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LAND SURFACE PROCESSES IN LARGE-SCALE HYDROLOGY

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CONTENTS

		Page
Foreword		V
Summary (English, French, Russian and Spanish)		VII
List of acronyms		IX
CHAPTER 1 — GENERAL INTRODUCTION		1
1.1	Objective and outline of the report	1
1.2	Climatic change and hydrological processes at the land surface	2
CHAPTER 2 — GENERAL CIRCULATION AND MESOSCALE CLIMATIC SYSTEMS		3
2.1	Introduction	3
2.2	General circulation aspects	3
2.3	Mesoscale climatic systems	4
2.3.1	Synoptically-induced mesoscale climate systems	4
2.3.2	Terrain-induced mesoscale climate systems	4
2.3.3	Rainfall	5
CHAPTER 3 — PHYSICAL CLIMATE MODELS		7
3.1	Introduction	7
3.2	Atmospheric general circulation models	7
3.3	Numerical mesoscale models	9
3.4	Embedded models	10
3.5	Model output enhancement	11
3.6	Predictability	11
CHAPTER 4 — ESTIMATION OF AREAL EVAPORATION		13
4.1	Evaporation/atmosphere interactions	13
4.2	Estimation methods for areal evaporation	13
4.2.1	The combination equation	14
4.2.1.1	Stomatal conductance models	14
4.2.1.2	Interception models	15
4.2.1.3	Application of the Penman-Monteith equation: scaling up	16
4.2.1.4	Simulation of soil moisture deficit	17
4.2.2	Priestley-Taylor expression	18
4.2.3	Complementary relationship	18
4.2.4	Atmospheric boundary layer methods	18
4.2.4.1	Bulk transfer method	19
4.2.4.2	Mixed layer models	19
CHAPTER 5 — USE OF REMOTE SENSING IN THE STUDY OF LAND SURFACE PROCESSES AT A REGIONAL SCALE		21
5.1	Introduction	21
5.2	Rainfall	21
5.3	Soil moisture	22
5.4	Evaporation	23
5.5	Radiation balance components	24
5.6	Land cover and biomass	25
CHAPTER 6 — MODELLING OF LAND SURFACE PROCESSES		26
6.1	Introduction	26
6.2	Simple hydrology models	26
6.3	Parameterization of land surface processes	28
6.3.1	Overview of current models	29
6.3.2	Land surface parameterization and subgrid-scale variability	31
6.4	Scale issues in large-scale hydrology: the drainage basin perspective	34

Page

CHAPTER 7 — INTERNATIONAL PROGRAMMES AND MAJOR FIELD EXPERIMENTS ON LAND SURFACE PROCESSES IN LARGE-SCALE HYDROLOGY 37
7.1 Major programmes 37
7.1.1 Global Energy and Water Cycle Experiment (GEWEX) 37
7.1.2 International Geosphere-Biosphere Programme (IGBP) 37
7.2 Recent and current international field experiments 37
7.3 Future field experiments 37

CHAPTER 8 — CONCLUDING REMARKS 39

Acknowledgements 41

References 41

APPENDIX I — Current international programmes on land surface processes of the hydrological cycle 57

APPENDIX II — Recent international field experiments on land surface processes 59

APPENDIX III — Current international field experiments on land surface processes 60

FOREWORD

The WMO Commission for Hydrology (CHy), at its eighth session in 1988, appointed Messrs J. D. Kalma (Australia) and I. R. Calder (UK) as joint Rapporteurs on Hydrological Interactions at the Land Surface. Their terms of reference included the preparation of a report on the potential interaction between climate and hydrological processes at the land surface.

The Commission also established a Working Group on Operational Hydrology, Climate and the Environment. Members of this Working Group included, among others, a Rapporteur on Hydrological Modelling for Climate Studies and a Rapporteur on Operational Hydrology and Climate Change. Considering that these rapporteurs had been asked to address the macroscale impact of climate change and macroscale interactions between climate and hydrological processes, it was resolved that the present report should address the interactions/coupling between land surface hydrological processes and climate. Clearly, some overlap exists between these reports, but this may serve as a valuable link between different perspectives.

I should like to place on record the gratitude of the World Meteorological Organization to Messrs Kalma and Calder for the time and effort they have devoted to the preparation of such an excellent report on what is a very complex subject. A word of appreciation is also due to all others who so willingly contributed in one way or another in the preparation of the report, including Messrs T. J. Lyons, N. Nunez and A. J. Pitman of Australia and C. J. Vörösmarty of the United States.

G.O.P. Obasi
(Secretary-General)

SUMMARY

This report addresses the major interactions between climate and land surface processes on a large scale. Particular emphasis is given to surface hydrology and to the role of vegetation in the context of global change. It reviews relevant recent work, provides an overview of various research strategies and defines important research areas.

The report consists of seven main chapters, three appendices and a list containing over 460 references. The first chapter introduces the subject matter. The second and third chapters consider, respectively, the general circulation and mesoscale climatic systems, and the physical models which are used in studying dynamic aspects of these systems. Chapter 4 reviews various estimation methods for areal evaporation. Chapter 5 describes recent advances in the use of satellite remote sensing in the study of land surface processes at a regional scale, while Chapter 6 reviews developments in land surface parameterization for climate modelling. In Chapter 7 and in Appendices I, II and III, an overview is given of major international research programmes and field experiments which aim at a better understanding of critical and surface processes and at improving their parameterization for large-scale hydrology and climate modelling.

RÉSUMÉ

Ce rapport porte sur les principales interactions, à une grande échelle, entre le climat et les processus de sol. L'hydrologie et le rôle de la végétation, considérés du point de vue de l'évolution du climat y occupent une place privilégiée. En outre, l'auteur se réfère aux travaux menés récemment dans ce domaine, passe en revue les différentes stratégies en matière de recherche et dégage les grands thèmes à retenir à cet égard.

Le rapport contient sept chapitres, trois appendices et une liste de plus de 460 références. Le premier chapitre constitue une entrée en matière. Les deuxième et troisième chapitres traitent, respectivement, de la circulation générale et des systèmes climatiques de méso-échelle ainsi que des modèles physiques utilisés pour étudier les aspects dynamiques de ceux-ci. Le chapitre 4 est consacré aux différentes méthodes d'évaluation de l'évaporation zonale. Le chapitre 5 retrace les dernières réalisations dans l'utilisation de la télédétection par satellite pour d'étude des processus de sol à l'échelle régionale, alors que le chapitre 6 fait état des progrès accomplis dans la paramétrisation relative aux phénomènes de sol pour la modélisation du climat. Le chapitre 7 ainsi que les appendices I, II et III donnent un tour d'horizon des principales actions menées sur le plan international — programmes de recherche et expériences sur le terrain — destinées à apporter une meilleure compréhension des processus de sol critiques et à améliorer leur paramétrisation aux fins de la modélisation du climat et de l'hydrologie à grande échelle.

КРАТКОЕ РЕЗЮМЕ

В настоящем докладе рассматриваются основные взаимосвязи между климатом и процессами на поверхности суши в крупном масштабе. Особый акцент делается на гидрологии поверхностных вод и роли растительного покрова в контексте глобального изменения. В докладе рассматриваются соответствующие последние работы и содержится обзор различных стратегий исследования, а также описываются важные области исследований.

Доклад включает в себя семь основных глав, три приложения и перечень, содержащий свыше 460 ссылок. Первая глава представляет собой введение в предмет исследования. Во второй и третьей главах соответственно рассматриваются общая циркуляция и мезомасштабные климатические системы, а также физические модели, которые используются при изучении динамических аспектов этих систем. В четвертой главе приводится описание различных методов оценки испарения по площади. Пятая глава описывает последние достижения в использовании дистанционного зондирования со спутников при изучении процессов на поверхности суши в региональном масштабе, в то время как в шестой главе рассматриваются достижения в области параметризации поверхности суши при моделировании климата. Глава 7 и приложения I, II и III представляют собой обзор крупных международных программ научных исследований и полевых экспериментов, направленных на более глубокое понимание критически важных процессов на поверхности суши и улучшение их параметризации для крупномасштабной гидрологии и моделирования климата.

RESUMEN

El presente informe versa sobre las principales interacciones entre el clima y los procesos de la superficie terrestre a gran escala. Se hace hincapié, sobre todo, en la hidrología de superficie y la función de la vegetación en el contexto del cambio global. Examina los trabajos pertinentes más recientes, ofrece una visión general de las diversas estrategias de investigación y determina importantes áreas en esta materia.

El informe está dividido en siete capítulos principales, tres apéndices y una lista con más de 460 referencias. En el primer capítulo se trata el tema propiamente dicho. El segundo y tercer capítulos examinan, respectivamente, la circulación general y los sistemas climáticos mesoescalares y los modelos físicos que se utilizan en el estudio de los aspectos dinámicos de estos sistemas. El capítulo 4 contiene varios métodos de estimación de la evaporación zonal. En el capítulo 5 se describen los últimos adelantos en el uso de la teledetección por satélite en el estudio de los procesos de la superficie terrestre a escala regional, mientras que en el capítulo 6 se analizan los logros en materia de parametrización de la superficie terrestre para la modelización climática. En el capítulo 7 y en los Apéndices I, II y III, se ofrece un panorama de los principales programas internacionales de investigación y experimentos prácticos con el fin de conocer mejor los procesos fundamentales de la superficie terrestre y de mejorar su parametrización para la modelización hidrológica y climática a gran escala.

LIST OF ACRONYMS

ABL	Atmospheric Boundary Layer
AET	Actual Evapotranspiration
AGCM	Atmospheric General Circulation Model
AMIP	Atmospheric Model Intercomparison Project
AVHRR	Advanced Very High Resolution Radiometer
AWC	Available Water Capacity
BAHC	Biospheric Aspects of the Hydrological Cycle
BATS	Biosphere – Atmosphere Transfer Scheme
BOREAS	Boreal Ecosystem – Atmosphere Study
CBL	Convective Boundary Layer
CCD	Cold Cloud Duration
CCM	Community Climate Model
CHy	Commission for Hydrology
CSIRO	Commonwealth Scientific and Industrial Research Organization
DEM	Digital Elevation Model
EFEDA	European Field Experiment in Desertification Threatened Area
ENSO	*El Niño*/Southern Oscillation
FAO	Food and Agriculture Organization
FIFE	First ISLSCP Field Experiment
GCIP	GEWEX Continental-scale International Project
GCTE	Global Change and Terrestrial Ecosystems
GEWEX	Global Energy and Water Cycle Experiment
GFDL	Geophysical Fluid Dynamics Laboratory
GHM	General Hydrologic Model
GISS	Goddard Institute for Space Studies
GLA	Goddard Laboratory for Atmosphere
GMCSS	GEWEX Multiregion Cloud System Study
GMS	Geostationary Meteorological Satellite
GOES	Geostationary Operational Environmental Satellite
GPCP	Global Precipitation Climatology Project
GRDP	Global Runoff Data Project
HAPEX	Hydrological Atmospheric Pilot Experiment
ICSU	International Council of Scientific Unions
IGBP	International Geosphere-Biosphere Programme
ISLSCP	International Satellite Land Surface Climatology Project
ITCZ	Inter-tropical Convergence Zone
JPL	Jet Propulsion Laboratory
MOBILHY	Modelisation du bilan hydrique
MSS	Multispectral Scanner
NASA	National Aeronautics and Space Administration
NCAR	National Center for Atmospheric Research
NDVI	Normalized Difference Vegetation Index
NMM	Numerical Mesoscale Model
NOAA	National Oceanic and Atmospheric Administration
NWP	Numerical Weather Prediction
OHCE	Operational Hydrology, Climate and the Environment
PBL	Planetary Boundary Layer
PET	Potential Evapotranspiration
PILPS	Project for Intercomparison of Land Surface Parameterization Schemes
RAMS	Regional Atmospheric Modelling System
SiB	Simple Biosphere Model
SMMR	Scanning Multi-frequency Microwave Radiometer
SPOT	Satellite probatoire d'observation de la Terre
SST	Sea-surface Temperature
SVAT	Soil Vegetation Atmosphere Transfer
TM	Thematic Mapper
WCRP	World Climate Research Programme
WMO	World Meteorological Organization

CHAPTER 1

GENERAL INTRODUCTION

1.1 Objective and outline of the report

Enormous progress has been made in the past decade in global climate modelling and in improving our understanding of global change processes. However, there has also been the realization that there are critical gaps in our knowledge about the coupled Earth-atmosphere system. Studies at the local scale have yielded considerable insight into hydrological processes and nutrient biochemistry. However, there is still a paucity of models capable of describing the distributed nature of land surface hydrological processes. Moreover, techniques still remain to be developed to extrapolate sensibly physically-based hydrological dynamics to regional and continental scales.

Hydrology and water quality at various scales are now significantly affected by human activities. The water cycle may also be subject to the impacts of greenhouse-induced climate change. With growing interest in how climate and land use change affect hydrology must come a range of formally integrated models of atmosphere and land-based water cycle.

Arnell (1993) recently explained why hydrologists have become interested in modelling at large (>10 000 km^2) scales. First, hydrologists want to improve the representation of hydrological processes in regional and atmospheric models (with particular emphasis on the treatment of variability within model grid cells and channel routing within and between grid cells). Second, there is increased interest in simulating river flows in large basins for a variety of operational and planning purposes.

This report addresses the major interactions between climate and land surface processes on a large scale. Particular emphasis is given to surface hydrology and the role of vegetation in the context of global change (i.e. changes in the physical, chemical and biological processes of the Earth's surface and atmosphere). The report reviews relevant recent work, provides an overview of various research strategies, and defines important research areas for hydrologists.

This report is concerned with the atmospheric and land surface processes at their interface and possible changes in those interactions. Changes in the atmospheric composition (CO_2 and trace gases) and in climate will modify vegetation and land surface processes (especially hydrological processes). Conversely, changes in vegetation and land use will influence climate (particularly through its role in the hydrologic cycle) at different scales.

In Chapter 2, key features of the general circulation are discussed and synoptic scale, mesoscale and microscale climate systems are described. The chapter reviews synoptically-induced and terrain-induced mesoscale (regional) climate systems.

Chapter 3 discusses numerical, physical models for studying dynamic aspects of all or part of the climate system. The chapter reviews atmospheric general circulation models (AGCMs) and numerical mesoscale models (NMMs) and discusses the role of the biosphere and biosphere-climate interactions in those models.

Areal evaporation is a key hydrological land surface process. Various approaches to its estimation are reviewed in Chapter 4. The review shows the problems associated with scaling-up, i.e. spatial aggregation of evaporation in non-homogeneous terrain. Following a section on evaporation/atmosphere interactions, the chapter summarizes the combination equation, the Priestley-Taylor expression, the Bouchet complementary relationship, and several atmospheric boundary layer methods.

Chapter 5 describes recent advances in the use of satellite remote sensing to obtain physical land surface parameters and information on the atmosphere used for simulating components of the radiation and energy balances of the land surface, for large-scale and meso-scale climate modelling, and for inventories of water resources.

Land surface–atmosphere interactions are embedded in regional and large-scale climate models through Soil-Vegetation-Atmosphere Transfer (SVAT) schemes of varying complexity (Henderson-Sellers, 1991). Although these models estimate time-varying run off, they seldom have been cast in a hydrological context (Miller and Russell, 1992; Kuhl and Miller, 1992; Dumenil and Todini, 1993). Regional and large-scale climate models involve the parameterization of land surface characteristics and processes. Following a brief description of several simple hydrology models used in climate models, Chapter 6 reviews recent developments in land surface parameterization for climate modelling, which involves the scaling-up of small-scale processes.

Several modelling approaches, including the Biosphere-Atmosphere Transfer Scheme (BATS) and the Simple Biosphere Model (SiB) are summarized. The chapter discusses scale issues in large-scale hydrology and, in particular, the issue of subgrid-scale variability in land surface parameterization. Linked atmosphere/hydrology models would require a shared SVAT scheme. However, such models would also generate and move runoff via overland and subsurface flow paths, groundwater/stream interactions and open channel networks. Such integrated models can describe the behaviour of drainage basins within the context of a dynamic climate system. Large-scale hydrological modelling requires both, a runoff-generation component and a routing component. Naden (1993) has pointed out that considerable attention has been given in recent years to the runoff generation component, to land surface processes, and to aggregation procedures for soil-vegetation-atmosphere transfer processes while relatively little attention has been paid to large-scale routing models, as discussed in papers by Naden (1993), Vorosmarty, *et al.* (1989) and Vorosmarty and Moore (1991).

Finally in Chapter 7 and Appendices I, II and III, an overview is given of major international research programmes and field experiments which aim at better understanding critical land surface processes and at improving their parameterization for large-scale hydrology and climate modelling.

1.2 Climatic change and hydrological processes at the land surface

The energy exchanges between the sun, Earth and outer space govern the general circulation (i.e. the large-scale motion) of atmosphere and oceans and, hence, control climate. There have been major changes over the last 30 000 years in the world's climate and water resources. It is not immediately apparent to what extent mankind can influence global climate processes. However, it is now clear that there will be a doubling of CO_2 concentrations by the year 2030, if recent and present trends in atmospheric CO_2 concentrations continue. Atmospheric general circulation models indicate that such increases in atmospheric CO_2 and other radiatively active gases will result in an increase in the mean air temperature over the globe of between 1.5 and 4.5°C (IPCC, 1990). These will cause major changes in the water resources and the agricultural potential of cool and temperate regions. Note that the stomatal mechanism is sensitive to CO_2 concentrations, solar radiation, leaf temperature, etc. (Jarvis and Mansfield, 1981; Zeiger, *et al.*, 1987). Surface warming in the tropics and subtropics will initially be much less and hardly of significance for agriculture and hydrology. However, some time in the next century, CO_2 effects are also going to be of major importance in the tropics.

Current research aims to predict the extent to which increases in atmospheric concentations of CO_2, methane, ozone, nitrous oxides, and other trace gases will cause changes in climatic variables, including rainfall, temperature, humidity, radiation, and wind speed. Such changes may have direct impacts on the various components of the hydrologic cycle. The amount and frequency of streamflow, groundwater recharge, and soil moisture storage may all be affected, with potentially important economic and environmental implications. Changes in climatic variables and soil moisture availability may affect vegetation as well as the energy and mass exchanges at the land surface. Thus, major reductions in available soil moisture have been predicted in a zonal belt centred at 42° latitude. Monsoonal precipitation may also increase significantly and there will be an overall increase in aridity in summer in the middle and higher latitudes. Rind (1988) has used a more realistic land surface scheme to predict the impact of a doubled CO_2 climate on the hydrological cycle.

Climatic changes include long-term climatic trends and changes in climatic variability. The rate of change is also practically very important. Climatic variability may have widely differing and complex causes operating at different scales. In the case of rainfall, the primary large-scale cause is variability in general circulation patterns. Important features of such variability are the synchronous occurrence of variations in climatic elements over large regions, which are often far apart, the extreme spatial coherence of rainfall anomalies, and their persistence over long time periods.

Large-scale causes of such climatic variability include those associated with the *El Niño*/Southern Oscillation (ENSO) phenomenon and interactions with middle-latitude atmospheric processes (Enfield, 1989). ENSO events and its resulting climatic variations over large regions are associated with disturbances in the Walker circulation resulting from variations in east-west sea-surface temperature (SST) differences and in the normal Hadley circulation.

There are strong links between variations in SST and rainfall, especially in many subtropical and tropical areas. Significant differences have been noted between hemispheres. Climatic anomalies associated with ENSO are highly persistent and nearly global in extent. Thus, drought in Australia, Indonesia, India, West Africa and north-eastern Brazil, as well as excessive rainfall in the central and eastern Pacific, Peru, Ecuador, and southern Brazil, are all related to the Southern Oscillation (Lockwood, 1988).

CHAPTER 2

GENERAL CIRCULATION AND MESOSCALE CLIMATIC SYSTEMS

2.1 Introduction

General circulation patterns define broad climatic conditions. Mesoscale variability is imposed on these broad patterns. Mesoscale systems are predominantly hydrostatic, with winds which are out of gradient wind balance, even above the planetary boundary layer. By contrast the synoptic flow of the general circulation is close to the gradient wind balance above the planetary boundary layer. Non-hydrostatic effects are accounted for within the microscale (Pielke, *et al.*, 1987; 1989).

This review is primarily concerned with interactions between hydrologic processes and climate as well as their parameterization for large-scale climate modelling and macro-hydrology. In the terminology of Orlanski (1975) this embraces mesoscale space scales. Orlanski distinguishes between meso-alpha (200–2 000 km) and meso-beta (20–200 km) space scales. The corresponding time scales are one to several days, and 1–24 hours, respectively. A good review of mesoscale systems is provided by Ray (1986).

The difference between both space scales may be illustrated with rainfall. Rainfall deficits over land areas in the 200–2 000 km range are the results of large-scale subsidence which inhibits the mechanisms producing precipitation (Landsberg, 1982). At the 20–200 km scale, rainfall deficits can result from minor changes in atmospheric circulation when systems which normally produce widespread precipitation are replaced by ones which produce isolated convection showers (Robinson and Fesperman, 1987). Such changes can be associated with changes in the monsoon regime (Adefolalu, 1986) or fluctuations in depression tracks (Ratcliffe, 1978; Namias, 1983).

It is also important to distinguish between short-term climatic variability (including changes in the frequency of occurrence of shorter-term events) and longer-term climatic change (including changes in the frequency of short-term anomalies).

2.2 General circulation aspects

The general circulation is driven by global radiation differences. The general circulation has zonal, meridional and vertical components. The three-cell general circulation models of Bergeron (1928) and Rossby (1949) show the meridional circulation associated with the Hadley cell (0–30°; easterly tradewinds), the Ferrell cell (30–60°; disturbed westerlies) and the polar cell (>60°; polar easterlies). The model of Palmen, *et al.* (1951) replaced the indirect Ferrell cell and the region of the polar front by a zone of large-scale eddy transfer, both at mid- and low-tropospheric levels.

Heating of the lower atmosphere in tropical regions causes rising air motion, because horizontal temperature gradients are small. Such heating originates from warm ocean surfaces, elevated plateaus, and equatorial rain forests. The oceans cover much of the tropics, and sea temperature changes are generally small. Summer-winter differences are not as pronounced as at higher latitudes. However, the spatial and temporal variations within tropical regions result in complex meridional and zonal patterns, and complicated annual and interannual cycles. There are also distinct teleconnections between different regions. Fronts within tropical regions are not well defined. Fronts entering from higher latitudes are very quickly dissipated, with decreasing temperature contrasts between air masses.

Meridional equatorial flow between 30°S and 30°N results from the pressure difference between the equatorial trough and the subtropical high pressure belt. The trade winds are tropical easterlies. Their return flow is shown as upper westerlies in the Hadley cell. The trade winds are strongest in winter. They are very persistent and steady, and reflect the permanence of the subtropical high pressure cells. They are present in all tropical oceans, except the Indian Ocean. Wind speeds in the subtropics, and especially at about 30° latitude are low. There is a low-level inversion in the trade wind zone which is generated by subsidence associated with the divergence of the trades as they flow toward the equator. This trade wind inversion acts as an effective lid on vertical cloud development. The height of this inversion increases towards the equator and towards the west over tropical oceans. The increase is therefore along the general direction of the trades. The inversion is strongest where the height of its base is lowest.

Mean zonal flow profiles at low latitudes show the important role of the westerlies aloft. Maximum speeds in the westerlies occur at a pressure level of about 200 hPa. Maximum speeds in winter are twice those in summer. These zonal flow profiles clearly show the presence of the subtropical jet stream, between 30° and 40° latitudes at about 12 km. The air subsiding from it forms the subtropical high pressure belt. More momentum than needed to sustain the jetstream is carried downward to maintain the eastward flowing surface winds of the middle latitudes against the opposing forces of surface friction.

At about 30° latitude in both hemispheres the subtropical belt of high pressure cells contains quasi-permanent high pressure cells separated from each other by cols. These cells have great permanence and appear to be locked into fixed positions.

At subtropical latitudes, the downward leg of the meridional Hadley cell represents persistent subsidence. Subtropical regions with maximum variability in the subsidence are semi-desertic, whereas regions with minimum variability are desert regions. The zonal Walker circulation is superimposed on the meridional circulation.

In general, low pressure prevails near the equator and high pressure dominates in the subtropics, whereas in the middle latitudes there are the eastward moving alternating high and low pressure systems. Large horizontal eddies cause transport of colder air to lower latitudes and warmer air to higher latitudes. This is overlaid by the meandering strong zonal current of the high level jet streams. Low pressure areas have convergence of air and ascending air motions with a likelihood of precipitation. The high pressure areas show divergence and descending air masses, which tend to be associated with clear skies and dry weather.

2.3 Mesoscale climatic systems

Mesoscale patterns in climatic variables, such as rainfall or temperature can be produced by mesoscale systems within the atmosphere or by the interaction of the atmosphere with mesoscale terrain features. Atmospheric mesoscale systems can be divided into systems that are primarily forced by instabilities of travelling larger-scale disturbances (synoptically induced) and those that are forced by surface inhomogeneities (terrain induced).

2.3.1 *Synoptically-induced mesoscale climate systems*

Synoptically-induced mesoscale systems include mesoscale convective clusters (Maddox, 1980), tropical cyclones, squall lines, and convective bands (Pielke, 1984). They have been reviewed by Atkinson (1981), Pielke (1984), and Ray (1986). The mesoscale convective cloud systems that produce most of the rain in the tropics consist of a combination of convective and stratiform clouds. The convective regions contain numerous deep cells that are often arranged in lines. Such regions are accompanied by mesoscale stratiform precipitation areas (Houze, 1989). The net heating of the system is dominated by condensation and evaporation associated with the vertical motions (Houze, 1989). Simple linear global models show that the general circulation is sensitive to the vertical distribution of heating associated with such tropical convection (De Maria, 1985).

2.3.2 *Terrain-induced mesoscale climate systems*

Terrain-induced mesoscale systems include mechanically-induced flows, such as lee waves and downslope winds as well as thermally-induced flows of land/sea breezes and valley winds, all of which may affect convective triggering in unstable systems (Atkinson, 1981; Clarke, *et al.*, 1981). Mesoscale circulations can lead to enhanced precipitation (Jiusto and Kaplan, 1972). Strong bio-physical feedback may exist with the underlying surface. Sellers, *et al.* (1989) noted that three main land surface properties govern the interactions with the atmosphere: the albedo (radiative transfer), surface roughness (momentum transfer), and the surface hydrological description (sensible and latent heat transfer). Over short time scales, soil moisture variations are more significant for the energy budget and the PBL structure than changes in roughness and albedo (Mahfouf, 1990).

Hare (1985) notes that drastic changes in (near) surface microclimates may result from changes in vegetation cover and changes in surface hydrology. He identifies several important feedback mechanisms including changes in surface albedo, surface roughness, and soil moisture status, which illustrate that human activities may increase drought.

Soil moisture availability is a major determinant of total evaporation. It also determines the relative roles of vegetation and bare soil in evaporation. There are also strong interactions between rainfall and evaporation. Numerical models are increasingly used to investigate the role of soil moisture and evaporation on climate (e.g. Charney, 1975; Charney, *et al.*, 1977; Walker and Rowntree, 1977; and Rowntree and Sangster, 1986).

It should also be noted that evaporation from the surface is a necessary, though not sufficient, condition for extratropical summer rainfall. For example, changes in soil moisture affect the albedo and thermal diffusivity of the soil as well as the Bowen ratio in the surface boundary layer. As moist soil dries out, a larger fraction of the absorbed energy is used to heat the air. The heat flow into the soil increases at first, then decreases as soil becomes very dry (Rose, 1966). Drought begets drought. Kunkel (1989) noted a marked decrease in evaporation during the recent 1988 drought in America. He further suggested that this may have played a role in the persistence of the drought by reducing the atmospheric water vapour supply and by increasing the flux of sensible heat to the atmosphere. For drying soils, soil moisture content tends to become patchy at spatial scales of individual storms and regional estimates of soil moisture and evaporation become less reliable.

Recent numerical studies illustrate the effect of soil moisture on local precipitation (Mintz, 1982; Shukla and Mintz, 1982). Fennessy and Sud (1983) examined monthly precipitation values over the contiguous USA, in relation to antecedent monthly precipitation, soil moisture, and evaporation. They found that large-scale droughts over extended periods may be partially maintained by the (negative) feedback influence of soil moisture on rainfall. These results agree with simulations of African droughts by Rowntree, *et al.*, 1985 and others. Numerical simulations by Ookouchi, *et al.* (1984) and Anthes and Kuo (1986) demonstrate the importance of soil moisture availability in generating mesoscale and synoptic-scale circulations.

Segal, *et al.* (1988) noted that mesoscale domains covered by extensive very dense, unstressed vegetation adjacent to bare soil areas, can generate significant mesoscale circulations. Their simulations suggest that if, after prolonged drought, the vegetation has access to groundwater reserves through deep tap roots, major differences will exist in the evaporation regime of the vegetated area compared to neighbouring bare ground. Such differences will invariably induce mesoscale circulations. Similar features have been suggested by Anthes (1984) and Segal, *et al.* (1989), amongst others.

There are many empirical studies of deforestation, surface heating of small islands, and the effects of irrigation schemes, especially in the former USSR, which indicate that differential heating at the mesoscale can result in increased rainfall and that vegetated surfaces are more likely to produce rainfall than bare soils. Anthes (1984) has hypothesized that planting bands of vegetation with widths of the order of 50–100 km in semi-arid regions could, under favourable large-scale atmospheric conditions, result in increases in convective rainfall. His hypothesis is illustrated in Figure 1. Anthes emphasizes that this effect is less likely in desert regions where moist convection is largely suppressed by synoptic scale subsidence. Yan and Anthes (1988) and Giorgi (1989) simulated the effect of alternating bands of dry and wet soil, some 144 km wide. The difference in evaporative cooling between dry and wet land generated horizontal gradients of surface temperature and sensible heat flux that lead to the

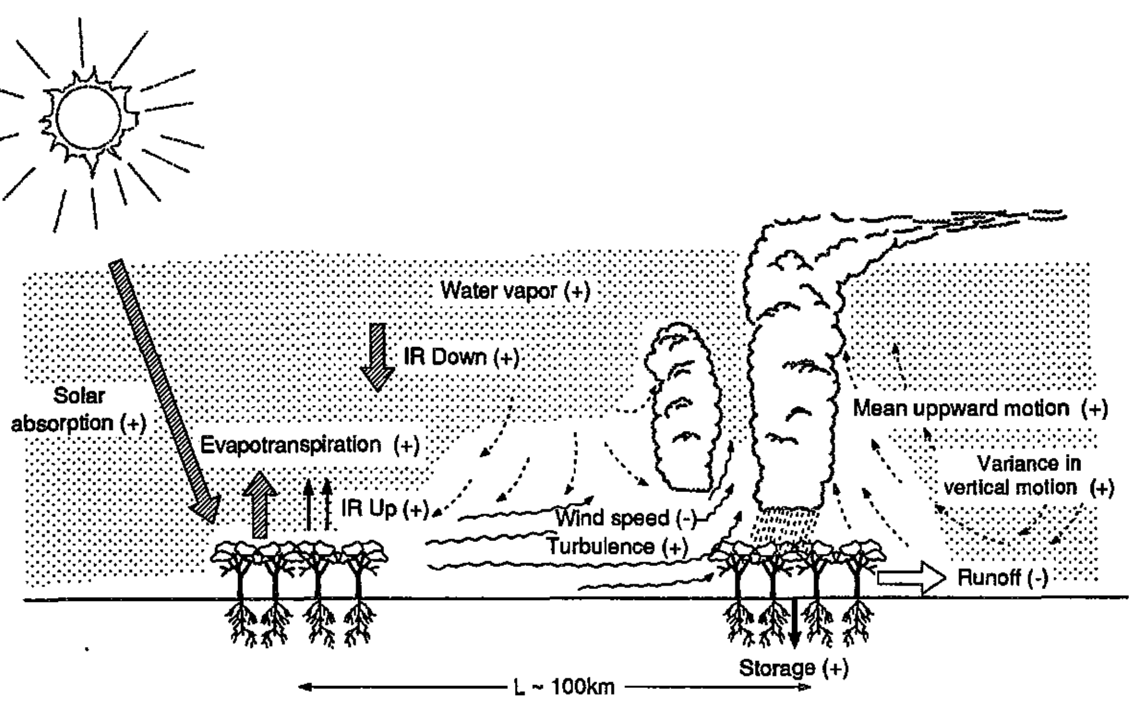

Figure 1 — Illustration of the hypothesized effect of establishing bands of vegetation in a semi-arid region of previously bare soil. Increases or decreases of an effect or process following the introduction of vegetation are indicated by pluses or minuses, respectively (after Anthes, 1984).

formation of a sea breeze type circulation at the dry/wet land boundary. As the circulation developed, moisture was advected from the wet region and the resulting convergence of inflow over the dry regions initiated vigorous convection and ultimately strong precipitation over the dry region.

Sud and Smith (1985) found that large-scale increases in surface albedo and reductions in surface roughness (as would result from the removal of forests) had a weakening effect on the Indian monsoon. Such large-scale reductions in surface roughness alter the convergence of horizontal water transport in the boundary layer and can lead to changes in the spatial distribution of rainfall (Sud, *et al.*, 1986). Charney, *et al.* (1977) have suggested that an increased albedo in the Sahel region would give rise to reductions in evaporation rate, cloudiness, and tropospheric moisture convergence, resulting in reduced rainfall.

Otterman (1989) and Otterman, *et al.* (1990) have argued that increases in precipitation resulting from land use changes are attributable to intensification of the dynamical processes of convection and advection resulting from plant-induced enhancement of the daytime sensible heat flux from a generally dry surface. They suggest that the enhancement results from the reduced surface albedo and reduced soil heat flux. Rabin, *et al.* (1990) also suggested that clouds form earliest over regions characterized by high sensible heat flux and are suppressed over regions characterized by high latent heat flux during relatively dry atmospheric conditions.

2.3.3 *Rainfall*

Rainfall depends on synoptic-scale processes (e.g. depressions and troughs leading to convergence, rising of air and cloud development followed by rain) and several mesoscale effects (such as the terrain forcing moist air to rise and the effect of surface heating). Here we use the example of rainfall to distinguish between general circulation aspects and the role of mesoscale atmospheric systems.

Rainfall is usually categorized into convectional, orographic and cyclonic rainfall. Convectional rainfall, which includes most tropical rainfall associated with the inter-tropical convergence zone, is characterized by a marked vertical ascent of air. When the convection is very vigorous it may be accompanied by thunderstorms.

Orographic rainfall is the forced ascent of moist warm air on the windward side of mountains resulting in a rain shadow on the lee side. Rain associated with forced low-level convergence upwind from a mountain barrier is also orographic rain. Cyclonic rainfall is associated with the steady ascent of air over a frontal boundary or the slow ascent of air within the core of a mature depression. In the case of active convergence within the depression, the rainfall pattern is convective. The term cyclonic rainfall is rarely appropriate for tropical rainfall.

Poleward transport of warm air results in the interaction between warm and cold air masses at higher latitudes. The migrant cyclonic storms which form along the boundaries between warm and cold air cause the cyclonic rainfall in more temperate regions. Most of the remaining rainfall in those regions is produced in irregular patterns by local thunderstorms, resulting from the vertical instability caused by intensive heating.

An important factor causing rainfall variability is the presence or absence of water vapour and/or rain inducing disturbances. Lockwood (1988) shows summer and winter transport fields of water vapour and source regions. Boucher (1975) notes that the global distribution of precipitable water shows land-sea effects, continentality, and the barrier effects of mountains. The lack of precipitation in deserts is largely due to stable, anticyclonic conditions.

Orography and the presence of substantial water bodies may influence rainfall at regional and local scales. Cool ocean currents may influence aridity of bordering land areas, except on the adjacent shore. Distance from the sea/ocean is thus an important factor because the cool onshore winds are heated quickly over the land areas causing drying conditions. Mountain ranges on the coast often cause rain on the windward side and very rapid change to semi-arid/arid conditions on their lee-side. Lakes may also influence rainfall at a regional scale.

Diurnal patterns accentuated by orography and terrain may be very important at a local scale. Steady onshore winds result in a small diurnal temperature range. Sea breezes re-inforce onshore winds and may combine with local anabatic effects. If air is moist and unstable, the sea breeze may lead to increased vertical motion, resulting in afternoon thunderstorm activity and rains. On the other hand, steady offshore winds cause more continental features. Hauze, *et al.* (1981), and Johnson and Priegnitz (1981) suggested that the early morning convective activity observed along the the coasts of northern Borneo during the winter monsoon was significantly affected by the land breeze.

The seasonal thermal contrasts of land and water are important. Winter outflow of dry air from continental high-pressure areas is associated with a lack of precipitation. Summer onshore winds from oceans may bring copious rain. However, mountainous (coastal) regions may show a rain shadow because of the ascending air.

In the tropics, moisture for rain is trapped under the trade wind inversion. Many tropical plateaus are, therefore, dry whereas nearby coastal lands may have high rainfall.

CHAPTER 3

PHYSICAL CLIMATE MODELS

3.1 Introduction

Physical models of all or part of the climate system have become widespread over the last 30 years. Physical models range from the energy balance models (e.g. Sellers, 1969), which treat a 'slice' through the atmosphere at any given point, up to atmospheric general circulation models (AGCMs).

Washington and Parkinson (1986) and Henderson-Sellers and McGuffie (1987) discuss the diversity of approaches in those models. Physical relationships between atmospheric variables and parameters are expressed with differential equations. Prognostic models contain at least one time-dependent equation. Dynamical models contain predictive equations for wind, temperature and moisture, as well as several diagnostic equations. The non-linear equations cannot be integrated analytically and the differential equations are, therefore, converted into a set of finite difference equations. Thus, continuous variables, such as temperature or pressure, are usually given as a finite three-dimensional set of grid-point values for a rectangular grid in the horizontal plane with multiple levels in the vertical (Avissar and Verstraete, 1990).

AGCMs attempt to describe the space and time variability of the atmosphere across the Earth's surface for periods ranging from a few days to perhaps a century. The Earth's surface is represented with a horizontal resolution of 2.8° to 8° and the vertical structure of the atmosphere is described with 2–11 layers. Avissar and Verstraete (1990) point out that numerical weather prediction models use a higher resolution of about 1° and can use up to 18 levels in the vertical. Mesoscale atmospheric models have a horizontal resolution of 1–20 km and use 5–20 layers for the first 3–15 km of the atmosphere.

Resolution is an important factor in modelling the Earth's surface. Within a single grid box, considerable heterogeneity should be represented (in roughness, albedo, vegetation and soil type, for example). In most land surface models this heterogeneity is ignored or simplified. More recently, Koster and Suarez (1992*a*, *b*; 1993) have tried to represent land surface heterogeneity and the effects of such heterogeneity on the surface atmosphere fluxes.

To increase the spatial resolution of AGCMs to less than 3° is computationally expensive. Other approaches to improve spatial resolution include using numerical mesoscale models (NMMs), embedding high resolution NMMs in AGCMs, and enhancing model output through integration of mesoscale observations and model results.

Large-scale and regional-scale models are needed for studies of the processes which link the terrestrial biosphere with the hydrological cycle and large-scale climate. The GEWEX Continental-scale International Project (GCIP) science plan (Schaake, personal communication) estimates that the resolution of AGCMs will be approximately 1° (or about 100 km) by the year 2000 or shortly after. At the same time, mesoscale models will have a resolution of about 10 km. The GCIP strategy is to develop improved land surface parameterization schemes at the 100 km scale and to do this by linking atmospheric and hydrologic models first at the mesoscale.

Becker (1991) points out that one of the major challenges in incorporating models for the land surface hydrology in numerical climate models "stems from the fact that the spatial resolution of the latter (10^4–10^6 km^2) is incompatible with the characteristic scales of surface hydrologic processes". Section 6.3.2 provides detailed discussions of approaches to considering subgrid scale variability. These include areal discretization (Entekhabi and Eagleson, 1991), and dynamic, statistical approaches to land surface process modelling.

The interactions between biosphere, atmosphere and hydrological cycle are complex and non-linear. Modelling these interactions requires integration of point data on state variables up to the grid sizes of NMMs and AGCMs. However, coupling of land surface and atmosphere models is very difficult because of the different spatial scales used in such models.

The following sections briefly review physical models with an emphasis on how resolution can be improved so that model output becomes more useful for catchment-scale studies. It will be shown that for such purposes, methods for improving resolution are of crucial importance. Where the user's scale of interest is significantly less than the model resolution, the model results may not be particularly useful, and can be highly misleading. The use of AGCM results is not valid if predictions of climate change at a catchment scale are required. However, if techniques to improve AGCM resolution prove successful, model results may become available at an appropriate resolution.

3.2 Atmospheric general circulation models

Atmospheric general circulation models are reviewed by Washington and Parkinson (1986) and Bach (1988). Individual AGCMs are described by Arakawa (1972); Corby, *et al.* (1972); Gates, *et al.* (1971); Hansen, *et al.* (1983, 1984); North (1975); North, *et al.* (1981); Sellers (1969); Washington and Williamson (1977); Williamson, *et al.* (1987); and McAvaney, *et al.* (1991).

AGCMs simulate the physical and dynamical processes that control the climate system. They represent, to varying degrees, the atmosphere, hydrosphere, biosphere, and cryosphere. Bach (1988) notes that AGCMs should also be based on realistic geography and topography, have a high spatial resolution and an adequate temporal resolution, incorporate a coupled model of atmosphere-ocean circulation, and simulate patterns of observed climate in a realistic manner.

Most state-of-the-art AGCMs deal with these elements adequately. They prescribe realistic geography and a smoothed topography field. The spatial resolution is usually approximately 5° x 5°, but may be as good as 3° x 3°, while the temporal resolution is usually 10 or 30 minutes. Few AGCMs incorporate a fully coupled ocean model due to problems of physically coupling the ocean model with the atmosphere because the atmosphere and the oceans operate at fundamentally different time scales (Washington and Parkinson, 1986). Most state-of-the-art AGCMs simulate the climate at the global scale remarkably well. It is at the continental and sub-continental scale, near to and at the land surface that AGCMs fail to simulate the climate system realistically. Partly, the problem seems to be the representation of topography in AGCMs, which has to be smoothed to the resolution of the model itself. However, the lack of a realistic parameterization of the land surface is also a major problem.

Many models have been used until now largely in time-independent equilibrium response experiments which are not fully realistic. However, at least three AGCMs have been coupled with adequate dynamic ocean models and have been used to predict equilibrium changes.

Many AGCMs have limited or no representation of the biosphere and biosphere-climate interactions. In almost all AGCMs the surface fluxes of radiation, sensible heat, latent heat and momentum, and key surface parameters are averaged over each grid cell.

AGCMs may be tested by simulating past climates using palaeo-environmental data which include pollen assemblages and geomorphic indicators. Such retrodictive tests provide a range of conditions and are invaluable in providing confidence in the use of such models for predicting future climatic changes. They can also be tested by simulating a known decade (e.g. AMIP).

The importance of the land surface in the modelling of the climate has been a matter of contention during the last decade. There is now an abundance of experimental evidence which suggests that the global atmosphere simulated by AGCMs is sensitive to parameterization and the state of the land surface. Although significant progress has been made with land surface parameterization, other parts of AGCMs, such as the cloud, aerosol, and ocean components still face serious problems.

Numerical experiments and sensitivity studies with AGCMs include the studies by Bhumralkar (1975); Hansen, *et al.* (1984); Hansen, *et al.* (1985); Mintz (1984); Sud and Molod (1986); and Sud, *et al.* (1986).

Mintz (1984) reviewed a variety of perturbation experiments using AGCMs. These showed that the climate simulated by these models was sensitive to soil moisture and evaporation (e.g. Charney, 1975; Miyakoda and Strickler, 1981; Shukla and Mintz, 1982; Carson and Sangster, 1981; Dickinson, 1984, 1986; and Rind, 1982), to albedo (e.g. Rowntree and Sangster 1986; Oerlemans and Van den Dool, 1978; Dickinson, 1983; Carson and Sangster, 1981; Charney, *et al.* 1977; and Chervin 1979) and to surface roughness length (e.g. Sud and Fennessy, 1984; Sud, *et al.*, 1985; and Sud and Smith, 1985). The radiation balance and infrared radiation were studied by Oerlemans and Van den Doel (1978); Rowntree (1975); and Short, *et al.* (1984).

However, the majority of these experiments were simulations of major perturbations to the land surface, which were far greater than those which would be induced by climatic variability on the scale found in the last 30 000 years. Shukla and Mintz (1982), for example, compared the effects of a permanently saturated land surface and a totally dry land surface on the simulation of the atmosphere and found a global response to the perturbation (Figure 2). This was not a test of the climate response to a natural phenomenon but an experiment to determine whether the climate simulated by AGCMs could be shown to be sensitive to the description of the land surface.

AGCMs have highlighted the role of the land surface on climate by showing that the surface fluxes of energy and moisture influence short-term weather systems and become increasingly dominant on longer time scales (Dickinson, 1989*a*). However, they show that the prediction of local effects is difficult because land surface models are still inadequate (Dickinson and Henderson-Sellers, 1988). Meehl and Washington (1988) note that soil moisture is the critical factor in determining the magnitude of summer drying. However, they show that the highly parameterized hydrology of current AGCMs, combined with the lack of appropriate data and the complexity of real hydrological processes, prohibits adequate model verification. Equally important was the observation by Rind (1982) that soil moisture deficits can be propagated year by year in an AGCM. This implies that for climate change experiments, the parameterization of soil hydrology is particularly important.

In regions where the atmospheric forcing varies on scales less than a few hundred kilometres, single grid boxes of AGCMs are inappropriate to describe regional climatic patterns.

Implicit (i.e. parameterized) convective schemes are also very important for predicting convectively-driven precipitation systems even with a grid size of the order of 10 km, because subgrid scale eddy fluxes

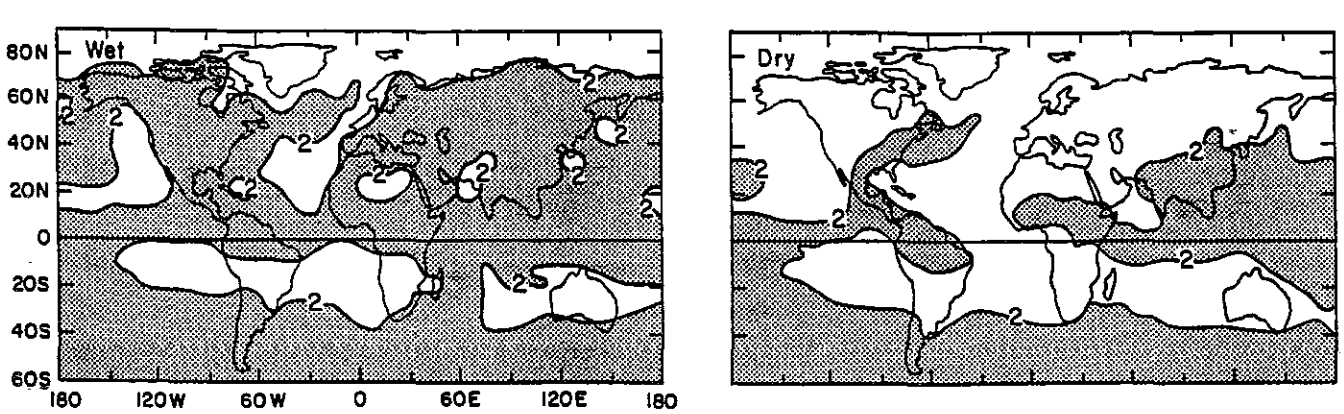

Figure 2 — Rate of precipitation (mm/day) for a wet and dry soil case experiment with a general circulation model (from Avissar and Verstraete, 1990) (after Mintz, 1984)).

should not be neglected on this scale and the convective rainfall does not necessarily occur in the presence of meso-beta scale (20–200 km) saturation (Zhang, *et al.* 1988). Neglect of latent heat in condensation and precipitation as well as sensible and latent heat exchanges at the surface produce significant bias in numerical mesoscale forecasts over periods of three days and less (Anthes, *et al.*, 1989).

Important features of the land surface climate may be relatively poorly simulated by AGCMs, because significant orographic features are often unrealistically absent or smoothed at the grid scales of current AGCMs. For example,the Great Dividing Range which strongly affects the climate of the East coast of Australia is almost absent from AGCMs with a horizontal resolution that is coarser than about 7° x 7°. Gutowski, *et al.* (1988) suggest that improvements in orographically-dependent fields may be achieved simply by increasing the resolution of AGCMs. However, this implies an unrealistic increase in the computational cost. Thus, at least a ten-fold increase in the horizontal resolution may be required to capture adequately regional-scale climates. For example, Ninomiya, *et al.* (1984) observed an increase in both the precipitation rate and the total amount of precipitation as the horizontal grid resolution increased. They attributed this to an increase in both the magnitude and the degree of concentration of moisture convergence in the lower layers.

While the climate predictions by AGCMs are generally very good, there are problems in specific regions. Krishnamurti, *et al.* (1984) evaluated an AGCM in tropical regions and found that it was able to predict the formation and motion of two major monsoon depressions. Such tropical forecasts with AGCMs have improved over the years through the enhanced parameterization of shallow and deep cumulus convection and increases in the horizontal resolution (Tiedtke, *et al.*, 1988).

Large-scale cloud fields in AGCMs are now reasonably realistic, but the models are unable to account for small-scale detail. The parameterization of the atmospheric boundary layer (Benoit, 1976) concentrates on the correct parameterization of the vertical fluxes of momentum, heat and water vapour, as they affect the upper atmosphere rather than the development of the boundary layer itself (Reiff, *et al.*, 1986; and Garratt, 1993). Recent analysis by Dickinson (1989*b*) of simulations of tropical deforestation by Dickinson and Henderson-Sellers (1988) and Henderson-Sellers, *et al.* (1993) indicate that the significant differences between the model results and observations were due to deficiencies in the simulation of rainfall and associated cloud (see also Shuttleworth and Dickinson, 1989).

AGCMs have had limited success in predicting rainfall because it often forms on scales smaller than those resolved by current models. This has been an important practical motivation in the development of NMMs (Keyser and Uccellini, 1987).

AGCMs can only produce uniform rainfall over a grid square, yet convective rainfall, especially in the tropics, is spatially restricted (Dickinson, 1989*b*). The neglect of the spatial variability of convective rainfall in current AGCMs is a potentially serious source of model error since it can reduce the effective water holding capacity of a uniform canopy by a considerable amount, as well as affect the surface energy balance (Pitman, *et al.*, 1993).

AGCMs are invaluable tools for investigating climate and climate change at continental scales. At finer spatial scales, the resolution becomes critical. It is not realistic to take a single AGCM grid square and use its predictions in isolation from other grid squares. Thus, for climate studies which are intended to be applied to problems with a spatial scale of less than about 3° x 3°, other approaches are required until AGCM resolutions can be improved.

3.3 Numerical mesoscale models

In regions where the atmospheric physical and dynamical forcings vary on a scale of less than a few hundred kilometres, current operational AGCMs are inadequate to describe detailed climatic patterns.

There is a need for at least a ten-fold increase in the horizontal resolution to represent adequately regional climates. This translates into a thousand-fold increase in the computational resources needed compared to AGCMs. These problems may be addressed with (independent) NMMs, with high-resolution NMMs embedded in AGCMs, and through enhanced output from AGCMs.

NMMs use numerical iterative techniques to solve the same physical equations as AGCMs, but they are specifically designed to simulate limited spatial areas. NMMs can provide diurnal variations in meteorological variables at various levels. They can incorporate variations in topography.

NMMs have been reviewed by Pielke (1982; 1984). They have been used to simulate sea breezes, frontal perturbations (Ross and Orlanski, 1982), tropical cyclones (Rosenthal, 1978), and squall lines (Pointin, 1985). In all cases, the NMM needs to be initialized with representative but dynamically consistent synoptic-scale meteorological fields. Segal, *et al.* (1982) used climatological averages, whereas Diab and Garstang (1984) used representative examples from a synoptic classification. The need to provide initial data restricts the use of NMMs to well instrumented regions.

NMMs have been useful for simulating individual events and their impact after the event, as well as predicting local changes in climate (Segal, *et al.*, 1983) and extending the mesoscale database through mapping phenomena such as frost (Avissar and Mahrer 1988*a*, *b*). NMMs have been used for both sensitivity and diagnostic studies of individual events. They have also provided mesoscale output for subsequent synoptic analysis (Keyser and Uccellini, 1987).

Pielke (1984) defines initialization in terms of objective analysis, dynamic initialization, and normal mode initialization. The first two have been reviewed by Bengtsson (1975) and the latter by Daley (1981). Within tropical NMMs the use of a backward implicit time difference scheme has been shown to work reasonably well (Satomura, 1988). These methods are effective if mesoscale observations are available to describe fully any mesoscale circulation. However, the density of data available is often insufficient to resolve mesoscale circulations. Without such observations, the slow spin-up of mesoscale circulations in a numerical model represents a significant problem for numerical prediction (Ninomiya and Kurihara, 1987).

Within the tropics, inadequate data on humidity and divergent wind fields result in poor rainfall distribution in the first two days of model integrations (Krishnamurti, 1989). This can be improved through improved objective analysis schemes as well as initialization schemes utilizing high resolution satellite, radar, and surface observations (Wolcott and Warner, 1981; Anthes, 1983; Wetzel and Woodward, 1987; and Austin, 1989). For example Illari (1989) and Austin (1989) showed that the inclusion of satellite-derived data on precipitable water content lead to improved representation of the monsoon onset as well as better forecasts of rain. Austin (1989) suggests that significant improvements in quantitative precipitation forecasts can be obtained when the numerical model is initialized with satellite/radar derived cloud and rain patterns. In the case of thunderstorms, where the time and space scales are relatively small, it is not clear that currently available mesoscale models could compute the development of such systems, even if adequate initialization was possible, as thunderstorm physics and dynamics are not normally incorporated in operational NMMs.

Zhang and Fritsch (1986) argued that mesoscale models need to use initial conditions which incorporate mesoscale features that are detectable on satellite and radar data and are apparent from careful subjective analysis. They generated artificial soundings in data-poor areas to provide a consistent initial field and were able to simulate the structure and evolution of the 1977 Johnstown flood. These results suggest that it may be possible to forecast the meso-beta scale structure and evolution of convective weather systems with useful skill for up to 18 hours in advance.

NMMs are currently used in France (Imbard, *et al.*, 1986; Durand, 1985; Imbard, *et al.*, 1987; Juvanon du Vachat, *et al.*, 1987), in Australia (Leslie, *et al.*, 1985), in Norway (Gronas, *et al.*, 1987*a*, *b*; Nordeng, 1986; Gronas and Midtbo, 1987*b*), in Japan (Tatsumi, 1987), in the United Kingdom (Golding, 1987*a*, *b*), and in Germany (Herzog, 1987). These models give improved predictions of mesoscale phenomena through their higher spatial resolution. However, they lack detailed mesoscale observations for verification and initialization. Their forecasts suffer from the delayed spin-up of mesoscale circulations as well as inadequate specifications of convection and boundary layer processes (Gronas, 1989).

Adrian (1987) determined initial conditions for a mesoscale model from both operational weather analysis and a forecast. Wetzel and Woodward (1987) developed a technique to obtain spatially- and temporally-averaged soil moisture data from satellite observations. Bougeault, *et al.* (1990) ignored the subgrid-scale variability of soil texture and water content and concentrated on the variability of vegetation and defined their surface parameters in terms of the dominant vegetation type.

For very short-range forecasting (e.g. flash floods), the spin-up time and data requirements of numerical models make them inappropriate (Pielke, 1982; Georgakakos, 1986).

Various one-dimensional atmospheric boundary layer (ABL) and planetary boundary layer (PBL) models are used in mesoclimatology. Such models include those of Blackadar (1976; 1978), Bodin (1979), Carlson and Boland (1978), and Carlson, *et al.* (1981).

Two-dimensional, linear mesoscale models have been developed, such as those used for land-sea breezes by Rotunno (1983) and many others. Anthes (1984) used Rotunno's general solution as the basis for a model of the circulations induced by periodic differential heating associated with variations of surface characteristics, in an effort to assess the impact of vegetation bands with widths of 50–100 km in semi-arid region (see section 2.3.2).

Three-dimensional numerical models include the regional atmospheric modelling system (RAMS) of Colorado State University. Its numerous applications include the studies by McCumber (1980) and McCumber and Pielke (1981) of the effect of heat and moisture fluxes on mesoscale circulations. Several diagnostic models have been developed to extend surface observations.

Nickerson and Smiley (1975) discuss surface layer and energy budget parameterizations for mesoscale models. Several regional-scale numerical prediction models have been developed specifically for soil moisture and evapotranspiration. These include the studies of Camillo, *et al.* (1983) and Wetzel and Chang (1987; 1988).

3.4 Embedded models

The idea of embedding an NMM for a comparatively small (sub-continental) region into an AGCM is attractive. In principle, the AGCM could simulate the global atmosphere at a low spatial resolution, with specific regions 'focussed in upon' to provide high resolution (e.g. 0.5° x 0.5°) simulations. At these higher resolutions, land use, soil type, vegetation type, and orography could all be represented realistically. The resulting data could be used for catchment models to assess the hydrological impact of climatic change.

However, there are some major problems. First, simulating a large region — the mesoscale domain must exceed around 3 000 x 3 000 km — (Dickinson, *et al.*, 1989) — is computationally expensive. Second, to date all embedding simulations have been done 'off line'. This means that the climate simulation by the AGCM is stored on magnetic tape, which is then accessed for each time step by the NMM. It is, therefore, a one-way procedure: the NMM is driven by the AGCM, but the AGCM's climatology is not affected by the simulation from the NMM. Thirdly, there is a problem with boundary discontinuity. The AGCM may have, for example, a horizontal resolution of 5° x 5°. The NMM, in contrast, has a resolution of about 0.5° x 0.5°. A numerical procedure is used to smooth this discontinuity across the domain boundary. The size of the NMM domain must be large enough so that boundary discontinuities do not affect the simulation. Finally, the AGCM has a time step of about 30 minutes. The NMM may use a one minute time step. Unless considerable care is taken with the choice of time step and resolution, problems arise if weather systems develop and move across the mesoscale domain at a speed which is different from that in the AGCM.

Most of these difficulties do not cause difficulties with simulations based on short (< six days) integrations (Giorgi and Mearns, 1991). However, extended simulations where the NMM is integrated simultaneously with the AGCM for multi-decade simulations are not practical at the present time. Furthermore, due to the problems outlined above, using embedded NMMs requires time, patience and considerable computing resources. It is not simply a matter of 'plugging in' the NMM in order to simulate a different region. Finally, due to the boundary discontinuity problem, the time when NMMs are embedded into AGCMs as a two-way process (i.e. the simulation from the NMM affects the climate of the AGCM) is a long way off.

Krishnamurti (1989) has suggested that the use of boundary conditions from a larger domain AGCM model leads to forecasts that are determined by the performance of the larger model and are a function of its resolution. Mesoscale models require accurate global models to provide the synoptic forcing or accurate large-scale analyses (Anthes, *et al.*, 1989). However, despite these problems, there are an increasing number of studies which suggest that the embedding technique is valuable.

Winter-time precipitation climatology of the western United States has been simulated by embedding an NMM within the coarse resolution of an AGCM. Results from embedding the Pennsylvania State University NCAR NMM (Anthes, *et al.*, 1987) in the NCAR community climate model (CCM) for the western USA have been described by Dickinson, *et al.* (1989), Giorgi, *et al.* (1989) and Giorgi and Bates (1989). These studies suggest that the high-resolution limited-area NMM is sensitive to the initial soil moisture regime, which is derived from AGCM forcing. However, Giorgi and Bates (1989) and Giorgi (1990) have all shown that the simulation by the NMM was more successful. Thus, appropriate-scale modelling of surface heterogeneities is essential (McCumber and Pielke, 1981; Wetzel and Chang, 1987) to represent adequately the initial surface regime and to apply appropriate forcing to the model.

Recent simulations by Pitman, *et al.* (1990) using the same model as Giorgi, *et al.* (1989), but for south-eastern Australia, have incorporated the BATS land surface scheme in order to address the problem of incorporating surface heterogeneities. Although only short simulations were performed, they showed that a much better precipitation climatology could be produced using the NMM embedded into the NCAR CCM, in comparison with the NCAR CCM alone. The NMM simulated the distribution of precipitation over south-eastern Australia realistically since the model tended to concentrate precipitation close to the coasts. This was despite the provision, by the NCAR CCM, of rather poor driving boundary conditions. It appeared that the NMM was able to cope with poor driving conditions to provide a marked improvement in the simulation of precipitation.

3.5 Model output enhancement

The AGCM model output has been enhanced by interpolating output from operational large-scale numerical models with digital terrain data. Kelley, *et al.* (1988) used this approach to simulate temperature fields and suggested that it may be used for precipitation. Such a process relies heavily on the parameterization of subgrid scale processes within the AGCM and has not been compared directly to equivalent NMM forecasts. It is not clear if the success depends on inherent physics of the NMM or on adequate parameterization of subgrid-scale processes in AGCM.

Alternatively, Heyboer, *et al.* (1989) integrate mesoscale observations with fine-mesh prognoses to produce very short-range gale warnings and tidal surge forecasts.

Short-range boundary layer forecasting has also been attempted by combining one-dimensional operational boundary layer models with a trajectory approach (Reiff, *et al.*, 1984). In this approach, backward trajectories are calculated from the place of interest using operational forecast wind fields to define source areas. The observed temperature and humidity profiles from the source region define the initial state of the boundary layer which is then modified over the trajectory using a one-dimensional boundary layer model. Although this approach has been shown to give reasonable results when advection is significant, it is limited to gently sloping terrain and does not work when large-scale dynamics interacts with mesoscale phenomena, such as the sea/land breeze circulation (Reiff, *et al.*, 1986).

3.6 Predictability

The limit of deterministic weather predictability is several days up to a week or so (Hoskins and Sardeshmukh, 1987) because of incomplete knowledge of the atmosphere's initial state (Gilchrist, 1986). Anthes and Baumhefner (1984) illustrated the relative contributions to predictability of initial data and models as a function of time. Improvement in longer-term forecasts appears to rest on the realistic simulation of the surface boundary conditions (Hoskins and Sardeshmukh, 1987), which include soil moisture and vegetation, snow cover, and the state of the sea surface.

Stone (1987) argued that dynamic models are incapable of handling a small mesoscale phenomenon, such as convection, and that most techniques available for forecasting mesoscale phenomena are applicable mainly in forecasting warm season convection and heavy rainfall. The prospects of better local rainfall forecasts are enhanced by digital processing and transmission of rainfall data from ground-based radars, frequent cloud imagery from geostationary satellites, and the development of interactive computer displays (Brown, 1987). This allows extrapolation of the current situation for very short-term forecasts and the necessary mesoscale detail to initialize a numerical mesoscale model for forecasts beyond the period of realistic extrapolation.

It is important to choose a physical climate model which was designed to investigate processes at the scale of interest. Thus, AGCMs are invalid tools for study at catchment (and perhaps sub-continental) scales, while NMMs cannot represent the full spatial extent of the climate system.

There is now an increasing tendency to work across scales. AGCM users are looking more and more at mesoscale processes, while hydrologists are beginning to investigate macroscale problems. There is a real hope that the embedding technique discussed above may provide the means for modelers to cross scales. In particular, there is an increasing interest about how climate change might affect specific areas. NMMs, driven by observed data cannot investigate these concerns, while AGCMs are spatially too coarse to resolve crucial topographic influences. The embedding of NMMs into

AGCMs, however, can provide the methodology for investigating climate changes at very high spatial resolutions for limited areas (Giorgi, *et al.*, 1989).

For agricultural meteorologists and hydrologists, AGCMs alone are probably a poor tool. They often need to know the climate for a small area and AGCMs cannot provide this information. NMMs can provide these data, but they require high quality initial conditions. Embedding NMMs in AGCMs overcomes this need, and allows full scale climate impacts to be investigated. Embedding techniques are still in the development phase, but results are encouraging. Feedback effects from the NMM to the AGCM cannot yet be included in the climate simulations. Eventually, two-way embedding procedures should be developed which will give modelers the opportunity to investigate climate and hydrological issues at high resolutions.

AGCMs cannot predict the effects of global changes on subregional scales. The embedding techniques discussed above could provide a solution to how to identify the effects on water resources and crop production at the catchment scale.

It is realistic to expect that, with the advance in computational power, the resolution used by AGCMs will be as fine as currently used by NMMs. Until then, embedding NMMs into AGCMs will provide a valuable methodology for linking the mesoscale with the macroscale in selected regions.

CHAPTER 4

ESTIMATION OF AREAL EVAPORATION

4.1 Evaporation/atmosphere interactions

There is increasing awareness of the importance of interactions between hydrological processes at the land surface and the atmosphere.

In the 1950s it was usual to consider that the evaporation at the land surface, when well supplied with water, was directly controlled by meteorological conditions and could be predicted if values for the net radiation, wind speed, temperature, and vapour pressure were given. Differences between vegetation types in terms of physical and biological properties were not considered to be very significant.

By the mid-1960s, the above methods had spread also to non-temperate climates. With the explicit description within the combination equation of the aerodynamic resistance (r_a) and surface resistance (r_s) terms and the realization that both terms could vary widely in time and, particularly for the aerodynamic resistance term, between vegetation types the possibility of very large variations in evaporation rate with time and space and of differences in evaporation rate between different plant and land surface types was recognized.

More recently, the existence of a number of feedback mechanisms, generally negative, which affect evaporation, have been recognized. There is also greater awareness that not only does the evaporation rate itself modify the near-surface meteorological conditions which are driving it, but the near-surface micro-meteorological conditions interact with the surface to enhance biological controls. Recent work by plant physiologists (Farquhar and Wong, 1984; Wong, *et al.*, 1985) suggests that plant water use may be governed by the rate of photosynthesis rather than being a passive response to environmental conditions. It has also been suggested (Cowan, 1982; Farquhar and Sharkey, 1982) that, as a result of selective adaption, many plants maintain water use efficiency more or less constant through stomatal control. Perception of evaporation — being controlled by atmospheric demands established with (nominally) fixed boundary conditions of temperature and humidity at the planetary boundary layer (PBL), the surface radiation budget, and the availability of water at the surface — is being reflected in the development of evaporation models which take into account these biological as well as physical interactions.

The most important interactions are:

(*a*) Water availability at the evaporating surface: A reduced water availability evidenced by a large surface resistance r_s (s m^{-1}) will ensure the partition of more of the available energy into sensible rather than latent heat, which leads to a rise in near-surface and PBL temperatures, increased vapour pressure deficit (VPD), and higher atmospheric demand;

(*b*) Changes in surface resistance: The surface resistance is affected by many biological and physical factors. A completely wet surface has zero surface resistance; dry vegetative surfaces have surface resistances which are largely controlled by stomata. Stomata respond primarily to light intensity, atmospheric vapour pressure, and leaf water status, which itself is often directly dependent on the availability of water to the plant roots;

(*c*) Stomatal responses: The stomatal responses exhibited by many plant species whereby stomata close as VPD increases, which represents a negative feedback on the evaporation rate but a positive feedback on VPD; with closing stomata and increasing surface resistance, r_s, sensible heat will increase, warming the near-surface atmosphere and the PBL and increasing VPD. Actual evaporation rates are, therefore, much more conservative than would be predicted with evaporation models, which do not take this feedback into account;

(*d*) The availability of water: Water availability is, for many parts of the world, the most important mechanism controlling evaporation. It is perhaps the least well understood and, arguably, the area where least resources are being made available for its study. Many evaporation models employ an empirical relationship which moderates evaporation as soil water status decreases. The moderator can be expressed in terms of stomatal resistance or as a factor which is multiplied by the 'potential' evaporation estimate; soil water status can be expressed as a deficit, a dimensionless availability which is the ratio of the soil profile water content divided by the total available water capacity (AWC) or as a water potential if the soil water retention characteristics of the soil are known. Many evaporation models assume, as was the case for the original Penman model (Penman, 1948), that transpiration is not limited until a threshold water content, dimensionless water availability, or water potential, is reached. However, it is realized that the rate at which water is demanded, as well as supplied, is as significant as the absolute amounts;

(*e*) Available water capacity (AWC): On a global scale, our knowledge of AWC and the moderator relationship is meagre. Current versions of AGCMs often assume simple formulations (see section 6.2) such as that of Holloway and Manabe (1971) where AWC is taken as 150 mm and transpiration is not limited until a deficit of 100 mm is reached. However, studies in the United Kingdom have shown that, for grass growing on different soil types, the AWC can range from 150–300 mm; for forests growing on peat soils in Wales, the AWC was in excess of 200 mm; for *eucalyptus* trees and indigenous forest growing in southern India, the AWC was 380 mm.

4.2 Estimation methods for areal evaporation

AGCM grid squares and large basins/catchments are rarely homogeneous in vegetation cover. Stewart (1989) has proposed that AGCM grid squares be subdivided into 'reasonably homogeneous' subareas and that evaporation be determined in each subarea with the Penman–Monteith combination equation, which incorporates biological responses and is driven by near-surface meteorological data generated by the AGCM. Grid square values would then be obtained by aggregating over all subareas.

The Penman-Monteith combination equation is discussed in section 4.2.1. Other approaches widely used at a regional scale include the Priestley–Taylor equation (section 4.2.2), the complementary relationship (section 4.2.3), and several atmospheric boundary layer methods (section 4.2.4).

4.2.1 *The combination equation*

Combination equations represent the energy budget at the land surface and the transfer of water vapour and heat between the surface and the atmosphere. The Penman–Monteith form uses aerodynamic and surface resistances to describe the effect of vegetation on transpiration. The aerodynamic resistance (r_a) describes the effect of surface roughness on heat and mass transfer between the surface and some reference level. The surface resistance (r_s) describes the resistance to the flow of water vapour from the evaporating surface to the ambient air. For a vegetative surface, r_s represents the biological control of transpiration and is largely controlled by stomatal resistance.

The Penman–Monteith equation is strictly true only in steady conditions. It may be used on an hourly or daily basis, provided that the values for the input parameters are determined in a similar way over the same period. The aerodynamic resistance (r_a) depends on stability, wind speed, and roughness. Surface resistance for a water surface is zero; for a drying soil, the surface resistance depends on the flow of water vapour to the surface and, thus, on the depth of liquid water and the permeability for vapour and water transport; for vegetation, the surface resistance depends on physical and biological factors controlling stomata. Submodels for r_s are determined in dry conditions using stomatal conductance and leaf area models (section 4.2.1.1) and in wet conditions using interception models (section 4.2.1.2).

In temperate climates with plenty of water, the actual evaporation for short vegetation may be estimated from the Penman potential evaporation or equilibrium evaporation as obtained with the Priestley-Taylor method (section 4.2.2). Other surface factors become important for trees and specific submodels are needed for such factors to use the Penman–Monteith equation successfully.

The Penman–Monteith combination equation in the form given by Stewart (1989) is given by:

$$E = [\Delta A + \rho c_p\, \delta q/r_a] / [\Delta + c_p \{1 + (r_s/r_a)\} / \lambda] \qquad (4.1)$$

where E is the rate of evaporation (kg m^{-2} s^{-1}), λ is the latent heat of vaporization of water (J kg^{-1}), Δ is the rate of change of the saturated specific humidity with temperature (kg kg^{-1} · C^{-1}), ρ is the air density (kg m^{-3}), c_p is the specific heat of air (J kg^{-1} · C^{-1}) and δq is the specific humidity deficit (kg kg^{-1}). r_a is the aerodynamic resistance to sensible heat and water vapour transfer between surface and reference height (s m^{-1}) and r_s is the surface resistance (s m^{-1}). If $r_s = 0$, then the water supply is unlimited and evaporation occurs at the potential rate. Ignoring the change in heat storage in biomass and canopy air and the net photosynthetic energy, the available energy A (W m^{-2}) is given by:

$$A = Q^* - G \qquad (4.2)$$

where Q^* is net radiation and G is soil heat flux. Stewart (1989) recommends that Equations 4.1 and 4.2 are used with hourly or half-hourly meteorological data.

The aerodynamic resistance r_a as a function of wind speed, surface roughness, and atmospheric stability is given for non-neutral conditions by:

$$r_a = [\ln\{(z - d) / z_0\} - \Psi]^2 / (k^2 u) \qquad (4.3)$$

where z is reference height (m), d is the zero plane displacement level (m), z_0 is surface roughness (m), k is the Von Karman's constant [–], Ψ is a stability function [–], and u is the wind speed (m s^{-1}).

If evaporation measurements are available, the surface resistance r_s may be calculated from Equation 4.1, such that:

$$r_s = [(\Delta\lambda\beta/c_p - 1)\, r_a] + [\rho\lambda\delta q/\lambda E] \qquad (4.4)$$

where β is the Bowen ratio, i.e. the ratio of sensible to latent heat flux.

4.2.1.1 Stomatal conductance models

Stomatal conductance, g_s, depends on radiation, CO_2 concentration in the intercellular space, vapour pressure deficit, leaf temperature, and leaf water status.

Typical responses of the stomata to these five environmental parameters are shown in Figure 3. Avissar and Verstraete (1990) discuss several parameterizations of stomatal resistance/conductance for application in numerical models.

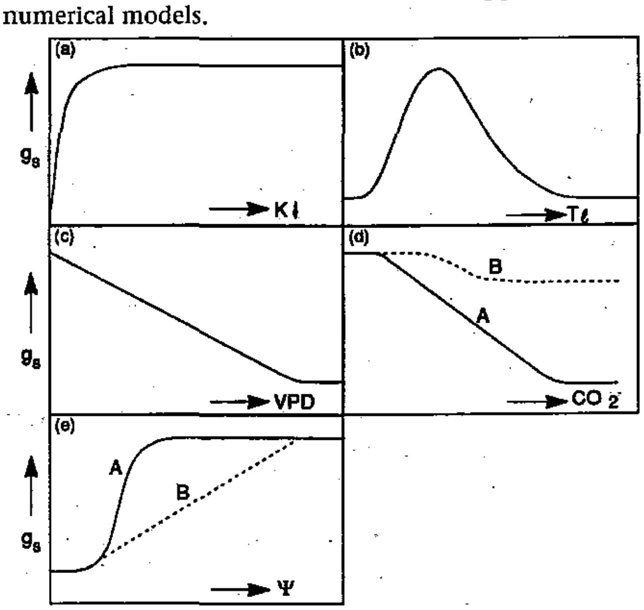

Figure 3 — Typical responses of stomatal conductance g_s to (*a*) solar radiation flux, $K\downarrow$; (*b*) leaf temperature T_l; (*c*) vapour pressure deficit (VPD); (*d*) intercellular CO_2 concentration; and (*e*) cell water pressure potential in leaves (Ψ) (from Avissar and Verstraete, 1990).

Jarvis (1982) has proposed functional relationships for this dependence. A range of parameters must then be determined by non-linear least square regression analysis using measured evaporation values.

Stewart (1989) proposes a model in which surface resistance is expressed in terms of measurements of solar radiation, atmospheric specific humidity deficit, air temperature, and soil moisture deficit. The surface resistance r_s may also be expressed as a function of the soil moisture availability m, the ratio of actual evaporation to potential evaporation. However, one should note the great spatial variability of surface resistance even for a single plant stand.

The Stewart (1989) formulation use a multiplicative relationship between moderators, which relate maximum stomatal conductance to each of the driving variables. For example, the Stewart formulation is:

$$g_s = g_{max}\, g(L)\, g(K_d)\, g(\delta q)\, g(T_l)\, g(\delta\theta) \qquad (4.5)$$

where g_s is the surface conductance and g_{max} is the maximum conductance when each of the moderators relating to leaf area (L), solar radiation (K_d), atmospheric specific humidity deficit (δq), leaf temperature (T_l), and soil moisture deficit ($\delta\theta$) has a maximum value of unity.

Price and Black (1989) propose using 'limited factor analysis' whereby stomatal conductance is calculated as the minimum of the separate moderator functions with the maximum value, i.e.:

$$g_s = minimum\ \{g1\ (W),\ g2\ (D),\ g3\ (S),\ g4\ (t)\} \qquad (4.6)$$

where $g1\ (W)$ relates to radiation, $g2\ (D)$ to vapour pressure deficit, $g3\ (S)$ to radiation and $g4\ (t)$ to time of day. This formulation implies a threshold relationship with the controlling variables whereby the conductance is only sensitive to changes in one of the variables at any one time.

Calder (1978, 1990) used a function which combined elements of the multiplicative formulation to describe the vapour pressure dependence and a seasonal relationship of r_s with a threshold relationship to describe the effects of radiation ($r_s = \infty$ when the net radiation $R_n < 0$).

The use of the different formulations, multiplicative or threshold, implies a fundamentally different sensitivity of r_s with the driving variables, and would predict very different sensitivities in any hydrological or meteorological model when used to investigate sensitivity of model predictions to changing input data (e.g. climate change data). Unfortunately, data sets may not be sufficiently accurate or extensive to resolve which of the formulations, or a mixed formulation, is most appropriate to describe the stomatal response from vegetation.

4.2.1.2 Interception models

Central to the calculation of evaporation rates with the combination equation is the determination of whether the evaporating surface is totally wet — in which case the surface resistance can be equated to zero — or dry, when an appropriate value for the surface resistance must be sought.

The wet surface determination is normally accomplished by using an interception model in which the change of storage of the interception store is calculated from the precipitation input to the canopy minus the evaporation and net rainfall (throughfall plus stemflow). Rutter, *et al.* (1971) were the first to formulate a model of the canopy water balance which described the rate at which the canopy wetted, the rate at which net rainfall took place, and the rate of evaporation. In the Rutter model, all of these components were directly linked to the amount of canopy storage. Inputs to canopy storage were assumed to be equal to the precipitation rate minus a constant p, the 'free throughfall fraction'. Drainage from the canopy was uniquely related to canopy storage via two drainage parameters (b and k), and evaporation was assumed to be proportional to canopy storage for storage less than the canopy storage capacity value S.

The model can be summarized in terms of the continuity equation:

$$dC/dt = Q - k\,(exp\,(bC) - 1) \qquad (4.7)$$

where:

$$Q = (1-p)\,R - EC/S \quad \text{for } C<S \qquad (4.8)$$
$$Q = (1-p)\,R - E \quad \text{for } C>S \qquad (4.9)$$

where:

C = canopy storage (mm)
R = precipitation rate (mm min^{-1})
b = drainage parameter (mm^{-1})
k = drainage parameter (mm min^{-1})
p = free throughfall fraction (dimensionless)
S = canopy storage capacity (mm)
Q = rate of precipitation plus evaporation (mm min^{-1})
E = Penman–Monteith evaporation rate in wet conditions (mm min^{-1})

The Rutter type models successfully predicted evaporation losses in United Kingdom conditions but were not able to describe the interception process in tropical conditions where, typically, rainfall intensity and raindrop size were larger. Calder (1986) described a stochastic interception model which, in contrast to the Rutter type models, related net rainfall rate to precipitation rate and to drop size. The model was able to account for observations, from both laboratory studies using rainfall simulators (Aston, 1979) and field studies (Calder, *et al.*, 1986) which were irreconcilable with the Rutter type estimates. It was found:

(*a*) That for the same canopy storage, net rainfall rates are higher in the early, wetting up stages of the storm as compared with the later, drying out phase; and

(*b*) That maximum canopy storage is approached in a 'gradual' manner and is only obtained after application of a considerably greater depth of rainfall than the capacity value.

The model uses the Poisson distribution function to estimate n, the mean number of drops retained per elemental area of the canopy, as a function of m, the mean number of drop strikes per element. It predicts an asymptotic wetting of the canopy at rates dependent upon the size of the drops, as indicated by the parameter q, which is the maximum number of drops which can be retained by a surface element. The basic equation of the model (see also Hall, *et al.*, 1992) gives the mean number of drops retained per element as:

$$n = q\,(1 - e^{-m}) + e^{-m} \sum_{i=1}^{r} \left[(i - q)\, m^i / i!\right] \qquad (4.10)$$

where r is the truncated value of q. The depths of canopy storage C and rainfall P are related to the model parameters through the relationships:

$$C = nvL \qquad (4.11)$$

$$P = mvL\, \kappa^{-1} \qquad (4.12)$$

where v is the mean volume of the raindrops, L is the mean number of elemental surfaces per unit ground area and κ is the projected area of the canopy per unit ground area. v and L are completely interdependent; the maximum storage capacity is given by:

$$Cmax = q(v)L \qquad (4.13)$$

Unlike the interception models developed by Rutter, *et al.* (1971) the stochastic model does not calculate net rainfall rate from the canopy in terms of a unique relationship with canopy storage; rather it predicts the ratio of the net rainfall rate to the rainfall rate, which is, for a given value of q, uniquely related to the proportional canopy storage, $C/Cmax$ (Figure 4).

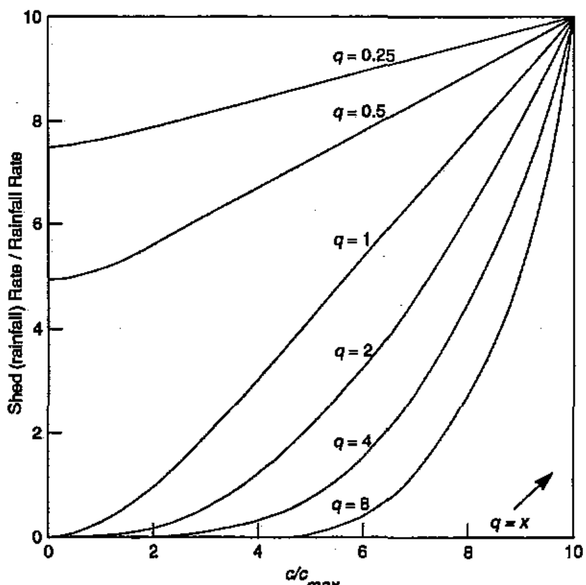

Figure 4 — The dimensionless relationship between the instantaneous proportion of rain shed from the canopy (the ratio of net rainfall rate — throughfall and steamflow — to the actual rate of precipitation input) and canopy storage expressed by q (the maximum number of raindrops retained by a surface element) (after Calder, 1986).

For operations within AGCM models, the description of the storm size and the spatial wetting of the canopy within the AGCM grid becomes important and probability functions have been suggested to take this spatial variability into account. Shuttleworth (1988) described a scheme, following a method proposed by Milly and Eagleson (1982) for distributing moisture and heat fluxes across a grid square, whereby the local rainfall rate in the rain-covered part of the AGCM grid was assumed to follow a negative exponential probability distribution:

$$f(R_1) = (\mu/R)\, exp\,(-\mu R_1/R) \qquad (4.14)$$

where R and R_1 are the model-generated grid-average rainfall rate and the local rainfall rate, respectively and μ is the fraction of the grid in which rain occurs. Assuming that net rainfall rate is independent of rainfall rate (i.e. a Rutter type model), Shuttleworth showed that the local net rainfall intensity, T_1, is given by:

$$T_1 = R_1 - (S - C)/\Delta t \;\text{ if }\; R_1 > (S - C)/\Delta t \qquad (4.15)$$

$$T_1 = 0 \;\text{ if }\; R_1 < (S - C)/\Delta t \qquad (4.16)$$

where Δt is the time step of the model. This scheme predicts that no net rainfall occurs locally until the canopy is saturated. The average net rainfall over the grid square is then found by integration to be:

$$T = R\, exp\,\{-\mu\,(S - C)\,/\,(R\Delta t^2)\,\} \qquad (4.17)$$

Dolman and Gregory (1992) report that a similar scheme, which incorporates an explicit dependence of throughfall on rain, has been incorporated in the U.K. Meteorological Office Unified Modelling System. The model formulation is:

$$T_1 = R_1 - (S - C)/\Delta t \;\text{ if }\; R_1 > S/\Delta t \qquad (4.18)$$

$$T_1 = R_1\,(C/S) \;\text{ if }\; R_1 < S/\Delta t \qquad (4.19)$$

Within the framework of the stochastic interception model, both these formulations are particular cases. They assume that the Shuttleworth formulation, which implies zero net rainfall until the canopy reaches saturation, would be described with a q value of infinity (applicable in conditions with small drop size, mist or low intensity rain) and the Unified Modelling System would be applicable in conditions where q is about 1.0, which would be appropriate for a tree species such as *eucalyptus maculata* with raindrops of about 2.5 mm in diameter (Aston, 1979). These two situations would be just part of the continuum of possible q values within the stochastic model framework.

Dolman and Gregory (1992) have investigated the predictive ability of these two schemes and have found reasonable agreement with interception loss measurements for tropical forests in Amazonia with a value of 0.1 for μ.

4.2.1.3 APPLICATION OF THE PENMAN–MONTEITH EQUATION: SCALING UP

The Penman–Monteith Equation 4.1 estimates evaporation as a function of vegetation characteristics and environmental conditions. The expression requires meteorological data at only one height. If the effect of vegetation on transpiration is expressed adequately in the surface resistance model, the Penman–Monteith equation is a physically and biologically realistic method. Stewart (1989) suggests that aggregation of individual subareas, which are homogeneous in vegetation, will then provide a reasonable estimate of areal evaporation.

The Penman–Monteith approach is also called a big leaf model. The model is based on the assumption of the validity of the K-theory (i.e. first-order closure or Fick's diffusion law); it assumes that the turbulent flux is linearly related to the mean gradient in the corresponding driving force in the canopy. First-order closure often fails to describe adequately turbulent transport in plant canopies. The 'effective diffusivity' is often more a function of the distribution of heat and water vapour sources and sinks within the canopy than of the turbulence level. The distribution of sources and sinks has implications for the calculation of fluxes within and just above the canopy based on the assumption of similarity of the heat and water vapour diffusivities (K_h and K_v). Higher-order closure theory should be included to deal better with within-canopy transport.

Theoretical approximations are possible for the scaling-up from leaf to canopy of r_a and r_s. Several processes associated with evaporation operate differently or differ in significance between leaf and canopy. These include the interrelationship between evaporation, humidity and stomatal conductance (canopy development, r_a/r_s and the coupling factor); the effect of leaf width on evaporation (change in radiation penetration); advection (problem of scale, different processes); canopy heat storage (woody biomass); and soil evaporation.

Higher-order closure models have been used successfully above canopies. Within canopies, they need extensive parameterization. Shortcomings of using K-theory in canopies, as is done with the Penman–Monteith approach, are quite fundamental; the diffusion across a plane at any level is not just a function of the concentration gradient and an effective diffusivity, determined by the turbulence field at that level. The diffusivity is also a function of the distribution of sources and sinks.

The assumption of similarity between the diffusivities for sensible heat (K_h, m^2 s^{-1}) and latent heat (K_v, m^2 s^{-1}) is only valid if the distribution of sources and sinks of heat and water vapour is the same below $z = 2H$, where H is canopy height. Above $z = 2H$ the assumption is always valid.

The simplest higher-order closure models are driven by the same number of above-canopy meteorological variables as needed for Penman–Monteith, but they need additional information on the leaf area distribution, radiative divergence in the canopy, stomatal conductance, and soil moisture conditions. The Penman–Monteith Equation 4.1 lumps all these factors in the canopy resistance, and the r_s term makes this method, therefore, rather unpredictable.

4.2.1.4 Simulation of soil moisture deficit

Soil moisture data are not routinely available. Simulated, rather than actual, soil moisture values are used in Atmospheric General Circulation Models (AGCMs). Dolman, *et al.* (1988) used a very simple water balance model to simulate soil moisture availability. They compared estimated and measured soil moisture deficit, found reasonable agreement and then used the simulated soil moisture values to estimate surface resistance r_s for use in the Penman–Monteith equation. A similar approach has been proposed by Stewart (1989) who expresses the surface resistance as a function of soil moisture availability, defined as the ratio of actual to potential evaporation.

Calder, *et al.* (1983) have compared the performance of different soil moisture deficit models with neutron probe observations of soil moisture deficits at six sites under grassland in the United Kingdom. All combinations of five estimates of potential evaporation (constant, mean climatological, Priestley–Taylor, Penman, and Thom-Oliver) and six different soil moisture regulating functions (no root constant, Penman root constant, layer, optimized linear, optimized exponential and optimized layer), giving 30 different model formulations, were investigated. As can be seen from Figure 5, they concluded that the most successful models, with respect to minimizing the sums of the squares of the residuals between observed and predicted soil moisture deficit, were those using the mean climatological or Penman potential evaporation estimates with the linear or layer regulating functions.

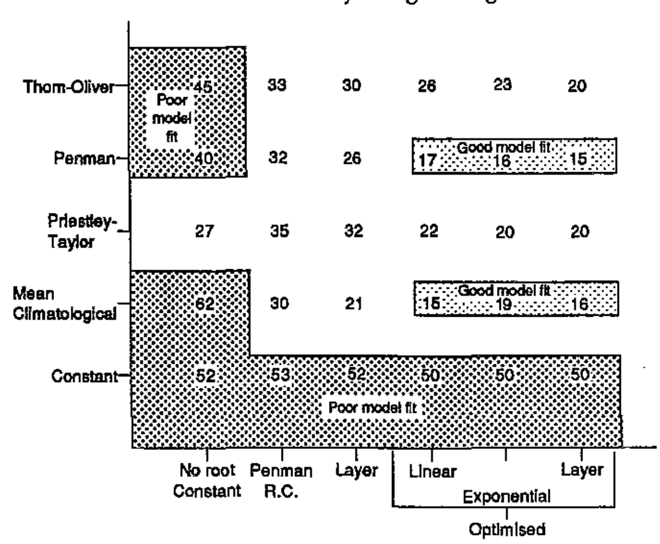

Figure 5 — The relative performance of a particular model over all the experimental sites in terms of the mean of the model error (mm) at all sites, shown for all combinations of root constant functions and evaporation estimates (from Calder, *et al.*, 1983).

The success of the mean meteorological formulation (a sinusoidal function with time of the year, which was fitted to published average Penman E_T data for Northamptonshire county, U.K.), over all sites and all years, was attributed to a negative feedback interaction between the stomatal response of the grass and the atmosphere which limits transpiration in times of high vapour pressure deficits. Stewart (personal communication) has also reported that during the course of the First ISLSCP Field Experiment (FIFE) in Kansas, a similar dependence of grass stomatal response on atmospheric vapour pressure deficit was inferred. The success of the simple mean climatological formulation, in comparison with more detailed models, which require daily observations of the meteorological variables, has advantages for many practical applications and has been incorporated into on-line flood forecasting models run by British water authorities.

Although some formulations of the regulating function were more successful at some sites than others, over all sites there was little difference in prediction capability between any of the optimized regulating functions. However, the success of the optimized regulating functions in comparison with those having fixed parameter values clearly demonstrates the importance of estimating this parameter accurately for each soil/crop combination that is of interest. The layer formulations, although not significantly better than the other formulations over all sites and all years, did have the advantage that they predicted higher evaporation rates following rainfall even though the whole profile remained with a high soil moisture deficit, a prediction which was generally borne out by observation.

4.2.2 *Priestley–Taylor expression*

Priestley and Taylor (1972) argued that, in the case of evaporation from large wet areas at the scale of the grid used in computer solutions of numerical weather prediction (NWP) models (i.e. 100 x 100 km), radiation, rather than advective controls, must dominate.

Priestley and Taylor suggested that equilibrium evaporation, i.e. the lower limit to evaporation from wet surfaces with a very long fetch, may be expressed as:

$$\lambda E = [\varepsilon /(\varepsilon + 1)]\,(Q^{*}-G) \qquad (4.20)$$

if the atmosphere remains saturated when in contact with a wet surface. In this expression, $\varepsilon = s\ \lambda/Cp$, with s being the slope of the saturation specific humidity curve, i.e. dq^*/dT; λ is latent heat of vaporization of water and C_p is specific heat of air at constant pressure.

For evaporation under average conditions, with minimal advection it was proposed that:

$$\lambda E = \alpha\,[\varepsilon/(\varepsilon + 1)]\,(Q^{*}-G) \qquad (4.21)$$

with α = 1.26, an empirical constant which represents the ratio between evaporation from a wet surface under conditions of minimal advection expressed in Equation 4.21 and the equilibrium evaporation given in Equation 4.20. This equation is often used as an estimate of potential evaporation in the absence of local advection. It can also give good estimates for evaporation in much smaller regions, from well-watered vegetation.

Regional evaporation implies that the area is large enough to ignore horizontal advection. Thus, the Priestley–Taylor approach minimizes local horizontal advection and assumes that the effects of vertical interactions between the air near the ground and the atmosphere, at large, are independent of normal variations in large-scale atmospheric conditions.

McNaughton and Spriggs (1986) have commented that Priestley and Taylor ignored the energy exchanges across the top of their very large box, i.e. at the upper limit of the air directly interacting with the surface.

4.2.3 *Complementary relationship*

Bouchet (1963) considered that potential evaporation is as much the effect of the actual evaporation as its cause. Heat and moisture released from the surface will modify T and q of the air above it. He suggested that the increase in potential evaporation observed above an area as it dries out, may be used as a measure of the real evaporation rate from the area.

If actual evaporation E is reduced below the potential rate E_{po} for an extensive wet region, an amount of energy Q would be released so that:

$$\lambda E_{po} - \lambda E = Q \qquad (4.22)$$

It is assumed in this approach that this energy release would not affect the net radiation much. Its effect would be on temperature, humidity and turbulence, and hence on potential evaporation E_p. If the area is big enough so that the change in energy does not result in changes in the transfer of energy between the modified air mass and that beyond, Q should just equal the increase in λE_p, the potential evaporation for the drying region. Thus:

$$\lambda Ep - \lambda E_{po} = Q \qquad (4.23)$$

therefore:

$$E + E_p = 2E_{po} \qquad (4.24)$$

Most applications of the complementary relationship have been concerned with finding appropriate expressions for E_p and E_{po}. Morton (1983) uses the Priestley–Taylor expression for E_{po} and a Penman expression for E_p. The complementary approach assumes that advection is not important. It also assumes that Q remains constant. If advective effects were to be reduced as evaporation decreases — perhaps because the air gets drier and more like the surrounding air, and the region itself is less of a sink for energy — then:

$$E + E_p < E_{po} \qquad (4.25)$$

This complementary approach is increasingly used in hydrological applications for large areas because it uses essentially only standard climatic data. However, the constancy of net radiation has never fully been tested and doubts have been expressed about the use of the Priestley–Taylor expression. In addition, this approach ignores vertical advection, i.e. air being blown down from large-scale air masses brought in by planetary weather systems.

4.2.4 *Atmospheric boundary layer methods*

Brutsaert (1988) has noted that improved parameterization at the regional scale will need a better understanding of the turbulent transport mechanisms in the atmospheric boundary layer (ABL). The ABL thickness is of the order of 10^3 m, tending to integrate at the regional scale; observed profiles in the ABL reflect surface conditions over upwind fetches of the order of tens of kilometers. The (daytime) convective boundary layer is of particular interest for regional evaporation studies because the surface fluxes tend to be greatest under unstable conditions.

There is a great need for better understanding of ABL processes and for the development of more generally applicable models. The current models do not cover a sufficiently wide range of atmospheric conditions for operational use. The many problems which need to be addressed include problems associated with spatial heterogeneity and non-steady conditions; condensation processes; stable night-time conditions; imperfect mixing; uncertainty about the entrainment processes; and the parameterization of surface layer transport processes.

4.2.4.1 BULK TRANSFER METHOD

Brutsaert and Mawdsley (1976) represent the surface fluxes of water vapour E and of sensible heat H over a extensive uniform surface with the following bulk transfer equations:

$$E = ku^* \rho\, (q_1 - q_h)\, [\ln \{(h - d)/(z_1 - d)\} + \Psi_v \{(z_1 - d)/L\} - D] \tag{4.26}$$

$$H = ku^* \rho C_p\, (\theta_1 - \theta_h)\, [\ln \{(h - d)/(z_1 - d)\} + \Psi_H \{(z_1 - d)/L\} - C] \tag{4.27}$$

where k is the Von Karman constant, u^* is friction velocity, d is zero plane displacement, C_p is volumetric heat capacity, h is height of ABL, z_1 is the level in the surface layer (<100 m), θ is potential temperature, q is specific humidity and L is the Monin Obukhov stability length. $\Psi_v\ (z/L)$ and $\Psi_H\ (z/L)$ are the Monin Obukhov stability functions for latent and sensible heat for the surface layer. The Monin Obukhov length L is given by:

$$L = -u^{*3}\, \rho\, /[kg\, \{(H/T\, C_p) + 0.61E\}] \tag{4.28}$$

where g is the acceleration of gravity.

The equation for friction velocity u^* is:

$$u^* = k\, V_h\, [[\ln \{(h - d)/z_0\} - B]^2 + A^2]^{-1/2} \tag{4.29}$$

where V_h is the mean wind speed at the top of the ABL and z_0 is the surface roughness. Empirical expressions have been derived for the similarity functions A, B, C and D as functions of $(h - d)/L$ (see Brutsaert, 1982; 1986).

If profiles of wind velocity, specific humidity and potential temperature are available, then the three fluxes may be determined simultaneously from Equations 4.26 to 4.29 without great difficulty.

In recent years, Brutsaert and Sugita (1991; 1992*a*; 1992*b*) have successfully obtained regional surface fluxes with a bulk similarity approach using remotely sensed surface temperature data in combination with temperature and wind profiles obtained with radiosondes.

4.2.4.2 MIXED LAYER MODELS

These models assume that the atmospheric boundary layer consists of a well-mixed layer above a very thin surface sublayer. During daytime, the sensible and latent heat fluxes are upwards, and the well-mixed boundary layer increases in height. The convective boundary layer (CBL) is capped by a thin inversion. It is assumed that the potential temperature θ and the specific humidity q are uniform throughout the well-mixed layer except in the surface sublayer.

The diurnal growth of the mixed layer takes place by erosion of the base of the capping inversion. This air is incorporated into the bulk air of the CBL. During the first hours after sunrise, the capping inversion is the remnant of the ground-based inversion formed during the night. The CBL quickly grows through this inversion layer. For the remainder of the day, the capping inversion is one which is caused by weather processes at a much larger scale. Its properties are independent of the surface energy budget of the local region.

The mixed layer model considers the rate of entrainment of air from above the mixed layer, its effect on the temperature and humidity of the bulk air in the mixed layer, and the changes in the surface energy balance. The convective boundary layer is thus represented as a well mixed layer of thickness h, potential temperature θ_m and specific humidity q_m. Below the CBL is a shallow surface layer. Above the CBL is the inversion with known temperature and humidity profiles $\theta_s\ (z)$ and $q_s\ (z)$.

The evolution of θ_m and q_m may be written with the following conservation equations for heat storage:

$$h\ d\ \theta_m/dt = H/(\rho Cp) - \langle w'\ \theta' \rangle \tag{4.30}$$

and for vapour storage:

$$h\ d\ q_m/dt = E/\rho - \langle w' q_h' \rangle \tag{4.31}$$

where h is the thickness of the CBL, ρC_p is volumetric heat capacity, H is sensible heat flux at the surface and E = actual vapour flux at the surface. The turbulent fluxes of sensible and latent heat resulting from entrainment of warmer and dryer air at the top of the ABL are given by $\langle w'\ \theta_h' \rangle$ and $\langle w'\ q_h' \rangle$.

Thus the upward buoyancy flux from the land surface and the downward buoyancy flux at the top of the ABL will result in growth of the CBL. Several entrainment hypotheses have been proposed (McNaughton and Spriggs, 1986; Raupach, 1991) which yield expressions for the change in the height of the CBL, *dh/dt*.

The lower boundary conditions at the land surface may be established with the general surface energy balance equation and the Penman–Monteith combination equation based on the big leaf model for:

(*a*) The water vapour flux across r_s between the vegetated surface at saturation specific humidity at canopy temperature T, and the bulk air in the canopy at actual q; and

(*b*) The water vapour and sensible heat fluxes between the canopy and the bulk CBL across r_a. In such expressions, r_s is the canopy resistance of the vegetation covering the uniform regional surface and r_a is the aerodynamic resistance of the surface layer. McNaughton and Spriggs (1986) and McNaughton (1986) used this model to predict latent and sensible heat fluxes at the surface with half-hourly time steps from initial temperature and humidity profiles obtained with balloons or radiosondes and time varying input data for net radiation, soil heat flux, aerodynamic resistance r_a and surface canopy resistance r_s.

Brunet, *et al.* (1991) describe a similar mixed layer model. It requires incoming solar radiation, surface roughness, surface albedo, geostrophic wind, and early morning temperature and humidity profiles as external inputs. The model calculates the growth in the height of the mixed layer and computes the surface fluxes in the turbulent surface layer. The only unknown variable is surface resistance. Brunet, *et al.* (1991) established an empirical relationship between surface resistance and surface temperature using ground data and then employed satellite derived surface temperature data to obtain regional estimates of surface resistance for use with the mixed layer model.

A combination of surface and mixed layer model is described by Diak and Stewart (1989). The model relates atmospheric measurements of the height of the planetary boundary layer obtained at a synoptic station and geostationary satellite measurements of the diurnal range in skin surface temperature. Diak (1990) has applied the method to determine the regional variation of fluxes and roughness in the central U.S.A. Preliminary error analysis indicates that expected errors in 12-hour flux totals are in the range of 1 MJ m^{-2}.

Raupach (1991) has recently addressed the problem of generalizing land surface models from homogeneous surfaces to heterogeneous surfaces in which local advection is prominent. He pointed out that land surface heterogeneity should be treated at the scale of the planetary boundary layer. He described a convective boundary layer (CBL) model for homogeneous conditions with various expressions for mixed layer growth. New expressions are proposed for the boundary conditions of latent and sensible heat at the surface. This CBL model is then extended to heterogeneous surfaces by introducing for the convective boundary layer a horizontal length scale $X = hU/w$, where h is the CBL height, U the horizontal velocity and w is the convective velocity, scale. When the length scale of individual land surface patches is much less than X, the heterogeneity is averaged by the CBL. If the surface patches exceed X, separate CBL developments occur over each patch.

CHAPTER 5

USE OF REMOTE SENSING IN THE STUDY OF LAND SURFACE PROCESSES AT A REGIONAL SCALE

5.1 Introduction

Remote sensing data obtained with satellites include images of the Earth's surface and information on the vertical structure of the atmosphere. Currently operating systems include geosynchronous satellites (GOES, GMS and METEOSAT), the NOAA polar orbiting satellites, the Landsat system, and the sun-synchronous SPOT satellites. The sensors on these satellites include radiometers, panchromatic and multispectral scanning radiometers, infrared sounders, and microwave sounding units. These sensors measure the electromagnetic energy impinging on their detectors. The detectors transform this energy into an electric signal which is amplified, digitized, transmitted to ground stations, archived, and recorded on magnetic tape. Becker, *et al.* (1988) describe how two types of models are required to convert satellite data into the desired information. These are geometric models to relate the position of the pixel in the recorded image to the correct position at the Earth's surface, and energetic models which relate the signal received at the top of the atmosphere to the relevant quantity on Earth.

Satellite data is used to obtain physical surface parameters (such as reflectivity, emissivity and surface roughness) and information on the atmosphere (e.g. temperature and humidity profiles, cloud cover, rainfall) as inputs into models for estimating the components of surface radiation and energy balances, for inventories of water resources, and for large-scale and mesoscale climate models. The use of remote sensing data for physical climate studies and for climate modelling is described by Becker, *et al.* (1988). Sellers, *et al.* (1990) have prepared a review of satellite data algorithms for use in global monitoring of important land surface-atmosphere interactions. Their review is based on a 1987 workshop held at the Jet Propulsion Laboratory (JPL) in Pasadena and addresses radiation balance components, vegetation index, soil moisture, and vegetation cover. Its aims were, *inter alia*, to assess the current status of the algorithms used to determine land surface parameters, and to review the methodologies that use these parameters in the estimation of surface energy balance. Schultz (1988) has discussed hydrologically-relevant remote sensing platforms and sensors and has provided several examples of the use of remote sensing information in hydrological modelling. These include the computation of historic monthly runoff for design purposes with the aid of a lumped system model using NOAA infrared satellite data as input, and real-time flood forecasting applying a distributed system model using radar rainfall measurements as input.

A wide range of hydrological and agricultural applications of remote sensing data are discussed by Fraysse (1980); Berg (1981); Goodison (1985); Barrett (1988); and Rango (in press). These applications include estimation of precipitation, remote sensing of snow, ice, surface water, soil moisture and ground water, land cover mapping, and the estimation of crop yield and biomass.

Van de Griend and Engman (1985); Barrett (1988); Schultz (1988); Kuittinen (1992); and Rango (in press) review hydrologically relevant platforms and sensors and discuss remote sensing applications in hydrological modelling. The development and testing of a remote sensing based hydrological model is described by Groves, *et al.* (1985). The use of remote sensing data for physical climate impacts studies and for climate modelling is described by Becker, *et al.* (1988). Applications in agronomy and agricultural meteorology include estimation of biomass and crop yield, determining water stress and soil moisture status, and estimation of evaporation.

Significant progress has been made in the use of remotely sensed thermal imagery and radar for obtaining regional-scale estimates of soil moisture availability and evaporation. Such techniques are based on the direct use of surface temperature measurements and on the use of radiation balance estimates obtained from thermal and visible imagery.

The use of remote sensing data for estimating precipitation is addressed below, as well as assessing soil moisture status, estimating evaporation, radiation balance components, and assessing land cover and biomass.

5.2 Rainfall

Rainfall distribution may be very non-uniform and non-structured due to small convective elements in large cloud systems, orographic effects, atmospheric stability effects, and wind conditions. Significant differences may occur over a few kilometres in large-scale cyclonic storms and over hundreds of metres in thunderstorms.

Precipitation data are required in a wide range of applications. For crop yield modelling and soil moisture evaluation, the required accuracy of daily rainfall is 10–30 per cent, with a horizontal resolution of 20–100 km. However, there is a serious deficiency of rainfall data at such a resolution over most of the globe and especially in semi-arid regions.

Satellite-based rainfall monitoring methods may yield improved estimates of areal rainfall in real or near-real time, especially when integrated with ground-based measurements.

Satellite-based remote sensing for rainfall estimation includes visible and thermal infrared imagery and microwave measurements.

Two quasi-operational methods based on visible and infrared satellite data are the cloud-indexing method, which uses visible and/or infrared data from polar-orbiting and/or geostationary satellites, and the life history method, which is based on infrared data from geostationary satellites (see also Kuittinen, 1992).

The cloud-indexing method (Barrett, 1981) uses satellite cloud images and assigns indices to each grid cell of a grid superimposed on the image which relate cloud type, cloud cover, amount and temperature to the probability and intensity of rain associated with those clouds. The method requires elaborate index calibration procedures which use detailed ground reports.

The life history method (Scofield and Oliver, 1977; Schofield, 1985) employs geostationary satellite data to track the life cycle of clouds. This method has been used widely in convective situations. The method is based on the assumption that rain will fall from the colder clouds and that rainfall intensity will depend on the size, altitude, and growth rate of the clouds. A fully automated process is described by Griffith, *et al.* (1978). Schofield (1985) presents techniques based on this method for estimating rainfall from thunderstorms, tropical cyclones, and extratropical cyclones.

Both methods have limitations. The use of visible imagery is limited by the fact that different cloud types may have similar brightness, the transparency of thin clouds, and reflection from sea surfaces. Thermal imagery is affected by water vapour absorption differences and complexities of cloud heights and temperature inversions. Methods based on both types of imagery are affected by the presence of several cloud layers, differences between sensors, and similar appearance of different surfaces.

Barrett (1985) summarizes the chief advantages and disadvantages of current and proposed satellite-based rainfall monitoring methods. Barrett notes that special emphasis has been given in recent years to developing interactive rainfall monitoring methods, which employ both manual and automated techniques. One such system provides improved rainfall data from the world's major crop-growing areas for input to the Agriculture and Resources Inventory Surveys through Aerospace Remote Sensing (AgRISTARS). The needs and opportunities for satellite monitoring methods in developing countries are discussed by Barrett (1983).

The Department of Meteorology at the University of Reading, United Kingdom has developed a rainfall estimation method for semi-arid regions, which is based on relating cloud-top temperatures below a certain threshold value to actual rainfall. This cold cloud duration (CCD) technique (Milford and Dugdale, 1987) was developed to estimate 10-day and monthly rainfall in the Sahel from METEOSAT thermal imagery.

Sahelian rainfall is largely produced by thunderstorm systems with clouds which extend high in the atmosphere. They have cold tops and can be recognized on METEOSAT images. Rainfall may, thus, be related to the duration in which an area is covered by cold clouds. The CCD is determined for each pixel from hourly images by obtaining the length of time over a 10-day period or a month in which the cloud top temperature is below a certain threshold temperature. These CCD values may then be mapped. The choice of threshold temperature is critical. It must be high enough to include all rain-bearing clouds and low enough to exclude those clouds not associated with rainfall. The success of this technique depends on determining the optimum threshold value (which depends on the season and latitude) and arriving at a relation between CCD and rainfall (through calibration with ground-based rainfall measurements).

The CCD method which was developed and tested for west African squall systems is now also used outside the Sahel. The ARTEMIS system of the Food and Agriculture Organization performs real-time acquisition of hourly METEOSAT data and processes the images to produce CCD maps and maps of 10-day and monthly rainfall. Huygen (1989) has recently reported on a study to establish CCD-rainfall relationships for Zambia, with rain bearing systems which differ significantly from those in the Sahel. He observed that the coefficients of simple linear regressions relating CCD to 10-day rainfall were time dependent. Huygen also describes the use of linear regression and (co)kriging methods for interpolating ground-based rainfall measurements with the help of METEOSAT thermal imagery.

The use of ground-based radars and satellite microwave radiometry are discussed by Rango (in press). The absorption and transmission of microwave radiation is affected by liquid water. Microwave radiation from the surface and the atmosphere has been measured with passive microwave systems on Nimbus satellites. Although microwave methods are more physically based than the methods referred to above, the results with passive microwave thus far have shown promising results over oceans where the elevation in brightness temperature is related to rainfall rate. Inference of rain over land is possible with microwave radiation at high frequencies because raindrops become strong scatterers of microwave at frequencies exceeding 80 GHz (Choudhury, 1991).

5.3 Soil moisture

Remote sensing techniques based on visible imagery, thermal imagery, and microwave radiometry have been developed in the last 10–15 years for estimating soil moisture at different scales (Vauclin, 1983; Kuittinen, 1992; and Rango (in press)). Soil moisture is a state variable which is the major determinant of crop production. It is, therefore, of crucial importance to monitor soil moisture in time, both at the local and the regional scale.

It is generally observed that solar radiation reflected from bare soil decreases at all wavelengths as the soil moisture content increases. Spectral reflectance measurements have, thus, provided a first-order estimation of soil moisture status. However, the reflectance also depends on chemical composition, structure, texture, and roughness of the soil. Thus, absolute values of soil moisture content cannot be obtained (Rango, in press). In the presence of vegetation, which obscures the soil, reflectance data provide important input in radiation and energy balance calculations for the vegetated surface.

Thermal infrared techniques relate surface temperature and (near) surface soil moisture. Reginato, *et al.* (1976) demonstrated the existence of inverse relationships for bare soil between $(T_{max}-T_{min})$ soil or $(T_{soil}-T_{air})_{max}$ and soil water content below field capacity. Jackson, *et al.* (1977) have shown that the concept of surface-air temperature differential shortly after solar noon as a indicator of surface soil water content appears promising for use on cropped surfaces as well.

Such observations relate to the thermal inertia of the soil, which is an expression of the resistance of the soil to a change in temperature. Thermal inertia P is defined is:

$$P = \sqrt{\rho \, c_p k} \tag{5.1}$$

where ρ is the density, c_p is the specific heat and k is the thermal conductivity of the soil (Mulders, 1987). Soil moisture increase will increase each of the right hand side terms and P will increase resulting in a decrease of the diurnal surface temperature amplitude. Thus, thermal imagery obtained near the time of the daytime maximum and the night-time minimum temperatures may be used to determine thermal inertia and, hence, soil moisture content.

Satellite-based thermal inertia mapping as a means of regional soil moisture monitoring has been addressed by numerous authors (Blanchard, *et al.*, 1974; Price, 1977; Carlson, *et al.*, 1985; Carlson, 1986; and others). Applications in semi-arid regions include work by Abdellaoui, *et al.* (1982) and Milford, *et al.* (1983).

The use of visible and infrared imagery is complicated by atmospheric effects due to humidity and clouds. Microwave observations on the other hand can be made through clouds (Rango, in press). Microwave sensors for soil moisture measurement in irrigated agriculture are discussed by Jackson, *et al.* (1980). Passive microwave sensors which measure thermal emission from the soil, (see Jackson, *et al.*, 1984) and active microwave sensors, which measure the back-scattered fraction of the emitted electromagnetic signal, (see for example Jackson and Schmugge, 1981) are also increasingly used for general soil moisture monitoring. Schultz (1988) and Choudhury (1991) describe several promising approaches based on active and passive microwave sensors. Yet most microwave applications are faced with the problem of space resolution versus sensor size. For this reason, microwave techniques have largely been restricted to airborne and ground-based platforms. Comparisons between active microwave-based soil water content determinations and surface moisture availability estimates based on thermal infrared measurements are described by Perry and Carlson (1988). Choudhury (1991) indicates that the effects of surface roughness and vegetation on the brightness temperature make the longer wavelengths (or lower frequencies) more suited for soil moisture estimation. He also reports on case studies done to estimate soil wetness using microwave sensors on board of satellites, including Skylab and Nimbus 5, 6 and 7. The coarse spatial resolution does not allow direct relationships between measured soil moisture and brightness temperature. However, these case studies have successfully evaluated the relationship between the regional antecedent precipitation index and brightness temperature.

Sellers, *et al.* (1990) report on the use of the polarization ratio P given by:

$$P = (TB_v - TB_h)/(TB_v + TB_h) \qquad (5.2)$$

where TB_v and TB_h are the brightness temperatures for vertical and horizontal polarization as obtained with the scanning multi-frequency microwave radiometer (SMMR) mounted on NOAA satellites. The results indicate that an increase in soil moisture will increase the value of P, whereas an increase in vegetation cover tends to lower the value of P. Becker and Choudhury (1988) have concluded from their experiments with a pair of 37 GHz SMMR-mounted radiometers that satellite-based measurements of microwave radiation emitted from the land surface have potential in monitoring soil moisture and vegetation density in marginal areas, such as the Sahel.

Choudhury (1991) shows several examples of using the difference between vertically- and horizontally-polarized brightness temperatures observed at 37 GHz frequency of SMMR on board the Nimbus 7 satellite in monitoring soil wetness and inundation.

5.4 Evaporation

The change in soil moisture equals the difference between precipitation and the sum of evaporation, net runoff and drainage.

The three 'loss terms' are dependent on soil moisture status, topography, soil physical characteristics, vegetation, climate, and soil depth. There is, thus, strong feedback between soil moisture and evaporation. Both terms are of great importance in runoff prediction, recharge, crop yield prediction, and land use planning. Yet, little progress has been made in separating their interacting roles over a range of space and time scales (Peck, *et al.*, 1983).

Significant progress has been made in the use of remote sensing for estimating evaporation on a local scale, especially in irrigated agriculture (Hatfield, 1983) using hand-held and mast-mounted infrared thermometers and airborne infrared line scanners. Thermal imagery obtained with satellites is used to obtain surface temperatures over large areas. Such temperature fields are related to thermal inertia and soil moisture. The general principles and hydrological applications of thermal infrared sensing were discussed by Becker (1980). Gurney (1986) and Carlson (1986) have reviewed methods which use thermal infrared imagery to obtain regional-scale estimates of soil moisture availability and evaporation.

Analytical models use the surface energy balance equation (i.e. the balance of incoming and outgoing energy fluxes at the Earth's surface) and an approximate simulation of the diurnal surface temperature cycle, which results from considering the periodic energy flux incident on a uniform homogeneous material. Price (1980, 1982*a*, 1989) shows how the 24-hour mean surface temperature and the amplitude of the diurnal temperature wave, which are closely associated with diurnal heat capacity and soil moisture status, may be used to estimate evaporation. Satellite data may also be used to improve the use of standard evaporation equations (Price, 1982*b*). Analytical models are attractive because of their simplicity. However, they oversimplify in general the complexities of the atmospheric processes.

Diagnostic models combine the surface energy balance expression with flux equations for sensible and latent heat which are based on the resistance concept. In analogy with Ohm's law, the resistance is defined as the ratio of the concentration difference to the flux density. A wide range of combination models have been used. Remotely sensed surface temperature may be combined with ambient air temperature and an estimate of the aerodynamic resistance for sensible heat transfer to determine the sensible heat flux from the surface.

As discussed in section 4.2.1, the surface energy balance expression may also be combined with flux equations for sensible and latent heat based on the resistance concept. Remotely sensed surface temperature, T_s, may be combined with air temperature, T_a, and the aerodynamic resistance for sensible heat transfer r_a (s m^{-1}) between the surface and the reference height of T_a to determine the sensible heat flux (H, W m^{-2}) with:

$$H = \rho Cp\ (T_s - T_a)/\ r_a \tag{5.3}$$

where ρCp is the volumetric heat capacity of the air (J m^{-3} K^{-1}).

If the net radiation to the surface, Q^*, and the soil heat flux, G, are known or can be estimated, the evaporative flux, λE, may be obtained from the difference between the available radiant energy, Q^*, and the estimated sensible heat flux, H. Thus:

$$\lambda E = Q^* - G - H \tag{5.4}$$

where all energy fluxes are in W m^{-2}.

This approach has been widely used with surface temperatures measured with ground-based infrared thermometers (Choudhury, *et al.*, 1986*a*; Choudhury, 1989; Brunel, 1989; Kalma and Jupp, 1990), airborne infrared line scanners (Jupp and Kalma, 1989), and satellite thermal sensors (Seguin, *et al.*, 1982*a*, 1982*b*; Seguin and Itier, 1983). The single-layer resistance model expressed in Equations 5.3 and 5.4 will be inadequate over land surfaces with partial cover (Kalma and Jupp, 1990; Kustas, 1990). The problem of obtaining daily totals from one-time-of-day flux estimates/measurements is discussed by Choudhury, *et al.*, 1986*b*.

Carlson (1986) notes that a variety of time-dependent, initial-value boundary layer models exist to predict the changes in time of surface variable and surface energy fluxes. Specific examples of studies which use thermal (and visible) satellite imagery with one-dimensional models are those of Rosema, *et al.* (1978); Soer (1980); Carlson and Boland (1978); Carlson, *et al.* (1981); Wetzel and Atlas (1983); Wetzel, *et al.* (1984); Gurney and Camillo (1984); Taconet, *et al.* (1986*a*, 1986*b*); and Serafini (1987). These studies illustrate a continuing trend towards better parameterization of land surface boundary conditions, energy transfer within canopies and heat and moisture transfer in the soil. Such studies have a wide range of agronomic applications, particularly in dry, marginal areas with frequent droughts. There is also a strong link with the mesoscale and large-scale numerical models referred to elsewhere in this report (see for example Rowntree and Sangster, 1986).

Rango (in press) notes the difficulty in applying such models over large areas because of the large number of input parameters relating to soil and vegetation, which are not readily available.

Various semi-empirical methods have been proposed with equations which replace complicated physical relationships. Such equations are based on field measurements or are the results of physical models. For example, Jackson, *et al.* (1977) used the following relation between daily total evaporation $(\lambda E)_d$ and net radiation $(Rn)_d$ which incorporated the difference between surface temperature T_0 and ambient air temperature T_a at about 1 pm:

$$(\lambda E)_d = (R_n)_d - B\ (T_0 - T_a)_{\ 1\,pm} \tag{5.5}$$

Here, B is an empirical constant which needs to be determined by local calibration. Similar approaches are described by Seguin (1984); Seguin and Itier (1983); Seguin, *et al.* (1986); Rosema (1986); and Brunel (1989).

Nieuwenhuis (1986); Nieuwenhuis and Bouwman (1986); and Nieuwenhuis, *et al.* (1985) describe a method to calculate the daily total of actual evaporation $(AET)_d$ as a fraction of the daily total potential evaporation $(PET)_d$. They use:

$$(AET)_d\ /\ (PET)_d = 1 - B'\ (T_o - T_o')_i \tag{5.6}$$

where T_o and T_o' are the surface temperatures at time i under actual and potential evaporation conditions, respectively. Time i is close to 1 pm. The calibration constant B' may be expressed as:

$$B' = a + b\ (u) \tag{5.7}$$

where a and b are empirical constants and u is the wind speed at a standard height.

Menenti (1984) and Nieuwenhuis and Menenti (1986) approach estimation of regional evaporation with an empirical expression relating spatial differences in evaporation *(dET)* to spatial differences in albedo *(da)* and in surface temperature (dT_o). Thus:

$$dET = p + q\ (da) + r\ (dT_o) \tag{5.8}$$

where p, q and r are empirically determined constants.

Semi-empirical models for estimating evaporation from remotely sensed data have been used in semi-arid regions in Africa by Rosema (1981, 1982, 1986) and Seguin, *et al.* (1987), and in Europe by Vidal, *et al.* (1987).

5.5 Radiation balance components

Interactions between land surface and the atmosphere are related to the surface radiation budget, which determines the amount of radiation absorbed by the land surface, and to the surface heat budget, which describes the proportions of this energy used for sensible and latent heat fluxes to the atmosphere and for heating the soil.

The surface radiation balance is defined as:

$$Q^* = K\downarrow - K\uparrow + L\downarrow - L\uparrow = K^* + L^* \tag{5.9}$$

where Q^* is net all-wave radiation, K^* is net short-wave radiation, L^* is net long-wave radiation, $K\downarrow$ and $K\uparrow$ are downward and upward (reflected) short-wave radiation, and $L\downarrow$ and $L\uparrow$ are downward and upward long-wave radiation, respectively.

Global radiation $K\downarrow$ at the surface may be estimated using brightness data as an index of atmospheric opacity, obtained from the visible channels of geostationary and polar orbiting satellites. Clear sky insolation values are parameterized or calculated. Clouds reduce these clear-sky values by an amount depending on their albedoes, which are determined from satellite observations of reflected radiation. Sellers, *et al.* (1990) have presented a review of algorithms for determining insolation at the land surface from remote sensing. They distinguish between statistical methods, physical methods, and hybrid methods.

Statistical techniques (see for example Tarpley, 1979; Justus, *et al.*, 1986; Nunez, *et al.*, 1984; Nunez, 1990*a*) generally rely on comparing satellite brightness data with simultaneous pyranometer measurements at the relevant pixel. Transmission functions are then developed which may be used to map solar radiation over the region of interest. Using three years of GMS imagery, Nunez (1990*a*, 1990*b*) applied such techniques to obtain maps of monthly solar radiation for regions of 2° longitude by 2° latitude surrounding each capital city in Australia. These maps were based on observations recorded three times a day in the visible band (0.55–0.75 μ).

Similar techniques may be used to obtain daily radiation for shorter periods.

A more general approach, with wider applicability, uses physical relationships to derive $K\downarrow$. This approach extracts surface albedo as a necessary step in differentiating cloudy from cloudless regions. Radiation theory is then used to estimate global radiation under clear and cloudy conditions, and to correct for atmospheric depletion between cloud top and satellite sensor (Dedieu, *et al.*, 1987). An alternative approach is to rely on an empirical relationship to trace direct and diffuse beams through the various layers (Diak and Gautier, 1983).

The accuracy of such relationships for estimating $K\downarrow$ depends strongly on the frequency of the satellite scans, the averaging period, and the amount of atmospheric opacity due to clouds and aerosols. Widely varying results have been reported. For daily totals, the rms values of estimated measurements for low zenith angles vary from 8 to 15 per cent of $K\downarrow$, but values as high as 36 per cent have been reported for winter-time conditions. Greater accuracies are obtained for periods longer than one day.

Satellite-derived estimates of the upwelling short-wave radiation $K\uparrow$ are based on the determination of broadband albedo. This technique needs two corrections. First, a correction is needed for the effect of the intervening atmosphere. This has been done with radiative transfer models and simple physical/empirical relationships (e.g. Chen and Ohring, 1984). Second, the effect of surface structure and texture on the satellite-derived bi-directional reflection function must be considered. Following these corrections, a hemispherical albedo is obtained from a single satellite value. Finally, a narrow-to-broadband correction is needed to obtain the required broadband surface albedo (see also Nunez, *et al.*, 1987; and Dedieu, *et al.*, 1987).

The incoming atmospheric radiation flux $L\downarrow$ has been calculated with radiative transfer models from TIROS operational vertical sounder (TOVS) data, which provide information on cloud cover as well as profiles of air temparature and humidity. Schmetz, *et al.* (1986) used screen level air temperatures to derive an empirical estimate of clear sky $L\downarrow$. These data were refined with cloud information derived from METEOSAT data to obtain the total (cloudy and clear) $L\downarrow$ value. Gupta (1989) used results from a radiation transfer model to relate L^* to simple TOVS parameters. The upward long-wave radiation flux may be obtained from apparent surface temperatures.

Simple linear relationships can also be established between the daily fluxes of K^* and L^* based on ground-based measurements. Such empirical expressions enable L^* and, hence, Q^* to be obtained from satellite derived $K\downarrow$ and $K\uparrow$ values. The estimates of net all wave radiation Q^* can then be used to estimate regional scale evaporation (see Kalma and Nunez, 1990).

Zhong, *et al.* (1990) have described the use of NOAA-7 infrared satellite data to test a simple relationship proposed by Nielsen, *et al.* (1981) between cloudiness, global radiation, and net radiation.

5.6 Land cover and biomass

For many years, visible, near-infrared and thermal imagery from various satellites have been used to map land cover, biomass and many other vegetation characteristics. Various vegetation indices have been employed which use data from different wave bands (see Sellers, *et al.*, 1990 for a review). They are generally based on the fact that plant materials are weak reflectors in the visible region of the spectrum and strong reflectors in the near-infrared. The use of the normalized difference vegetation index (NDVI) in small-scale studies, continental-scale studies, and global studies is discussed by Fung and Tucker (1986). They note that remote sensing can provide quantitative information on the changes in land cover and in the functioning of the global biosphere, provided field measurements and ground truth are obtained at least at an ecosystem level. The measurements must address the heterogeneities of species and dynamics within an individual ecosystem as well as heterogeneities within the same ecosystem at different geographic locations.

Thermal imagery is used to obtain canopy temperature, which is linked to water stress and general crop health. Work with satellite data to obtain vegetation indices has centred on NOAA/AVHRR (Townshend and Tucker, 1984), Landsat multispectral scanner (MSS) and thematic mapper (TM). Choudhury and Tucker (1987) describe the use of Nimbus 7 scanning microwave radiometer data for vegetation monitoring. In recent years, SPOT satellite data are also used for vegetation surveys.

As noted earlier, the work of Becker and Choudhury (1988) has shown that polarized microwave radiation measurements in the 37 GHz band obtained with NOAA-SMMR provide an indication of vegetation density in marginal zones, such as the Sahel. In zones with sparse vegetation the polarization ratio *P*, defined earlier, is more sensitive than the AVHRR vegetation index, while the opposite appears true in areas with dense vegetation. Choudhury (1991) showed examples of using the difference between vertically- and horizontally-polarized brightness temperatures observed at 37 GHz frequency of SMMR on board the Nimbus 7 satellite to indicate vegetation density for three regions in Africa. He also described a relationship between annual evaporation computed from a water balance equation and annual mean values of the polarization difference for pooled data from four continents, which include a wide range of different surface types.

Examples of land cover mapping and assessments of dry matter production in semi-arid regions include work by Carnegie, *et al.* (1974); Prince and Tucker (1986); Tucker, *et al.* (1983); and Van Dijk, *et al.* (1989).

CHAPTER 6

MODELLING OF LAND SURFACE PROCESSES

6.1 Introduction

The realistic prediction of regional and large-scale climates depends on adequate representations of the topography and the land surface processes. Parameterizing land surface processes and surface characteristics in atmospheric general circulation models are described by Carson (1982); Hunt (1985); Dickinson (1986); Eagleson (1982); Henderson-Sellers, *et al.* (1986); Laval, *et al.* (1984); and Rowntree (1983).

Most AGCMs treat the fluxes of radiation, sensible heat, latent heat, and momentum at the surface as independent processes. The key surface parameters usually specified are albedo, surface roughness, and soil moisture availability (see section 3.2). Brutsaert (1988) notes that general climate models are quite sensitive to the parameterization of land surface processes in general and of evaporation, in particular.

Large-scale models to predict hydrological land surface conditions play an important role in climate prediction with AGCMs. These hydrology models will also be essential in the validation and calibration of the AGCMs and in the assessment of the hydrologic impact of predicted climate change in specific river basins and large watersheds.

Land surface hydrology models are used with global climate models to achieve the following objectives (Becker, 1991):

(*a*) To provide the lower boundary moisture and energy fluxes for the AGCM grid cells needed in climate simulation;

(*b*) To compute basin discharges in gauged river basins for given AGCM outputs for the purposes of verification, validation, and prediction;

(*c*) To compute fresh water inflows into oceans in coupled land-ocean models.

Becker and Nemec (1987) introduced the concept of hydrological modelling at two levels ('domains'). The first level concerns all vertical processes of moisture exchange, storage and flux — including precipitation, evaporation, soil moisture recharge, infiltration, percolation, and groundwater recharge — which can be applied to any land area of interest, ranging from AGCM grid cells to river basins. The second level comprises modelling of the lateral flows of overland flow, groundwater flow, and interflow at river basin level. If river discharges for a real basin are to be used for validation of climate model predictions, then the first domain model is used with AGCM grid cells or, more appropriately, with subgrid cells, whereas the second domain model is related to the gauged river basin and is coupled to the appropriate AGCM subgrid cells (Becker, 1991). The same approach can also be used with very large river basins.

6.2 Simple hydrology models

Information on large-scale evaporation from land surfaces is important for climate modelling and weather forecasting. It is also essential for many hydrological purposes, such as the prediction of floods and droughts. Brutsaert (1988) observed that "there is a critical need in hydrology and climate modelling for sound parametric formulations for land surface evaporation from areas with typical length scales of 10–10^2 km; these are characteristically the scales of a region or of the grid size for integration of the general circulation models".

Aggregation in AGCMs, which have a spatial scale in the range of 10^4–10^6 km^2, involves the use of very simple parameterizations of the surface hydrology. Many current AGCMs (such as the GFDL model described by Gordon and Stern, 1982) treat surface hydrological processes in terms of grid cell water storage based on the very simple 'bucket' formulations described by Manabe, *et al.* (1965) and Holloway and Manabe (1971). The amount of water in the bucket controls evaporation and surface runoff. The Goddard Institute for Space Studies (GISS) model of Hansen, *et al.* (1983) uses the two-layer approach of Deardorff (1977). However, these AGCMs do not consider the role of vegetation or inter-cell transfer of water.

In the simple hydrology models used in climate modelling, the water available for evaporation is obtained with a water balance equation whereby the amount stored in ground or on surface is a function of precipitation, evaporation, and runoff. Most models have one or more reservoirs for storage. The main difference is the number of stores and how they are managed — interactively or prescribed.

The resistance to water transfer is due to the intermediate layer of vegetation between the free atmosphere and the soil. The aerodynamic resistance of the atmospheric boundary layer is taken care of by an eddy exchange coefficient. The vegetation resistance is represented by an aridity factor to obtain actual from potential evaporation. The aridity factor is often equated to the dimensionless amount of available water in the upper soil layer.

The Budyko "bucket" approach described by Manabe, *et al.* (1965) and Holloway and Manabe (1971) is based on a very simple soil moisture availability function and does not consider explicitly the role of vegetation. In the bucket model, the potential evaporation is modified by a factor that depends on moisture availability. When evaporation exceeds precipitation, the water level in the bucket is lowered. It is raised when precipitation exceeds evaporation. When the bucket overflows, runoff occurs. Field capacity is assumed to correspond to a stable soil moisture storage of 15 cm. The change in moisture content is obtained from:

$$dW/dt = P - E - Q \tag{6.1}$$

where P is rainfall, Q is runoff and E is evaporation. A bulk aerodynamic formula is used to calculate potential evaporation E_0 from wind velocity and the difference between the saturated mixing ratio at surface temperature and the mixing ratio at a reference level. A wetness parameter w' is used to reduce the actual evaporation below the potential value as soil dries. This yields:

$$E = w' E_0 \tag{6.2}$$

where $w' = 1$ if $W > 10$ cm and $w' = (W/10)$ if $W \leq 10$ cm

The Deardorff (1977) method is based on a surface layer of thickness $d_1 = 0.5$ cm and a reservoir of thickness $d_2 = 50$ cm, with volumetric moisture content values of WG and WB, respectively. The temporal variation in these variables is given by:

$$dWG/dt = -\{C_1(E - P)/\rho_w d_1\} - \{C_2 (WG - WB)/\tau\},\ 0 \leq WG \leq WMAX \tag{6.3}$$

and:

$$dWB/dt = -\{(E - P)/\rho_w d_2\} \tag{6.4}$$

Evaporation is given by:

$$E = E_0 \text{ if } WG > WSAT \tag{6.5}$$

and:

$$E = (WG/WSAT)\, E_0 \text{ if } WG \leq WSAT \tag{6.6}$$

In this approach, E_0 is potential evaporation, $WMAX = 0.40$ and $WSAT = 0.75WMAX$ and representative initial values need be selected for WB and WG. τ is 1 day and ρ_w is the density of liquid water. Deardorff (1977) has provided empirical values for C_1 and C_2, which will depend on soil type.

Hansen, *et al.* (1983) used a two-layer model based on Deardorff (1977) in which the upper soil layer can respond immediately to rainfall or evaporation, while the lower soil layer is a store. Water contents W_1 and W_2 are the ratios of available water to field capacity in the upper and lower layers, which have depths of f_1 and f_2 cm, respectively. Soil depths and water holding capacities need to be set.

The rate of change in the upper layer is:

$$dW_1/dt = \{(P - E - Q)/f_1\} + (W_2 - W_1)/\tau \tag{6.7}$$

where τ is the time constant for diffusion between the two layers, usually taken as one day. For the lower layer:

$$dW_2/dt = (f_1/f_2)\,(W_1 - W_2)/\tau \tag{6.8}$$

When $W_1 \leq 1$, the runoff is calculated from:

$$Q = (W_1)\,(P) \tag{6.9}$$

and evaporation is obtained from:

$$E = (W_1)\,(E_0) \tag{6.10}$$

If $W_1 \geq 1$:

$$E = E_0 \tag{6.11}$$

where E_0 is potential evaporation rate.

Hunt (1985) has used a one-dimensional radiative-convective model to compare the soil hydrology formulations of Holloway and Manabe (1971), Hansen, *et al.* (1983), and Deardorff (1977), which do not consider vegetative cover explicitly. He concluded that the methods of Holloway and Manabe and the GISS model of Hansen, *et al.* have very long response times for a drying surface. The Deardorff method was preferred because of its realistic response time, simplicity, physical basis, and the appropriateness for the time scale of most AGCMs.

Russell and Miller (1990) ran a four-year simulation with Climate Model II of Hansen, *et al.* (1983) to compute runoff for 33 of the world's major rivers. They observed good agreement for rivers with annual runoff exceeding 200 km^3, but overpredicted for rivers with less than 200 km^3/yr. Some of the inaccuracy is due to poor model precipitation fields. However, the model's evaporation and groundwater storage parameterizations are also suspect. They used the model with a horizontal resolution of 4° latitude by 5° long and with nine vertical layers. At the surface, grid boxes are divided into land and ocean fractions. The runoff in each grid box depends on precipitation, evaporation, and water storage within the land portion. Evaporation is computed from:

$$E = \beta \rho\, CV\, (q_s - q_a) \tag{6.12}$$

where β is a dimensionless efficiency factor (see below), ρ is the surface air density, C is a dimensionless drag coefficienct dependent on stability, V is the surface wind, q_s is the saturated specific humidity at the surface, and q_a is the specific humidity of the surface air.

Each grid box has two layers for groundwater storage: the upper layer responds directly to evaporation and precipitation, and the lower layer acts as a seasonal reservoir. There is a two-day time constant for diffusion between the two layers. If the soil is unsaturated, runoff R is computed from:

$$R = 1/2\, PW_1 \tag{6.13}$$

where P is the precipitation and W_1 is the ratio of water in the first layer divided by its field capacity. β in Equation 6.12 is equal to W_1, unless the ground is snow covered, in which case $\beta = 1$. For saturated conditions:

$$r = p \tag{6.14}$$

Wood, *et al.* (1991) have reported on the hydrological performance of a five-year simulation with the Geophysical Fluid Dynamics Laboratory (GFDL) model (Gordon and Stern, 1982) using a grid resolution of 3.75° longitude by 2.25° latitude, and using validation data on runoff, precipitation, and temperature for a large number of locations in the continental U.S.A. The traditional bucket model reproduces the major east-west structure of seasonal variation in temperature and precipitation, but the model gives results which are too wet and cold in winter and too dry and hot in summer, with only crude agreement for runoff. An interesting feature of their study is the use of a new statistical water balance model, which yields more dynamic and realistic short-term variations in soil moisture than the traditional bucket hydrology used in most current AGCMs. The most significant differences were due to the use of base flow in their model to simulate between-storm runoff.

Koster and Eagleson (1990) recognize that sensitivity studies are needed to test AGCM soil hydrology parameterizations. Off-line testing involves subjecting a proposed land surface model to an AGCM generated time series of precipitation, radiation, temperature, and humidity conditions. Such testing ignores the potentially significant feedback between surface and atmosphere. The alternative, testing land surface models within AGCMs, is limited because of computational constraints. They describe a one-dimensional interactive soil-atmosphere model based on the structure of the soil-atmosphere column at a single grid square in the Model II version of the GISS AGCM (Hansen, *et al.* 1983). The model consists of nine atmospheric layers and two soil layers. Forced by prescribed seasonal cycles of solar radiation, moisture convergence, and heat convergence, the resulting 1D climate is then controlled in part by the formulation of the soil hydrology. This model was successfully used for testing soil hydrology parameterizations, which are variations from the GISS model as described by Hansen, *et al.* (1983) and referred to above.

A wide range of water balance models have been used for large-scale hydrological studies especially to assess the sensitivity of water resources to climatic change (Klemes, 1985; Nemec and Schaake, 1982; Wigley and Jones, 1985; and Gleick, 1987). The simplest models operate by monthly time steps and solve the water balance of a drainage basin, which may be represented by:

$$P = E + Q + dS/dt \tag{6.15}$$

where P is precipitation, E is evaporation, Q is runoff and dS/dt is the rate of change in storage. A hydrological model with daily time step was used by Bultot, *et al.* (1988*a,b*) to study the impact of climatic change induced by doubling of CO_2 on the water balance of three basins in Belgium.

Schaake and Chunzhen (1989) have described two simple models for use in large-scale studies of the potential impact of climatic change on water resources. They developed a simple linear, analytical model which assumes that:

$$Q = Q_s + Q_g = \alpha P + kS \tag{6.16}$$

where Q_s is storm runoff expressed as a constant proportion of precipitation P, and Q_g is the groundwater runoff given as a constant proportion k of storage S. In reality α and k are not constant. If S^* is the maximum value (limit) of S, then maximum Q_g will occur when $Q_g = a\,S^*$. The parameter k indicates how fast groundwater flows to streams. It can be determined from the exponential decay of streamflow hydrographs during dry spells. If E_0 is the average monthly potential evaporation, then the actual evaporation E is determined from:

$$E = (S/S^*)\, E_0 \tag{6.17}$$

Substituting Equations 6.16 and 6.14 into Equation 6.15 yields a model which emphasizes the forcing of storage and evaporation by precipitation. If P is given as a function of time, i.e. $P(t)$, an analytical solution may be found for storage as a function of time, i.e. $S(t)$. Equation 6.16 may be used to generate the hydrograph $Q(t)$. This simple linear model performed well in a humid basin, but failed in a semi-arid environment. The main reason is that α is not constant but a variable which depends on P and the moisture deficit $D = S^* - S$. Clearly P and D will vary throughout the basin.

Schaake and Chunzhen (1989) also tested a non-linear monthly water balance model. If D_t is the moisture deficit in month t and D_{max} is the maximum deficit when $E = 0$, then the actual evaporation E_t in month t may be calculated from the potential evaporation E_{pt} in that month with:

$$E_t = E_{pt}\,((D_{max} - D_t) / D_{max}) \tag{6.18}$$

The monthly time step used was due to data availability, although storm runoff occurs over much shorter periods. A parameter Θ was introduced to account for processes operating at time steps shorter than one month. ΘE_t is then the proportion of E_t that must be satisfied from precipitation P_t in month t before infiltration or runoff can occur. The precipitation in excess of ΘE_t is then split between infiltration and storm runoff. First a proportion of z of the moisture deficit D_t must be satisfied by infiltration before any storm runoff. The precipitation component contributing to storm runoff is then:

$$PXX = P_t - \Theta E_t - zD_t \tag{6.19}$$

If PXX is positive, storm runoff can occur. An asymptotic relationship between storm runoff Q_{st} and PXX is introduced to account for the diminishing effect of the moisture deficit applying at the beginning of the month. This is given by:

$$Q_{st} = [PXX/(PXX + D_t)]\, PXX \tag{6.20}$$

Groundwater runoff varies with the deficit D_t as:

$$Q_{gt} = k\,(S_{max} - D_t) \tag{6.21}$$

If D_t equals, or is greater than S_{max}, then:

$$Q_{gt} = 0 \tag{6.22}$$

Finally, the change in moisture deficit for the month is:

$$D_{t+1} = D_t - P_t + E_t + Q_{st} + Q_{gt} \tag{6.23}$$

This monthly, non-linear water balance model was calibrated for several selected basins in humid and semi-arid climates. Values for the five lumped parameters (D_{max}, z, S_{max}, Θ and k) were selected and the model was then run for 52 basins in the southeastern U.S.A. Simulation results for 49 out of 52 basins were within 20 per cent error bounds. Mean simulated annual runoff for the 52 basins is presented in a contour map in Figure 6(*a*). This map may be compared with the map of observed mean annual runoff shown in Figure 6(*b*).

6.3 Parameterization of land surface processes

Avissar and Verstraete (1990) and Avissar (1991) describe the various surface processes occurring at the lower boundary of the atmosphere. The extent to which exchanges of heat, mass, and momentum between the land surface and the air are incorporated in atmospheric models will depend on their relative contributions compared to dynamic processes, such as advection. The need for detailed parameterizations of surface processes will become less when the space scale of interest increases and the time scale becomes shorter.

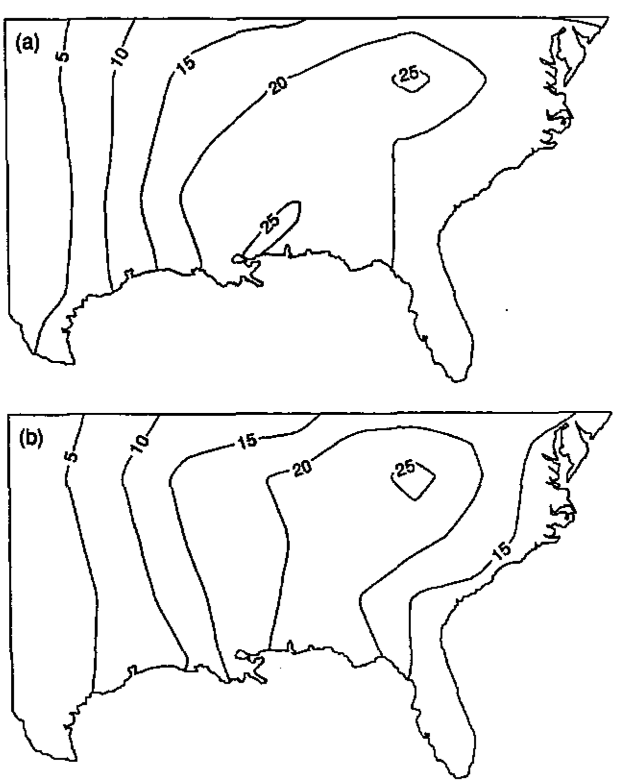

Figure 6 — Annual runoff maps based on (*a*) mean simulated and (*b*) observed annual runoff (inches) obtained for 52 basins in the contiguous U.S.A. (from Schaake and Chunzhen, 1989).

The realistic representation of the physical land surface processes is difficult because of the number and the complexity of the processes involved and because the processes occur widely at different time scales. The spatial resolution of the atmospheric models is also much larger than the characteristic scale of the relevant surface processes. Thus, climate models require that the highly variable surface fluxes must be approximated in terms of the larger-scale processes explicitly described in the climate model (Avissar and Verstraete, 1990).

Wood (1991) defines parameterization as the parametric representation of small-scale processes at large scales. Small-scale processes are, thus, scaled up in some appropriate fashion and parameterization represents their influence at the large scale.

The need to represent the land surface in AGCMs, and the realization that the state of the land surface affects the simulation of the atmosphere, has led to detailed land surface models, which consider vegetation explicitly and which are linked with AGCMs (Avissar and Verstraete, 1990).

6.3.1 *Overview of current models*

Land surface parameterization schemes are based on estimating the atmospheric source/sink strength for momentum, heat and mass, and the energy and/or mass available at the surface to drive the transfers and/or to provide the quantities being transferred.

The parameterization for atmospheric demand is based on formulations of the eddy fluxes in the turbulent boundary layer, exchange coefficients, and vertical gradients of the mean quantities. Most climate models use a simple formulation for atmospheric demand based on a drag coefficient — relating drag force to bulk wind speed squared — assumed to be the same for momentum, heat and mass, or on turbulent similarity theory by which fluxes are obtained from gradients. This means that turbulent transfer must be averaged to account for spatial diversity in surface conditions within AGCM grid squares.

The main surface budgets are those for radiation and water. The energy available for latent and sensible heat transfer is important. Many schemes do not really consider the amount of energy available for vertical exchange. This is tantamount to allowing the exchange of unlimited amounts of heat. This may possibly be permissible for short-term numerical weather prediction, but would introduce serious errors when studying long-term climate phenomena. The solution clearly lies in using a combination approach, such as the Penman-Monteith formulation to parameterize evaporation. The problem lies in using such an approach for non-homogeneous grid cells with all the diversity and complexity existing in the land surface processses.

Carson (1982) has reviewed early land surface models used in AGCMs. Early parameterizations of soil and vegetation layers include the work of Deardorff (1978), McCumber (1980); and Yamada (1982).

The vegetation was considered explicitly by Deardorff (1978) who used an efficient one-layer foliage parameterization which includes the effects of soil and leaf albedos and emissivities, total leaf area, stomatal resistance and intercepted water. The energy budget equations at the ground level and for the canopy are solved simultaneously with semi-empirical expressions for aerodynamic resistance, canopy resistance, and soil resistance. The canopy resistance expression includes incident solar radiation, soil moisture content in the surface layer and in the root zone, leaf area index, and a shelter factor.

In recent years, several land surface models have been designed for AGCMs, which include an explicit parameterization of vegetation. The models represent significant improvements in process description, but need many parameters describing the state of the vegetation and the underlying surface. Their incorporation in AGCMs is not without problems (Verstraete and Dickinson, 1986; Henderson-Sellers, 1991; and Garratt, 1993).

Various sensitivity studies have been carried out with soil-canopy schemes to assess the importance of the land surface properties—albedo, roughness, surface wetness and vegetation density. Kowalczyk, *et al.* (1991) note that these properties vary considerably across land surfaces in space and in time. Figures 7 and 8, taken from their paper, show summaries of the variation of albedo and roughness with surface type.

Dickinson (1983, 1984) and Dickinson *et al.* (1986) developed the biosphere atmosphere transfer scheme (BATS). BATS incorporates two soil layers, a vegetation canopy, a variable albedo and several important hydrological parameters. In BATS, the depths of the upper and lower soil layers and the partitioning of roots between them are variable. Energy and moisture fluxes are calculated as a function of soil type. Ten soil parameters are required to specify the soil properties. The rooting ratio between upper and lower layers is also required. The latent and sensible heat fluxes in and

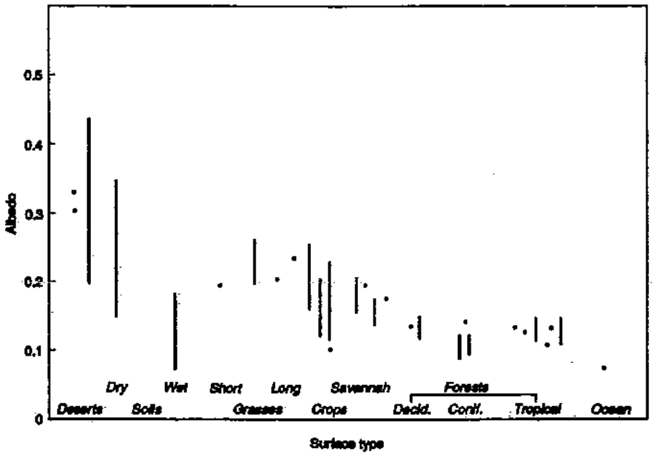

Figure 7 — Variation of albedo with surface type, based on observations, literature values, and model specified values. Vertical bars indicate a range of values (after Kowalczyk, *et al.*, 1991).

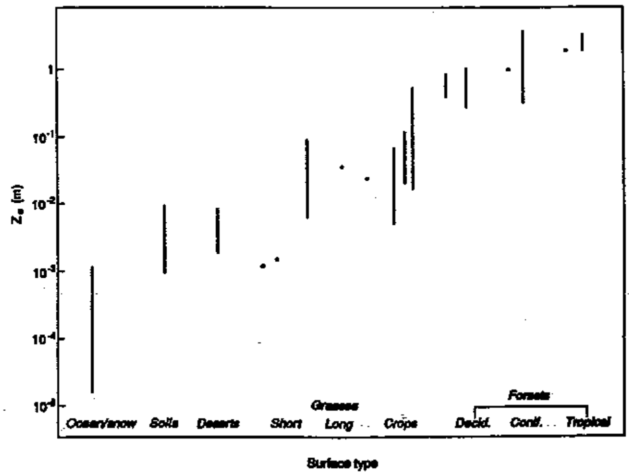

Figure 8 — Variation of surface roughness length with surface type, based on observations, literature values, and model specified values. Vertical bars indicate a range of values (after Kowalczyk, *et al.*, 1991).

above the canopy depend on the characteristics of the vegetation canopy specified by eleven vegetation/land cover parameters.

Wilson, *et al.* (1987) tested the stand-alone BATS surface parameterization package of the National Centre for Atmospheric Research (NCAR) Community Climate Model (CCM) described by Dickinson, *et al.* (1986) for several ecotypes and for a series of parameters. BATS was found to be particularly sensitive to the parameterization of soil texture. However, these experiments were conducted with prescribed atmospheric forcing. This prevented a full test of the results, since feedbacks could not occur.

In a recent study, Pitman (1993) describes the sensitivity of BATS to values chosen for the individual parameters by linking the land surface scheme to a single column model, which include surface-atmosphere feedbacks. It is observed that BATS is most sensitive to parameters which describe the availability of moisture, the availability of energy, and the efficiency of the linkage between surface and atmosphere. Pitman notes that, in particular, the lack of information on root distribution may have important implications for the utility of many advanced land surface models.

Sellers (1986) and Sellers, *et al.* (1986) have described a simple biosphere model (SiB) which has important features in common with BATS, but is considerably more complex. Despite the term 'simple', SiB represents a very complex land surface model for simulating the land surface-atmosphere interaction in an AGCM. The general strategy of the SiB model has been to model the vegetation as the major determinant of the land surface-atmosphere interactions. The vegetation canopy in each grid area is represented by two layers which may be present or absent at any point and at any time. The upper layer comprises trees and shrubs and the lower layer represents grasses and other herbaceous species. Each vegetation type has morphological, physical and physiological properties assigned to it, which are assumed to extend uniformly throughout the AGCM grid cell. Some 50 parameters are required to specify the biological and physical properties of every grid cell.

A number of SiB submodels use these parameters to describe the interception and partitioning of visible, near-infrared and thermal radiation; the interception of rainfall and its subsequent evaporation; infiltration, drainage and storage of the residual rainfall; the environmental and biological control of transpiration; and aerodynamic transfer of sensible heat, water vapour, and momentum from soil surface and vegetation to the atmosphere. The three most important model components are those for radiative transfer, and for aerodynamic and surface resistances.

Radiative transfer within the canopy and between canopy and soil surface is simulated in SiB with a modified two-stream radiative transfer approximation to describe the effects of multiple scattering within canopies. The model reproduces realistic surface albedos which change with solar zenith angle. The SiB model uses flux gradient relationships to model turbulent transfer.

Derived profiles of wind speed and transport coefficients are integrated to describe aerodynamic resistances to turbulent transfer between ground and canopy air space, vegetation and canopy air space, and canopy air space and the free atmosphere above the vegetation.

The surface resistances to water vapour transport from within the canopy, the ground cover, and the upper soil layer include a soil resistance, a ground cover resistance, and an upper story canopy resistance. Stomatal resistance, is made dependent on incident radiation, air temperature and water vapour pressure, and leaf water potential. The last factor is dependent on soil moisture content.

The SiB model originally had seven prognostic state-variables, namely, canopy temperature and ground temperature, interception storages for canopy and ground cover, and soil moisture storages for the two classes of vegetation and one for the soil recharge layer (Sellers, *et al.*, 1986). Wood (1991) notes that recently an eighth variable was added to follow the deep soil temperature.

Sato, *et al.* (1989) have studied the effects of implementing the SiB model in a general circulation model and determined the ability of SiB to predict the diurnal behaviour of surface fluxes in different parts of the world. They analysed the impact of SiB on the hydrological and energy cycles of the monthly forecast and compared SiB with a land surface parameterization based on the bucket model.

Sellers and Dorman (1987) have described the testing of SiB with micrometeorological and biophysical measurements from surface experiments conducted over arable crops in Germany and the U.S.A. and for a forested site in England. Sellers, *et al.* (1989) discuss the operation and calibration of SiB using micrometeorological and hydrological measurements taken in and above tropical forests in the central Amazon basin. These calibrations resulted, *inter alia*, in changes in the specified interception store of the canopy and in the SiB model for turbulent heat transfer. Finally, Sud, *et al.* (1990) report on the use of the NASA/Goddard Laboratory for Atmosphere (GLA) AGCM to assess the influence of including SiB. Monthly simulations of the AGCM with and without SiB show that SiB simulates much lower evaporation rates over land and generates significantly different values of surface fluxes for both vegetated and bare soil regions. These differences are accompanied by large changes in simulated rainfall.

Although BATS and SiB assume homogeneous conditions and oversimplify the land surface, they are extremely valuable in AGCMs and mesoscale models. They also enable interpretation of remote sensing data and field measurements because they provide a physical link between surface physical quantities and radiation fluxes.

Pitman (1988) and Pitman, *et al.* (1991) have described a new simple land surface scheme called the bare essentials of surface transfer (BEST), which has been designed to retain those elements which are considered essential in representing the space and time variability of the land surface-atmosphere interaction. BEST models resistances below, within, and above the canopy. Two versions of BEST are compared with BATS and with observations from the Amazon in a recent paper by Yang and Pitman (1993). The implementation of BEST into the AGCM of the Australian Bureau of Meteorology Research Centre is discussed by Yang, *et al.* (1993)

Kowalczyk, *et al.* (1991) have described a one-dimensional soil canopy scheme for use in the four-level and nine-level AGCMs used by the Commonwealth Scientific and Industrial Research Organization (CSIRO) in Aspendale, Australia. Their model includes details on soil type, albedo, roughness length, canopy resistance and degree of surface wetness. It considers canopy interception of rainfall, runoff, deep soil percolation, snow accumulation, and melting. The canopy parameterization allows for full cover (with the same vegetation and soil surface temperatures) or for fractional cover with vegetation and soil surface temperature calculated separately. Their two-layer soil moisture scheme uses the force-restore scheme of Deardorff (1977), as modified by Noilhan and Planton (1989).

Raupach (1991) has identified serious problems with several existing land surface models. For homogeneous land surface conditions, the most important problems are the specification of stomatal resistances and the description of turbulent transfer. Raupach notes that current comprehensive multilayer canopy-atmosphere models have become very complicated; BATS requires some 20 and SiB some 50 input parameters. Raupach proposes a simplified canopy-atmosphere model with four surface parameters: canopy albedo, canopy height, an optimal canopy resistance, which depends on surface type, and leaf area index. This model includes a vegetation treatment which is claimed to be more realistic than the current practice in AGCMs. It also avoids the need to iterate for surface temperature.

In 1992, the WMO CAS/JSC Working Group on Numerical Experimentation (WGNE) and the Science Panel of the GEWEX Continental-scale International Project (GCIP) launched a joint WGNE-GCIP Project for the Intercomparison of Land Surface Parameterization Schemes (PILPS). The principal aim of PILPS is to achieve greater understanding of the capabilities and potential applications of existing and new land surface schemes in atmospheric models. Its focus is to improve simulations of energy and water exchanges at the continental surface in climate models and to extend simulations to carbon uptake and release. The goals of PILPS are to document current models, acquire and disseminate appropriate data sets, identify data gaps and propose means of acquisition, and to identify parameterization inadequacies and propose solutions (PILPS, 1992).

A brief report on Phase 1 of PILPS has been presented by Pitman, *et al.* (1993). They report that 22 land surface models have been run using identical data obtained from the NCAR CCM1 climate model. These data were representative of locations in tropical forests (Amazonia), grassland (U.K.), and tundra (northern Canada). Each location was described in terms of soil and vegetation characteristics. Initial results show reasonable agreement in the annually averaged surface temperature and sensible and latent heat fluxes, and less agreement in the monthly data and negligible agreement in the instantaneous data. No agreement is reached over any time scale in predicted runoff or snow cover.

The physics which underly the exchange of momentum, heat, and water at the land surface are fairly well understood. Yet, schemes which are complete enough to account for the major processes and simple enough for use in climate models need a great deal more development. Most land surface models are lumped and are 'single leaf — single stoma' oriented (Avissar, 1991). Such an approach is conceptually inappropriate in view of the natural landscape variability and this problem needs to be overcome. The next section discusses current approaches to considering significant areal heterogeneities in climate modelling.

6.3.2 *Land surface parameterization and subgrid-scale variability*

Most AGCMs use spatially fixed land surface parameterizations because of data availability problems. Information on soil and vegetation parameters can be obtained from a number of basic sources, most of which are provided at 1° x 1° resolution (e.g. Henderson-Sellers, *et al.*, 1986). Pitman (1991) notes that many AGCMs use a "most common" procedure for aggregating such 1° x 1° information to the resolution of the AGCMs (see also Dickenson, *et al.*, 1986).

It is important to determine how sensitive climate simulations are to spatial variations in the important surface parameters. The importance of accurate land surface parameterizations is emphasized by the studies of Delworth and Manabe (1988, 1989) and Oglesby and Erickson (1989) who analysed the impact of soil moisture and surface wetness on climate. Wetzel and Chang (1987, 1988) and Wilson, *et al.* (1987) have observed the important effect of soil hydraulic parameters.

In recent years, numerous papers have appeared which describe attempts to develop land surface parameterizations incorporating subgrid-scale terrain, soil, and vegetation heterogeneity. Wetzel and Chang (1988) have demonstrated the fundamental weakness of simple averaging. They show the strongly non-linear relationship between soil moisture and evaporation, which results in a relationship between regional evapotranspiration and area-averaged soil moisture that is fundamentally different from the relationship that applies at a point. In order to predict evaporation adequately on the scale of most numerical weather prediction models, Wetzel and Chang have described an explicit model for computing grid cell average evaporation in non-homogeneous landscapes. The model incorporates a statistical distribution of near-surface soil moisture on the subgrid scale, the variance of which is obtained from observed distribution of soil moisture and precipitation. The authors emphasize that large errors may result from failure to account for the fundamental differences between the local soil moisture evaporation relationship and the area-averaged relationship. They report that for numerical weather prediction models with a grid spacing in excess of 100 km, the expected subgrid-scale variability of soil moisture may be as large as the total amount of potentially available water in the soil.

In an unpublished presentation at the 1991 Hahn Symposium, Wood described three modelling experiments concerned with horizontal scaling. The first experiment investigated the hydrologic response of a catchment in which spatial variability in topography, soils, and hydrologic inputs resulted in spatially variable responses. In the second experiment, the SiB was applied to a non-homogeneous domain based on data from Amazonia to assess the impact of variability in initial soil wetness, vegetation density, and rainfall on simulated evaporation. The third experiment addressed the scaling in the normalized difference vegetation index (NDVI), and sensible and latent heat fluxes as derived from high resolution thematic mapper (TM) data.

The major result of Wood's studies was that fluxes and surface characteristics essentially scale linearly. Wood points out that these results suggest that effective, equivalent parameters can be used in macro-scale models for the calculation of spatially-averaged quantities such as large-scale fluxes "provided that the equivalent parameters reflect the statistical characteristics of the subscale variability". The one exception to these results concerned temporal averaging of rainfall in the SiB experiment. The hydrologic response experiment suggested that at scales larger than about one kilometre there has been enough sampling of the smaller scale heterogeneities so that the average response is well represented by a macroscale model with average parameters.

Avissar and Mahrer (1987) developed a pseudo-three-dimensional numerical mesoscale model of the surface which represents agricultural fields by patches of dry bare soil and vegetated areas. They computed energy balance equations and surface energy fluxes as well as soil moisture and temperature profiles separately for each patch. Assuming that horizontal fluxes between different patches are small in comparison to vertical fluxes, the vertical fluxes are horizontally averaged to compute global or total fluxes to the atmosphere from the grid cell.

Avissar and Verstraete (1990) note that heterogeneity increases with the horizontal scale of the domain of interest, but that large heterogeneities can also be found on a scale of a few metres. Milly (1991*a*) and Van Genuchten (1991) emphasize the importance of soil heterogeneity at a variety of scales. The sensitivity of climate simulations to subgrid-scale variability is not very well known.

Two types of heterogeneity need to be considered according to Avissar (1991). These are landscape patchiness — associated with different types of surface areas/land use — and intra-patch heterogeneity associated with small-scale variability in soil, vegetation and terrain parameters. Patches are often called 'hydrological units or zones' with 'uniform, or at least similar, hydrological behaviour' (Becker, 1991). Avissar (1991) proposes a parameterization based on landscape patchiness. Probability density functions are applied to each landscape type for the key parameters of the soil-plant-atmosphere system.

The current aproach to dealing with landscape patchiness within a grid cell area is to combine all the single patches of a certain type into the one unit and to model any of such aggregated patch units with the appropriate component model. Total fluxes from the entire grid cell are then obtained by area weighting. Such 'semi-distributed' approaches ignore the real spatial distribution of a particular landscape type and any interactions with adjacent types. It provides a compromise between the 'big leaf — big stoma' approach and the detailed distributed methods (Becker and Serban, 1990).

A generalization of the semi-distributed modelling approach was described by Avissar and Pielke (1989). Each grid cell is subdivided into homogeneous subregions. They assume also that horizontal fluxes between subgrid areas are small in comparison to the vertical fluxes and that variation in atmospheric conditions across a grid area is small at the top of the atmospheric surface layer. Patches of the same surface type in different locations are then regrouped into one subgrid surface class and energy balance formulations are applied to each homogeneous subgrid surface class. They noted that parameterization of subgrid-scale processes in AGCMs was even more difficult, since their horizontal resolution is less than that of mesoscale climate models. Approaches similar to that of Avissar and Pielke (1989) are described by Becker and Pfuetzner (1987) and Famiglietti and Wood (1991). In a related development, the macroscale has been considered in the analysis of rainfall fields through the testing of stochastic and scaling (fractals and multifractals) models in conjunction with remote sensing (Lovejoy and Shertzer, 1991; Olson, *et al.*, 1992).

Several papers at the 1993 IAHS Symposium on Exchange Processes at the Land Surface for a Range of Space and Time Scales, held in Yokohama (Bolle, *et al.*, 1993) provide an excellent overview of recently developed strategies and approaches for aggregation and upscaling over varied, heterogeneous terrain.

Avissar and Chen (1993) assume that atmospheric variables can be separated into large scale, mesoscale and turbulent scales. Noting that AGCMs have hitherto largely ignored mesoscale fluxes, they describe a set of prognostic equations for use in AGCMs which include both mesoscale and turbulent fluxes. They demonstrate the use of mesoscale kinetic energy for parameterizing mesoscale fluxes generated by subgrid-scale landscape discontinuities in simulations of the impact of patchy deforestation in the Amazonian rain forest. Their simulations have been carried out with the Colorado State University regional atmospheric modelling system (RAMS) on a 50 x 50 km deforested area in a simulated rain forest domain of 250 x 250 km.

Blyth, *et al.* (1993) describe the use of the detailed 3D numerical model PERIDOT and a simple 1D model to estimate evapotranspiration components from a partially wet 100 x 100 km HAPEX area using several aggregation schemes. Whereas the prediction of total evaporation can be accurate using simple averaging schemes, the partitioning of evaporation into transpiration, evaporation of intercepted water, and evaporation from bare soil requires more complex aggregation methods.

The Colorado State University RAMS was used by Pielke, *et al.* (1993) to simulate the impact of real world variability in topography and albedo. They present an analytical framework for parameterization of landscape — driven mesoscale fluxes for use in AGCMs and numerical weather prediction models (NWPMs) using frequency and length scales of land surface features.

A strategy for identifying micro, meso, and macroscale heterogeneity has been described by Raupach (1993) who presents modelling results obtained with a slab model to illustrate the use of his spatial averaging schemes over contrasting land surfaces with and without energetic interdependence.

Koster and Eagleson (1990) note that all land surface models need a certain amount of parameterization, because of the differences in spatial scale of land surface and atmospheric processes. The grid spacing of AGCMs cannot resolve individual catchments or vegetation stands. Yet, the land surface component of AGCMs must parameterize the effects of subgrid-scale variability on the water and energy balances of the AGCM grid cell.

The importance of accounting for subgrid-scale precipitation variability is discussed by Entekhabi and Eagleson (1989) and Shuttleworth (1988). Shuttleworth described a method of retaining the spatial extent and intensity of precipitation for AGCMs. This method assumes that rain falls over a proportion μ of the grid cell, where μ is arbitrarily chosen as 1 for so-called "large-scale dynamical rain" and 0.3 for "convective rain". The AGCM computes the relative proportion of these two types of rainfall at each place and time as it operates. The local rainfall rate is represented by a specified probability density function and can be computed from the average rainfall rate and μ. Shuttleworth also describes the use of this method to compute total surface runoff. Pitman, *et al.* (1990) and Pitman (1991) used BATS to investigate the Shuttleworth method. They showed the importance of subgrid-scale parameterization of rainfall and indicated that great care must be taken unless μ is predicted by the AGCM.

Pitman (1991) has described several methods for accounting for subgrid-scale heterogeneity in the representation of the land surface in AGCMs. He observed that aggregating data leads to different results compared to choosing the most common ecotype as representative of the grid cell.

Intrapatch heterogeneity (Avissar, 1991) occurs on a microscale, i.e. on a much smaller scale than landscape patchiness, and is usually more random. Avissar has illustrated intrapatch heterogeneity with examples based on aspect of the slopes of small furrows and stomatal resistance. Becker (1991) notes that it will be more appropriate to assess intrapatch heterogeneity with statistical approaches based on distribution functions for the relevant land surface characteristics. He describes an application of the model of Becker and Pfuetzner (1987) which simulates evapotranspiration with a distribution function for the water holding capacity of the root zone for unsaturated areas in a river basin.

Abramopoulos, *et al.* (1988) have developed a parameterization scheme for land surface-atmosphere interactions which considers spatial subgrid-scale variability through an area-weighted compositing scheme for vegetation and soil parameters based on 1° x 1° grid data. Their model includes explicitly the processes of transpiration, evaporation of intercepted precipitation and dew, evaporation from bare soil, infiltration, soil water flow, and runoff. The compositing scheme is evaluated by off-line tests comparing GISS-AGCM data obtained for 8° x 10° regions in Brazil, Sahel, Sahara and India with surface hydrology model calculations for the constituent 1° x 1° cells. The sensitivity of the surface hydrology model was greatest for fractional ground cover and least for soil hydraulic conductivity and matric potential.

Statistical parameterizations, which take into account intrapatch heterogeneity, have been suggested by Freeze (1980), Entekhabi and Eagleson (1989, 1991), Famiglietti and Wood (1991), and Avissar (1991). Entekhabi and Eagleson (1989) have derived expressions for the processes of bare soil evaporation, transpiration and runoff by assuming statistical distribution functions for precipitation and soil moisture at the subgrid scale. Their study attempts to account for both non-linearity of evaporation in relation to soil moisture and for the interaction between vertical and lateral fluxes in determining surface moisture distribution. In their expressions, the runoff ratio is a function of the relative saturation of the soil at a point, the unsaturated soil hydraulic properties, and the precipitation intensity. Soil capillarity and the strength of the atmospheric evaporative demand are incorporated in the bare soil evaporation model, whilst a simplified root soil moisture extraction model is used to obtain the transpiration efficiency. Mean values of the calculated fluxes are used in AGCM water balance calculations.

Famiglietti and Wood (1991) developed a macroscale hydrological model which incorporates subgrid-scale variability in topography, soils, vegetation, and rainfall. They describe the spatial distribution of soil moisture with the help of a gamma distribution of a topography-soils index and several soil parameters and compute the spatial distribution of infiltration and evapotranspiration. However, the derivation of their topography-soils index requires several assumptions, in particular with relation to rainfall distribution, which affect the predictive basis of their method.

Despite the progress described above in tacking subgrid-scale variability, there remains in practice a clear gap between the scales applied in atmospheric models (especially the AGCMs) and advanced land surface hydrological models. This is due in part to the current re-appraisal of existing physically-based models and to the difficulties in their application to the scale of a drainage basin (Beven, 1989, 1992; Bathurst and O'Connell, 1992). It is also due to the fact that large basins are a complex mosaic of land cover domains, each with characteristic water and energy dynamics that are difficult to quantify.

6.4 Scale issues in large-scale hydrology: the drainage basin perspective

The hydrological cycle provides the crucial link between climate, vegetation and the soil. The constituent processes of the hydrological cycle may show considerable spatial and temporal variability. On a local, short time scale there is the contrast between humid and dry surfaces, which may induce mesoscale circulations, trigger convection, and enhance rainfall. This has been studied for dry and humid strips of land (Anthes, 1984), as well as for the contrast between forest and agricultural lands (André, *et al.*, 1986). Such contrasts may be important in the link between the global energy cycle and local climate.

At a larger spatial scale and over longer time periods, soil moisture anomalies can induce long-lived responses in atmospheric circulation patterns. Positive (wet) anomalies would result in increased evaporation, increased precipitation, and reinforcement of such positive anomalies. On the other hand, negative (dry) anomalies may persist for weeks or months.

At a very large spatial scale, for very long time periods, one must consider changes in land surface processes in response to global climate dynamics and climate changes in response to natural or man-made changes at the land surface. On the one hand, global warming may affect water storages, spring flooding, and drought dynamics. On the other hand, there is the likely impact on climate of massive land use changes associated with deforestation and large irrigation schemes.

Theories of many processes including infiltration, evaporation, overland sediment transport, and subsurface water movement have been, and continue to be, developed for small space/time scales. The extrapolation of the theory of such non-linear hydrological processes to large-scale natural systems, such as drainage basins or flood plains and to AGCM grid cells poses serious problems. New analytical formulations, experimental verification and methods for parameterizing soil-water-vegetation interactions are needed for the prediction of hydrological processes at large scales. Such prediction tools are required in regional water resources planning and management, in the calibration and validation of AGCMs, and in the prediction and interpretation of the hydrologic effects of climate change.

The rationale for a grid-based global hydrologic model (GHM) is discussed by Harrison, *et al.* (1991). They propose a structure for such a GHM, which ultimately should be coupled with an AGCM. They review the spatial and temporal scales of important hydrological processes and consider whether data was available at a particular scale. Their review suggests a GHM grid size of 10 km and a daily time step.

GHMs and other large-scale hydrological models aggregate energy and mass fluxes over large grid cells, river basins, sub-basins or catchments. They use lumped parameters to describe the land surface over such areas. Validating such models for large basins will involve comparing monthly values of simulated runoff and streamflow with measured river discharge.

Such hydrological models (with grid sizes of 10 km and more) will be based on soil moisture accounting (water balance) formulations as used in the Stanford watershed model and the Sacramento model. The models must incorporate submodels for vertical water and energy fluxes, water storage on the surface and in the soil, and horizontal routing of surface and subsurface water. Harrison, *et al.* indicate that the soil-vegetation-atmosphere-transfer (SVAT) models now available, including SiB, BATS and BEST, are not immediately suitable for incorporation in a GHM and that a simplified, low-resolution SVAT model needs to be developed. The development of routing models to channel surface and subsurface runoff through appropriate drainage networks is conceptually straightforward using simple electrical analogue methods, yet computationally demanding. The daily weather data required for the GHM may be obtained with a stochastic spatial weather generator using monthly climate data or with suitable interpolation techniques from grid data generated with the GHM.

Becker (1991) has noted the desirability of having uniform 'external systems inputs' (i.e. rainfall, snowfall, radiation) over the large land area of interest. This would require areal disaggregation of AGCM grid cell, region or river basin into meteorological zones where this condition is fulfilled.

The concept of macroscale hydrology is addressed in a number of recent papers (Becker and Pfuetzner, 1987; Shuttleworth, 1988; Becker and Nemec, 1987; Vorosmarty, *et al.*, 1989).

The recent IAHS Symposium on Macroscale Modelling of the Hydrosphere (Wilkinson, 1993) held in Yokohama showed a wide range of approaches to representing land surface processes in macroscale hydrological modelling. There are two approaches to macroscale modelling according to Arnell (1993). The first involves a top-down approach in which a simple conceptual hydrological model is applied to coarse gridded data over large areas. Grid cells are treated as single lumped catchments and simple routing procedures are used to simulate flows. The second approach is bottom-up and involves the use of detailed physically-based hydrological process models at high spatial and temporal resolution, i.e. in each catchment in a large area.

Vorosmarty, *et al.* (1993) enlarged the definition of the macroscale hydrology model to include a variable resolution structure bounded by the relatively fine (hillslope/small catchment) scale and the coarse (mesoscale; AGCMs) resolution, with a focus on the mesoscale, as shown in Figure 9. This concept relies on the simultaneous treatment of the dynamics of both the atmosphere and land surface over large domains, whilst preserving hydrological processes occurring in small catchments. The spatial domain, and indeed the spatial discretization of the mesoscale is much larger than the traditional experimental drainage basin.

Jakeman, *et al.* (1993) note that detailed physically-based distributed models are only partially successful in their representation of the hydrological cycle because many of the key biological and hydrological

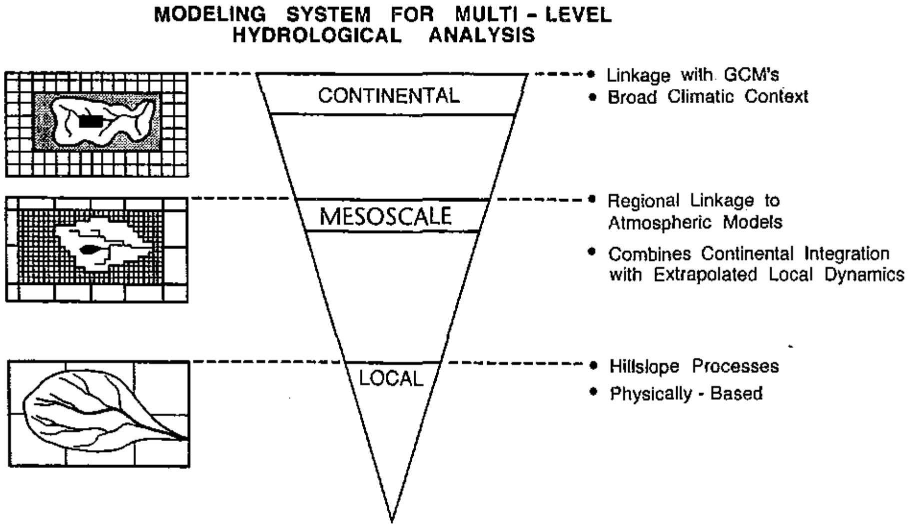

Figure 9 — Variable resolution strategy to develop linked atmosphere-hydrology models for regional-scale studies (after Vorosmarty, *et al.*, 1993).

interactions of the soil-vegetation-atmosphere system are only partially understood. They describe a conceptually simple, lumped unit hydrograph model which may be used to estimate daily runoff and evaporation from a few years of precipitation, temperature, and discharge data. The model is transferable across a wide range of hydroclimatological regimes and catchment sizes where reasonable quality rainfall, streamflow, and temperature data are available. It can be applied, in principle, at a continental scale.

The paper by Bobinski, *et al.* (1993) describes a successful comparison of a rainfall-runoff model based on the partial area concept and an atmospheric-land surface mesoscale model for a 760 km^2 mountainous basin in Poland. Their results point at the feasibility of developing a hydrologic model linked to an atmospheric model, with distributed parameters accounting for areal distribution of precipitation, soil properties, vegetation, and the river network topology.

Jolley and Wheater (1993) have described the performance of a grid-based SVAT scheme — the U.K. Meteorological Office rainfall and evaporation calculation system (MORECS) — with respect to runoff generation for a 40 x 40 km grid-based model of the 10 000 km^2 Severn River basin. They achieved reliable monthly and annual runoff figures but noted the importance of improved rainfall distribution which require improved techniques for generating realistic rainfall fields across large basins.

There is a need to integrate hydrological processes operating at finer scales to regional and continental scales and to couple hydrological and atmospheric models. Drainage basins are aggregations of small watersheds into large drainage networks that encompass regional and continental scales that are similar in scope to what is simulated in atmospheric models. They are clearly circumscribed landscapes comprising both fluvial and interfluvial areas, but they can also be thought of in terms of coupled, ordered river networks. The drainage basin facilitates the closure of water and material cycles, permitting regional and continental-scale water inventories to be obtained by careful tracing of fluxes into and out of the atmosphere, land-based ecosystems, and river systems. Discharge records and inert tracers can provide a reliable indication of the status of the land-based water cycle. The drainage basin approach encourages the validation of linked models of atmospheric, terrestrial, and aquatic ecosystems in which complex feedbacks among the component systems can be identified and quantified. Such a drainage basin approach is increasingly possible by combining improved modelling techniques, field-scale monitoring and remote sensing tools.

Within the drainage basin context, coupled atmospheric-hydrologic models may be viewed as a hierarchical structure with finer-scale, site-specific submodules interacting with simulations over broader domains. Such a configuration seeks to distil the impacts of global change down to regional and local levels, while simultaneously propagating physically valid dynamics up to regional and synoptic scales — a coupled "top-down" and "bottom-up" approach. At the coarse scale, the spatial and temporal scales for atmospheric calculations are typical of current AGCMs, (approximately 200 km length scale and subhourly time steps). The macroscale hydrology component could operate at length scales from 10 to 50 km or larger, and at weekly-monthly time steps. At the other end of the spectrum, hill slope catchment models could be operating with Digital Elevation Models (DEMs) at grid spacings well below one kilometre in conjunction with subdaily meteorological observations.

Mesoscale atmospheric models with 60 km resolution have already been used to predict climate variability to scales which are much smaller than those of the AGCMs (see Giorgi and Mearns, 1991). Mesoscale models such as MM4 developed at the National Center for Atmospheric Research (NCAR) and Pennsylvania State University have successfully simulated the general atmospheric dynamics down to scales as small as 25 km (Anthes, *et al.*, 1987; Anthes, 1990). Important feedbacks between atmosphere and land surface are now studied at the mesoscale. Non-hydrostatic mesoscale models should bring this limit down to a few kilometres. These scales coincide with those applicable in regional models of river basins 10^3 to 10^4 km^2 in area, with grid resolutions of the order of 1–10 km.

Although both, coarse and fine scale analysis offer advantages, an intermediate scale of resolution is needed to provide a process-based view of hydrology at the regional scale. A mesoscale atmosphere-hydrology model will provide the intermediate scale of resolution between fine and coarse scales. The model will simulate a series of subregional basins characteristics of the overall regional drainage system. The regional domain atmospheric models will operate on the order of minutes and a 10–50 km spatial resolution and land surface hydrology models will maintain daily time steps and a resolution of 1–10 km.

Vorosmarty, *et al.* (1993) point out that such a mesoscale atmosphere-hydrology model arises from the application of reciprocal scaling functions between the local and continental-scale endpoints. These take the form of spatially-correlated statistical distributions that can be based on observed local data (Eagleson, 1978). Once a classification of all subregional basins is applied to the entire domain at the GCM scale, a characteristic set of scaling functions is used to interpret GCM derived climatic forcings for each subregional mesoscale drainage basin. Simultaneously, dynamics developed at the local hillslope or small catchment resolution (i.e. for energy, momentum, water vapour, and runoff) are scaled back to the mesoscale, integrated and later passed to the synoptic scale for incorporation into the basin-wide AGCM-hydrology model. In this way, an area-integrated flux of latent heat can be computed as a lower boundary condition within an AGCM grid and local distributed runoff can be aggregated to produce large river discharges. Such a multi-tiered approach requires an integrated nesting of models and associated data sets within an interactive geographic information system.

CHAPTER 7

INTERNATIONAL PROGRAMMES AND MAJOR FIELD EXPERIMENTS ON LAND SURFACE PROCESSES IN LARGE-SCALE HYDROLOGY

7.1 Major programmes

7.1.1 *Global Energy and Water Cycle Experiment (GEWEX)*

GEWEX was initiated by the Joint Scientific Committee of WMO and the International Council of Scientific Unions (ICSU) as the third major project-oriented activity in the World Climate Research Programme (WCRP). Its aim is to observe, understand, and model the hydrological cycle, together with the energy fluxes in the atmosphere, at the land surface, and in the upper parts of the oceans. GEWEX aims to determine global distributions of water and energy fluxes from observations and to compute their values from predicted atmospheric properties, for the purpose of quantifying the energetic processes of the Earth's climate system and the atmospheric forcing on the ocean, land, ice, and vegetation.

More details on GEWEX are given in Appendix I.

7.1.2 *International Geosphere-Biosphere Programme (IGBP)*

IGBP is a worldwide programme of the International Council of Scientific Unions (ICSU) for the 1990s and is concerned with human-induced global change. It has strong interactions with the World Climate Research Programme (WCRP) and the Global Energy and Water Cycle Experiment (GEWEX).

The objective of IGBP is to describe and understand the physical, chemical, and biological processes that regulate the entire Earth system, the changes that occur in this system, and the manner in which they are influenced by human activities.

IGBP has core projects which focus on important components of the Earth system (IGBP, 1990). These include the biospheric aspects of the hydrological cycle (BAHC) and the global change and terrestrial ecosystems (GCTE).

The BAHC core project has been set up to study how vegetation interacts with the physical processes of the hydrological cycle. Its main aim is to gain an understanding of the role of terrestrial vegetation, in combination with the topographic and pedologic structure of the land surface layer, in the hydrological cycle. Such predictive understanding must be for scales compatible with global observing systems and climate models and should form the basis for global climatic and ecological studies.

Appendix I provides further details on the IGBP and BAHC core project.

7.2 Recent and current international field experiments

In Appendix II an overview is given of two major programmes of field experiments, the Hydrological Atmospheric Pilot Experiment (HAPEX) and the International Satellite Land Surface Climatology Project (ISLSCP).

The HAPEX programme was initiated in the early eighties by the World Climate Research Programme (WCRP) of WMO and ICSU. Its major objective is to conduct experiments aimed at improving the land surface hydrology in AGCMs for representative biomes. Appendix II also describes the recent HAPEX-MOBILHY experiment in south-western France. The current HAPEX-Sahel experiment is noted in Appendix III.

The aim of ISLSCP is to encourage and conduct research towards the use of satellite-based remote sensing for global climate modelling over land surfaces and calibrating land surface descriptions in global climate models. Appendix II briefly describes the First ISLSCP Field Experiment (FIFE) conducted in Kansas in 1987 and 1989.

Finally, in Appendix III a brief description is given of three international field experiments currently in progress. These are HAPEX-Sahel in Niger, the European Field Experiment in Desertification Threatened Area (EFEDA) in Spain and the Boreal Ecosystem-Atmosphere Study (BOREAS) in North America.

7.3 Future field experiments

It has been noted in GEWEX planning discussions (ICSU/WMO, 1990*d*) that field experiments are required to obtain insight into the significant interactive biophysical processes in soil-vegetation-atmosphere systems, to develop and validate simple one-dimensional models, and to provide data at appropriate spatial and temporal scales for parameterization of land surface processes. Particular attention should be given to the hydrological and biophysical impact of tropical deforestation and of drought in semi-arid regions and desert margins.

Planning for IGBP (IGBP, 1990) has resulted in the recommendation that future land surface experiments should maintain long-term monitoring over several years in selected subareas nested within experimental regions. IGBP has also identified tropical deforestation and desertification as priority issues and has emphasized the need for land surface experiments for a range of critical biomes.

Shuttleworth (1991*a*, *b*) emphasized the value of longer term measurements at single sites for the calibration of surface parameterization schemes and for assessing AGCM performance, provided such sites are large, uniform and well watered with a weak precipitation gradient. He noted that the monthly average of weather variables and fluxes at a site may be about equal to the monthly average values for the relevant AGCM grid box. Thus, the time average may be used as a surrogate for the area average. Shuttleworth also concluded from HAPEX and FIFE that there is no place for complacency regarding instrumental performance and accuracy in measurements of net radiation and surface energy fluxes.

Hydrological data for catchments, synthesized to the scale of the AGCM grid mesh, will be very useful in validating long-term AGCM predictions of present rainfall and runoff with existing surface cover. However, the greatest calibration needs of AGCMs are at the daily and hourly time scale when hydrological models are least applicable. Yet the need remains to underpin surface and above-surface measurements with the best available hydrological model. The calibrated AGCM response should be capable of simulating both short- and long-term responses. This test may involve upgrading the surface component in the hydrological model. The hydrological model, on the other hand, can assist in timing catchment variables. The representativeness of the AGCM calibration must be established by testing its ability to represent the more slowly varying catchment measurements over time.

HAPEX, FIFE and other experiments have shown significant discrepancies between airborne- and ground-based flux measurements. Until these discrepancies are explained and corrected, the aircraft-based sensors cannot be seen as the primary source of data to test aggregation hypotheses. However, aircraft data are valuable in showing that mesoscale models can provide realistic simulations of land use interactions at the 10 km scale. Shuttleworth (1991*a*) concludes that aircraft should be more seen as providing input data for boundary layer models rather than reference data on areal flux measurements.

Future land surface experiments should provide data which are appropriate for the validation of soil-vegetation-atmosphere-transfer (SVAT) schemes which are being developed at a range of scales, especially those to be incorporated into AGCMs, and the interpretation and use of remote sensing data.

Future experiments should include long-term monitoring of the physical environment and calibration of remote sensing data in different biomes. It is important that such experiments are implemented in critical biomes, such as those facing desertification and deforestation.

Climate–land surface interactions must be studied over a range of spatial and temporal scales. Spatial variability of land surface processes can be important at different scales. Field experiments and SVAT model development, which can take place at the local scale, the landscape scale, and the regional (river basin) scale, should be nested. The continental scale should also be recognized for modelling studies.

First, at the local (patch) scale of 1 x 1 km, vegetation, soil, and terrain are uniform and one-dimensional SVAT models can be developed and evaluated for single vegetation patches. At this scale, surface inhomogeneities greater than several hundreds of metres must be accounted for. This is the scale of agricultural field experiments which address plant control of transpiration as a function of leaf area and phenology. The SVAT model must also consider the roles of surface, canopy, and aerodynamic resistances in transpiration and evaporation, soil physical characteristics affecting water holding capacity of soil, infiltration and deep percolation, and radiation and canopy microclimate.

Second, for small catchments, i.e. at the landscape scale of 10 x 10 km, vegetation and soil are patchy but the terrain is uniform. This requires one-dimensional SVAT models based on spatially-integrated values for surface properties and environmental data. Such aggregated values may be obtained from statistical treatment of surface heterogeneities and measurements, or from the use of remote sensing imagery in which the pixel is treated as the integrating entity.

Third, at the meso or regional scale of 100 x 100 km, vegetation and soil are patchy and the terrain and microclimate are non-uniform. Mesoscale atmospheric circulation, such as that associated with land-sea breezes, slope circulations, and convective motions, can be triggered by inhomogeneities in land cover, terrain or soil properties, at scales of beyond 10 km. A similar scale of about 10 km applies to the atmospheric forcing of land surface processes by frontal systems or convective storms. Field studies and numerical mesoscale models must be used at this scale to develop relationships for the interactions between vegetation and atmosphere. Mesoscale models are very powerful tools to investigate the effects of spatial integration.

Finally, at the continental scale, the representation of land surface hydrology and mesoscale atmospheric phenomena in continental and global models must be addressed.

Variability of land surface processes will require field observations and measurements with appropriate time and space resolutions in order to develop and test parameterization schemes. Such field experiments should take place over relatively short periods, in relatively small areas, in carefully selected regions, and under representative climatological conditions. There is also a need to determine the variables which drive land surface processes on a large scale from satellite remote sensing and large-scale monitoring networks.

Shuttleworth (1991*a*) observed that proper representation of land surface and atmospheric processes over continental areas will require mesoscale models with a spatial resolution of about 10 km, despite the fact that grid resolution of AGCMs will remain greater than 100 km for a long time to come. He also noted that 5–10 km is also the pixel size of satellite remote sensing imagery which can be used effectively to study land surface processes.

In general, field experiments must take place at two levels. First, at a scale of 1–10 km (i.e. local patch and landscape scales) for testing one-dimensional formulations of atmospheric processes and for the detailed evaluations of SVATs. Second, at a scale greater than 100 km, to carry out three-dimensional studies of mesoscale variability in the atmospheric boundary layer (ABL) and at land surface, and for the development of ABL codes used in AGCMs. At both scales, field experiments must be aimed at collecting aggregated data at the appropriate space and time scales to compare with the results of mesoscale models and satellite remote sensing for complex land surfaces.

CHAPTER 8

CONCLUDING REMARKS

This report recognizes that global-, continental- and regional-scale activities in climate modelling, hydrology, and ecology require study and description (parameterization) of the interactions between climate and land surface processes.

Chapter 2 shows that broad, large-scale climatic conditions are defined by general circulation patterns with zonal, meridional and vertical components. Mesoscale variability is imposed on these broad patterns. This variability is caused by synoptically-induced or terrain-induced mesoscale systems within the atmosphere. The latter are clearly linked to topography and land surface processes, as evidenced by the studies reviewed in this chapter. Such mesoscale variability is expressed in spatial patterns of rainfall, surface fluxes (including evaporation), and state variables (e.g. soil moisture content).

The review in Chapter 3 indicates that land surface parameterization in atmospheric general climate models is still inadequate despite much experimental evidence that simulations of global climate are sensitive to prescribed states of the land surface. Considerable work needs to be done on improved descriptions of large-scale hydrological regimes as well as on suitable parameterizations of the important land surface processes. This will also assist prediction of the impact of major changes in land use/cover on climate. It is shown that numerical mesoscale models and embedding procedures will give improved predictions of mesoscale phenomena through their greater spatial resolution.

Difficulties in large-scale parameterization of critical land surface process and evaporation, are discussed in Chapter 4. Evaporation at the land surface can vary widely with time and space, especially between different plants and land surfaces. A review of several methods of estimating areal evaporation, including the combination equation, climatological techniques based on the Priestley-Taylor expression and the complementary relationship, and atmospheric boundary layer methods, illustrate the problems of coupling atmosphere and biosphere and of predicting evaporative fluxes and soil moisture across non-homogeneous land surfaces.

Major advances have been made in recent years on the use of remote sensing for a wide range of climatological, hydrological and agricultural applications because it provides a time series of instantaneous, spatially-integrated values over non-homogeneous land surfaces. Chapter 5 outlines how satellite-based remote sensing is used in regional-scale studies of land surface processes (including surface radiation budget, evaporation, and precipitation) and in descriptions of land surface characteristics/state variables (including soil moisture status, land cover, and biomass) in regions which vary in terrain, soil and cover.

Models of land surface processes predict the lower boundary conditions for climate prediction with atmospheric general circulation models. Chapter 6 reviews several hydrology models based on one-layer and two-layer water balance models with simple evaporation parameters which have been used in current AGCMs since the early eighties. Reference is made to recent sensitivity studies which show that such simple models do not realistically represent short-term dynamic variations in surface hydrology. Two simple hydrology models are described which are used in the assessment of large-scale hydrological impacts of climate change. A review of parameterizations of land surface processes currently used in AGCMs shows that most models are lumped, i.e. they assume homogeneity at the land surface and few models have been validated on a large scale.

Components of the regional water balance, such as evaporation and runoff, will depend strongly on spatial patterns in rainfall and on the relative importance of water surfaces, impervious areas, flat areas with deep or shallow groundwater, and hillslopes with shallow or deep soils. Land use (e.g. urban areas, forests or agricultural fields) and soil characteristics are also major factors in spatial variability. A distinction is made between the succession of major terrain and land use types ('landscape patchiness') and small-scale variability in soil and vegetation ('intra-patch heterogeneity'). There has been considerable effort in recent years in the development of land surface parameterizations which incorporate subgrid-scale terrain, soil and vegetation heterogeneity as well as climatic variability (particularly in precipitation). This is a critical problem facing climate modellers as well as those involved in the development of large-scale hydrology models.

Current approaches in hydrological modelling to subgrid-scale variability and land surface non-homogeneity include the use of area-weighted compositing schemes and semi-distributed modelling techniques for landscape patchiness, and the use of statistical probability distributions for rainfall and for intra-patch heterogeneity due to small-scale variability in terrain, soil, and vegetation attributes.

To assess the response of hydrological systems to global change and major land use changes requires continuous coupled atmospheric-hydrological models with manageable surface hydrology parameterizations. These models must consider magnitude, frequency, and distribution of rainfall events and evaporation.

Theories for many non-linear processes, such as evaporation, infiltration, and overland transport, have been developed for small space/time scales and their extrapolation to large scales poses serious problems. There is conflicting evidence as to whether surface fluxes and surface characteristics scale linearly.

Most current large-scale hydrology models concentrate on the channel network despite considerable process in hillslope hydrology and modelling of small catchments. These models generally treat land surface processes in a very simplistic manner. New models are needed to predict streamflow and groundwater recharge as well as continuous soil moisture accounting techniques. One approach could be the development of a hybrid, semi-distributed model which uses a distributed model based on the natural drainage network and lumped parametric or physically-based conceptual models for individual subcatchments. Such an approach could be used to investigate interactions and relationships between inputs, outputs, and state variables at various scales.

Ecologists and agronomists are very interested in the interactions between atmosphere, ecosystems and crops. Climate models have been used to study the impact of major changes in land use and land cover, such as tropical deforestation and climate desertification interactions. Such hydro-ecological studies require realistic land surface process representations, sensitivity tests, and evaluation of current biosphere models against ground truth at various scales. In addition, the direct effects of increases in CO_2 on photosynthesis and transpiration need to be assessed by studying climate interactions at various scales. This will involve the development of appropriate plant-environment models for considering plant productivity and water use, which incorporate radiative transfer processes.

In Chapter 7 and Appendices I to III an overview is given of several international programmes and major field experiments on land surface processes. An assessment of their overall objectives, the rationale for site selection, scope, and cost-effectiveness goes beyond the scope of this report. Recent experiments have provided significant insight into land surface processes. Field experiments currently being planned aim to:

(*a*) Validate hydro-meteorological (SVAT) models;
(*b*) Develop procedures and suitable algorithms for calibrating and using remote sensing;
(*c*) Study climate-land surface interactions over a range of spatial and temporal scales;
(*d*) Assess spatial variability in surface fluxes; and
(*e*) Test approaches for incorporating subgrid-scale variability in hydro-meteorological models.

In conclusion, this report has emphasized that the spatial resolution of existing climate models and large-scale hydrology models is very much larger than the characteristic scale of the relevant land surface processes. Small-scale processes need to be scaled up in some fashion so that they can be adequately represented at the large scale. Further work is needed to determine what level of detail is needed in specifying the surface hydrology in terms of vertical resolution and horizontal heterogeneity, and to model the interaction between suitably aggregated surface water, groundwater and atmospheric processes.

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APPENDIX I

CURRENT INTERNATIONAL PROGRAMMES ON LAND SURFACE PROCESSES OF THE HYDROLOGICAL CYCLE

I.1 Global Energy and Water Cycle Experiment (GEWEX)

GEWEX was initiated by the Joint Scientific Committee of WMO and the International Council of Scientific Unions (ICSU) as the third major project-oriented activity in the World Climate Research Programme (WCRP). Its aim is to observe, understand, and model the hydrological cycle, together, with the energy fluxes in the atmosphere, at the land surface and in the upper parts of the oceans. GEWEX aims to determine global distributions of water and energy fluxes from observations and to compute their values from predicted atmospheric properties, for the purpose of quantifying the energetic processes of the Earth's climate system and the atmospheric forcing on the ocean, land, ice, and vegetation.

GEWEX has the following four scientific objectives:

(*a*) To determine the hydrological cycle and energy fluxes by means of global measurements of observable atmospheric and surface properties;

(*b*) To model the global hydrological cycle and its impact in atmosphere and oceans;

(*c*) To develop the ability to predict the variations of global and regional hydrological processes and water resources, and their response to environmental change;

(*d*) To foster the development of observing techniques, data management and assimilation systems suitable for operational applications to long-range weather forecasts, hydrology, and climate predictions.

GEWEX is, thus, concerned with the role of physical and hydrological processes in climate systems. Its tasks include the development of weather and climate simulators. It places great emphasis on understanding heat and water exchange at the land surface. Evaporation from the land surface is on a global basis equivalent to 12 W m^{-2} and a large proportion of the evaporated water returns to the neighbouring land as precipitation, because locally-enhanced convective systems are relatively short lived.

The scientific plan for GEWEX was published in August 1990 (ICSU/WMO, 1990*a*). The plan identifies two phases, a build up phase (1991–1998), and a global observing phase beginning with the deployment of Earth observing space platforms from 1998 onwards. The major tasks for GEWEX include:

(*a*) Preparation of the global data and information system for the observing platforms;

(*b*) Investigation of various key land surface and atmospheric processes;

(*c*) The refinement and improvement of models at all scales;

(*d*) Contributing to the development of the space component; and

(*e*) Conducting major international field projects.

WMO and the International Association of Hydrological Sciences have established a joint working group on GEWEX to develop scientific initiatives to be undertaken in the field of large-scale hydrology in support of GEWEX (see Klemes, 1990). This working group has decided on several international projects, including the GEWEX Multiregion Cloud System Study (GMCSS), the Global Precipitation Climatology Project (GPCP), the Global Runoff Data Project (GRDP), and a project on macroscale hydrological models. The latter project will study the capability of hydrological models to predict the impact of climatic changes on streamflow. This project has developed into the GEWEX Continental-scale International Project (GCIP), to be carried out between 1992–1998 in the Mississippi Basin.

The scientific plan for GCIP was published by the World Climate Research Programme in December 1991. The GCIP goal is "to bridge the gap between those scales significant for modelling discrete processes of the hydrological and energy cycles over land and those scales that are practical for modelling the global climate system and predicting regional impacts of climate change" (Schaake, 1991). The plan notes that the scientific aims of GCIP derive from four fundamental questions (WCRP, 1991):

(*a*) How do water and energy budgets vary in time and in space on a continental scale?

(*b*) How can surface and groundwater processes at the catchment scale be aggregated interactively with the subgrid-scale atmospheric processes in general circulation models?

(*c*) How can the diverse measurements fundamental to the determination of hydrological and energy cycles be best incorporated into analyses and coupled hydrological/atmospheric models in a consistent fashion?

(*d*) What are the best approaches for assessing the effects of climate variability and change on water resources?

The GCIP plan notes that it must assemble the necessary data sets to support energy and water budget analyses as well as to develop coupled hydrological/atmospheric models.

GEWEX will also cooperate actively with the International Geosphere-Biosphere Programme (IGBP) of ICSU, in particular through its Core-Project on the Biospheric Aspects of the Hydrological Cycle (BAHC) (see section below). GEWEX and BAHC have established a joint working group to advise on the organization and coordination of field experiments on land surface processes, especially those involving the interactions between atmospheric boundary layer, hydrology, vegetation and soils (see ICSU/WMO, 1990*b*, 1990*c*).

I.2 International Geosphere-Biosphere Programme (IGBP)

IGBP is a worldwide programme of the International Council of Scientific Unions (ICSU) for the 1990s and is concerned with human-induced global change. It has strong interactions with the World Climate Research Programme (WCRP) and the Global Energy and Water Cycle Experiment (GEWEX).

The objective of IGBP is to describe and understand the physical, chemical and biological processes that regulate the entire Earth system, the changes that occur in this system, and the manner in which they are influenced by human activities.

IGBP has core projects which focus on important components of the Earth system (see IGBP, 1990). These include the Biospheric Aspects of the Hydrological Cycle (BAHC) and the Global Change and Terrestrial Ecosystems (GCTE).

The BAHC core project has been set up to study how vegetation interacts with the physical processes of the hydrological cycle. Its main aim is to gain an understanding of the role of terrestrial vegetation, in combination with the topographic and pedologic structure of the land surface layer, in the hydrological cycle. Such predictive understanding must be for scales compatible with global observing systems and climate models and should form the basis for global climatic and ecological studies. The objectives of the BAHC Core Project are:

(*a*) To determine the biospheric controls of the hydrological cycle through *in situ* and remote sensing measurements, for the purpose of developing land surface process models at a range of temporal and spatial scales. Such models may range from soil-vegetation-atmospheric transfer models (SVATs) for small vegetation patches to atmospheric general climate models (AGCMs) and general hydrologic models (GHMs) with grid cells of hundreds of kilometres;

(*b*) To develop appropriate databases for the description of the interactions between the biosphere and the physical Earth system and for the testing and validating of the simulation models.

BAHC, thus, envisages observational studies of mass and energy fluxes in soil-vegetation atmosphere and hydrological systems at different scales. Such studies are necessary for several important reasons. First, they serve to improve our ability to integrate the understanding of small-scale hydrological processes, so that interactions at a large scale can be adequately described. Second, they are needed to test and validate modelling studies of global change in relation to the hydrologic cycle.

It has been noted (IGBP, 1990) that there are two main approaches to the scaling up problem:

(*a*) Physical reasoning is used to develop scaling-up models to proceed from validated small-scale models to models at increasingly large scales;

(*b*) Methods are developed for validating large-scale parameterizations by obtaining appropriate area averages of key surface parameters from field experiments.

BAHC recognizes the important role of land surface processes and the biosphere in the hydrological cycle. It is particularly concerned with evaporation as a source of moisture for the atmosphere and subsequent precipitation, and addresses the interactions between vegetation and the hydrological cycle at time scales of 10–100 years.

The GCTE core project studies how global changes affect the hydrological cycle. The objective of GCTE is to develop the capability of predicting the effect of changes in climate, land use and CO_2 on terrestrial ecosystems, and the feedbacks of changes in ecosystems to climate.

Several specific tasks are identified in these core projects. These include:

(*a*) Long-term monitoring and documentation of the Earth system;

(*b*) Process studies and field experiments;

(*c*) Studies to analyse and synthesize global data sets;

(*d*) The development of comprehensive models.

There are many interactions between the BAHC and GCTE projects. They have a common interest in the links between water availability and ecosystem distribution, and in the development of models for energy and mass transfer between soil, vegetation, and atmosphere.

A key question which needs to be addressed jointly by BAHC and GCTE may be formulated as follows: "As climate and land use change, how will the biotic response affect the exchange of water and energy between the terrestrial ecosystems and the atmosphere, thus affecting the feedback to further climate change?"

Dyck (unpublished) has pointed out that the specific effects of the hydrological regime on vegetation which require study depend on the time scales under consideration. The IGBP time scale — decades to centuries — implies that the main issues are species adaptation or replacement, and the associated changes in leaf area index, rooting depths, and primary productivity. Understanding such long-term effects requires to study observable components of medium- and short-term effects.

RECENT INTERNATIONAL FIELD EXPERIMENTS ON LAND SURFACE PROCESSES

This Appendix gives a brief overview of two major programmes of field experiments, the Hydrological Atmospheric Pilot Experiment (HAPEX) and the International Satellite Land Surface Climatology Project (ISLSCP).

This section draws heavily on unpublished notes and published papers by Shuttleworth (1988, 1991*a*, 1991*b*).

II.1 HAPEX

The World Climate Research Programme (WCRP) of WMO and ICSU initiated, in the early eighties, the Hydrological Atmospheric Pilot Experiment (HAPEX), with as its major aim to conduct experiments "addressing the need to improve the land surface hydrology in AGCMs for representative biomes" (Shuttleworth, 1991*a*). The important features of HAPEX are that they involve an area of the order of at least 10 000 km^2, i.e. big enough to encompass mesoscale meteorological phenomena. The area must contain an array of climatological stations and surface flux monitoring sites with measurements covering at least one year, which are underpinned by classical water balance measurements. In addition, there should be intensive experimental periods with ground-based, airborne and space borne measurements, as well as turbulent boundary layer measurements using aircraft and radiosondes (Shuttleworth, unpublished).

Shuttleworth (1988, 1991*a*, unpublished), and Noilhan, *et al.* (1991) have provided summary descriptions of the HAPEX-MOBILHY experiment carried out in south-western France in 1985–1987.

The Hydrological-Atmospheric Pilot Experiment/ Modelisation du bilan hydrique (HAPEX-MOBILHY) took place in south-western France in 1985–1987, in a 100 x 100 km^2 region. The experimental programme was coordinated through WCRP. Emphasis was on calibrating the surface hydrology parameterization in AGCMs. Its aim was to compare area-averaged evaporation fluxes obtained with hydrological techniques, surface energy budget methods, direct aircraft measurements, and mesoscale hydrolological and atmospheric models.

André, *et al.* (1986, 1988) have reported that neutron moisture metre soil moisture measurements gave evaporation data over periods of two weeks and more, which compared well with atmospheric measurements with spatial resolutions from metres to hundreds of metres. Mesoscale atmospheric models were used to test the validity of using the surface characteristics of a single hypothetical equivalent vegetative cover over regions with soil and vegetation heterogeneity at scales of one to two kilometres in comparisons with the weighted averages of component surface fluxes and the above canopy weather variables. These assumptions were found to be adequate when using mesoscale meteorological models. It appears that for vegetation with characteristic length scales of 10 km or more, local boundary layer development may invalidate this aggregation process. Three-dimensional hydrological models based on stream gauge data gave realistic areal evaporation data over periods of at least three months. Those models were found to be inadequate for shorter time periods. Numerical simulations of the surface moisture budget are described by Mahfouf (1990).

Schmugge and André (1991) have published the proceedings of a 1988 workshop held to discuss the results of the HAPEX-MOBILHY measurement campaign.

The HAPEX-MOBILHY data set is increasingly used in validating new methodologies for modelling surface energy fluxes over complete and partial canopies (see for example Jacquemin and Noilhan, 1990; Noilhan, *et al.*, 1991; Mascart, *et al.*, 1991; Ben Mehrez, *et al.*, 1992; Pinty, *et al.*, 1992; and Siebert, *et al.*, 1992).

II.2 ISLSCP

The aim of the International Satellite Land Surface Climatology Project (ISLSCP) is to stimulate and conduct research oriented towards evaluating and implementing the use of satellite-based remote sensing data to provide routine global monitoring of climate over land surfaces, and techniques for calibrating the land surface description in models of global climate (Shuttleworth, 1988).

The First ISLSCP Field Experiment (FIFE) was carried out in Kansas, U.S.A., during 15–20 days in both 1987 and 1989, in a 15 x 15 km^2 region with relatively uniform pasture. FIFE involved a large number of surface and airborne flux measurements and used remote sensing obtained with aircraft and satellites. It was conducted in close cooperation with space agencies. Its aims were to study the various biophysical processes which control soil–vegetation–atmosphere transfer (SVAT) processes at scales of a few metres to several kilometres, and to develop techniques for applying remote sensing data to SVAT models.

Hall, *et al.* (1989) and Sellers, *et al.* (1989) have reported on some of the FIFE results. Surface radiation and heat fluxes were calculated and non-linear energy balance and vegetation process models were used to integrate over a range of spatial scales. Heat and vapour fluxes from soil and vegetation were successfully separated. Various algorithms were tested for determining the components of the surface radiation budget and for estimating the photosynthetic capacity and maximum canopy conductance from remote sensing data. Differences between sites were generally small. FIFE yielded good illustrations of important errors in micrometeorological measurements.

FIFE also provided a warning about the use of radiometric surface temperature in the estimation of sensible heat flux. Evaporation estimates using surface temperature showed significant systematic error and spatial and temporal variability in differences between radiometric and aerodynamic surface temperatures.

APPENDIX III

CURRENT INTERNATIONAL FIELD EXPERIMENTS ON LAND SURFACE PROCESSES

Three international field experiments are in progress. First, the Hydrological Atmospheric Pilot Experiment in the Sahel (HAPEX-Sahel) studied land surface processes in Niger in an arid region of 100 x 100 km, some two months after the rainy season in 1992. There were also three smaller sites, each 15 x 15 km, with detailed vegetation and flux measurements. At one of these sites this experiment will continue over three years.

Second, the European Field Experiment in Desertification Threatened Area (EFEDA) is conducted in a semi-arid well-developed area in Spain, which is threatened by desertification if major changes take place in the region's rainfall. The 1991–1996 study will addresss meteorological and hydrological issues in a study area of 86 x 130 km.

Third, a major experiment has been initiated for the boreal forest biome in North America. The Boreal Ecosystem-Atmosphere Study (BOREAS) is planned for 1993–1997 as a contribution to the International Satellite Land Surface Climatology Project (ISLSCP) and to the International Geosphere-Biosphere Project (IGBP). Its goal is to obtain an improved understanding of the interactions between the boreal forest biome and the atmosphere in order to clarify their roles in global change. The specific objectives of BOREAS are described by NASA (1992):

"To improve understanding of the processes which govern the exchanges of energy, water, heat, carbon and trace gases between boreal ecosystems and the atmosphere, with particular reference to those that may be sensitive to global change; and

To develop and validate remote sensing algorithms for transferring our understanding of the above processes from local to regional scales" (NASA, 1992).