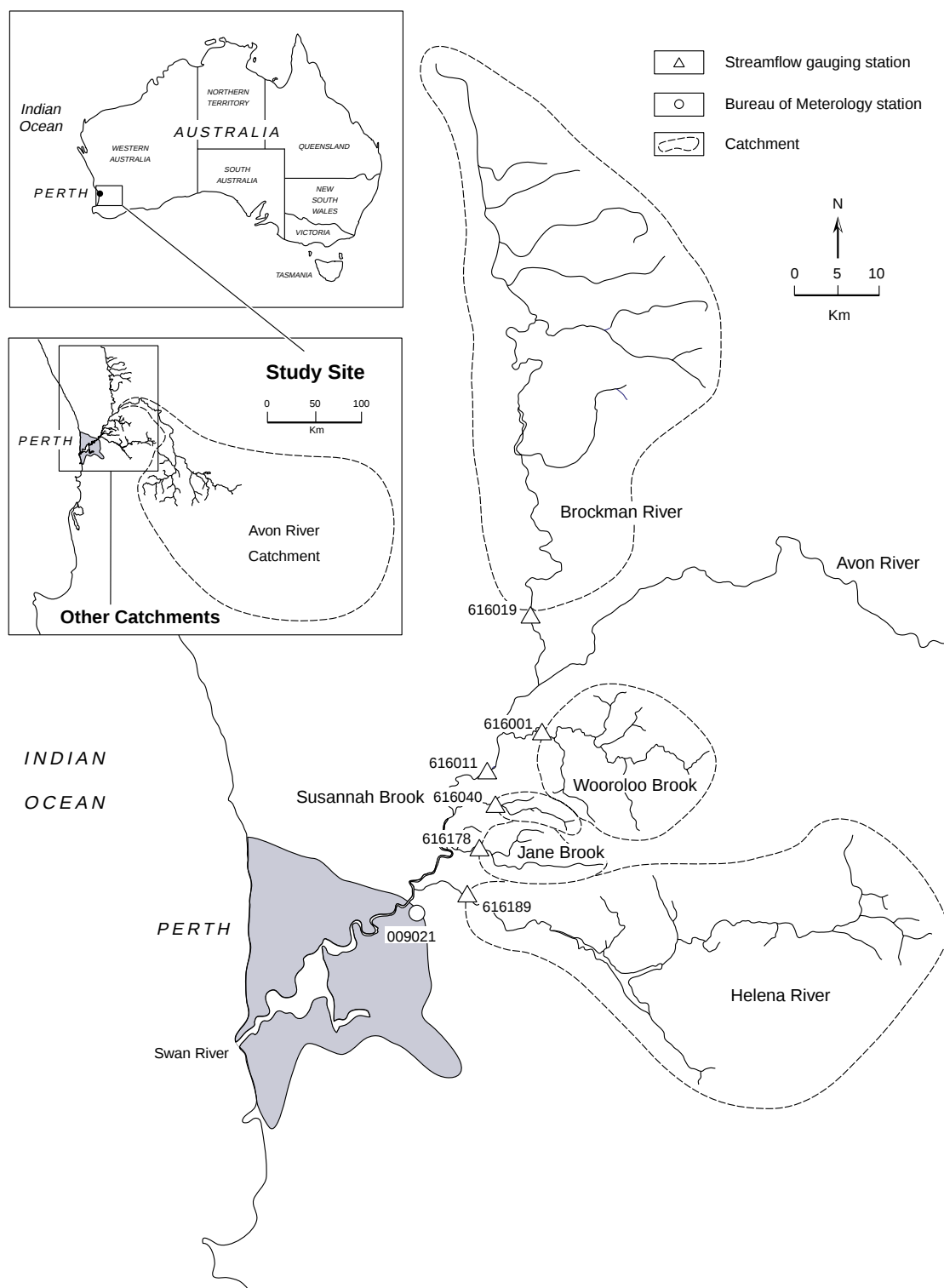


Figure 9.2 Schematic of study area

Table 9.2 Description of catchments used in this study

Catchment	Area	Land Uses	Streamflow Gauging Site	Comments
Avon River	<i>Including above Northam 120,000 km²</i>	<i>Going up the Avon there is 30-40km of forest, then it emerges onto the wheat belt. Low intensity grazing and cropping.</i>	616011	<i>Terminated at Beverley as not much streamflow above that point. Drains the wheatbelt area and then cuts through the escarpment.</i>
Brockman River	1512 km ²	Some forest, then sheep grazing.	616019	<i>No significant water extractions North-East of the study area. OUTFLOWS to this river (which then joins the Avon outside the study area)</i>
Wooroolo Brook	525 km ²	Pasture, forest, mixed farming	616001	No significant water extractions
Susannah Brook	25 km ²	Pasture, forest, not much cropping	616040	No significant water extractions
Jane Brook	73 km ²	Urban, pasture, not much cropping	616178	No significant water extractions
Helena River	1600 km ²	Urban, forest, pasture below Mundaring Weir; mostly forest above the weir	616189	<i>Forms lower boundary of study area. Heavily regulated by Mundaring Weir</i>

9.4 Modelling

Using streamflow data provided for each of the streamflow gauges shown in Table 9.2, and rainfall and temperature data from Belmont (station no. 009021), CMD-IHACRES was calibrated and the results are shown in Table 9.3 for the periods indicated. Four year periods were chosen for calibration as a compromise between using the longest period possible and using a period which doesn't contain significant land use change. Rainfall from just one station was used since it was the station for which the stochastic weather generator was run, producing 1000 years of climate data under $1\times\text{CO}_2$, $1.5\times\text{CO}_2$ and $2\times\text{CO}_2$ conditions. The methodology used to stochastically generate synthetic series for future climates is described in Bates et al. (1993), Charles et al. (1993) and Bates et al. (1994). The altered climate scenarios were derived from the CSIRO9 Mark 1 GCM runs (circa 1992). However, the $2\times\text{CO}_2$ output from this model is now thought to over predict the climate change. For this reason, we have included a ' $1.5\times\text{CO}_2$ ' scenario. This scenario was derived from interpolated parameters that generate about half the climate change of the ' $2\times\text{CO}_2$ ' scenario.

Rainfall records from one rainfall station are unlikely to capture the actual rainfall over a catchment, especially if the rainfall station lies outside the catchment boundary and/or the catchment is large. This reliance on a single rainfall station would be expected to cause significant problems in reproducing the current streamflow regime using a rainfall-runoff model. This problem is not as severe as it may seem for two reasons: first, the model appears to be robust enough to perform well even with just one rainfall station (see Table 9.3); and secondly, we are concerned with the changes that occur rather than the accuracy with which we can reproduce current streamflows. In particular, the interest is in changes in the average recurrence intervals (ARI) of flood events up to an ARI of 1000 years by using the simulations from the stochastic weather generator.

The Avon River presents a particular challenge since it is extremely large in areal extent. It extends well beyond the coastal region for which rainfall data has been provided, and it contains several lakes within the catchment boundaries. It does

however, represent a significant proportion of the total volume of streamflow reaching the Swan River and hence is important for any urban flooding study of Perth. Of the other catchments, the Helena River presents a unique challenge in terms of streamflow prediction since it is heavily regulated by Mundaring Weir. While the weir acts like a small dam, heavily attenuating small flows, it has little effect on large flow events which is the main concern here.

Streamflow entering the Perth urban area from the studied catchments represents some 90% of the total flow of the Swan River. As such this is a fairly comprehensive study of the likely impacts on the urban flooding regime for Perth. The Swan River flows through the centre of the city, as can be seen in Figure 9.2,

9.4.1 Rainfall-runoff model calibrations

Each of the six catchments was calibrated over a four year period of the historical record. The calibration period was restricted to four years since the catchments investigated have been subject to ongoing land use change and this is the longest period within which the land use can be considered static. The calibration results can be seen in Table 9.3 and Figure 9.3. The Avon River, Brockman River and Helena River were all modelled with a one store linear module structure, while the rest used two stores in parallel. The one store structure reflects the lack of an identifiable base flow component in the data, and reduces the number of model parameters from seven to five (Jakeman et al. 1990).

Figure 9.3 shows the modelled streamflow and the observed streamflow for each of the six catchments calibration periods. In general, the modelled flow agrees quite well with the observed, despite the problems associated with the use of a single rainfall station. Note that in Figure 9.3 (a) the modelled flow generally underestimates the observed

Table 9.3 Calibration results for the six catchments

Catchment	Period	NSE	B cumecs/d	Linear module structure	Model Parameters						
					c_1	c_2	c_3	c_4	τ_q	τ_s	V_s
Avon River	1/1/76 – 31/12/79	0.71	1.90	1 store	10	1	0.63	0.12	7.6		
Brockman River	1/1/84 – 31/12/87	0.74	0.00	1 store	18	1	0.68	0.10	24.1		
Wooroolo Brook	1/1/79 – 31/12/82	0.74	0.00	2 stores in parallel	13	7	0.46	0.03	1.3	35.9	0.77
Susannah Brook	1/1/82 – 31/12/85	0.78	-0.01	2 stores in parallel	13	8	0.36	0.04	1.0	34.6	0.72
Jane Brook	1/1/80 – 31/12/83	0.78	-0.02	2 stores in parallel	23	13	0.32	0.02	1.2	39.3	0.70
Helena River	1/1/88 – 31/12/91	0.61	0.03	1 store	84	13	0.64	0.02	2.5		

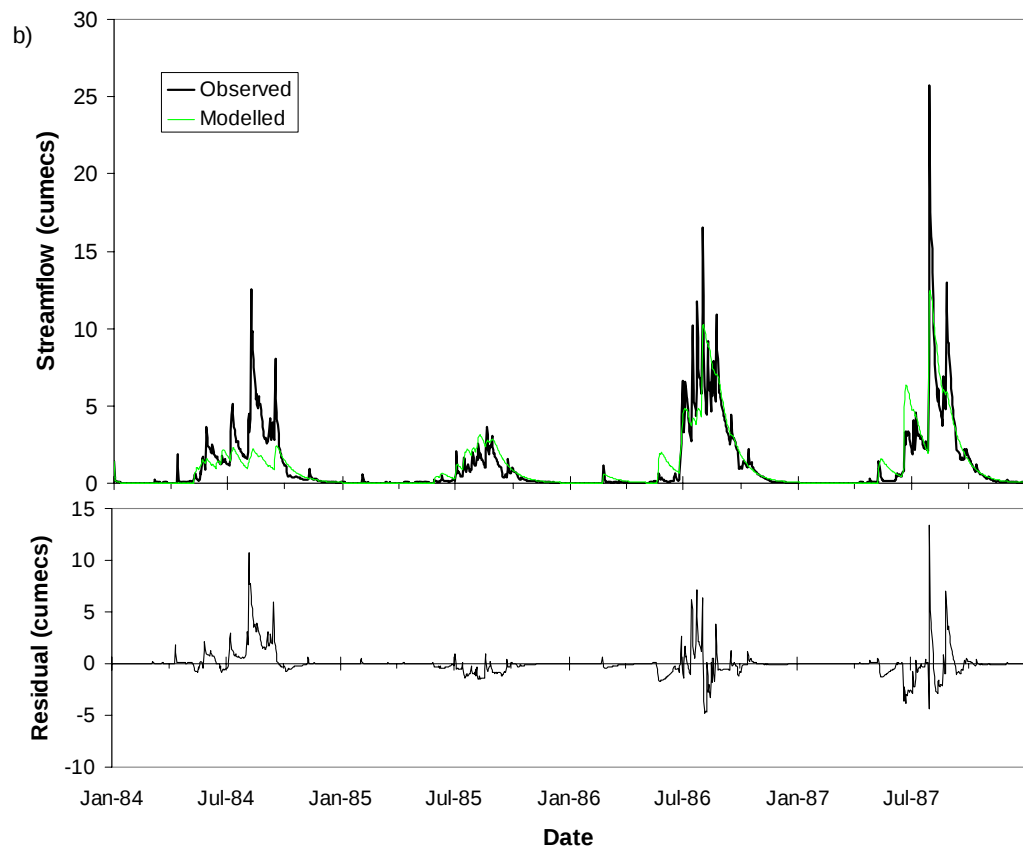
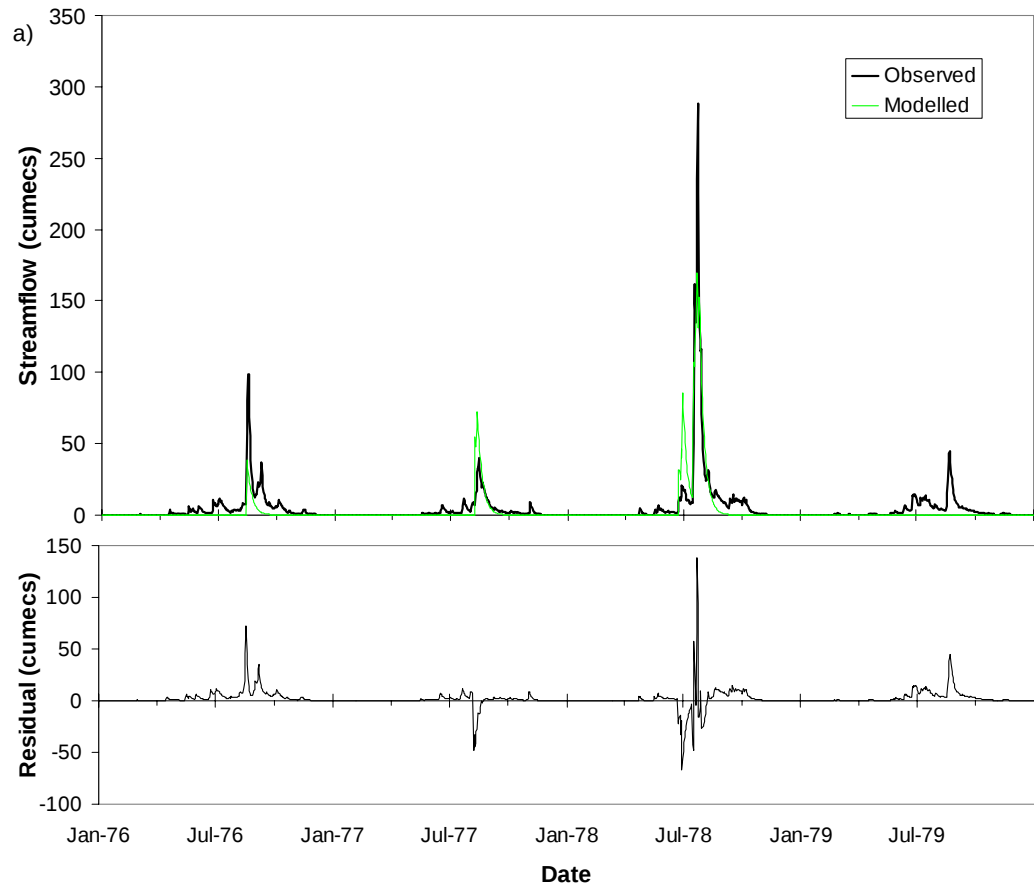


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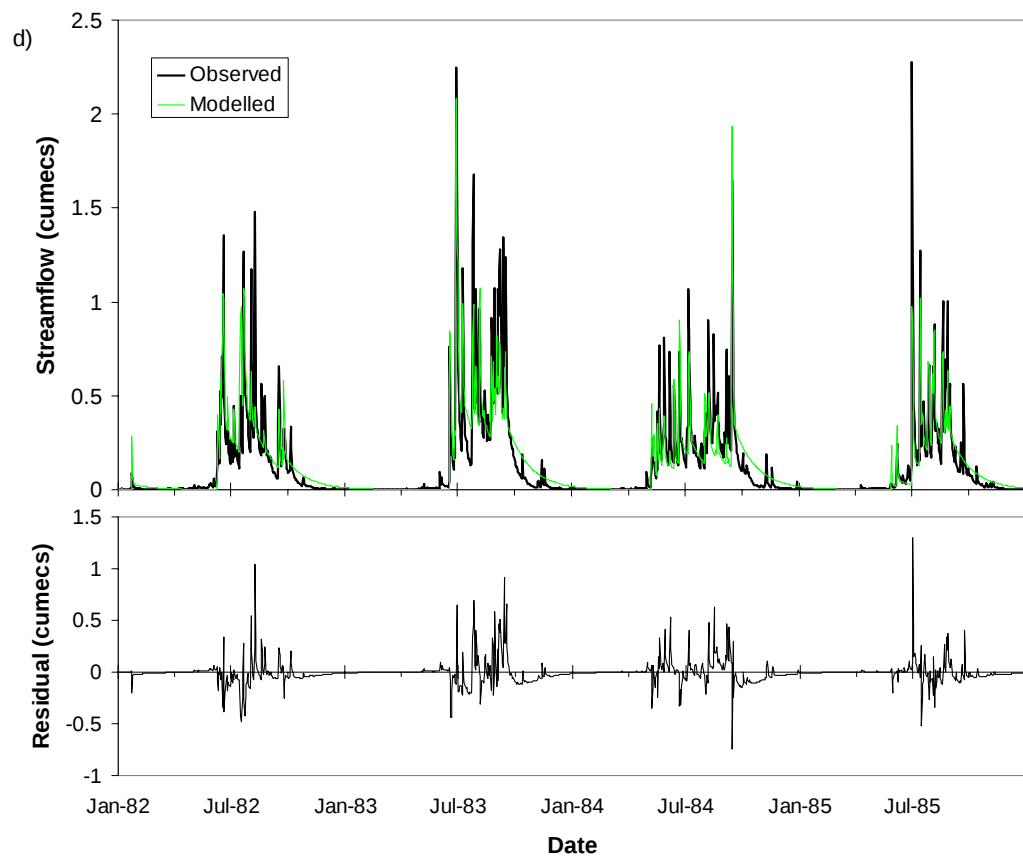
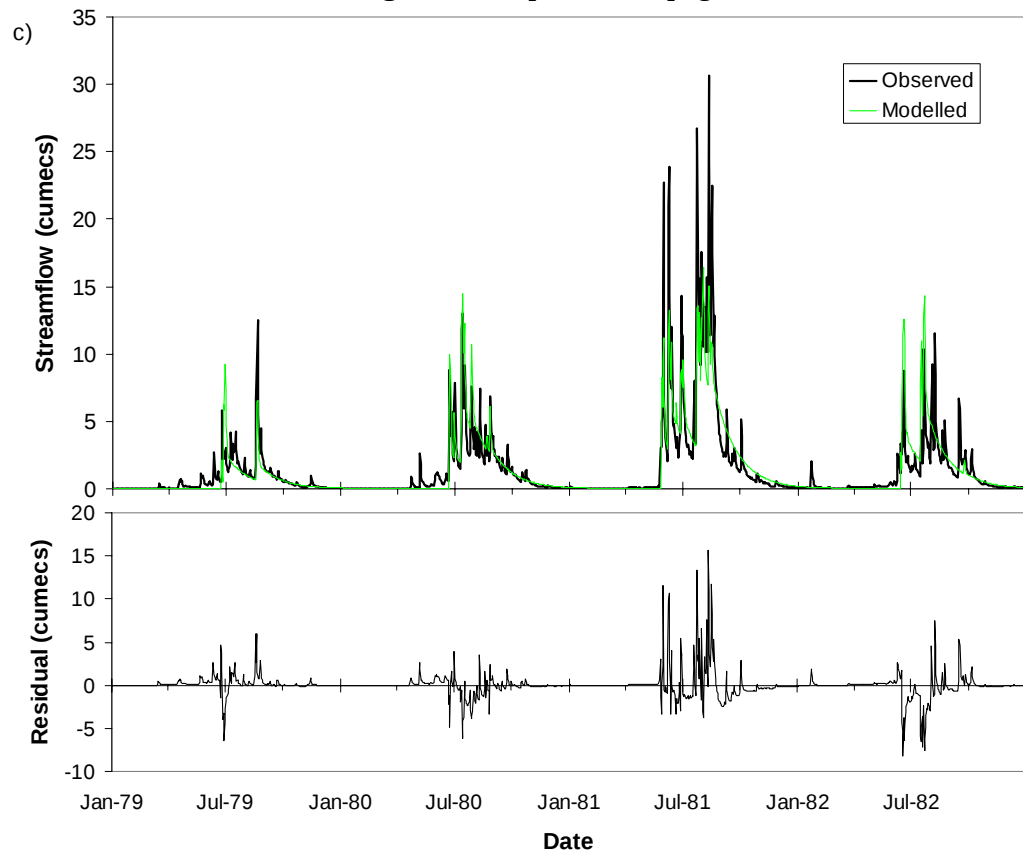


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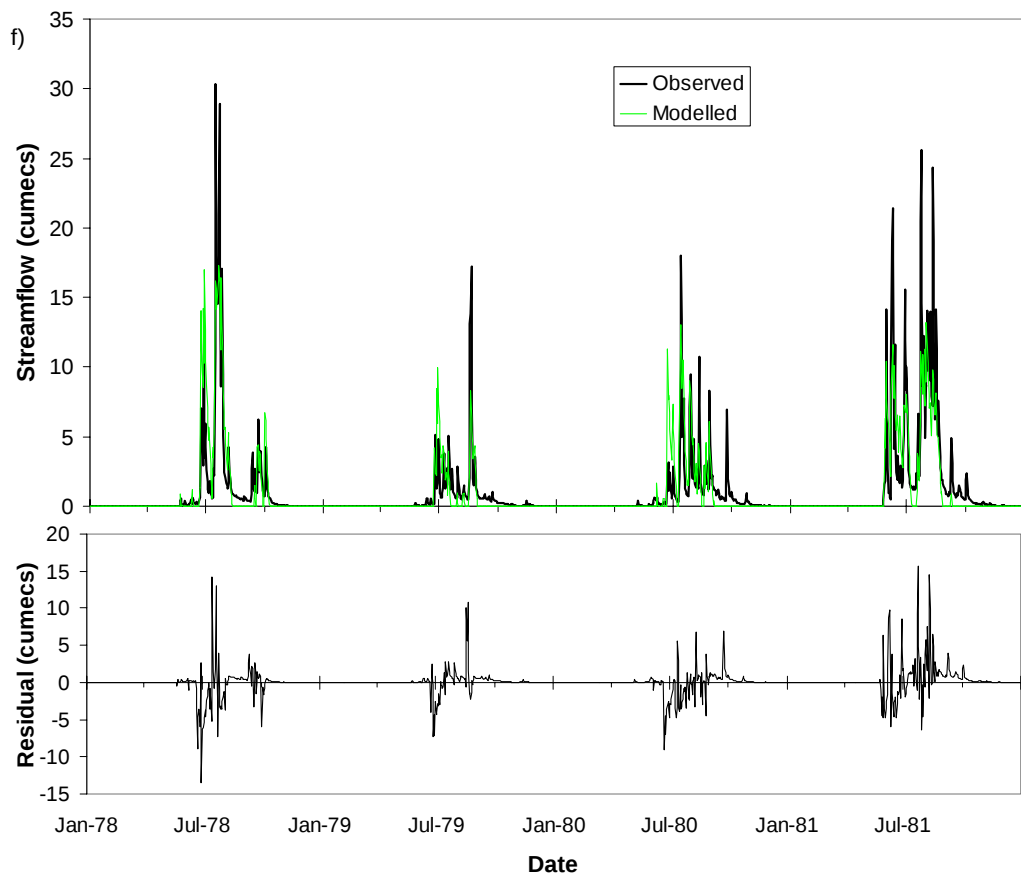
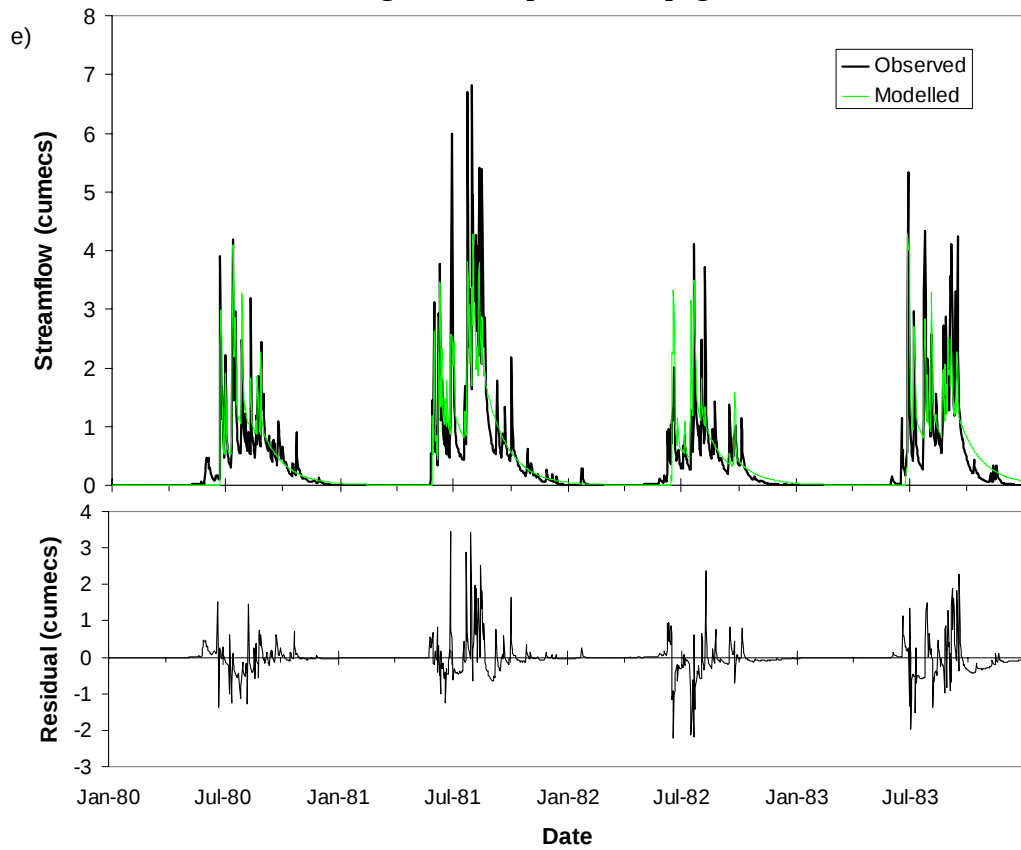


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Figure 9.3 caption Observed and Modelled streamflow for the calibration period for the: **a)** Avon River; **b)** Brockman River; **c)** Wooroolo Brook; **d)** Susannah Brook; **e)** Jane Brook; and **f)** Helena River.

flow. This is quantified in Table 9.3 where the Avon River has significant bias. This bias of 1.90 cumecs per day, while large compared to the other catchments modelled, is still only a small fraction of the total flow from the Avon River. Figure 9.3 shows the Avon River producing around ten times as much streamflow as the other catchments.

In figure 9.3 (f) it is clear that the Helena River streamflow is strongly regulated by Mundaring Weir, since the model fails to capture most of the small releases from the weir. It does get the timing of the large releases correct though their magnitude may be understated. This error is not of paramount importance since the bulk of the study is focused on the comparative performance under different CO₂ conditions rather than absolute values.

Figure 9.4 shows how evapotranspiration changes with CMD given a unit temperature change (ie. a 1°C change) for each of the six catchments. The differences between these relationships can be explained principally in terms of vegetation and its distribution within the catchment, as well as catchment size. Two of the largest catchments, the Avon and Brockman Rivers, produce very little ET once the CMD has reached 50-60mm while for low CMD they display a high gradient and high saturation ET. This is fairly typical of catchments dominated by farming with forest remaining along the highest parts of the catchment. It has been well established that forests transpire at a significantly greater rate than grasslands (Holmes and Sinclair 1986). In Figure 9.4, the forest dominates the ET at first but since the forest only exists in the catchment highlands it is the first area to dry out, subsequently the lowland grassed areas begin to be the dominant ET source and a sharp fall occurs in the total ET.

The Wooroolo Brook and Susannah Brook catchments display a similar but less pronounced change in ET. This could be explained by the forest not being restricted to

the upland areas of the catchments as they were for the Avon and Brockman catchments, making the transition from forest dominated ET to grassland dominated ET less severe. The ET for Jane Brook displays the smallest gradient for high catchment moisture (low CMD); this is due to the relatively uniform land cover in the catchment as shown in Table 9.2. In particular there is no forest present and hence no transition from forest dominated to grass dominated ET. In fact the ET displays an almost linear fall with increasing CMD.

The Helena River streamflow is strongly affected by the presence of Mundaring Weir. Much of the ET losses come from the open water in the reservoir above the weir. As such, even as the higher forested parts of the catchment begin to dry out the ET losses remain high.

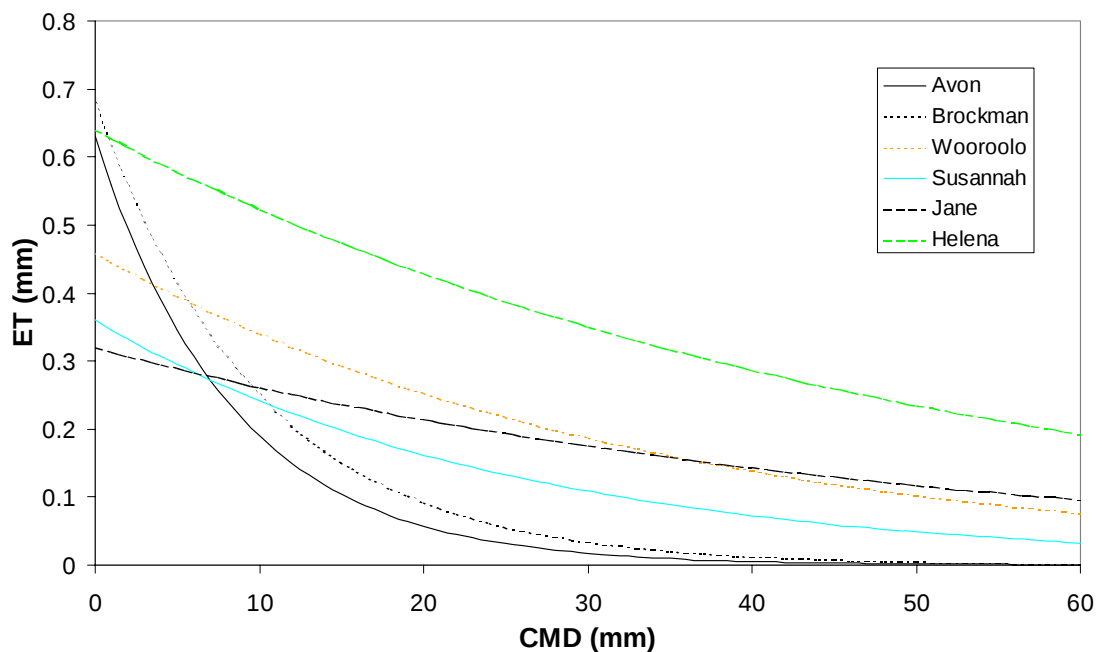


Figure 9.4 Variation in ET with Catchment Moisture Deficit given a unit temperature input.

9.5 Results

The models given in Table 9.3 were then used, in conjunction with the 1000 year precipitation and temperature records, to produce 1000 years of streamflow data under $1\times\text{CO}_2$, $1.5\times\text{CO}_2$ and $2\times\text{CO}_2$ conditions.

Figures 9.5 and 9.6 display properties of the temperature and precipitation time series under different CO_2 conditions. Clearly seen in Figure 9.5 is a fairly uniform increase in temperature with increasing CO_2 . In isolation, this increase in temperature would be expected to lead to an increase in evaporation, and hence a decrease in streamflow. Figure 9.6 (a) shows a slight increase in precipitation with increasing CO_2 across the broad range of precipitation occurrence. Figure 9.6 b) shows significant increases in the

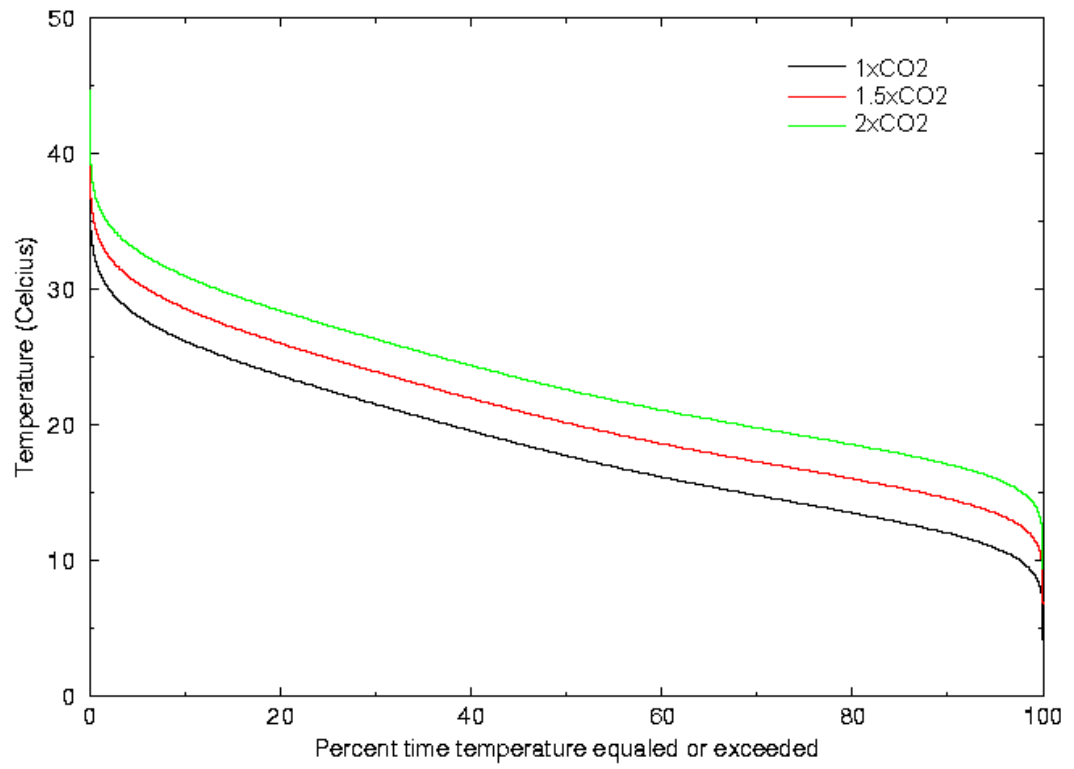


Figure 9.5 Temperature duration curves for 1xCO₂, 1.5xCO₂ and 2xCO₂ conditions

a)

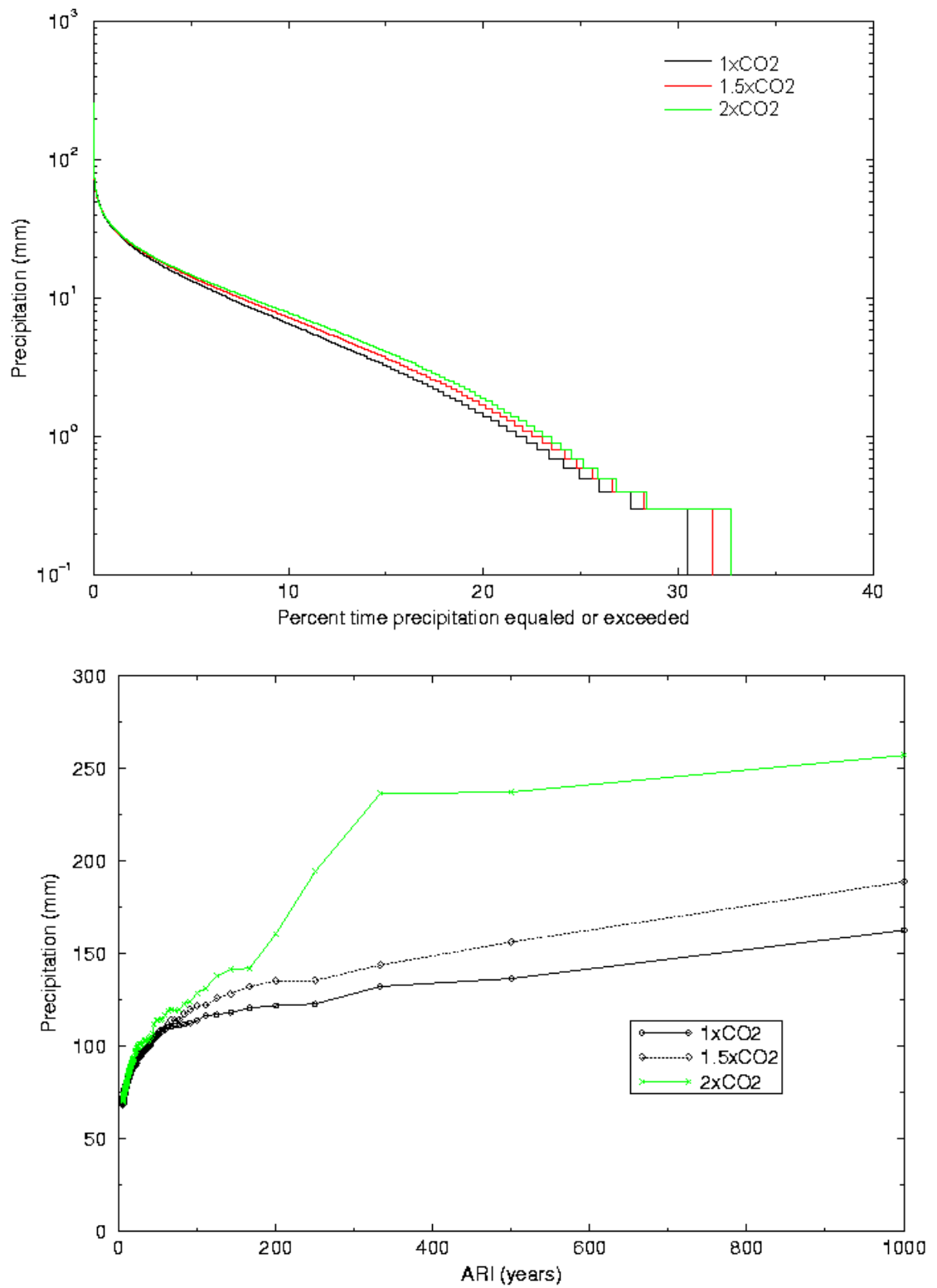


Figure 9.6 Precipitation under 1xCO₂, 1.5xCO₂ and 2xCO₂ conditions.

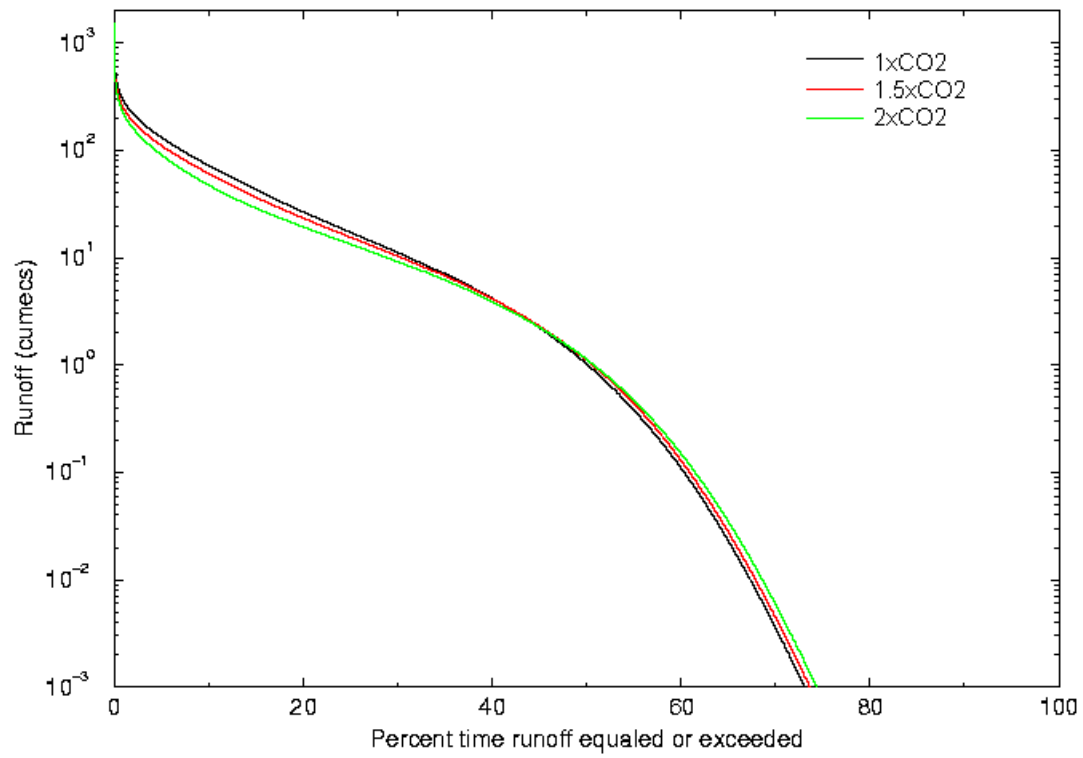
a) precipitation duration curve; **b)** average recurrence interval.

precipitation of the few largest events, ie. events with very large ARIs. In isolation, this change in precipitation would be expected to have little effect on the streamflow except for the few largest events when significant increases in streamflow could be expected.

Figure 9.7 shows flow duration curves for each catchment under the three CO₂ scenarios. In general, the potential impacts outlined above have occurred, with flow decreasing as CO₂ increases. The 1.5×CO₂ scenario streamflow is consistently lower than present streamflow conditions, while the 2×CO₂ scenario streamflow consistently produces even less streamflow. Deviations from this general outcome occur in the Brockman River where Figure 9.6 b) shows an increase in the lowest flows with increasing CO₂.

In Figure 9.8 ARI curves for all six catchments are shown. The magnitude of flood events under 1.5×CO₂ conditions is generally below those under 2×CO₂ conditions and almost always lower than present day conditions. For the Brockman and Helena Rivers, flood events under 2×CO₂ are always smaller than those under present day conditions. Wooroolo Brook and Jane Brook have only the rarest flood events under 2×CO₂ conditions being greater than those under current conditions, while Susannah Brook and the Avon River have larger flood events under 2×CO₂ conditions for ARIs greater than 125 years and 84 years respectively.

a)



b)

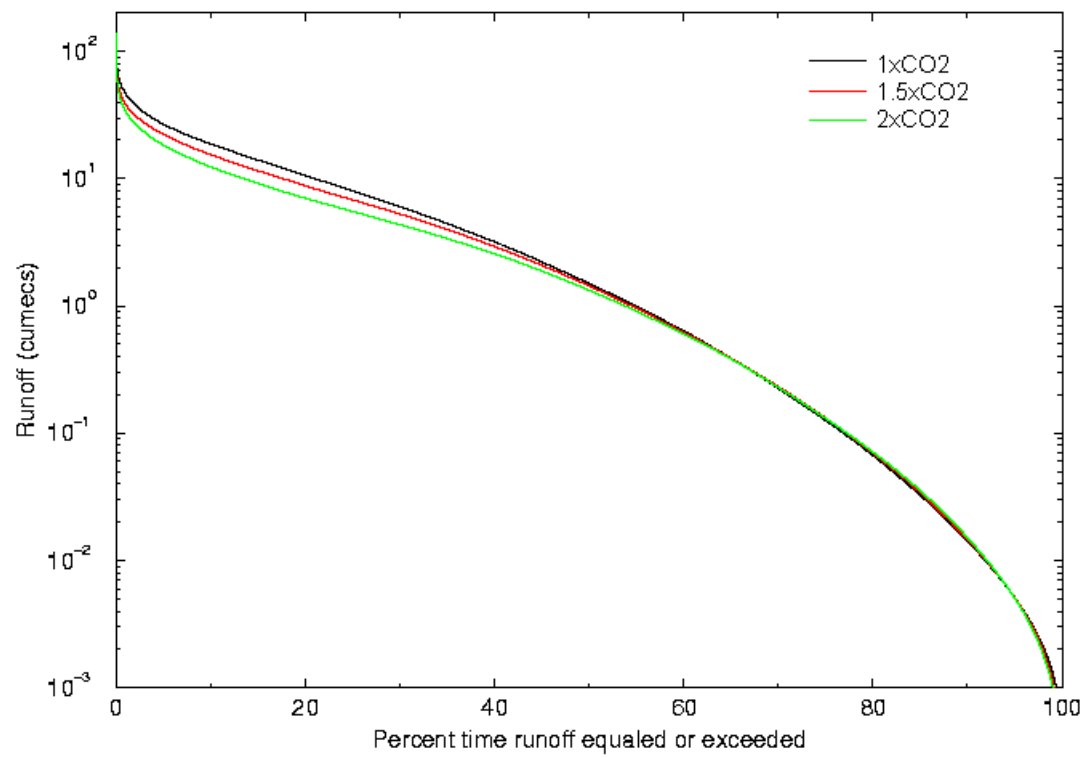
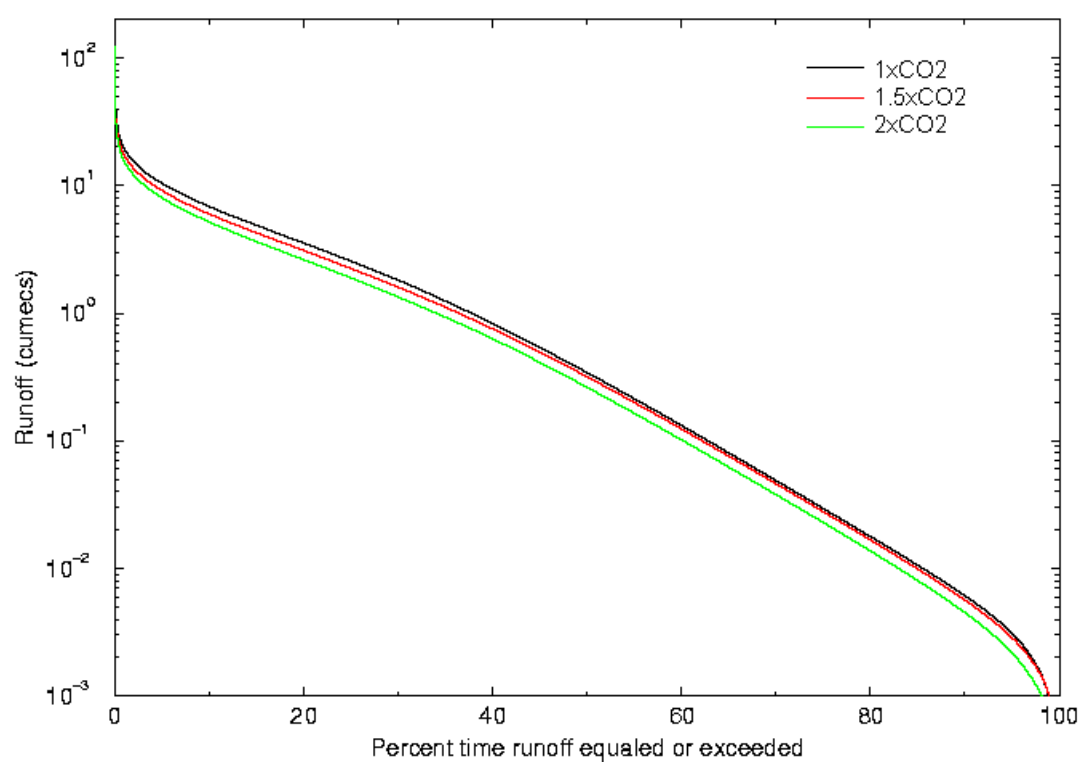


Figure 9.7 caption over page

c)



d)

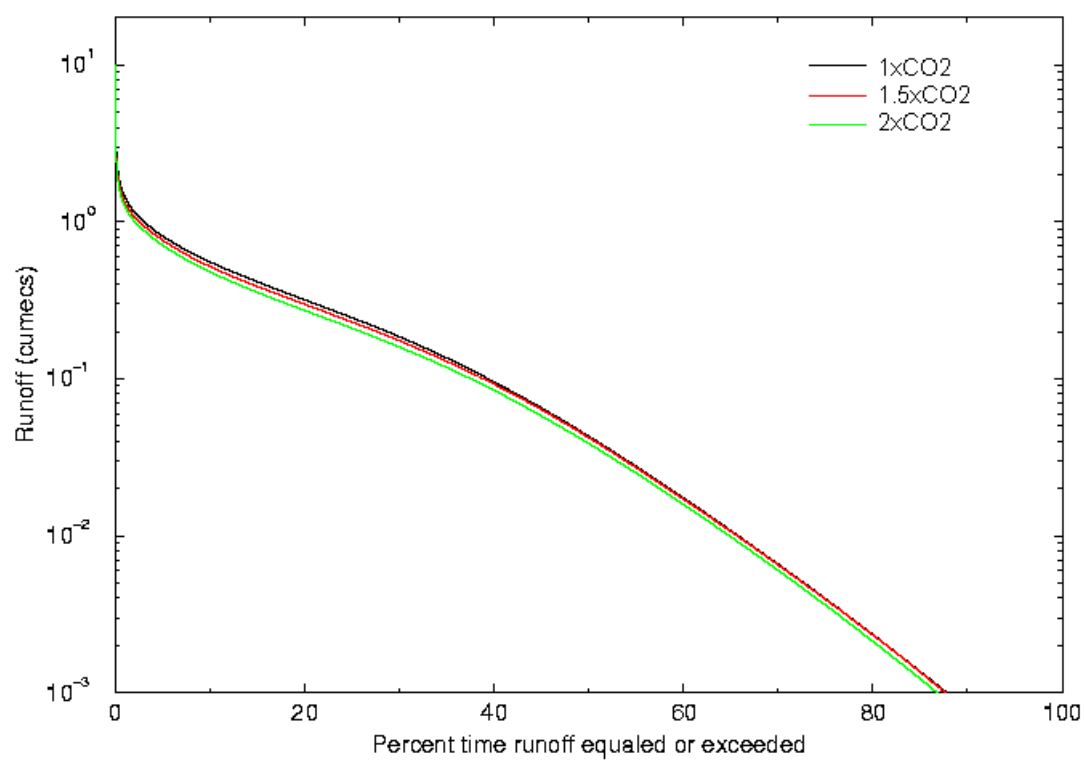


Figure 9.7 caption over page.

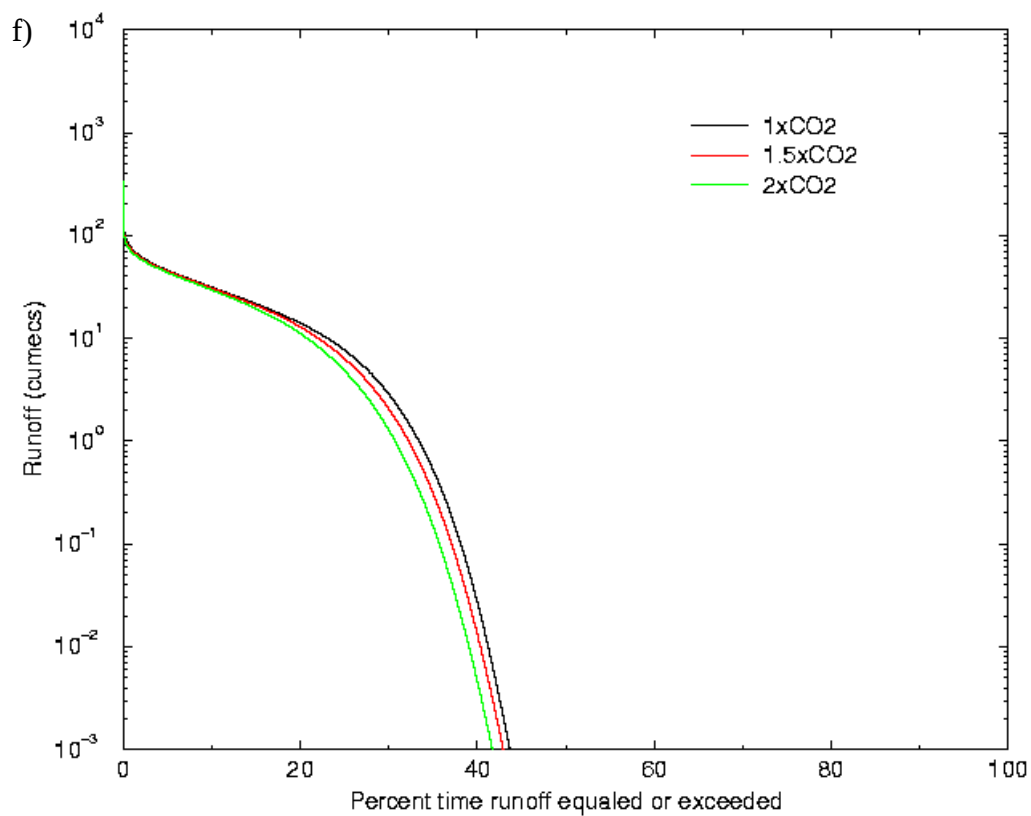
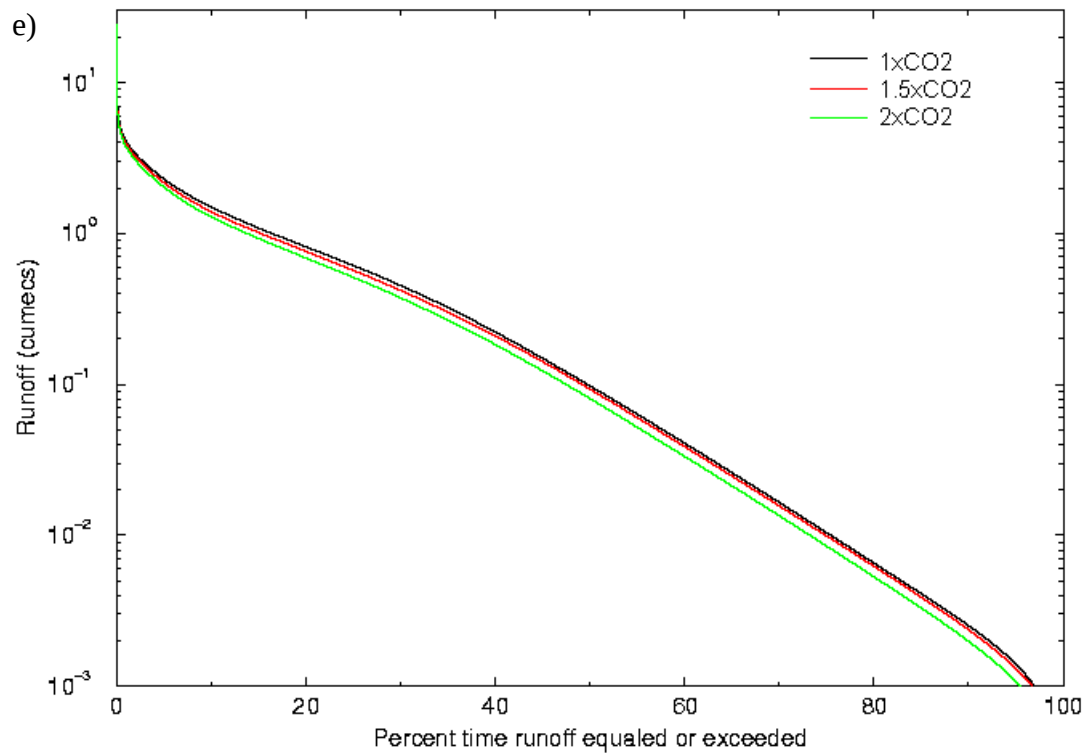
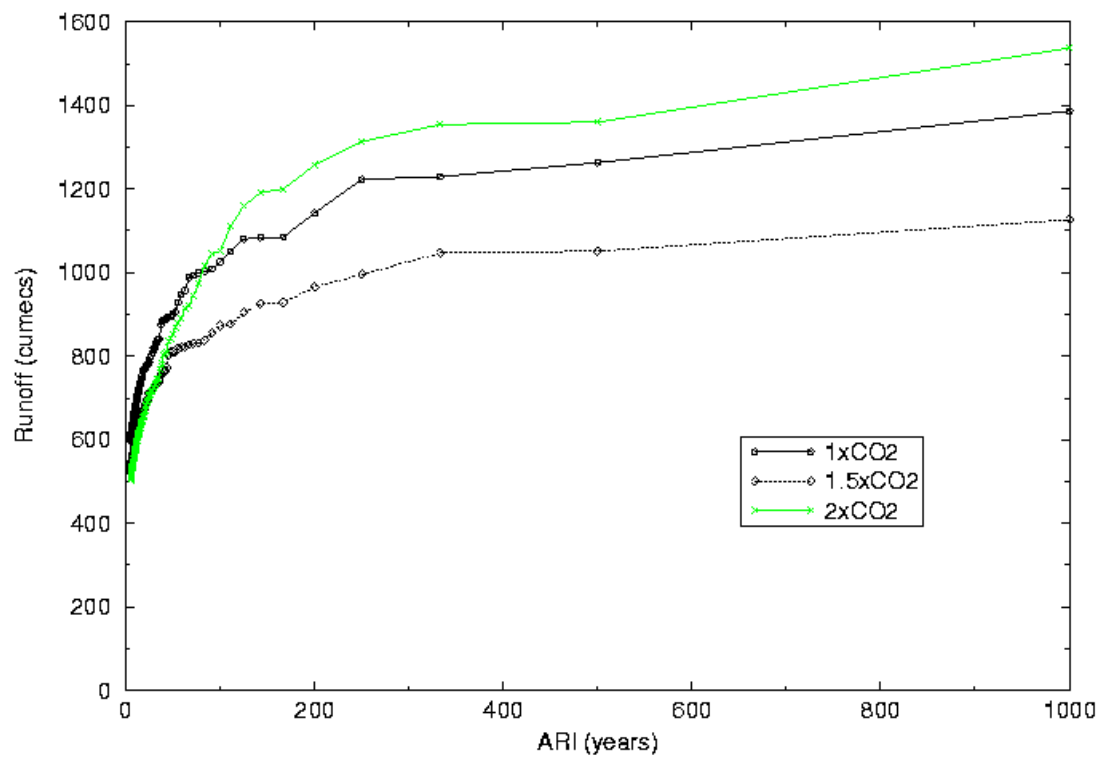


Figure 9.7 Flow duration curves for the three CO₂ scenarios: **a)** Avon River; **b)** Brockman River; **c)** Wooroolo Brook; **d)** Susannah Brook; **e)** Jane Brook; and **f)** Helena River.

a)



b)

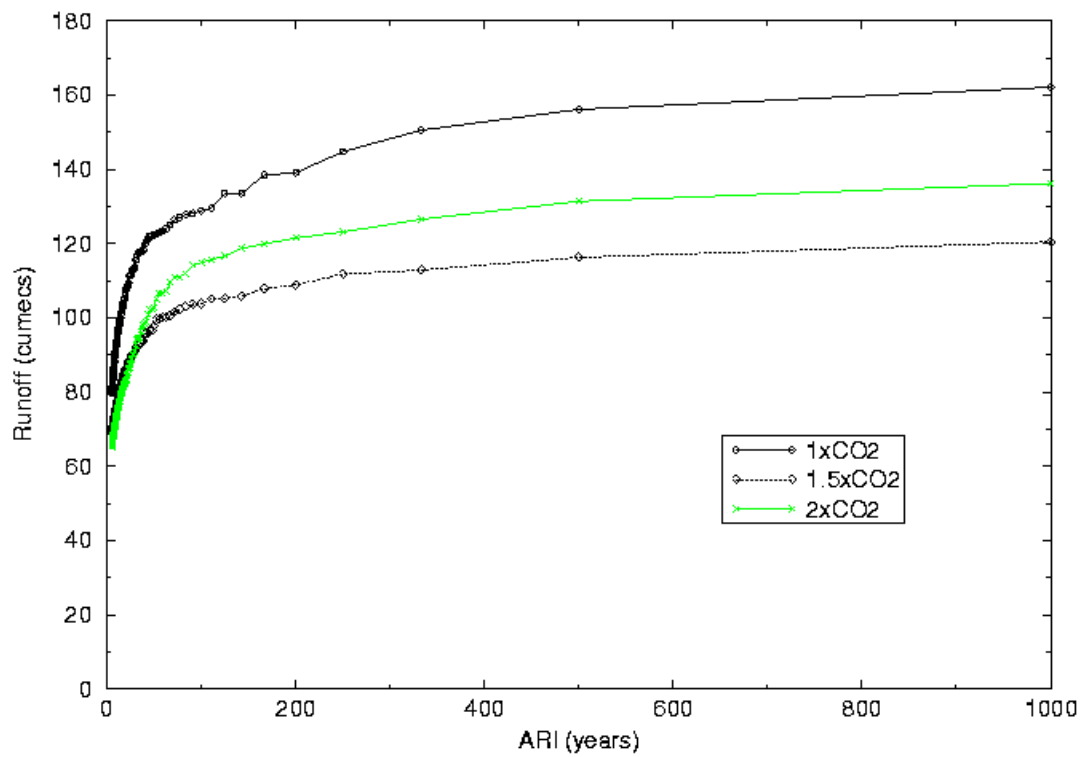
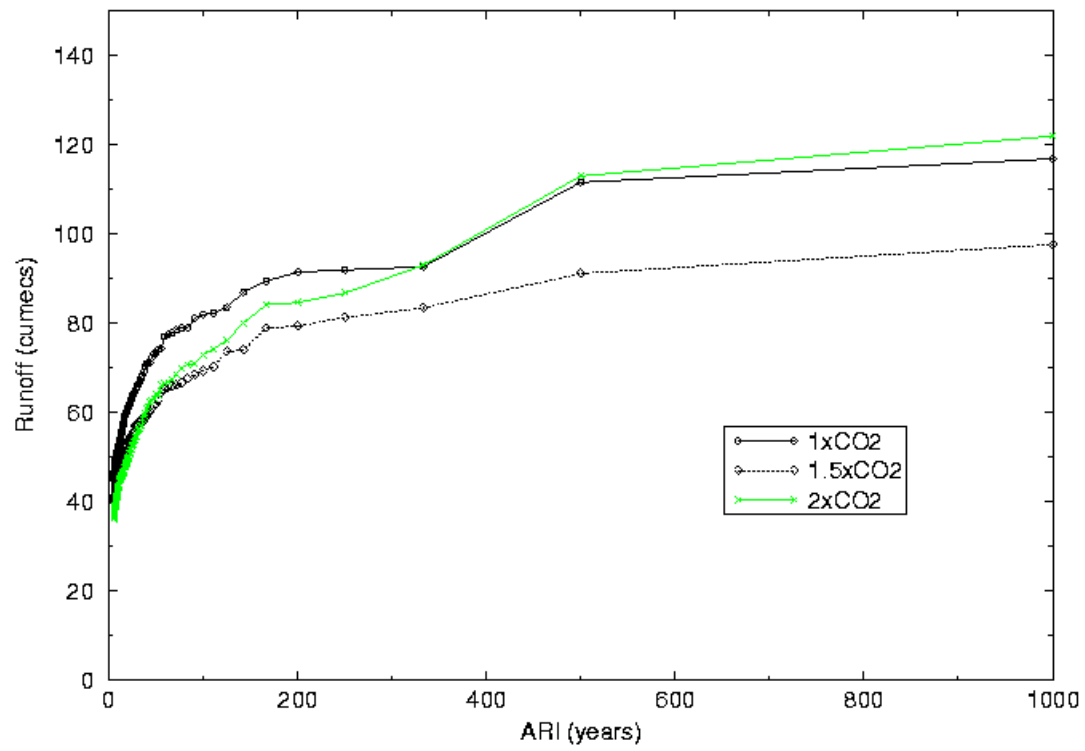


Figure 9.8 caption over page.

c)



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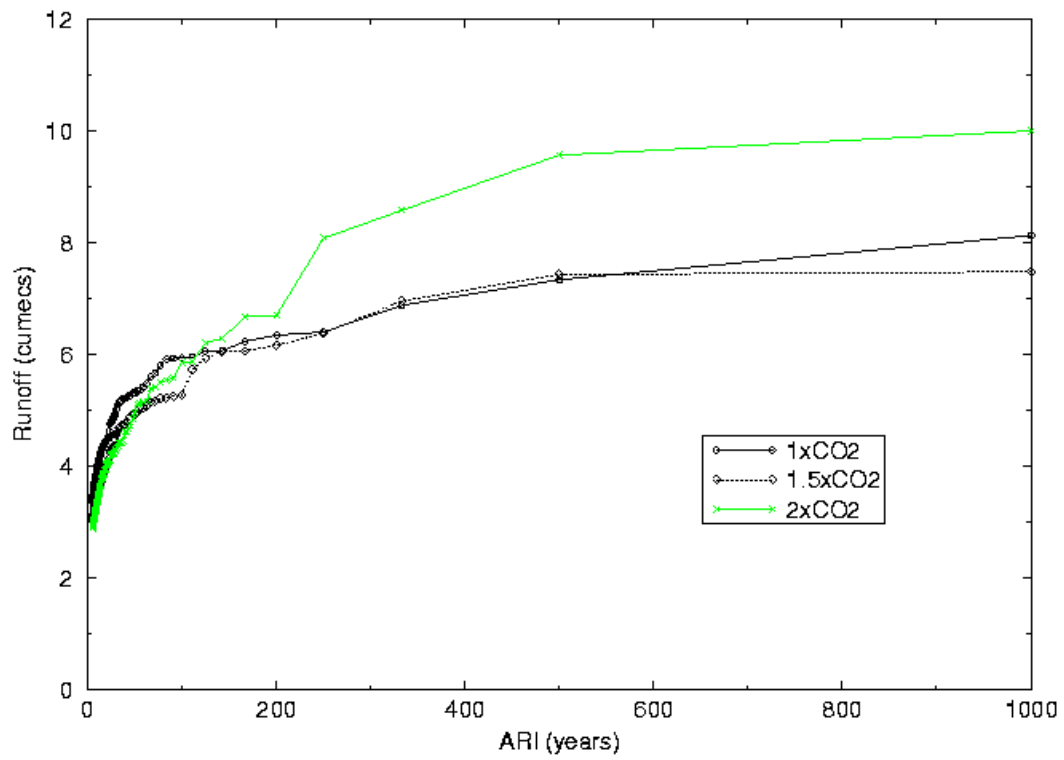


Figure 9.8 caption over page.

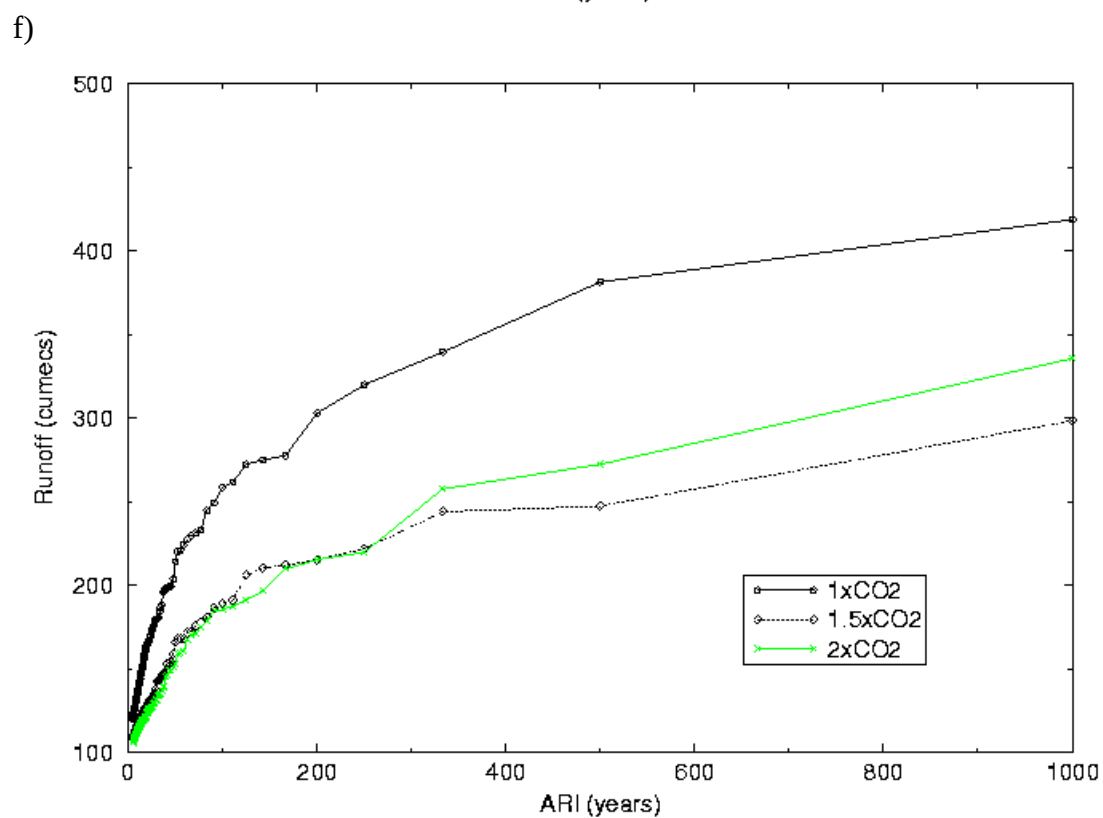
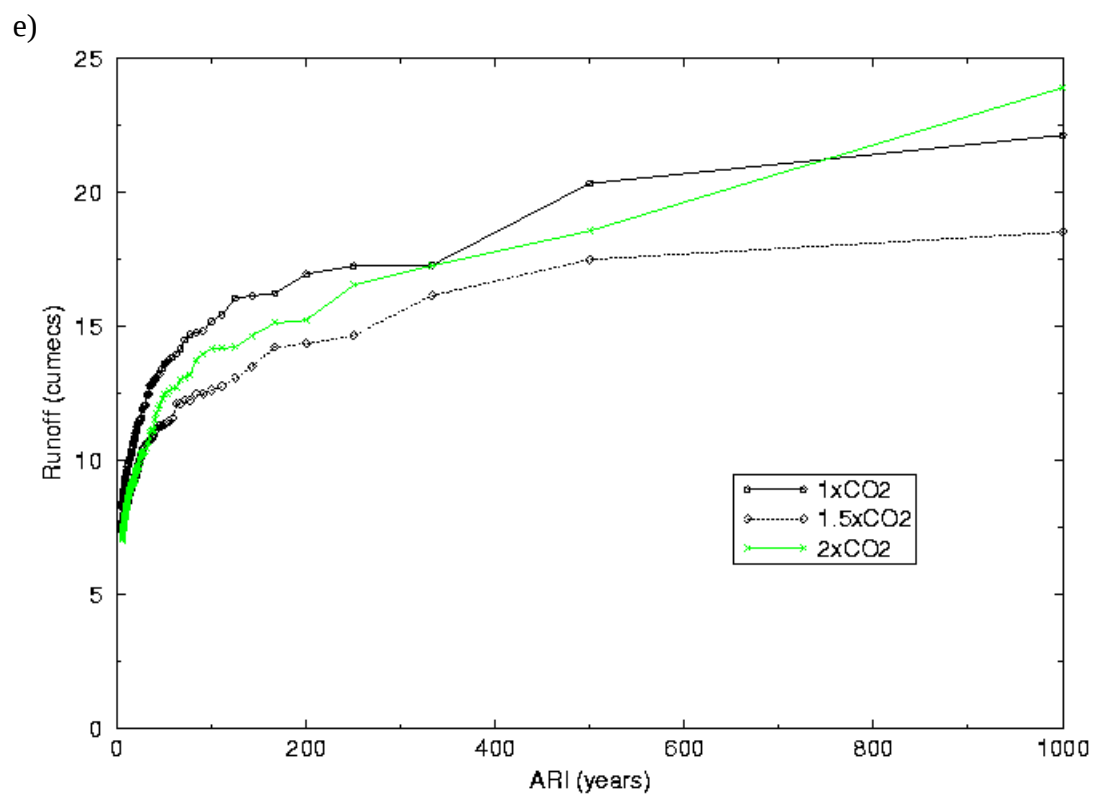


Figure 9.8 ARI curves for: **a)** Avon River; **b)** Brockman River; **c)** Wooroolo Brook; **d)** Susannah Brook; **e)** Jane Brook; and **f)** Helena River.

9.6 Discussion and Conclusions

Figure 9.8 provides the most insight into the changing flood regime with increasing CO₂ levels in the atmosphere. It demonstrates the influence of catchment landscape characteristics on the streamflow response changes with a significantly different response to global warming for each catchment. The two increased CO₂ scenarios can be thought of as one way of attempting to encapsulating the uncertainty in the global warming prediction, hence the associated streamflow and ARI curves, in some sense, capture the uncertainty in the streamflow response.

Figure 9.9 shows the ARI curves for the total flow entering the Swan River from these tributaries. It would seem that the most likely outcome of a period of global warming in this region will be a reduction in streamflows, even for the very rare flood events. The extent of this reduction is somewhat uncertain, however, it does not appear likely that the future flooding regime will be much more severe than under present conditions.

This outcome may be typical of relatively dry areas where current catchment moisture is rarely high enough to allow evaporation at the potential rate. Under globally warmed conditions any increase in precipitation will allow evaporation to reach this potential rate more often. It should also be noted that an increase in potential evaporation is associated with global warming. To account for this increase in evaporative demand precipitation would have to increase substantially. In the current study the precipitation does not increase by a substantial enough factor and streamflow levels in general decrease. Put simply, in the 1.5×CO₂ scenario the increase in evapotranspiration is considerably greater than any increase in precipitation thus leading to a reduction in the magnitude of the streamflow. While in the 2×CO₂ scenario the increase in evapotranspiration is matched by the increase in precipitation only for rare large precipitation events (ie. greater than 100 year ARI), however for more common events (less than 100 year ARI) the evapotranspiration has increased more than the precipitation leading to lower streamflows than observed under present day conditions.

Urban storm water planning in Australia often focuses on 100 year ARI levels (ACT government 1994). Figure 9.9 shows a definite fall in the magnitude of the 100 year ARI event. Using the two CO₂ scenarios to provide our error margins this fall is between 4% and 17%. The 1xCO₂ 100 year ARI event now falls between the 110 year ARI and 330 year ARI events under increased CO₂ conditions.

This implies that if the current storm water system in Perth is adequate for the present hydrological regime, it could prove adequate under globally warmed conditions. This may be good news for urban storm water system planners in the Perth area, but the general conclusion that dry areas may get drier could prove to be a much larger problem for issues associated with the water supply to the same urban area.

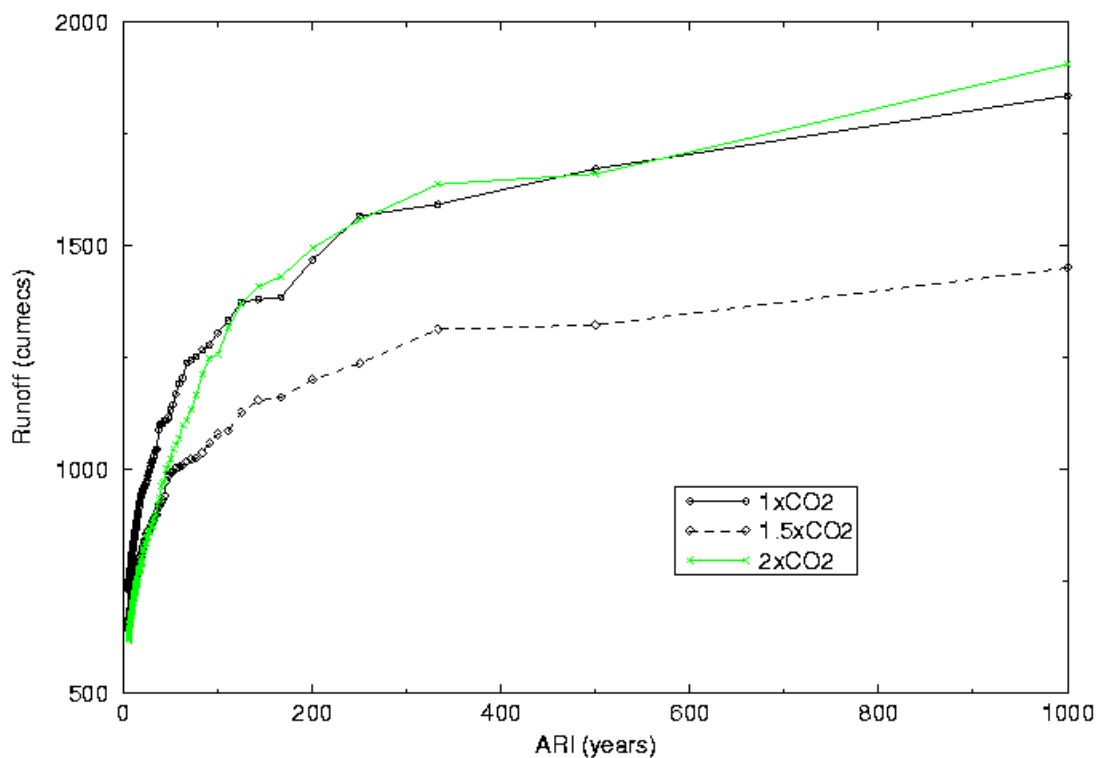


Figure 9.9 Average recurrence interval for flood events in the upper Swan River

Chapter 10

Conclusions

The main objective of this dissertation is the development of tools and methodology to perform assessments of the potential impact of climate change on the surface hydrological regime in data sparse areas. Currently there is significant uncertainty related to climate science and potential future global warming. This uncertainty, related to a large extent to uncertainties within climate models themselves, appears to be especially large in relation to land surface-atmosphere energy-water interactions, and the surface hydrology. Rainfall-runoff models on the other hand, can simulate the surface hydrology with reasonable confidence but they require time series of climatic variables to drive them. This dissertation demonstrates the potential for improvement in climate model simulations of surface hydrology by incorporating the metric/conceptual rainfall-runoff model CMD-IHACRES. The practical use of these models, along with a stochastic weather generator, for conducting climate change impact studies is also demonstrated.

Previous studies have attempted to improve the streamflow simulations of climate models using models and knowledge from hydrological research. Famiglietti and Wood (1991) developed a land-surface hydrology model in which the runoff response was controlled by variations in topography and soil properties, as characterised by the topography-soil index suggested by Beven and Kirkby (1979). Several others have used variations of the Nanjing model (Zhao 1977), including Dumenil and Todini (1992) and Wood et al. (1992), to provide the runoff formulation. In each case they found the runoff simulation was an improvement over that produced by the bucket model of Budyko (1956). However these studies did not always make comparisons with observations, and when they did it was on time-scales of a month or longer. Since many hydrological events of interest, such as floods, occur on time-scales much shorter than a month these previous studies fail to address issues of significant importance to water resource planners. Most of the work in this dissertation has been performed on a daily time-step, with the potential to be used on sub-daily time-steps, thus maintaining greater applicability to water resource planning as well as to the short time-step requirements for feedback of evapotranspiration responses to climate models.

10.1 Streamflow Modelling

A review of the current state of rainfall-runoff modelling reveals several factors which must be considered when choosing an appropriate rainfall-runoff model for a particular purpose. These include the output data required from the model, the input data required by the model, the level of parameterisation within the model, whether the model is conceptual or physically based, lumped or distributed, and that the model is suitable for the temporal and spatial scale of interest. For the purposes of this dissertation the rainfall-runoff model was required to output both streamflow and evapotranspiration at catchment scale, on a daily time step, with the possibility of using shorter time steps in conjunction with climate models. The model was restricted to input data requirements of precipitation, temperature and streamflow. This rather severe input data requirement means that only a handful of parameters can be used before over-parameterisation begins to reduce the confidence in a models results. This over-parameterisation problem also limits model selection to be of the lumped conceptual type as physically-based and/or distributed models generally require significantly more parameters. While no rainfall-runoff model could be found which satisfied all the above criteria, the rainfall-runoff model developed by Jakeman and Hornberger (1993) and Jakeman et al. (1994), IHACRES, satisfied all but the output of evapotranspiration at the appropriate time scale. This model was chosen as the basis for further development such that the model would output evapotranspiration as well.

In order to develop an evapotranspiration formulation for use in the rainfall-runoff model, a review of current evapotranspiration modelling was conducted. It should be noted that significant uncertainty is still associated with the modelling of evapotranspiration. This uncertainty can be at least partially explained by the difficulties involved in attempting to measure evapotranspiration and hence validate any model of it. This measurement difficulty is exacerbated by the fact that evapotranspiration is generally modelled as an areal average even though the spatial heterogeneity present means that it is unclear how to construct such an average (Raupach 1995). Evapotranspiration is usually modelled by estimating the evapotranspiration which would occur under the current conditions given no water availability restrictions at the surface. This is referred to as potential evapotranspiration. Once the potential

evapotranspiration has been established it is then adjusted to provide an estimate of the actual evapotranspiration. This adjustment can be made in two ways: the potential evapotranspiration can be considered limited by the available soil moisture (Monteith 1965) or; the potential evaporation is thought to display a complementary relationship to actual evapotranspiration around a pre-defined equilibrium evapotranspiration (Bouchet 1963).

Use of the complementary relationship requires significantly more data than precipitation and temperature, which is the only input data available for the model. Thus the evapotranspiration is modelled by adjusting the potential evapotranspiration according to the available soil moisture. This method has the advantage of requiring soil moisture accounting, which is also important for the modelling of streamflow, and is the most commonly used method in climate models. Unfortunately, potential evapotranspiration formulations which are applicable on a daily to sub-daily basis such as Penman (1956), Priestley and Taylor (1972) and Monteith (1981), also require significantly more data than the precipitation and temperature which is available. As a result a new method for determining the potential evapotranspiration was developed that depends only on the temperature and the available soil moisture.

Development of the rainfall-runoff model centered around the catchment moisture store accounting scheme. This scheme performs soil moisture accounting in terms of the catchment moisture deficit (CMD). The CMD is a measure of how far from saturation the soil is. CMD is zero when the catchment is saturated and increases as the catchment dries out. Using CMD has the advantage of not requiring a parameter to represent the maximum moisture a catchment can hold. The model estimates evapotranspiration using the modified temperature approach mentioned above. The model uses a maximum of seven parameters and is referred to as CMD-IHACRES.

CMD-IHACRES was successfully tested and validated on several catchments covering a wide range of hydroclimatologies. It was found to perform well in both humid and semi-arid areas with streamflow prediction being as good, if not better, than the original IHACRES model. Testing of the evapotranspiration formulation reveals that while the modified temperature approach is able to model the overall trend in the data, it does not

capture the day-to-day variation well. More complex formulations such as Priestley and Taylor (1972) or Penman (1956) are much better at capturing this variation, suggesting that when data availability allows the use of one of these more complex formulations is advisable.

Given that practical use of CMD-IHACRES requires the calibration of parameters against streamflow data, a question remains about the validity of the predicted evapotranspiration time series. To investigate this the model was calibrated using both evapotranspiration data (used to calibrate the parameters in equation 4.9) and streamflow data (used to calibrate the rest of the parameters), as well as being calibrated using streamflow data alone, and the results compared. Calibration using both evapotranspiration and streamflow data provided the best estimate of evapotranspiration while the streamflow displayed some initial discrepancies which lasted around 20 days. Similarly, calibration against streamflow data alone produced the best estimate of streamflow while evapotranspiration estimates were not biased substantially after an initial period of around 40 days.

10.2 Climate modelling

The climate system is made up of many complex, interacting processes. In attempting to model this system, climate models themselves have become very complex. They contain many parameters but do not rely on site specific data for calibration, instead they depend on several global datasets to determine their parameter values. Several intercomparison studies have been conducted using global climate models, which indicate uncertainties within, and discrepancies between the models. Regional climate models (or limited area models) on the other hand, have had relatively few intercomparison studies performed. Since regional climate models are a significant focus of this dissertation a clearer picture of the model uncertainties was required. In order to obtain a quantitative feel for these uncertainties an intercomparison experiment was performed using three regional climate models: MM5/BATS, MM5/SHEELS and RegCM2.

First, the surface energy budget terms simulated by the models were compared. In terms of the net incident energy all three models are within 15% of the observed total. In particular the MM5 based models overestimate the downward longwave radiation. RegCM2 comes closest to reproducing the observed net incident radiation on a daily basis. All three models produce reasonable simulations of daily latent heat in the normal/wet year, however they significantly underestimate this in the dry year. The sensible heat flux is simulated relatively poorly by all three models with the worst agreement again occurring in the dry year. The observations indicate a strong seasonality in the net surface heating which the models do not capture. However the observations were insufficient to allow comparison with a full annual cycle such that the true magnitude of this seasonality is unknown. In terms of the overall energy balance there was agreement among the models and observations in the normal/wet year and considerable disagreement in the dry year. In almost all the surface energy related variables investigated the models simulate more variance than is observed, in particular the MM5 based models. This indicates that the modelled ground heat and moisture stores are able to change much faster than is observed. In the overall simulation of the energy balance RegCM2 is the best performer of the regional climate models.

Secondly, the surface water budget terms simulated by the models were compared. While none of the models demonstrate very much skill in reproducing daily values of precipitation, RegCM2 is best able to reproduce the observed annual totals. In terms of the evapotranspiration, the MM5 based models simulate much larger variance than is observed on a daily basis, as well as a much larger change in the evapotranspiration regime between the normal/wet year and the dry year. The runoff simulated by the models exhibits the largest relative error of all the variables investigated, indicating that the runoff formulation is in urgent need of improvement. Looking at changes in the soil moisture again reveals that the models display too much variance, however observations were taken at irregular intervals making direct comparison difficult. In terms of the overall water balance there was general agreement amongst the models and observations in the dry year, while there is considerable disagreement in the normal/wet year. None of the three models performs better in the simulation of all the water fluxes though MM5/SHEELS does perform better in terms of the overall water balance.

In comparing these regional climate models it was found that under dry conditions energy balance related parameterisations were responsible for the majority of the differences between the models, while under wet conditions water balance related parameterisations began to dominate these differences.

10.3 Climate – Surface hydrology interactions

In an effort to improve the runoff simulation in the climate models, which was found to be particularly poor, the effect of including CMD-IHACRES as the runoff component in the three regional models was investigated. Offline simulations indicate that significant improvements in runoff are obtained through the use of CMD-IHACRES. That is, incorporating CMD-IHACRES into regional climate models improves not only the daily estimates of streamflow but also the characteristics of the runoff time series as revealed in flow duration curves and double mass plots, along with improvements in the seasonal cycle. Comparison of two non-dimensional quantities that are indicative of the interplay between evapotranspiration and runoff again reveals significant improvement when CMD-IHACRES is included with the regional climate models. In particular, the combination of RegCM2 and CMD-IHACRES almost precisely reproduces the observed values of these parameters. This suggests that the combination of a CMD-IHACRES type of model and a regional climate model may well prove to be of practical use in investigating climate change effects on streamflows in data sparse areas. These results suggest that further experiments with CMD-IHACRES run on-line with a LAM are warranted.

Finally, the use of CMD-IHACRES to perform an assessment of the hydrological impact of climate change was demonstrated in a case study near Perth, Western Australia. In this study, climate change scenarios for use with CMD-IHACRES were developed using a combination of output from a global climate model (CSIRO9) and a stochastic weather generator (WGEN). CMD-IHACRES was then run on the six major tributaries of the Swan River, which flows through Perth. The study demonstrates that

although the same climatic scenarios were used for all the catchments, varying catchment attributes leads to significant differences in catchment response to the climate change scenarios. Overall, a fall in the magnitude of runoff events is the most likely outcome in this region under globally warmed conditions. For example, the magnitude of the one in one hundred year flood is expected to decrease by between 4% and 17%.

10.4 Future work

Several avenues of future research have become apparent through the course of this dissertation. Further development of CMD-IHACRES is required, particularly in respect to the form of the discharge equation (4.10). While it is not clear what an effective rainfall time series should look like, this could be estimated using a statistical model such as DBM (Young 1993), thus obtaining a statistically ‘optimal’ effective rainfall time series. This could then be used to validate various discharge formulations. Further study into the implications of calibrating on a daily timestep and then simulating on shorter timesteps is also required to facilitate the on-line incorporation of CMD-IHACRES into a regional climate model.

Further investigation into and development of the regional climate models are also clearly warranted. In particular the longwave radiation scheme in MM5 and the runoff parameterisations in all the regional climate models need attention.

Treating the climate model-simulated runoff as effective rainfall and including the linear component of CMD-IHACRES to convert this to runoff appears to show promise. This still requires the estimation of the three parameters in the linear component of CMD-IHACRES. This could be achieved in three ways: the non-linear component of CMD-IHACRES could be changed such that it produces an effective rainfall time series in a similar fashion to the regional climate model, the CMD-IHACRES could be calibrated as before to determine the parameter values; the parameters could be related to landscape attributes, in the style of (Post 1996), in order to determine them without

calibration; or these parameter values could be determined by direct analysis of the streamflow time series (Clarke 1996).

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