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**The estimation of groundwater recharge by
soil water balance in semi-arid regions**

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soil water balance in semi-arid regions**

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Finally, to my wife Carla and Arthur, the smiley baby.

DEDICATION

To CARLA and ARTHUR

ABSTRACT

Quantification of groundwater recharge is a crucial prerequisite for sustainable groundwater resource management, particularly in semi-arid areas where there are large demands for groundwater supplies. This research presents an alternative approach for recharge estimation based on the soil water balance technique. The purpose is to develop a model which provides a suitable balance between physical credibility and data which realistically can be gathered.

A spreadsheet model was written based on the conceptual representation of the principal physical processes which actually affect recharge in a semi-arid area. Alternative procedures were included in order to represent: (a) the estimation of runoff, (b) the inclusion of the period with predominant bare soil evaporation and (c) the accounting for evapotranspiration following rainfall on dry soil.

The model was tested using real data from a semi-arid region (Northeast Nigeria) making use of selected periods of days and years in order to illustrate the principal model characteristics. The results were presented in the form of diagrams and graphs helping to visualise the interactions between the physical components and the effect of the additional procedures on recharge estimation.

The credibility of the model was investigated using an alternative concept of “analysis of plausibility”. This concept makes use of as wide as possible a range of quantitative and qualitative information from the hydrological system in order to verify the robustness of the model when extensive datasets required by conventional validation techniques are not available. The results suggested that the modelled recharge is physically sound and it is in line with the overall determination of recharge in semi-arid areas by a range of methods.

The soil water balance model was utilised to explore important aspects of recharge in semi-arid regions showing the effect of the field variability on the model’s output. The preliminary results show that the developed concept reasonably represents the inherent field variability, thus corroborating the strength of the approach for recharge estimation in semi-arid regions.

Table of Contents

CHAPTER ONE.....	1
INTRODUCTION.....	1
1.1. GROUNDWATER ESTIMATION IN SEMI-ARID REGIONS.....	1
1.2. ESTIMATING RECHARGE BY A SOIL WATER BALANCE MODEL.....	2
1.3. CREDIBILITY OF THE ESTIMATION METHOD WITH LIMITED DATA....	3
1.4. SEMI-ARID CONDITIONS.....	3
1.5. AIM AND OBJECTIVES.....	4
1.6. THESIS STRUCTURE.....	5
CHAPTER TWO.....	6
GROUNDWATER RECHARGE ESTIMATION.....	6
2.1. GROUNDWATER RECHARGE.....	6
2.2. SOIL WATER HYDROLOGICAL PROCESSES.....	8
2.2.1. Rainfall.....	9
2.2.2. Interception.....	10
2.2.3. Surface Runoff.....	10
2.2.4. Soil evaporation.....	12
2.2.5. Transpiration.....	13
2.2.6. Root growth and root distribution.....	14
2.3. MOVEMENT OF WATER IN SOIL DUE TO ALL HYDROLOGICAL PROCESSES.....	15
2.4. DISCUSSION.....	18
2.5. PRINCIPAL METHODS FOR ESTIMATING RECHARGE.....	20
2.5.1. Empirical rainfall-recharge expression.....	20
2.5.2. Lysimeters.....	21
2.5.3. Water table fluctuation method.....	22
2.5.4. Chemical methods.....	22
2.5.5. Zero flux plane.....	24
2.5.6. Darcy's law.....	24
2.5.7. Methods based on the numerical solution of Darcy's law.....	25
2.5.8. Soil water balance technique.....	26
2.6. AN ARGUMENT FOR SOIL WATER BALANCE AS A RECHARGE ESTIMATION TECHNIQUE.....	29
2.7. REVIEW OF EXISTING SOIL WATER BALANCE MODELS.....	31

2.7.1.	Conventional single layer model.....	31
2.7.2.	The CROPWAT model.....	33
2.7.3.	The BALANCE model.....	34
2.7.4.	The Four Root Layers Model (FRLM).....	35
2.7.5.	Discussion.....	36
CHAPTER THREE.....		37
THE SAMBA MODEL.....		37
3.1.	INTRODUCTION.....	37
3.2.	THE SAMBA MODEL.....	38
3.3.	A PRACTICAL APPROACH FOR THE INCLUSION OF RUNOFF.....	39
3.4.	THE POTENTIAL EVAPORATION FOR A CROPPED AREA INCLUDING THE PERIOD WHEN BARE SOIL IS PREDOMINANT.....	41
3.4.1.	The crop coefficients.....	42
3.4.2.	The Kc coefficient for periods when soil evaporation is predominant.....	44
3.5.	OVERALL PROCEDURES FOR ACTUAL EVAPOTRANSPIRATION DETERMINATION.....	46
3.6.	THE ACTUAL EVAPOTRANSPIRATION FOR WATER STRESS PERIODS..	47
3.7.	SOIL WATER BALANCE FOR A SERIES OF CLIMATIC AND SOIL MOISTURE CONDITIONS.....	52
3.7.1.	Soil moisture deficit less than RAW.....	52
3.7.2.	Soil moisture deficit greater than RAW.....	53
3.7.3.	Soil moisture deficit greater than TAW / TEW.....	54
3.7.4.	Near surface storage NSS.....	55
3.7.5.	Potential recharge.....	57
3.8.	THE SPREADSHEET SOIL WATER BALANCE MODEL.....	57
3.9.	COMPARISON BETWEEN THE METHODOLOGY PRESENTED IN THIS STUDY AND ALTERNATIVE SOIL WATER BALANCE TECHNIQUES.....	58
3.9.1.	The conventional single layer model.....	58
3.9.2.	The CROPWAT model.....	59
3.9.3.	The BALANCE model.....	60
3.9.4.	The Four Root Layers Model (FRLM).....	60
CHAPTER FOUR.....		62
SELECTED MODEL OUTPUTS.....		62
4.1.	INTRODUCTION.....	62
4.1.1.	Field data.....	62
4.2.	MODEL INPUTS AND OUTPUTS FOR A REPRESENTATIVE YEAR.....	64

4.2.1.	Rainfall.....	64
4.2.2.	Surface runoff.....	64
4.2.3.	Near surface storage.....	65
4.2.4.	Potential and actual evapotranspiration.....	65
4.2.5.	Soil moisture deficit and available water.....	66
4.2.6.	Overall view of the water balance and the resulting potential recharge.....	68
4.3.	THREE IMPORTANT ASPECTS OF THE SAMBA MODEL.....	69
4.3.1.	The planting date and the onset of rains.....	69
4.3.2.	Surface runoff.....	71
4.3.3.	Near surface storage.....	74
4.4.	EFFECT OF RUNOFF AND NEAR SURFACE STORAGE ON POTENTIAL RECHARGE FOR A SEQUENCE OF YEARS.....	79
4.4.1.	Effect of surface runoff on recharge estimation.....	79
4.4.2.	Effect of the near surface storage approach on recharge estimation.....	80
CHAPTER FIVE.....		86
PLAUSIBILITY ANALYSIS OF THE MODEL RESULTS.....		86
5.1.	INTRODUCTION.....	85
5.2.	AN EXPERIMENTAL FIELD TRIAL IN SEMI-ARID NIGERIA.....	86
5.2.1.	Field trial data.....	87
5.2.2.	The SAMBA model output.....	89
5.2.3.	Comparative analysis between modelled and measured soil water content.....	91
5.2.4.	Comparative analysis of modelled and estimated drainage below a maximum root depth (potential recharge).....	93
5.3.	QUALITATIVE ANALYSIS OF ALTERNATIVE INFORMATION FROM HISTORICAL RECORDS.....	95
5.4.	INSIGHTS FROM OTHER STUDIES IN SEMI-ARID REGIONS.....	100
5.4.1.	Surface runoff.....	100
5.4.2.	Potential recharge.....	101
CHAPTER SIX.....		106
VARIABILITY OF POTENTIAL RECHARGE.....		106
6.1.	INTRODUCTION.....	106
6.2.	EFFECT OF TIME-AVERAGING OF REFERENCE EVAPOTRANSPIRATION ON THE SOIL WATER BALANCE.....	107
6.3.	WATER BALANCE FOR THREE SURFACE RUNOFF SITUATIONS.....	109
6.4.	ANALYSIS OF SPATIAL VARIABILITY.....	113

6.4.1.	Analysis using the long-term rainfall series (1962-97).....	114
6.4.2.	Recharge variability for a wetter period (1974-79).....	119
6.5.	PRELIMINARY ANALYSIS OF SPATIAL VARIABILITY DUE TO RAINFALL DISTRIBUTION.....	121
6.6.	DISCUSSION.....	127
CHAPTER SEVEN.....		129
CONCLUSIONS AND RECOMMENDATIONS FOR FURTHER INVESTIGATIONS.....		129
7.1.	THE SOIL WATER BALANCE TECHNIQUE FOR RECHARGE ESTIMATION IN SEMI-ARID REGIONS.....	129
7.2.	THE SOIL WATER BALANCE MODEL AND THE RESULTS USING ACTUAL FIELD DATA.....	130
7.3.	CREDIBILITY OF THE APPROACH.....	131
7.4.	INVESTIGATING RECHARGE USING THE SOIL WATER BALANCE MODEL.....	132
7.5.	FURTHER INVESTIGATIONS.....	133
REFERENCES.....		135
APPENDICES.....		143

List of figures

Figure 2. 1. Groundwater recharge definitions.....	7
Figure 2. 2. Soil water processes.....	9
Figure 2. 3. Definitions of surface runoff including typical scale of occurrence.....	11
Figure 2. 4. Daily soil evaporation from a bare soil surface.....	12
Figure 2. 5. Rate of transpiration as a function of soil water content.....	14
Figure 2. 6. Redistribution of soil water following a single rainstorm of 72.2 mm.....	15
Figure 2. 7 Hydraulic conductivity as a function of water content.....	16
Figure 2. 8. Profile water distribution at three sites in Niger.....	17
Figure 2. 9. Schematic representation of the soil water balance technique.....	26
Figure 2. 10. Computational representation of the soil water balance in three typical situations.....	27
Figure 2. 11. The faction F relating AE and PE in absence of additional water supply according to the conventional method.....	32
Figure 2. 12. The stress coefficient as function of the soil moisture deficit.....	34
Figure 2. 13. Inputs and outputs of the BALANCE model.....	35
Figure 2. 14. Schematic representation of the FRLM model.....	36
Figure 3. 1 Two examples of runoff calculation.....	41
Figure 3. 2. Growing stages including the period before and after the crop when the surface is predominantly bare.....	43
Figure 3. 3. Daily potential evapotranspiration PE calculation for a crop season.....	45
Figure 3. 4. Physical situation after a day with a significant rainfall (40 mm).....	48
Figure 3. 5. The conventional soil water balance during a rainy day following a dry period.....	49
Figure 3. 6. The alternative near surface storage approach included in the SAMBA model in order to estimate AE following a heavy rainfall when SMD is $> RAW$	50
Figure 3. 7. NSS storage after a rainfall event of 40 mm and actual evapotranspiration 5 mm for two different values associated with soil type.....	51
Figure 3. 7. Actual evapotranspiration when $SMD < RAW$. All values in mm/day.....	53
Figure 3. 8. Actual evapotranspiration when $SMD > RAW / REW$. All values in mm/day....	54
Figure 3. 9. Actual evapotranspiration when $SMD > TEW$. All values in mm/day.....	55
Figure 3. 10. Example of daily water balance including the near surface storage.....	56
Figure 3. 11. Soil water balance and the resulting potential recharge.....	57
Figure 4. 1. Yobe river system. Northeast semi-arid zone of Nigeria.....	63
Figure 4. 2. Daily rainfall and estimated runoff from the SAMBA model for the representative year of 1964 at Nguru, Nigeria.....	64
Figure 4. 3. Estimated water available near surface from the SAMBA model . 1964.....	65
Figure 4. 4. Estimated potential and actual evapotranspiration from the SAMBA model. 1964, Nguru, Nigeria.....	65
Figure 4. 5. Estimated soil moisture deficit and available water for evapotranspiration from the SAMBA model. 1964, Nguru.....	66
Figure 4. 6. Water balance components and the resulting potential recharge from the SAMBA model. 1964, Nguru, Nigeria.....	68
Figure 4. 7. Selecting arbitrary planting date for a crop during the rainy season.....	69
Figure 4. 8. Surface runoff determination for a climatic average year in Nguru using the SAMBA model.....	71
Figure 4. 9. Relationship between annual rainfall and estimated annual runoff for 36 years in Nguru.....	72
Figure 4. 10. Year with similar annual rainfall but different estimated annual runoff than 1990.....	73
Figure 4. 11. Soil water balance from SAMBA model for one particular day in Nguru.....	75

Figure 4. 12. The near surface storage and soil evaporation for a sequence of ten days in Nguru (SMD is larger than TEW).....	76
Figure 4. 13. The soil water balance components during 15 days when $SMD > RAW$	77
Figure 4. 14. Recharge calculated with and without the runoff component.....	79
Figure 4. 15. Effect of surface runoff on potential recharge estimation. Nguru, 1995.....	80
Figure 4. 16. Comparing the effect of the inclusion of NSS on annual potential recharge..	81
Figure 4. 17. Comparing the effect of the inclusion of NSS on annual actual evapotranspiration AE.....	81
Figure 4. 18. The effect of NSS on recharge and actual evapotranspiration for a particular period of 1974.....	82
Figure 4. 19. Example of how the conditions from the previous rainy season affects the AE and recharge estimation in the following growing season (1967).....	83
Figure 5. 1. Representative soil moisture profile under a millet crop.....	89
Figure 5. 2. Predicted values from the SAMBA model using the information from the field trial at Maduguri. Crop season of 1992.....	91
Figure 5. 3. Observed and predicted water contents in a sandy soil under a millet crop field experiment in semi-arid Nigeria.....	93
Figure 5. 4. Predicted soil water content from the SAMBA model using a factor of near surface storage 0.75.....	94
Figure 5. 5. Cumulative drainage below the root zone (1.3 m) in a millet crop plot estimated by the model and by the soil water content measurements.....	96
Figure 5. 6. Relationship between two alternative outputs from the SAMBA model and field observations from Mortimore (1989).....	99
Figure 5. 7. Modelled crop season in Nguru for two years.....	100
Figure 5. 8. Rainfall and modelled potential recharge for Nguru.....	103
Figure 6. 1. Reference evapotranspiration estimated for the 1992 season in Maiduguri....	108
Figure 6. 2. Soil water balance for a millet crop and the resulting soil moisture deficit curves from the SAMBA model using daily estimated ETo and monthly mean Eto.....	108
Figure 6. 3. Water balance for three net surface runoff scenarios.....	112
Figure 6. 4. Variations of the mean recharge for the period 1962-97 from a mean reference value (14 mm/year) as function of the single variations of selected parameters....	115
Figure 6. 5. Water balance partitioning resulting from the variation of the crop coefficient K_c mid.....	116
Figure 6. 6. Water balance partitioning resulting from the variation of the maximum root depth.....	117
Figure 6. 7. Water balance partitioning resulting from the variation of TAW.....	118
Figure 6. 8. Variation of the mean recharge from a mean reference value for the wet period of 1974-79.....	120
Figure 6. 9. Spatial distribution of the mean annual rainfall for 1992-1995.....	122
Figure 6. 10. Mean annual potential recharge (1992-95) estimated for a reference surface using the daily rainfall data from 18 rainfall stations.....	122
Figure 6. 11. Correlation between the total rainy days with rainfall bigger than 20 mm and the mean annual recharge for 1992-95 for the study area.....	123
Figure 6. 12. Correlation between the annual rainfall and annual recharge for 1994.....	124
Figure 6. 13.. Correlation between rainfall during the month of August 1994 and estimated annual recharge for the study area.....	125
Figure 6. 14. Spatial distribution of the total rainfall for the month of August 1994.....	126
Figure 6. 15. Monthly potential recharge for August 1994 estimated for a reference surface using the rainfall daily data from 18 rain gauge stations.....	126

List of tables

<i>Table 2. 1 Typical steady infiltration rates.....</i>	<i>11</i>
<i>Table 3. 1. Example of coefficients for runoff determination for a sandy soil.....</i>	<i>39</i>
<i>Table 3. 2. Near surface storage for actual evapotranspiration estimation</i>	<i>56</i>
<i>Table 4. 1. Definitions for onset of rains in the region of semi-arid West Africa.....</i>	<i>70</i>
<i>Table 5. 1. Summary of crop yields and rainfall pattern in Dagaceri and Nguru</i>	<i>98</i>
<i>Table 6. 1. Water partitioning for the three surface conditions.....</i>	<i>111</i>
<i>Table 7. 1. Input parameters of the soil water balance model.....</i>	<i>130</i>

CHAPTER ONE

INTRODUCTION

1.1. GROUNDWATER ESTIMATION IN SEMI-ARID REGIONS

Accurate quantification of groundwater recharge rates is a crucial prerequisite for efficient groundwater resource management, particularly in semi-arid areas where there are large demands for groundwater supplies. The correct estimation of recharge is a key factor to determine sustainable yields of regional aquifers and to avoid undesirable effects, such as decline in aquifer water tables and water quality deterioration. Recharge originating from the direct percolation of rainfall over large areas can provide a substantial volume of water for domestic and agricultural uses, as illustrated by several studies cited in this thesis.

Groundwater recharge rates are difficult to derive with confidence (Simmers, 1998). One of the principal difficulties is the association between a reliable estimating method and the availability of information from the physical system. The estimation of recharge requires the knowledge and understanding of the complex soil-plant-water relationships. Therefore, an ideal recharge estimation technique should fully reflect the individual physical processes present in the soil system.

However, information about soil, vegetation and climatic properties is frequently limited in semi-arid areas. Many of the dry regions in the world are situated in developing countries where environmental record keeping is poor and the capacity to collect information is fraught with financial problems. Therefore, under this reality the use of techniques which require a large number of parameters, such as complex models, or the use of methods which demand an extensive field measurement campaign may not be justified. The quantification of groundwater recharge has to be based on the appropriate balance between the knowledge of the physical processes and the data that can be actually gathered.

In summary, there is a need for a groundwater recharge estimation technique matching data availability with the adequate representation of the physical processes which affect recharge. This is particularly relevant since much understanding of spatial and temporal relationships can be revealed through the conceptual and numerical development of the technique.

1.2. ESTIMATING RECHARGE BY A SOIL WATER BALANCE MODEL

This thesis aims to demonstrate that the use of techniques based on the soil water balance of the hydraulic components is a plausible method for estimating recharge in semi-arid areas.

Soil water balance techniques account for all water entering and leaving the system based on the quantification of the individual physical processes (the inputs and outputs), without representing all the hydraulic processes and their interactions which describe the movement of water within the soil. Consequently, this approach is based on fewer physical processes and it is not subject to the uncertainties of the mechanisms of a full soil physics analysis.

Soil water balance methods are usually criticised because they are too simplistic or crude for estimating recharge. However, the work presented in this thesis shows that this approach can be functional if the key physical hydraulic processes, and their interactions, are adequately represented. Many conventional soil water balance models are designed for different purpose than recharge estimation. Usually they represent different field conditions than the actual conditions observed in semi-arid areas. Therefore, some key processes that affect recharge determination are often neglected.

This study identifies a series of processes which need to be included in the soil water balance, based on insights and observations from real situations present in those regions. They are (i) the determination of soil evaporation during the period preceding the crop season; (ii) the determination of actual evapotranspiration following a heavy rainfall event and; (iii) the determination of surface runoff losses.

1.3. CREDIBILITY OF THE ESTIMATION METHOD WITH LIMITED DATA

The issue of data limitation, as mentioned in the Section 1.1, affects not only the modelling processes but also the judgement of the model validity. When data are limited, the conventional parameterisation and calibration approaches lose their relevance. Conventional calibration is usually established through a series of statistical matches which require extensive data from field measurements. Moreover, since recharge cannot be directly measured, validation has to be based on indirect assumptions.

However, it is still necessary to make judgments about the credibility of the model structure and the reliability of the model results. For this purpose, a new approach for testing the model reliability has been devised based on the concept of “plausibility”, as reported at an international conference (Carter et al., 2002). The term plausibility is here defined as the reasonable or probable outcome from the model, including the judgment about the structure, as well as about the model results.

This concept is based on a more lateral approach where selected model outputs, representing not only recharge but other components, are analysed using as wide a range of quantitative and qualitative observations as possible.

1.4. SEMI-ARID CONDITIONS

The conditions prevailing in a typical semi-arid climate are incorporated in this thesis through the utilisation of long-term climatic data and field observation from a representative region in West Africa (Northeast Nigeria), and from data in the literature.

Recharge in semi-arid areas is the result of an irregular and sporadic rainfall distribution concentrated in one season, occurring at the same time as the vegetation growth. These extreme characteristics affect all the physical hydrological processes and prove to be a challenge for any methodology utilised for estimating recharge.

The focus of this study is on areas with agricultural activities where the interaction among the components of the soil water balance is accentuated. The reasons for this assumption are based on the fact that in those areas, groundwater is a potential source of water for crop development and a correct assessment of the rates of water replenishment to the aquifer system is necessary. Second, in those areas the land use changes in a short period of time allowing the investigation of the effect of different land use patterns and agricultural practices on recharge.

Therefore, the conditions assumed as representative for a semi-arid area are related to a non-irrigated cropped plot planted at the beginning of the rainy season. The soil characteristics are based on typical sandy soils as found in the literature.

1.5. AIM AND OBJECTIVES

The aim of the study is to develop a procedure for estimating direct recharge in semi-arid areas based on the soil water balance technique. The objectives are:

1. to develop a model which is physically credible and adequately reflects the principal physical processes which affect recharge,
2. to develop a model which makes use of readily available monitored data instead of specialised information which requires expert and expensive measurements,
3. to develop a model which makes use of a small number of key parameters to describe the principal physical processes,
4. to apply the model to real conditions and present the results in diagrammatic form to help understand the interactions of the physical processes,
5. to demonstrate the credibility and plausibility of the model results when applied to typical semi-arid conditions, and
6. to investigate the variability and sensitivity of recharge to a series of selected factors.

1.6. THESIS STRUCTURE

The thesis contains seven chapters, with chapter one introducing the subject, aims and objectives of the research.

The overview of the principal soil water physical components involved in the soil water balance of a semi-arid area is presented in chapter two. It also reviews some usual methods and soil water balance models for determining groundwater recharge and their applicability regarding the guidelines adopted in this research.

Chapter three describes the fundamentals of the methodology developed for estimating recharge based on the soil water balance technique. The conceptual and the computational modelling of the new components introduced in this study are detailed using a series of diagrams derived from the insights of the physical systems explored in chapter two.

The incorporated procedures of the soil water balance model described in chapter three are examined in chapter four through the analysis of selected outputs. These outputs are derived from real field data using representative periods.

Chapter five explores the credibility of the soil water model conceptualisation using the concept of plausibility analysis developed in this research. Selected model outputs are compared to a range of available quantitative and qualitative information from semi-arid areas in order to assess the robustness of the model.

Chapter six investigates some functional aspects of the soil water balance model regarding the sensitivity of the model outputs resulting from the variability of selected parameters.

Finally, chapter seven presents the summary of the work, the overall conclusions about the contribution achieved and introduces some recommendations for further research.

CHAPTER TWO

GROUNDWATER RECHARGE ESTIMATION

This chapter presents an overview of groundwater recharge estimation starting with definitions of groundwater recharge and the investigation of the principal physical processes which influence recharge. Approaches utilised for estimating recharge are presented, followed by an argument for the use of the soil water balance technique as a suitable method for semi-arid regions. The chapter concludes with a brief discussion of alternative soil water balance models.

2.1. GROUNDWATER RECHARGE

Recharge can be defined in different ways according to its sources and according to the hydraulic processes involved. In general, groundwater recharge may be defined as the downward flow of water reaching a groundwater system, forming an addition to the groundwater reservoir (Lerner et al, 1990).

Recharge may have various sources such as rainfall, percolation from streams, canals and lake beds, return flow from irrigation and artificial injection of water. These sources are not exclusive and many combinations of different sources can occur in a specific location.

The principal recharge mechanisms have been defined (Lloyd, 1986; Lerner et al., 1990) as:

- *Direct recharge*, defined as the recharge derived from rainfall over large areas that enters the soil and, in excess of soil moisture deficits and evapotranspiration, moves downward by direct percolation through the unsaturated zone;

- *Indirect recharge*, resulting from the water infiltrating through the beds of surface water courses or lakes; and
- *Localised recharge* resulting from the near-surface concentration of water in the absence of well-defined channels.

The focus of this thesis is on the determination of direct recharge. However, in particular situations indirect and localised recharge can be important sources of water, mainly in areas where an interaction between a permanent river and the groundwater system exist.

Figure 2.1 presents diagrammatically a simplified concept of groundwater recharge showing the flow of water from surface to the saturated zone when the water table is deep.

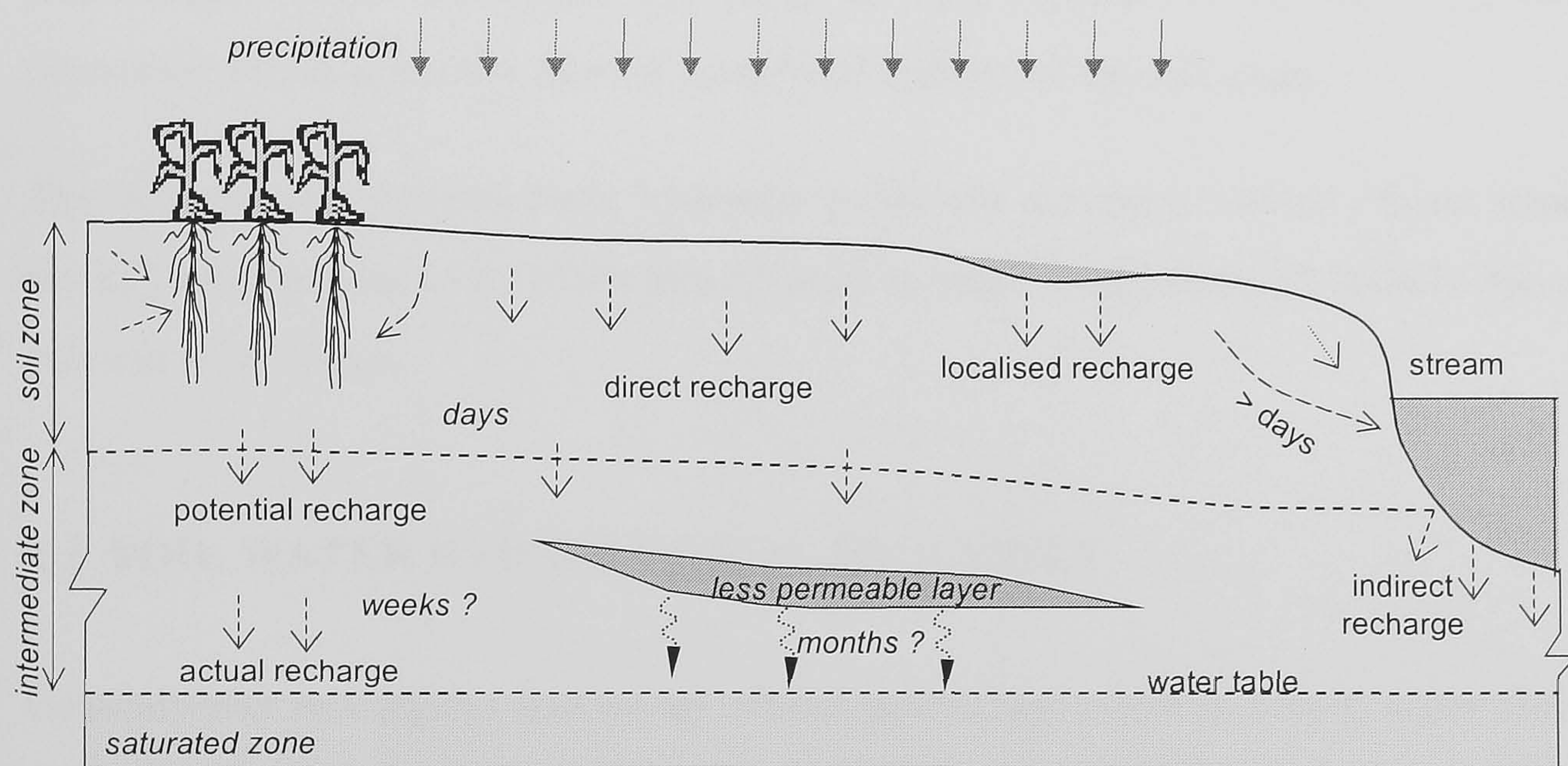


Figure 2. 1. Groundwater recharge definitions.

The unsaturated upper part of the soil is sub-divided into two main zones: the soil zone and the intermediate zone. After a significant rainfall period, water infiltrates into the soil, moves through the soil zone and is potentially available for recharge. The water that leaves the bottom of the soil zone is defined as *potential recharge* (Rushton, 1988). This water will move vertically to the saturated zone provided that the underlying rock formation is permeable. The amount of water that indeed reaches the water table is defined as *actual recharge*.

The period taken by water to reach the saturated zone may be highly variable. Usually, the water front reaches the bottom of the soil zone in a matter of days after a wet period. However, from potential to actual recharge many factors may delay the water movement for weeks or even months. For instance, delay due to the presence of less permeable layers may affect the time taken by water to reach the saturated zone.

Moreover, the recharge/leakage mechanisms are complex and in particular situations not all potentially available water reaches the aquifer. Various processes such as routing to the river system or removal by deep roots may divert a proportion of the surface infiltrated water.

This study assumes that the bulk of the potential recharge will in the end contribute to actual recharge, regardless of the time taken by water to percolate through the intermediate zone. Therefore, the focus of this research is on the main physical processes which affect the flow of water into and out of the soil zone.

The next section outlines these hydraulic processes aiming to identify those which are the key mechanisms to be taken into account in order to represent physically the overall process of recharge.

2.2. SOIL WATER HYDROLOGICAL PROCESSES

Groundwater recharge is part of an overall hydrologic cycle in which water flows are multidirectional and the relationships between the components are complex. Although they occur simultaneously in the field, they can be easily visualised as a sequence of processes that affects the flow of water through the soil zone.

Figure 2.2 shows the principal hydrological processes acting on the soil surface during a particular period of time.

Water distributed on a soil surface area by *rainfall*, is partially *intercepted* by vegetation, if any. Surplus rainfall reaches the ground and begins to *infiltrate* into the soil. Some water may not be absorbed and accumulate in surface depressions from where it will evaporate or infiltrate after the rain ceases.

However, if the rainfall continues at greater than infiltration rates, part of this water stored on the surface may start to flow overland and leave the area as surface *runoff*. From the water infiltrated into the soil, some *evaporates* directly from the soil surface, some is taken up by plants for growth or *transpiration*, some *drains* downward beyond the soil zone (as potential recharge) and, finally, the remainder accumulates within the soil zone and adds to the soil moisture storage.

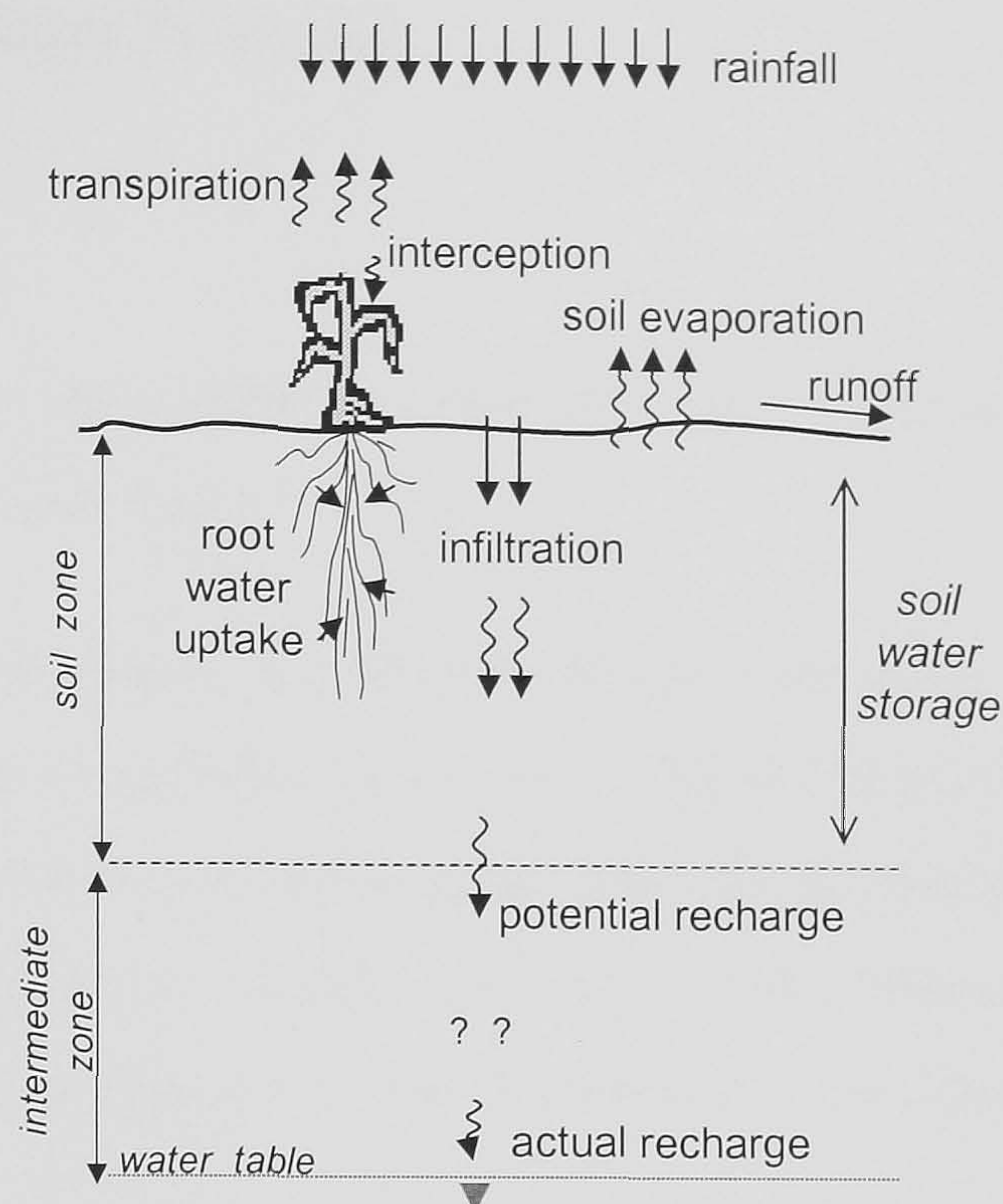


Figure 2. 2. Soil water processes (based on Hillel, 1998).

The following sections investigate the role of each process in the vertical, one-dimensional, water flux into the soil zone and their principal characteristics in a semi-arid climate area.

2.2.1. Rainfall

Rainfall is the primary source of water to the soil water cycle. Rainfall distribution and rainfall intensity are key factors affecting all the other hydrologic processes such as infiltration, runoff and soil water storage.

In semi-arid regions, rainfall is characterised by a high temporal variability with a significant year-to-year variation. The annual distribution is characterised by a concentrated period of few months when rainfall occurs in the form of sporadic high intensity short duration events. The effect of these features on recharge is significant, as is discussed during the development of this study.

The use of daily rainfall for recharge estimation is strongly recommended by several authors (e.g. Howard and Lloyd, 1979; Gee and Hillel, 1988). Averaged values, such as

monthly means or decadal periods, underestimate the effect of occasional highly intense rainfall events on recharge, leading to misleading conclusions.

2.2.2. Interception

A proportion of the falling rainfall may be intercepted by vegetation and evaporated directly to the atmosphere by the process of interception.

Kirby et al. (1991) point out that interception losses may be the largest component of the total evaporation in areas where the canopy vegetation is wet for a significant part of the year such as in regions with temperate climate. In these areas, most of the rainfall occurs at regular low intensity and vegetation canopy is dense. For instance, interception losses can range from 15-40% of annual rainfall in temperate coniferous forests (Rosenzweig, 1998).

However, for short vegetation with low leaf area indices, the inclusion of the interception process is questioned based on the argument that the effect of interception on the overall water balance is counteracted by the fact that interception reduces the amount of soil water that is lost via plant transpiration and soil evaporation and, therefore, the net effect of interception is negligible (Walker and Langridge, 1996).

In hydrological studies in semi-arid areas the inclusion of interception is usually neglected due to sparse vegetation cover and the highly intense storms (Kruseman, 1997, Walker et al, 2002). Therefore, in this study the process of interception has not been considered.

2.2.3. Surface Runoff

The proportion of water which becomes runoff is directly related to the infiltration capacity of the soil. The rate of infiltration depends among others on soil texture, initial wetness, surface cover, agricultural practices and structural factors such as the presence of cracks or joints. Table 2.1 shows typical infiltration rates for different soil textures.

Table 2. 1 Typical steady infiltration rates (from Hillel, 1998)

Soil type	Steady infiltration rate (mm/h)
Sands	> 20
Sandy and silty soils	10 – 20
Loams	5 – 10
Clayey soils	1 – 5
Sodic clay	< 1

Surface runoff is also governed by rainfall characteristics (Rockström and Valentin, 1997). Consequently, the conditions for the occurrence of runoff in semi-arid areas are extremely favourable despite the common presence of sandy soils (with higher infiltration rates). This is due to the combination of intense rainfall events and soils with relatively weak structure and tendency to form surface crusts (Lal, 1991). For instance, Owonubi et al. (1991) cite studies in the Sahelian region of Africa where, apart from early in the season, all rainstorms of more than 20 mm in 24 h resulted in runoff.

The term surface runoff needs to be defined from the point of view of the spatial scale of the study.

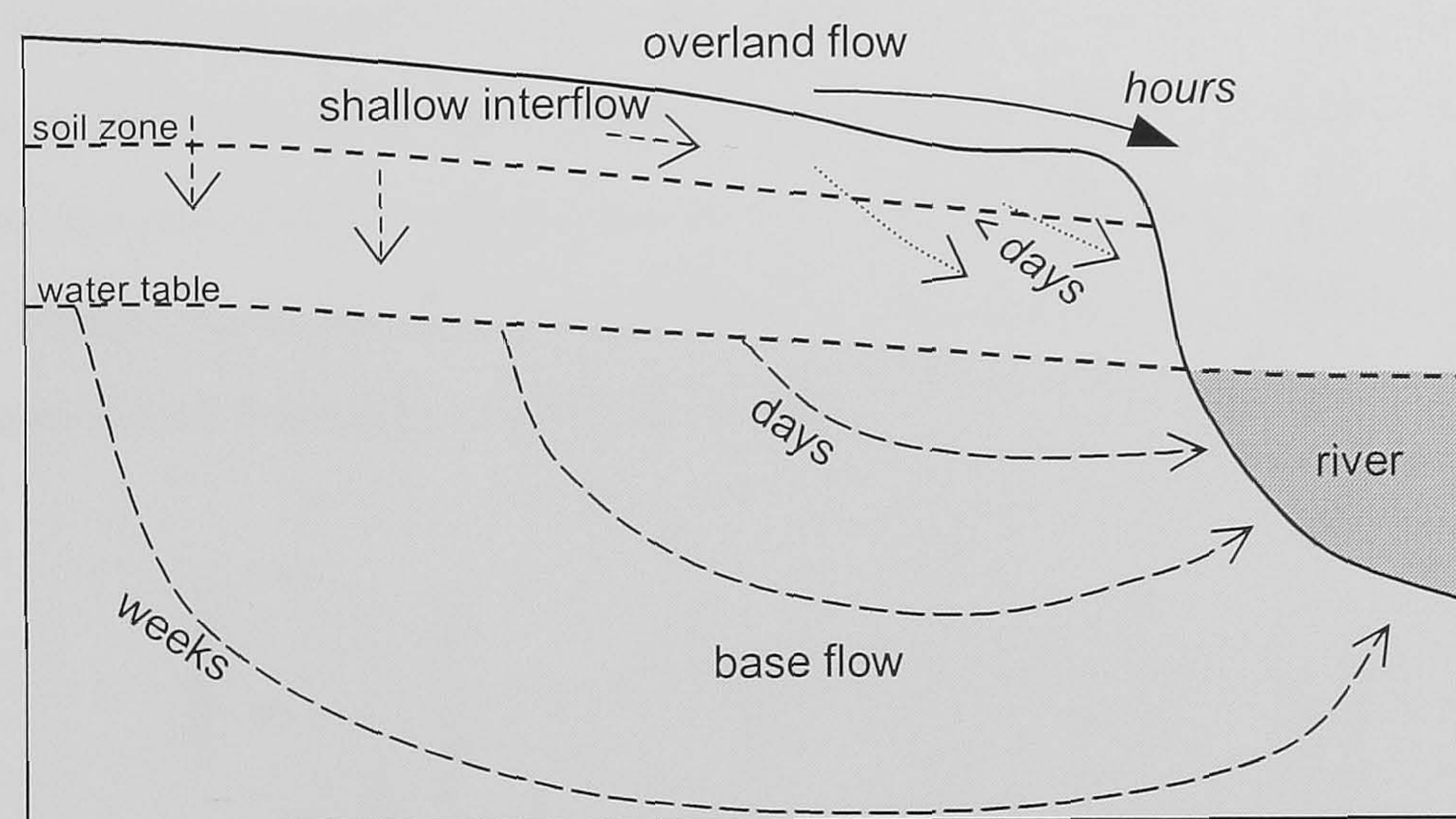


Figure 2. 3. Definitions of surface runoff including typical scale of occurrence.

From the point of view of small plot studies, the term *overland flow* may be more appropriate, including the water captured in surface pathways plus the *shallow interflow* during a short time period (figure 2.3 above). On the other hand, from the point of view of catchment studies, surface runoff is the *river discharge*, that over a long term is the total of *overland flow*, *interflow* and *baseflow* (Carter, pers. com.).

The overall water balance usually considers a small area or plot. In this case, surface runoff is in fact the net loss from the difference between the outflow of water and the incoming overland and interflow. In particular situations, depending on localised factors such as slope, the resulting net surface runoff may be negative or zero such as observed by Rockström et al. (1998) and Gaze et al. (1997) in semi-arid Niger.

2.2.4. Soil evaporation

The soil drying process can be described in three distinct and consecutive stages (Hillel, 1980): (a) in the first stage the soil surface is wet and evaporation occurs at potential rates, that is, the only limiting process is the atmospheric demand; (b) when the soil becomes dryer, water cannot be supplied to the surface fast enough to meet evaporative demand. In this case, the rate of evaporation decreases progressively as the depth of the dry layer increases; finally (c) the rate of evaporation becomes small and can persist for a long period. Usually, this last stage is not considered due to the small rates compared to the potential demands.

The process of evaporation from a bare soil can be visualised in figure 2.4, based on a field experiment in southwest Niger (Wallace et al., 1993; Wallace and Holwill, 1997). The authors estimated evaporation from a bare sandy loam soil using a Bowen ratio energy budget method. This method calculates evaporation from the measurements of temperature and humidity at different levels.

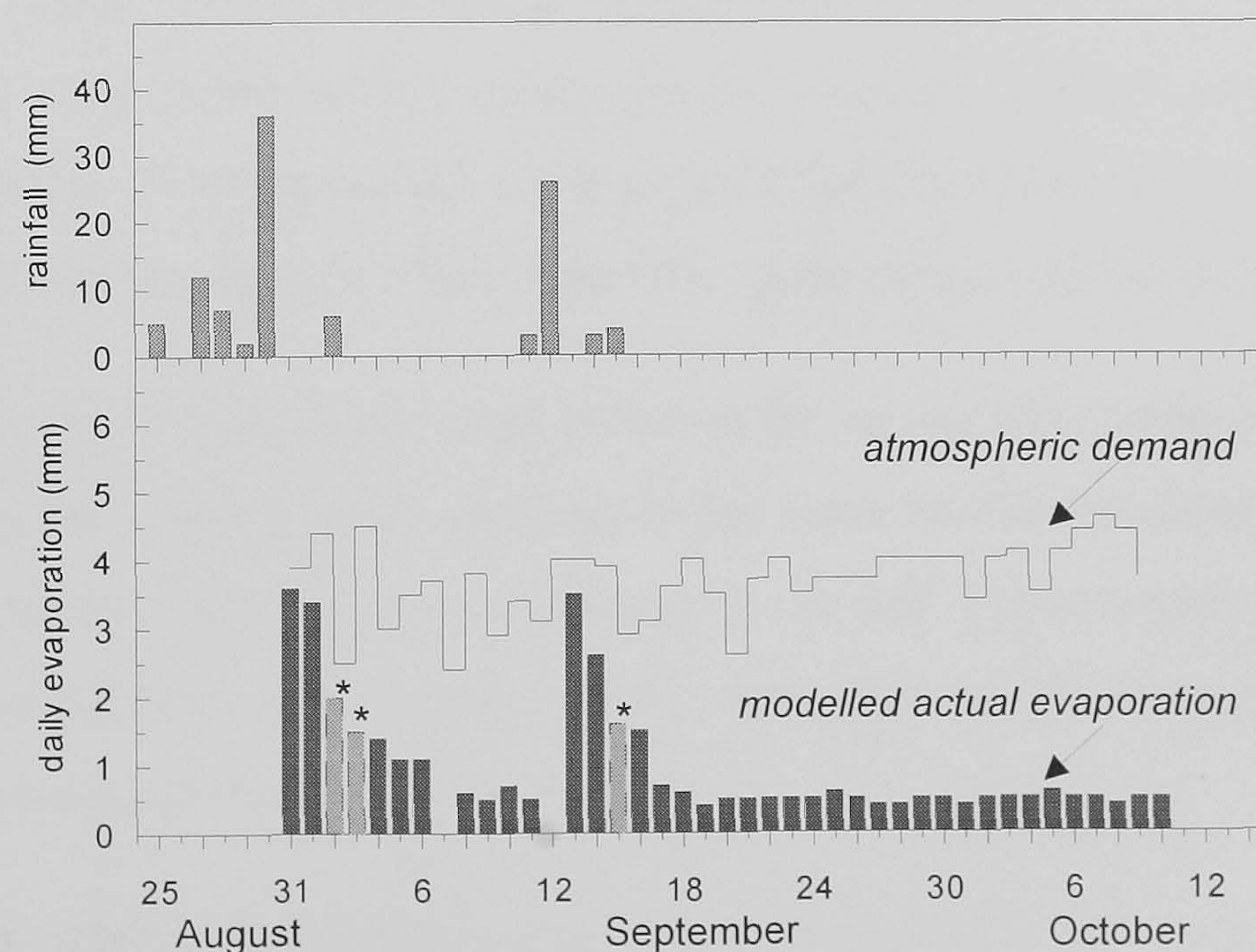


Figure 2. 4. Daily soil evaporation from a bare soil surface (from Wallace and Holwill, 1997).
*interpolated value.

Note that the term modelled actual evaporation represents the actual evaporation computed by a particular estimation method, e.g. the Bowen ratio approach or the linear model of Doorenbos and Pruitt (1977) as shown in figure 2.12. Throughout this thesis the modelled actual evapo(transpi)ration is simply named actual evapo(transpi)ration and the evaporative atmospheric demand named as potential evapo(transpi)ration.

The direct evaporation from the soil surface is an important component of the overall soil water balance of a cropped semi-arid area. Soil evaporation is the dominant evaporative process during the fallow period and during the earlier crop season when the soil is practically bare (Allen, 1990). Moreover, evaporation from the soil beneath a crop during the mature stages of the crop can also lead to a considerable loss of water in semi-arid areas, because crops are sparse and may not provide complete ground cover (Wallace, 1991).

2.2.5. Transpiration

In a cropped field, evaporation from the soil surface is accompanied by transpiration from the crop. Transpiration is the process of loss of soil water from vegetation due to the extraction of water by its root system.

Transpiration is caused by the evaporative demand of the atmosphere through the processes of vapour exchange between the leaves and air. As with soil evaporation, it depends on a large number of climatic factors such as radiation, air temperature, air humidity and wind speed, which determine the pressure gradient between the plant tissues and the air. In addition, soil water content and the ability of the roots to extract water from soil are key factors which affect the transpiration rate (Allen et al., 1998).

The water movement from inside plant leaves to the air outside can be controlled by the plants opening and closing small apertures in the leaf surface (stomata) in response to atmospheric demand and the amount of water in the soil. Consequently, different kinds of plants have different transpiration rates, depending on their number of stomata, among other characteristics.

The relation between actual transpiration rates (represented by the ratio actual/potential transpiration) and soil water content is illustrated by figure 2.5.

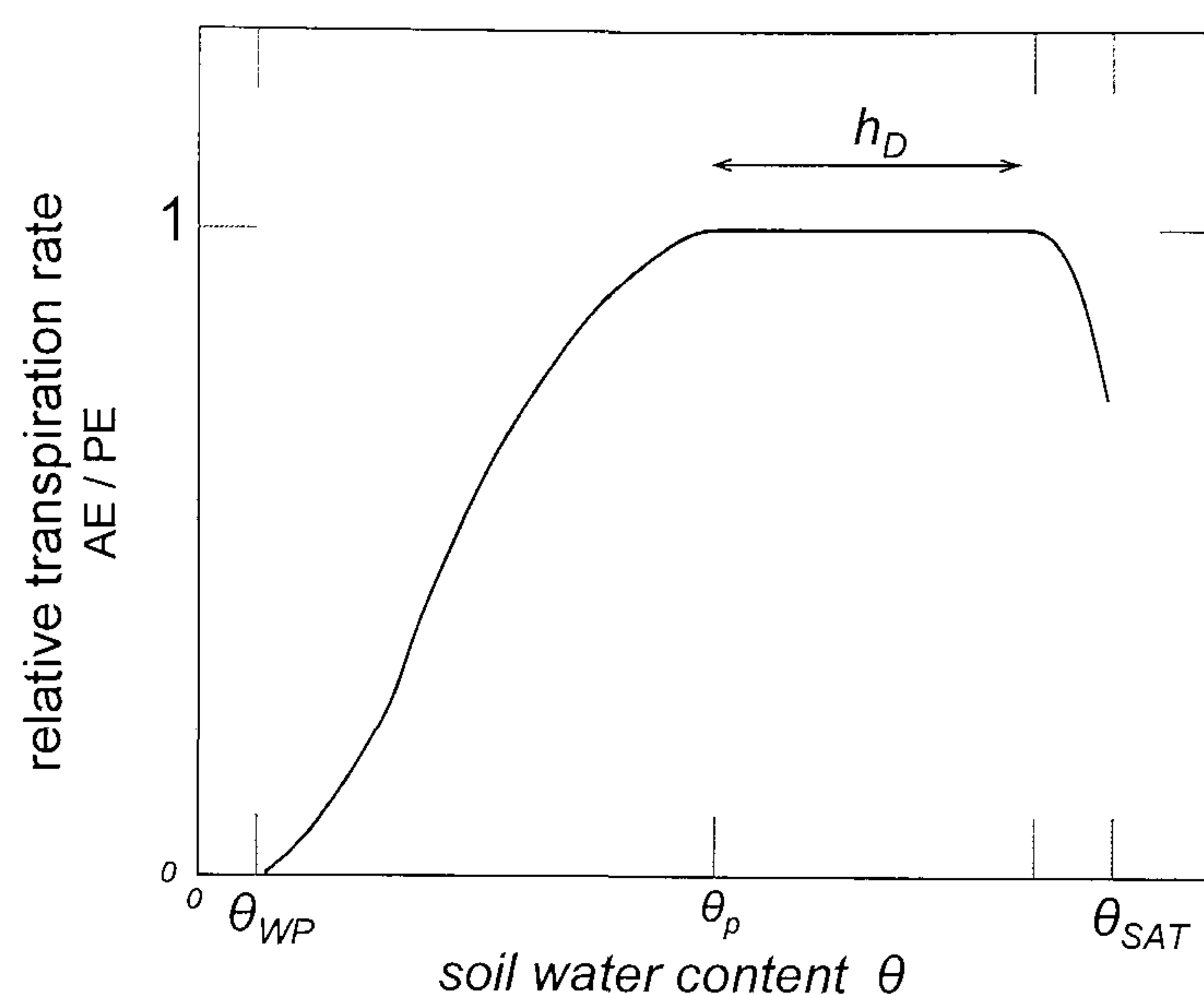


Figure 2. 5. Rate of plant transpiration as a function of soil water content (modified from Kutilek and Nielsen, 1994).

The plants transpire at potential rates when there is no limitation of water supply and the stomata are fully open. When the soil water content decreases, the evaporative flux is maintained by the gradient of the water potential until a critical value at which the stomata start to close and the transpiration rate is gradually reduced (θ_p). Transpiration is negligible when the soil water content is at *permanent wilting point* θ_{WP} .

Another interesting aspect shown in figure 2.5 is that near the point of saturation the rate of transpiration also decreases due to the effect of waterlogging on the roots.

The process of plant transpiration is directly associated with the process of root growth within the soil profile.

2.2.6. Root growth and root distribution

The process of root growth secures access to new sources of water in the soil profile, increasing the volume of soil water available for plants.

Root extension into soil is time and space dependent. During the crop development stages, roots grow rapidly in order to meet the plant water demands. Then, the rate of growth decreases and becomes negligible at maturity.

Root development is severely inhibited by a variety of factors such as soil moisture content, high bulk density, high water table, low fertility, low pH, soil compaction, shallow bedrock and horizontally stratified layers of shale or clay (Borg and Grimes, 1986; Canadell et al., 1996). Consequently, maximum rooting depth and distribution vary considerably.

Canadell et al. (1996) summarised the maximum rooting depth of 253 species worldwide including natural vegetation and crops. The maximum depth of the samples varied from 0.3 m to 68 m. However, approximately 80% of the species were within of 2 m depth. For crops the average was 2.1 ± 0.1 m.

The distribution of roots within the profile varies with depth. For example, Rockström et al. (1998) observed that at maturity 95% of the roots were concentrated in 0-1.4 m soil depth with the greatest density of roots up to 0.3 m in a pearl millet crop in Niger. In terms of water uptake, the uneven concentration of roots within the soil profile may suggest that roots utilise primarily the water localised near surface for transpiration. However, as cited by Li et al. (2001), roots are able to make progressively their way to available water at deeper soil layers when water content near the surface approaches the wilting point.

2.3. MOVEMENT OF WATER IN SOIL DUE TO ALL HYDROLOGICAL PROCESSES

Water distribution in the soil zone is complex since all physical processes described are strongly interlinked and simultaneous.

Figure 2.6 shows the modelled soil water content curves in a sand soil profile, after a single rainstorm of 72.2 mm lasting 6 hours. The soil is not vegetated; therefore, direct evaporation from soil surface and drainage are the only present hydraulic processes. Note that the soil water content before the storm was at wilting point.

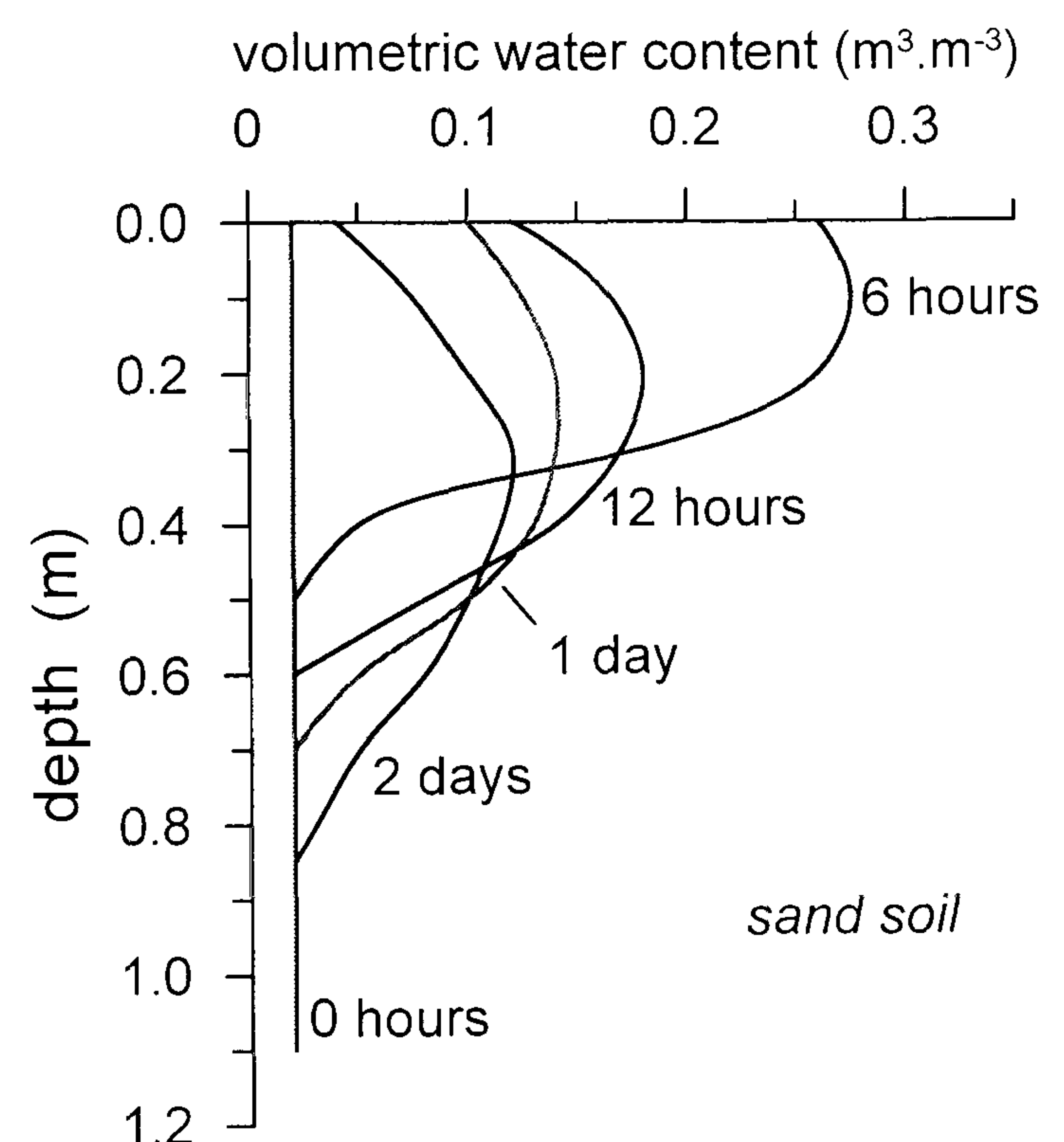


Figure 2. 6. Redistribution of soil water following a single rain storm of 72.2 mm lasting 6 hours. The periods indicate time after the onset of the storm (from Hillel, 1977).

During the storm, the water content in the upper layers of the soil profile increased rapidly. Approximately one day after the end of the rain event, a proportion of the infiltrated water had moved downward and the soil water content had increased at deeper layers. At near surface the water content gradually decreased due to the combined effect of evaporation and vertical water movement. However, a significant proportion of the infiltrated water remained in the upper parts of the soil profile.

Initially, the movement of water at layers near surface was rapid due to the high hydraulic conductivity. As the layers became drier the movement gradually slowed down.

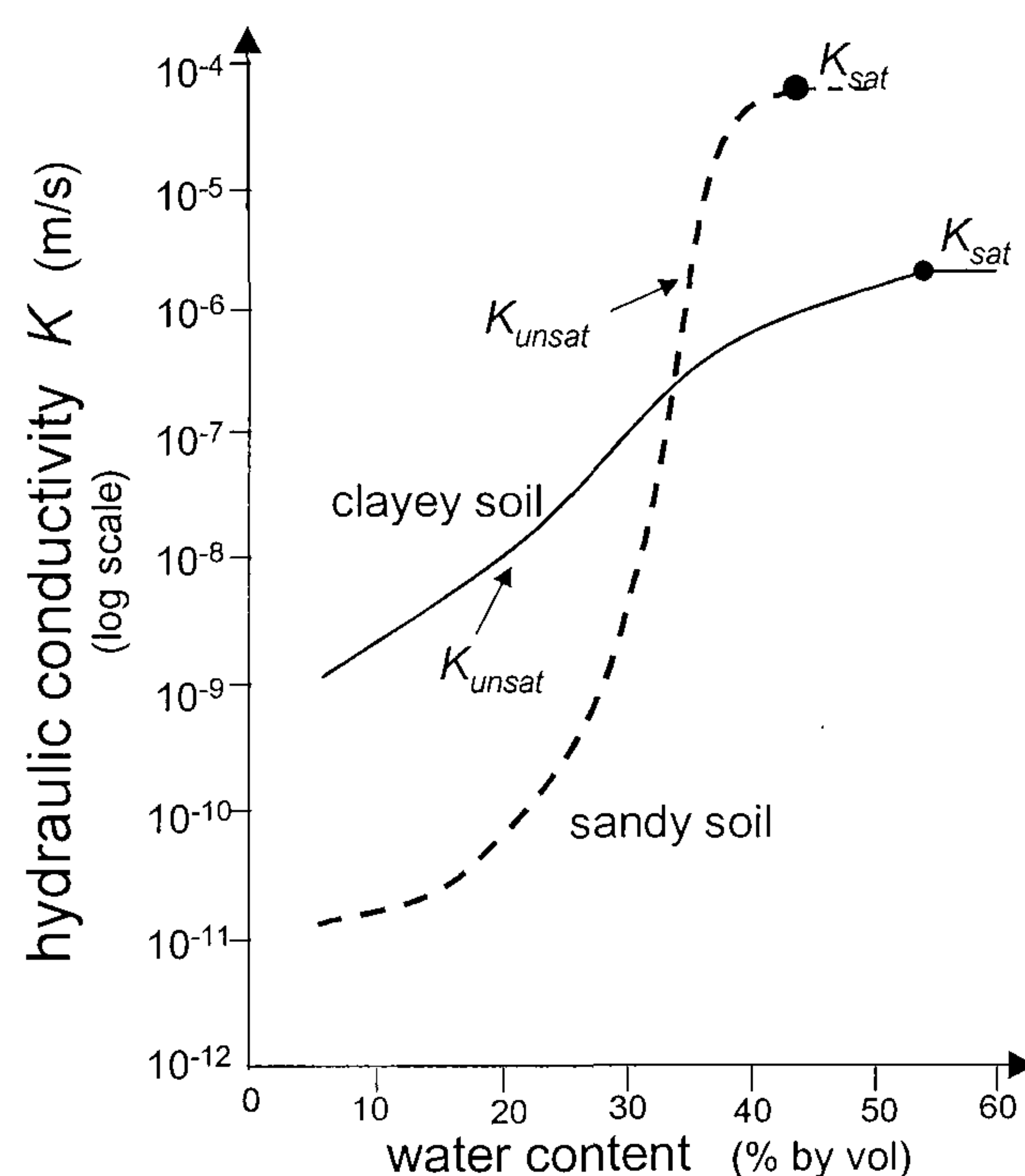


Figure 2. 7 Hydraulic conductivity as a function of water content (modified from Hanks, 1992 and Hillel, 1971)

Figure 2.7 shows the relationship between soil water content and the unsaturated hydraulic conductivity K_{unsat} for two types of soils.

The hydraulic conductivity for sandy soils is relatively low when the soil is dry. However, it changes abruptly and becomes significantly higher as the soil becomes wetter. For clayey soils, the increase of K_{unsat} is more gradual.

The soil water content above which the hydraulic conductivity becomes sufficiently high to permit the rapid transmission of water to deeper layers is defined as the field capacity point. At this point, water is held at such low suction that drainage occurs rapidly. For fine textured soils, such as clayey soil, field capacity is not as distinct as in sandy soils. Therefore, at low water contents, clayey soils can have higher K_{unsat} than sand and consequently water movement may persist for longer.

Field capacity can be defined in terms of a soil-water holding property as it represents the maximum water content that the soil can hold following free drainage under gravity (Landon, 1991). This definition is useful in terms of the definition of the

amount of water which the soil can hold against the gravitational forces and make available for plants for a longer period. Therefore, during a period with no infiltration, water held by the soil particles is depleted by the demand of the vegetation.

The flow of water into a soil profile during a longer period such as a crop season involves all the hydraulic processes concurrently as illustrated in figure 2.8.

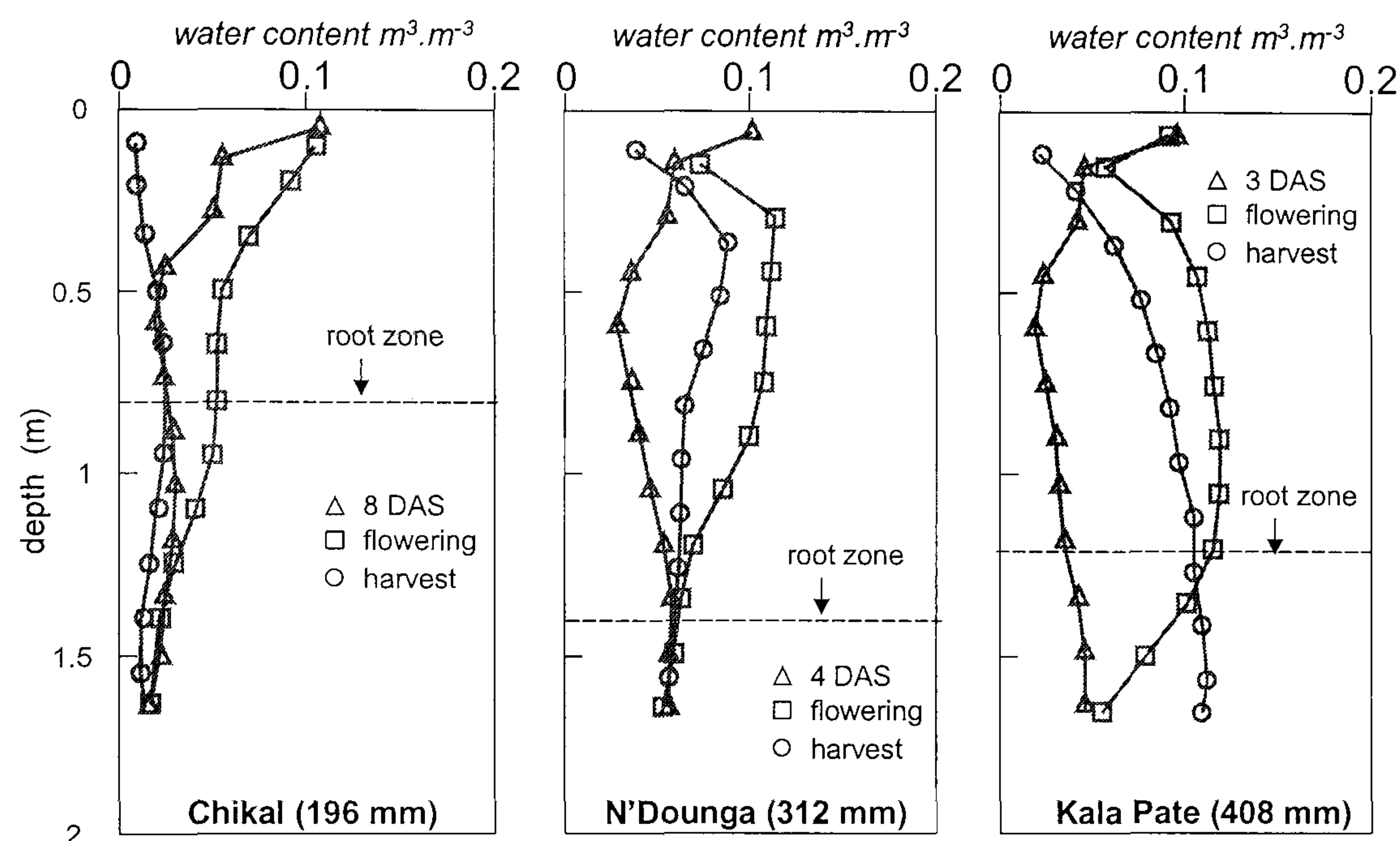


Figure 2. 8. Profile water distribution at three sites, Chikal, N'Dounga and Kala Paté (Niger) with the respective total rainfall. DAS is the days after sowing (from Payne et al., 1991).

Figure 2.8 shows an experiment carried out by Payne et al. (1991) with the purpose of quantifying the root zone water balance of a cropped millet plot, non-irrigated, planted in deep sandy soil at three different sites in semi-arid Niger. As in typical rainfed crops in semi-arid areas, the planting of the crop coincides with the onset of the rainy season.

The soil water content measured a few days after sowing (DAS) shows the water content increasing in the near surface layers. Thus, although water content in deeper soil layers remains low, water is available for young plants in the topsoil. At the time of flowering, water is distributed to deeper layers and the roots reach their maximum depth. Interesting to observe is that at Chikal (fig. 2.8.a), the water shortage resulting from the low rainfall (196 mm) is a limiting constraint to root development.

From flowering to harvest, the water content decreases at all sites due to root water uptake. However, at N'Dounga and Kala Pate water is still available for plants at

harvest. In Kala Pate, downward water drainage is significant during the period between flowering and harvest, as result of the water excess during the period.

This example shows that any attempt to explain the water flux processes into a soil profile have to take into account the principal hydrologic processes which affect the overall soil water balance, hence recharge.

2.4. DISCUSSION

Figure 2.8 illustrates the soil water distribution during a crop season in semi-arid climates showing the result of the physical processes on the soil water content within typical profiles.

From the review of the principal mechanisms present in the water flow through the soil system (figure 2.2), we can conclude that in the field conditions previously defined as the subject of this research (Chapter one), the following mechanisms have to be represented in the modelling and estimation of recharge in semi-arid areas. They are:

- *Rainfall*. Rainfall is the main input of water to the system. Due to its localised and irregular distribution in time as well as space, daily time steps need to be utilised. Summing or averaging over longer periods hides the effect of extreme precipitation events (those most responsible for recharge events).
- *Surface runoff*. The conditions for the occurrence of surface runoff in semi-arid areas are extremely favourable. Generally, runoff is observed in most field experiments in semi-arid regions. For example, Lal (1991) reviews some previous studies in West Africa where runoff was observed in a range of field conditions. Rockström et al. (1998), Rockström and Valentin (1997), Peugeot et al. (1997), Grema and Hess (1994) are examples of field investigation in semi-arid areas where runoff has been observed (Chapter five presents some results from these studies).

Therefore, it is expected that a proportion of water may be lost through runoff and, consequently, the neglect of this process might result in the overestimation of recharge.

- *Soil evaporation.* In a semi-arid area, the soil surface is bare for a great part of the year during the dry season and at the beginning of the rainy season. Therefore, when the soil surface is not covered by vegetation, the direct upward evaporation from the surface is the principal process. Moreover, soil evaporation is directly related to crust formation which increases surface ponding of rain and enhances surface runoff.
- *Crop transpiration.* During the rainy season, crops such as millet, groundnuts and cowpeas are planted in the semi-arid regions. Therefore, crop transpiration is coincident to rainfall and concurrent to the soil water storage and recharge. The representation of crop transpiration in semi-arid climates has to take the conditions of soil water stress into account.
- *Root growth.* The process of root growth affects the pattern of water extraction and water movement in the soil profile. Roots have the ability to continuously reach the moist regions of the soil in an attempt to avoid water stress due to the deficit of water within the profile. Therefore, the process of root growth expands the volume of water available to the plant system. This ability of roots to access water wherever it is in the soil profile needs to be reflected in any representation of the soil-plant-water system.
- *Soil water distribution following a rain event.* Figures 2.6 and 2.8 illustrate the soil water movement and redistribution of water within the soil profile following a rainfall event and a wet period. The soil retains a proportion of infiltrated water which is therefore available for evapo(transpi)ration. This process is important because in semi-arid regions crops survive the frequent dry spells due to the water holding properties of the soil.
- *Potential recharge.* The excess of water after the processes of runoff, evapo(transpi)ration, and soil water distribution will drain downward through the bottom of the root zone. Therefore, deep drainage (or potential recharge) is included

in the water balance as the residual of the soil water accounting of the main processes involved.

2.5. PRINCIPAL METHODS FOR ESTIMATING RECHARGE

The physical processes detailed in the sections 2.2 and 2.3 are the basis of the principal approaches designed for estimating recharge. Section 2.5 outlines these approaches and discusses their main implications for this research.

This section does not set out to review in detail all methods commonly used for estimating recharge. There are many published reviews with valuable information about the applicability and problems associated with each method such as Allison (1988), Gee and Hillel (1988), Rushton (1988), Lerner et al. (1990), Allison et al. (1994), Hendrickx and Walker (1997) and Scanlon et al. (2002).

The methods here described can be broadly classified as simple empirical expressions relating rainfall to recharge, methods based on direct measurements of determined processes, and physical models which attempt to represent the water flow mechanisms.

2.5.1. Empirical rainfall-recharge expression

This method relates precipitation (P) empirically to recharge (R_{ech}) as

$$R_{ech} = k_1(P - k_2) \quad 2.1$$

where k_1 and k_2 are constants relating to a particular area or catchment.

This is the simplest method utilised to estimate recharge and it provides a rough estimation based on observed rainfall. The use of this approach is sometimes accepted for making ‘first-guess’ estimates of recharge (Allison, 1988). However, the empirical method estimates recharge from precipitation alone and it does not consider the principal hydraulic processes actually affecting recharge. Therefore, the lack of physical significance of the method can lead to completely misleading results.

2.5.2. Lysimeters

A lysimeter is a container in which a volume of soil, with or without vegetation, is isolated so that the flow through it can be collected at its bottom.

Usually lysimeters are utilised for the investigation of crop evapotranspiration rates. In this case, a lysimeter needs to be instrumented to measure the input of water and the changes of water content inside it. Kirby et al. (1991) give an example of a well instrumented lysimeter utilised to estimate transpiration and interception in a temperate forest in Wales.

Applications of lysimeters to estimate recharge are illustrated by the studies of Kitching and Bridge (1974), Kitching et al. (1977) and Kitching and Shearer (1982) in England. In semi-arid areas there are the lysimeters constructed by Kitching et al. (1980) in the semi-arid region of Cyprus. Also, the studies carried out by Gee et al. (1994) in three semi-arid regions of the USA using different soils and vegetation types in order to investigate the variability of the water balance and recharge.

Lysimeters provide the only direct method for estimating recharge fluxes and they are important devices to understand the recharge mechanisms in a specific site and usually utilised to validate soil water flux models. However, there are several problems associated with their use.

First, they are not routinely used because they are expensive with high cost of construction and maintenance. Second, problems associated with edge effects, water collection at the bottom and use of undisturbed soil for filling may affect the results. It may take several years for a lysimeter filled with disturbed soil to plausibly represent the surroundings conditions. For example, in the cited study of Gee et al. (1994) one of the lysimeters took two years to be filled and around five years to reach equilibrium. Third, a considerable number of lysimeters may be required to represent the high spatial variability of climate, soils and vegetation in semi-arid areas. Finally, the surface runoff is not taken into account by lysimeter studies since the lysimeter container prevents the overland flow. Therefore, in conditions when surface runoff is an important component of the soil water balance, recharge may be overestimated.

2.5.3. Water table fluctuation method

The basic assumption of the water table fluctuation method is that rises in the water table are the result of recharge arriving at the water-table. So, if the rise in a water table is Δh , the recharge R_{ech} during the period Δt is indirectly estimated as

$$R_{ech} = S_y \cdot \Delta h \pm \frac{Q}{A} \cdot \Delta t \quad 2.2$$

where S_y is the specific yield of aquifer, A is the area of the aquifer and Q represents the net flow from the aquifer caused, for instance, by abstraction.

This method provides valuable evidence of the occurrence of recharge, as pointed out by Barnes et al. (1994) since the fluctuation of the water table suggests the addition of water from the unsaturated zone above. Carter and Alkali (1996) also support the observation of seasonal level fluctuations as an important indicator of the occurrence of recharge.

However, external components can affect the interpretation of the water level measurements and lead to misleading results. For instance, effects of pumping (or recovery after cessation of pumping) on water table fluctuations are difficult to estimate and can be confused with recharge. The presence of natural lateral flow is another element not directly identified by the water levels and a likely source of errors. Moreover, the determination of the specific yield S_y in the boundary between the saturated and unsaturated zone is difficult and can lead to significant errors (Sophocleous, 1991).

Another problem associated with the method is that a large number of observations from many different points during a relatively long period are necessary to characterise the spatial and temporal regional variability (e.g. Leduc et al., 2001).

2.5.4. Chemical methods

Chemical methods consist of the analysis of specific elements from water chemical samples in order to indirectly estimate recharge rates. These techniques are based on

the broad assumption that rainfall transports the tracer element to depth and it remains as an indirect record of the vertical flux of water within the soil profile.

Allison (1988), Allison et al. (1994), Hendrickx and Walker (1997) and Edmunds and Tyler (2002) present a detailed review of the tracers methods and list several studies where chemical methods have been used to estimate soil water fluxes.

Examples of the use of applied tracers for recharge estimation are the study carried out by Rangarajan and Athavale (2000) using the tritium injection method for estimating potential recharge during the period of 25 years in several Indian basins. Edmunds et al. (2002) have obtained estimates of recharge using the chloride (Cl) mass-balance method in semi-arid Nigeria. After establishing a conceptual model of Cl concentration, the spatial variability was evaluated through an extensive sampling at 360 regional shallow wells. Another example of hydrogeochemical analysis for estimating recharge is the work of Abd El Samie and Sadek (2001), who made use of isotopic oxygen-18 and deuterium to identify flow characteristics in the Sinai Peninsula.

The problems associated with the chemical technique are the uncertainties in measuring rainfall chemistry in order to establish a correlation between rainfall and solute concentration. Moreover, several assumptions need to be adopted in terms of water and solute movement. Models of solute transport utilise assumptions that limit the fluid flow to a one-dimensional piston-flow mechanism with uncertainties caused by the heterogeneity of the soil system.

Another shortcoming of the technique is the assumption of zero surface runoff in order to establish the concentration balance between solute and rainfall. As discussed in the section 2.2.3, the conditions for runoff in semi-arid areas are highly favourable.

Finally, chemical sampling in semi-arid areas can be a complex process which requires costly analysis and specialised field and laboratory practice.

2.5.5. Zero flux plane

Recharge can be estimated from changes in soil water content measurements below a plane where the vertical hydraulic gradient is zero (the zero flux plane ZFP). The ZFP divides the movement of water within the soil profile into upward (due to evapotranspiration) and downward (recharge). The rate of soil water content change below the ZFP plane is assumed to be equal to recharge. This technique was utilised by Wellings and Bell (1980) to estimate the soil water balance and recharge to the Chalk aquifer in the UK.

This technique requires soil matric potential measurements in order to locate the ZFP and soil water profile measurements to estimate water changes below ZFP.

The principal shortcoming of the method when applied to semi-arid conditions is the impossibility of determining the ZFP when water flows downward throughout the entire soil profile. This condition often occurs during the rainy season. Some authors such as Klaij and Vachaud (1992) suggest the use of the Darcy's law to estimate recharge in this situation.

2.5.6. Darcy's law

Darcy's law is utilised to estimate recharge by the following equation:

$$R_{ech} = -K(\theta)dH/dz \quad 2.3$$

where $K(\theta)$ is the hydraulic conductivity as a function of the water content θ ; H is the total head and z is elevation. This technique requires the determination of the function $K(\theta)$ and makes the assumption that for homogeneous porous media the matric pressure gradient is often nearly zero, and water movement is due to gravity. Therefore, the total head gradient dH/dz is assumed to be equal to 1. Thus, from equation 2.3, recharge is equal to the hydraulic conductivity at the water content θ .

However, as is shown by figure 2.7, hydraulic conductivity K varies many orders of magnitude with soil water content. Consequently, the resulting estimated recharge also varies over a wide range.

2.5.7. Methods based on the numerical solution of Darcy's law

The Darcy equation combined with a mass conservation equation results in a differential equation utilised to describe the process of soil moisture content changes from storage, named the Richards' equation (Lerner et al., 1990). The numerical solution of the Richards' equation associated with the insights provided by field measurements, allow the development of numerical flow modelling to represent the complex physical mechanisms and their interactions within the unsaturated zone. The main advantage of modelling is that models permit the simulation of the effects of the variability of the hydrological components on recharge, which would take a longer time to be observed through direct field experiments.

There are several methods to solve the Richards' equation. For instance, Kutilek and Nielsen (1994) and Campbell (1985) present an extensive review of analytical and numerical solutions under different boundary conditions. All the methods involve the discretisation of the soil in a number of homogeneous layers. The soil water flow in each layer is then the product of the balance between the flow computed in the surroundings layers and at the top and bottom of the soil profile.

Numerical models require the knowledge of the relation of the unsaturated hydraulic conductivity K_{unsat} to soil water content θ and matric potential for each of the discrete layers. In order to represent the natural soil profile variability and meet the computational requirements, a large number of discrete layers are necessary (each one with their respective hydraulic parameters). This increases the number of parameters and the consequent need for input data.

In addition, the process of root water uptake is usually included as a sink term S associated with the flow equation. This term is determined for each layer and the sum is considered to represent the crop transpiration component. Therefore, the root growth process is fractionated into discrete soil layers.

Hendrickx and Walker (1997) list some recent model codes examples. Usually, they were originally designed for agricultural purposes where recharge is the derived water flow through the bottom of the last discrete layer. Examples of model application in

semi-arid areas are the studies of Rockström et al. (1998) who utilised the numerical model SOIL in order to quantify the soil water balance partitioning in sparsely cropped fields in semi-arid Niger; Zhang et al. (1999a,b) utilised the model WAVES to determine the impact of different agronomic practices on recharge in southeastern Australia; and Arora and Gajri (1996) applied the SWAP model for assessing the water balance components under maize.

Problems associated with the use of numerical models are the uncertainties in the relationship between hydraulic conductivity and matric potential or water content (Scanlon et al., 2002) and problems with the reliability of estimation of the input parameters (Cuenca et al., 1997).

2.5.8. Soil water balance technique

The soil water balance technique considers the hydrologic processes as inputs and outputs of an overall mass balance (figure 2.9.a). The inputs and outputs are based on the main physical processes controlling the water content of a given volume of soil during a period of time.

The water content of the representative soil volume changes due to the addition and withdrawal of water as a result of hydrological processes. These processes are simultaneously quantified (as illustrate by figure 2.2 and described in the Section 2.3).

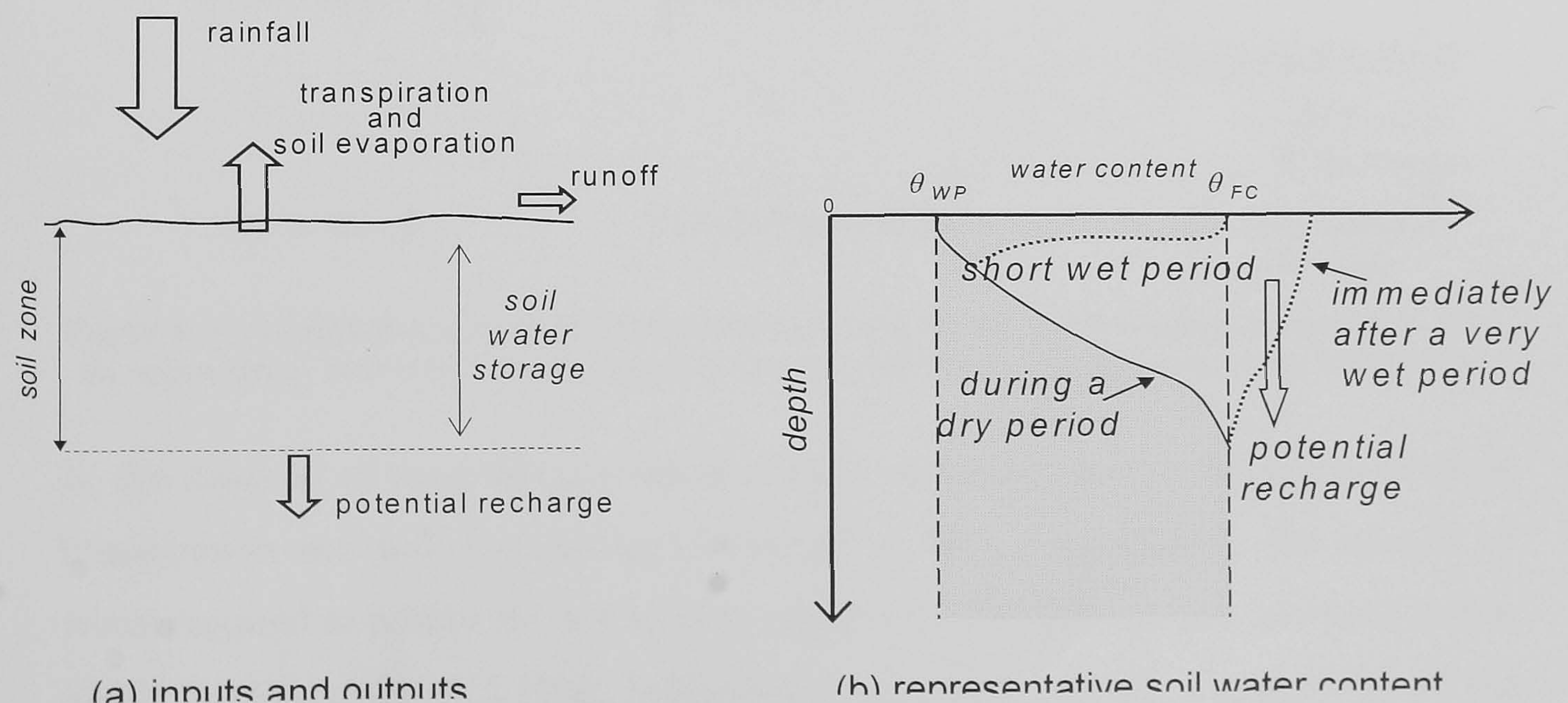


Figure 2. 9. Schematic representation of the soil water balance technique

Figure 2.9.b shows a representative soil water content profile which underlies the soil water balance concept. It represents a soil unit with a crop during the rainy season. During a prolonged dry spell the water held by the soil against gravity (field capacity point θ_{FC} as defined in section 2.3) is depleted by the demands of vegetation and soil evaporation. Note that the shaded areas in figure 2.9 show the soil water content available for plants above wilting point. The figure shows a situation immediately after a short wet period, when the water content at near surface increases but still remains below field capacity. In the absence of any following rain event, this additional water is directly evaporated from bare soil and/or utilised by the plants. However, if the rain continues or there is an occurrence of a very significant rain event, the water content increases beyond field capacity and hydraulic conductivity is high enough to allow for the rapid downward flow of water (potential recharge).

Figure 2.10 illustrates a standard computational concept of a soil water balance model over a time period of one day.

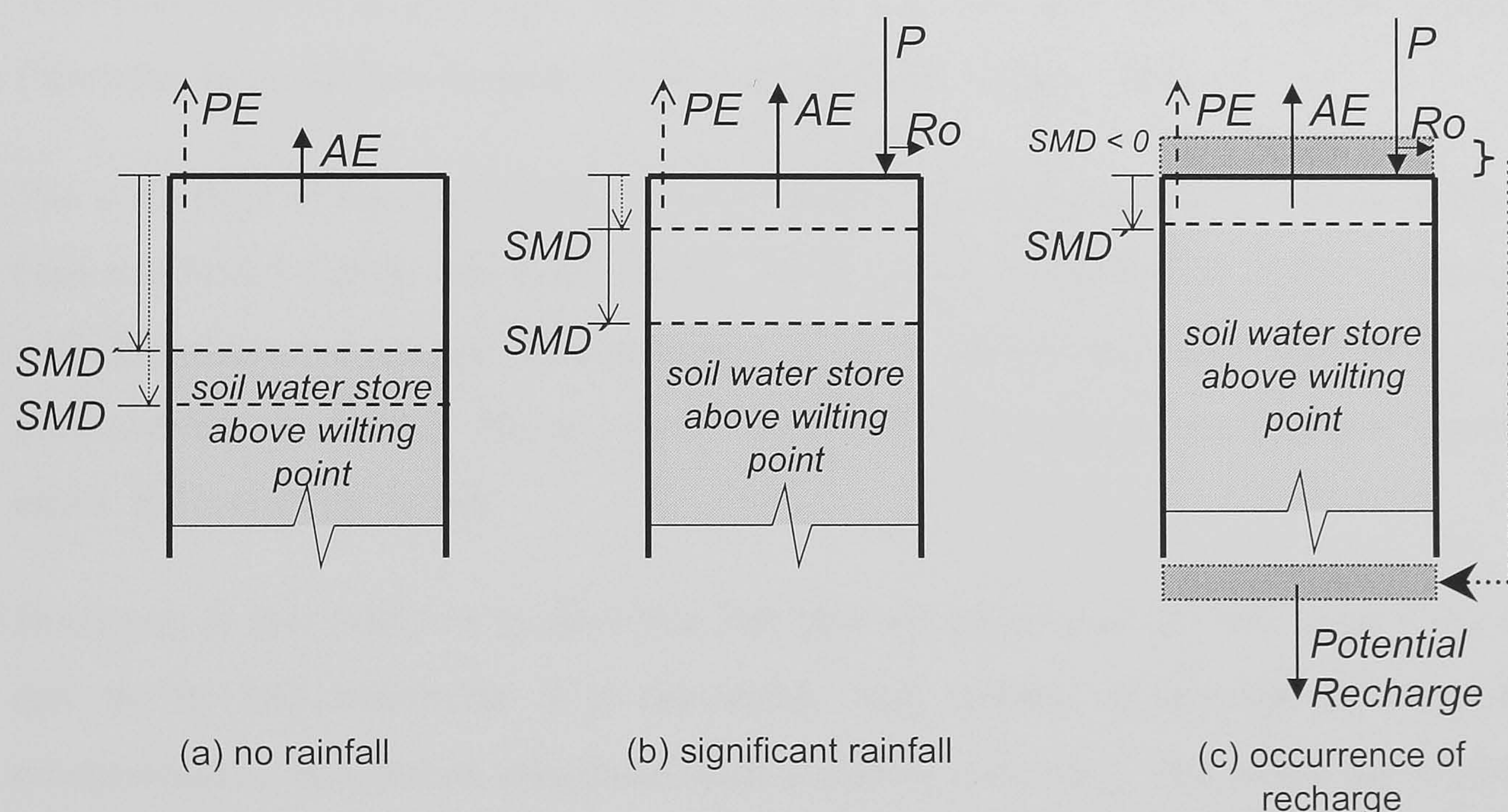


Figure 2. 10. Computational representation of the soil water balance in three typical situations. SMD' is the soil moisture deficit at the beginning of the period, and SMD is that at the end of the same period.

In the absence of rain, the soil water content decreases due to the effect of plant transpiration and soil evaporation (combined as AE). Consequently the amount of water required to restore the soil to field capacity (defined as the soil moisture deficit SMD) increases (figure 2.10.a). Following a significant rainfall, a proportion of the rainfall P may runoff over the surface R_o ; the infiltrated water may allow for plants to

transpire at potential rates PE ; the water surplus is stored in the soil thus reducing the soil moisture deficit (figure 2.10.b). When the infiltrated water is enough to bring the soil water content beyond field capacity, the excess water (negative SMD) moves downward as potential recharge (figure 2.10.c).

The soil water balance technique is normally utilised in agricultural studies for crop yield prediction and determination of crop water requirements during a crop season. In the literature there are many computational water balance models such as the approaches described by Pereira et al. (1992) and Smith et al. (1996).

Abdulrazzak et al. (1989) presents an example of the application of the soil water balance technique for estimating recharge in a Saudi Arabian basin. The authors concluded that the soil water balance was a useful instrument for water management policies in an arid region. Other examples of soil water model application in semi-arid areas are the studies of Lloyd et al. (1967) in northeast Jordan and Odigie and Anyaeche (1991) in northeast Nigeria. In the UK this approach is widely utilised following the studies of Penman (1949, 1950) and Grindley (1967).

The utilisation of this method in semi-arid areas is criticised by some authors such as Gee and Hillel (1988) and Lerner et al. (1990). They suggest that errors associated with the measurement and determination of large components, such as rainfall and evaporation, can lead to faulty values since recharge is the residual of two almost equal, but uncertain values.

However, in this study we believe that this type of criticism of the soil water balance can be in fact overcome if a reasonable and reliable determination of these components is introduced associated with a correct time step. The following section addresses some critical issues for the application of soil water balance techniques in semi-arid areas.

2.6. AN ARGUMENT FOR SOIL WATER BALANCE AS A RECHARGE ESTIMATION TECHNIQUE

The determination of groundwater recharge under the guidelines presented in Section 1.1 points the need for a method which reflects insights into the complexity of the real physical system, but balances model complexity with the detail of the data which can be realistically gathered.

The following issues provide an argument for the soil water balance method in semi-arid studies of recharge:

- ***Physical credibility.*** The soil water balance technique, as briefly described in section 2.5.8, is based on the knowledge of the physical processes and their interactions. It accounts for all water entering and leaving the system (the inputs and outputs of figure 2.10.a). The technique exhibits insights of the key components of the system based on observation and/or experience (e.g. figure 2.2) capturing the key processes in potential recharge. Therefore, if all the important processes are included, the soil water balance model can be physically sound and robust.

However, sometimes soil water balance models do not include some important processes that would be necessary. For example, although in Section 2.2.3 surface runoff is identified as a very likely process in semi-arid areas, conventional soil water balance techniques usually do not distinguish between runoff and deep drainage from the bottom of the root zone, or simply assume that runoff from a plot is zero.

Moreover, using a sufficient short time step (daily or less) to represent the rainfall characteristics in semi-arid regions overcomes the usual criticism that this approach is not suitable in semi-arid areas because rainfall and actual evapotranspiration are similar. This type of criticism is derived from the utilisation of rainfall summed or averaged over longer time periods. However, rainfall in a single day may greatly exceed evapotranspiration even in arid climates.

- ***Modelling complexity and uncertainties.*** The soil water balance technique does not fully incorporate the knowledge of water movement within the soil, as the

methods based on the physical laws (Darcy and Richards) attempt to do. As described in sections 2.5.6 and 2.5.7, these laws represent the physics of flow in response to hydraulic or potential gradients, as a function of unsaturated hydraulic conductivity.

However, due to the uncertainties in hydraulic conductivity determination caused by the natural variability of the real soil system and the uncertainties in the non-linear relationships between hydraulic conductivity and matric potential, or water content, recharge estimates based on the Darcy and Richards equations may be highly uncertain. For instance, K_{unsat} is highly sensitive to water content θ (figure 2.7), therefore, measurement errors in θ (even within a fraction of percent) lead to significant variation in calculated K_{unsat} , hence in recharge.

The inclusion of computational equations which better represent the complex real system results in an increasing model complexity and a consequent increase in parameters and input data. However, adding complexity does not necessarily increase model accuracy, even when additional and reliable data are available (Walker et al., 2002).

Similar problems associated with K_{unsat} determination can be found in the determination of the specific yield S_y utilised in recharge computation by the water-table fluctuation method (equation 2.2). Specific yield is highly variable and difficult to determine due to boundary assumptions.

Moreover, field measurements such as in lysimeters, soil water profile content (zero flux plane) and chemical techniques have inherent problems due to the localised nature of the observations and disturbance of the system they are attempting to represent.

- **Data availability and practicability.** As the number of parameters increases and detailed measurements of the hydraulic soil properties for each discrete layer are required, the approach in question may become impractical. Moreover, there is a need for even more information to carry out the validation of the model.

Usually, in semi-arid areas the availability of complex information about the soil-vegetation system is poor and there are financial and technical limitations involved in gathering a large amount of data.

Soil water balance techniques make use of a small number of assumptions derived from the system. Consequently, this approach usually requires fewer key parameters than the approaches based on the numerical representation of the flow equations.

Usually soil water balance requires readily available data such as standard climatic observations and conventional soil and vegetation properties. Soil water balance can thus provide a better balance between the conceptual representation of the physical processes, the computational capacity and data availability.

2.7. REVIEW OF EXISTING SOIL WATER BALANCE MODELS

This section presents a brief review of some existing water balance models describing the main characteristics and giving examples of application. At the end of Chapter three they are compared to the methodology developed in this study.

Four existing methods were select for this review based on their different representation of the physical mechanisms. They are: The conventional single layer model, the CROPWAT model, the BALANCE model and the Four Layer model.

2.7.1. Conventional single layer model

The conventional single layer model refers to the methods based on the studies of Penman (1949, 1950) and Grindley (1967). Originally, these authors were concerned with the determination of actual evaporation and soil moisture deficits and not directly with the estimation of recharge. However, their studies precipitated much other literature regarding soil water balance techniques, since their concept of soil moisture accounting became common to many subsequent models (Lerner et al., 1990).

In this approach, the soil profile is treated as a single reservoir filled by rainfall and emptied by evaporation and drainage through the bottom of the root zone (runoff is

not taken into account). In order to represent the actual situation when vegetation is under water stress and no water supply is present, Penman (1950) introduced the concept of *root constant* (C). The root constant defines the amount of water in mm that can be extracted from a soil at potential rates by given vegetation. It is assumed that actual evaporation AE is equal to potential evaporation PE until soil moisture deficit SMD reaches the root constant C . Then, AE is a fraction F of PE to the permanent wilting point D (figure 2.11). Assuming a precipitation P at the start of the day, the final equation is:

$$AE = P + F \cdot (PE - P) \quad 2.4$$

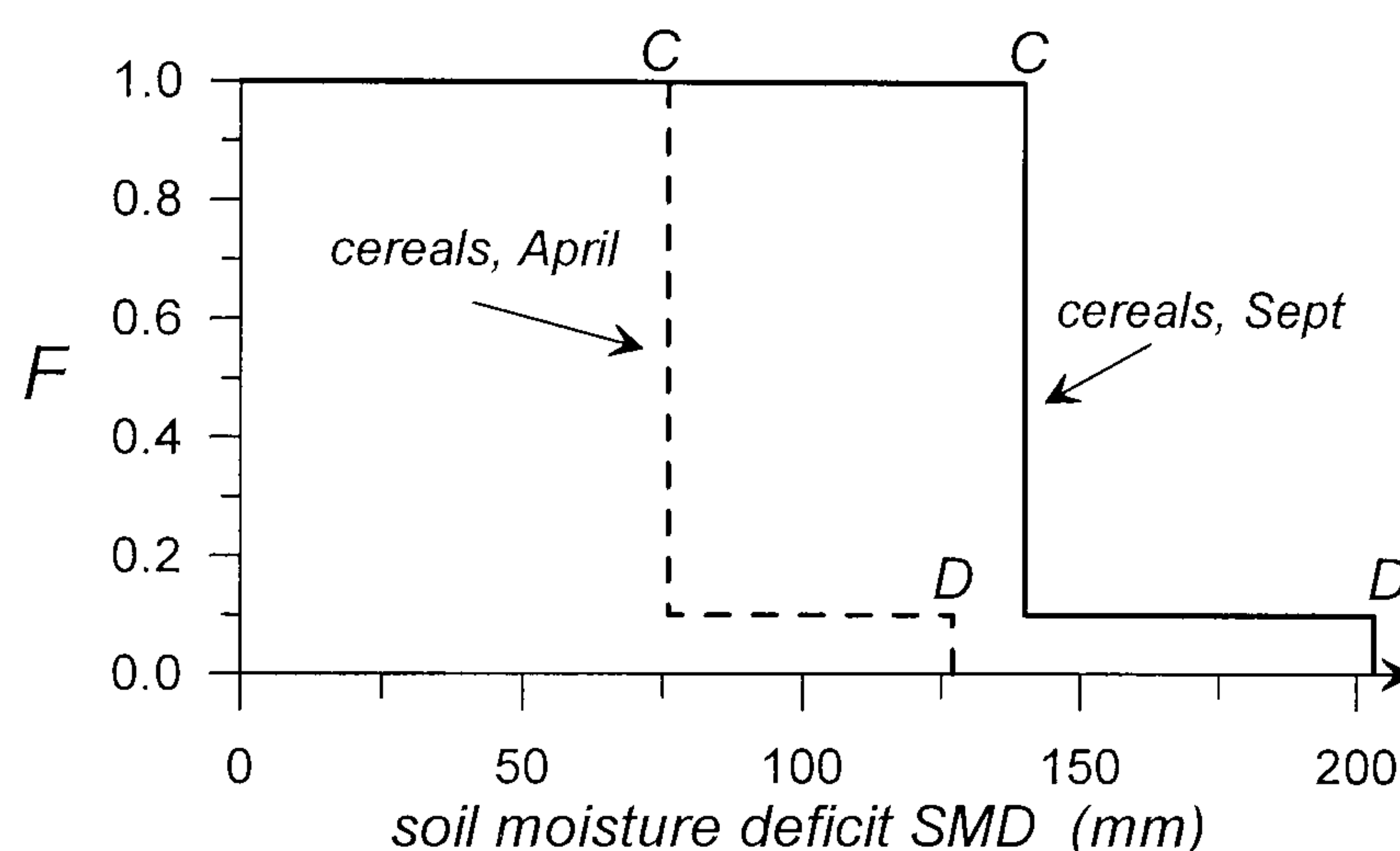


Figure 2. 11. The fraction F relating AE and PE in absence of additional water supply according to the conventional method (from Lerner et al., 1990)

The parameters C and D are related to the crop type and stage of crop development. Figure 2.11 shows the parameters at the initial and late stages of a cereal crop. Lerner et al. (1990) list a series of monthly C and D values for different crop types utilised in the UK.

This method relies on accurate estimates of the parameters C and D . However, they have a strong empirical characteristic and, consequently, they are not easily determined. Moreover, they fail to represent some particular field situations such as the period immediately after harvest, when D has to be reduced to reflect changes in vegetation cover and root depth, but the soil moisture deficit is still large. The other main shortcoming of this approach is that it takes no explicit account of soil type.

Example of studies for estimating recharge based on this approach are the cited studies of Lloyd et al. (1967) and Odigie and Anyaeche (1991) (Section 2.5.8); Rushton and Ward (1979) for estimating recharge to the Chalk aquifer in the UK; and Seymour et al. (1998) in assessing the potential recharge to the Fylde aquifer also in the UK.

2.7.2. The CROPWAT model

The CROPWAT model is a computer program developed by the FAO (Food and Agriculture Organization of the United Nations) to carry out standard calculations for crop water and irrigation requirements (Smith, 1992). The procedures for the calculation are based on the methodologies presented in the FAO Irrigation and Drainage Paper No. 24 (Doorenbos and Pruitt, 1977) and in the FAO Irrigation and Drainage Paper No. 33 (Doorenbos et al., 1978). These techniques were later revised and presented by Allen et al. (1998) in the FAO Irrigation and Drainage Paper No. 56.

The method is based on a daily soil water balance for the determination of actual evapotranspiration from a crop plot. It begins with the estimation of the evapotranspiration from a reference surface ET_o . The reference evapotranspiration ET_o is estimated using the modified Penman-Monteith equation in which are assigned particular values for the parameters related to a specific grass reference surface, well supplied with water and with a complete surface cover (Allen et al, 1996).

The determination of the potential evapotranspiration from a particular crop PE is made through the use of crop coefficients K_c . The crop coefficients represent an integration of the effects of crop characteristics which differentiate the crop from reference surface. A more detailed discussion about crop coefficients is carried out in Chapter three.

Actual evapotranspiration AE is related to potential crop evapotranspiration PE by a similar relationship to that presented in section 2.7.1. That is, AE is equal to PE as long as the soil moisture deficit SMD has not reached a critical level. Beyond this level, the crop is under water stress and actual evapotranspiration is reduced proportionally to a stress coefficient K_s (figure 2.12).

The depth of soil water which can be used effectively by the crop, defined as Total Available Water (TAW), depends directly on the rooting depth of the crop and on the moisture holding properties of soil. That is, TAW (in mm) is the difference in soil moisture content between field capacity (θ_{FC}) and wilting point (θ_{WP}) at a particular root depth Z_r . The fraction of the TAW extractable by the roots without water stress is

the Readily Available Water (*RAW*) obtained by multiplying *TAW* by a depletion factor *p*. Figure 2.12 illustrates the relation between *AE* and *PE* for two different types of soils, as well as the equations utilised for the parameters calculation. Note that for a sandy soil, the availability of water for evapotranspiration *TAW* is smaller due to the lower field capacity point.

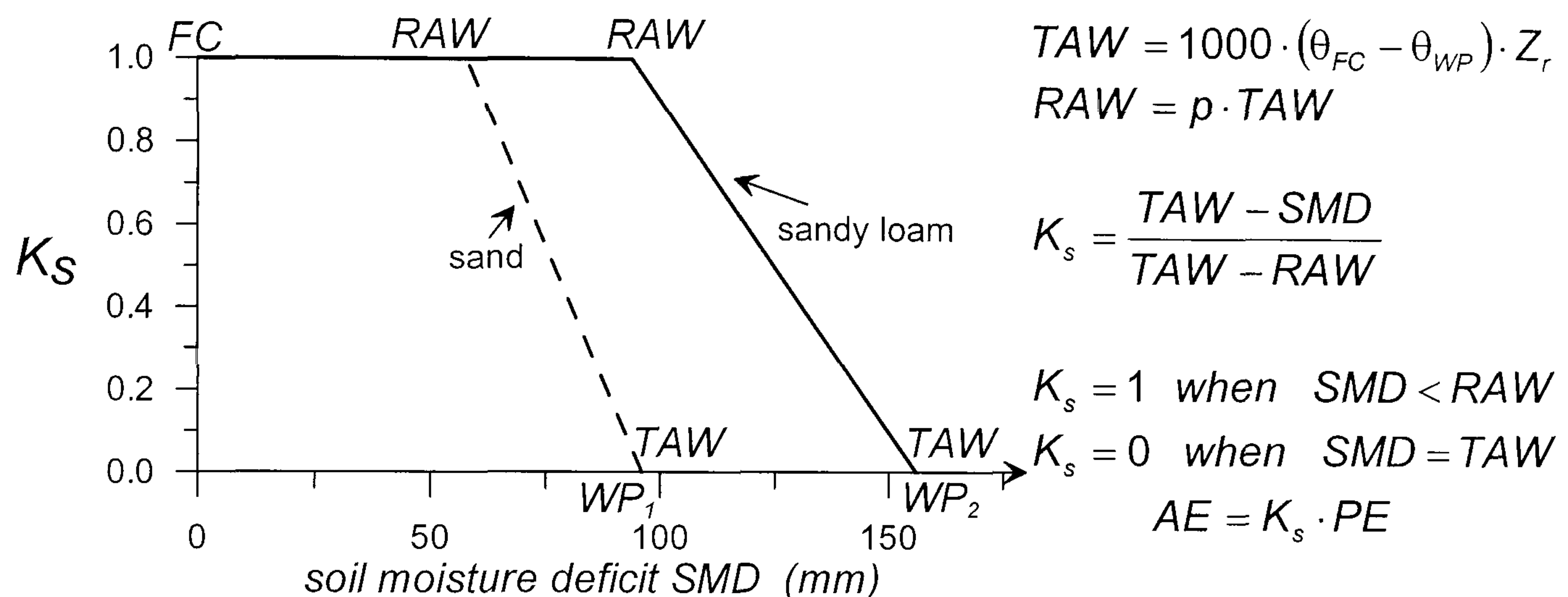


Figure 2. 12. The stress coefficient as function of the soil moisture deficit – FAO approach.

The methodology present in the CROPWAT model represents an improvement to the conventional single layer method in terms of actual evaporation determination, including important physical mechanisms such as the root growing process and soil water holding properties. Moreover, the concept of crop coefficients and reference evapotranspiration has been extensively used worldwide in many agricultural, hydrological and environmental studies (Smith et al., 1992).

2.7.3. The BALANCE model

The BALANCE model (Hess, 1994) is a comprehensive program written to estimate daily soil water balance for a cropped or un-cropped surface. It was originally written for agricultural purposes in the UK. Meanwhile, the model has been tested and applied on sandy soils in semi-arid Nigeria (Hess and Grema, 1994). This model presents two main differences when compared with the two previous approaches. They are the inclusion of multiple soil layers and the determination of soil evaporation and plant transpiration separately.

The soil profile is divided into three stores (figure 2.13). The first store consists of a fixed upper layer of approximately 15 cm from where runoff and bare soil evaporation are calculated, followed by the root zone and the remaining soil zone beyond the root depth. The movement of water between these layers occurs when water content exceeds field capacity. Soil evaporation is calculated by the method presented by Ritchie (1972) and crop transpiration by a method similar to the FAO approach.

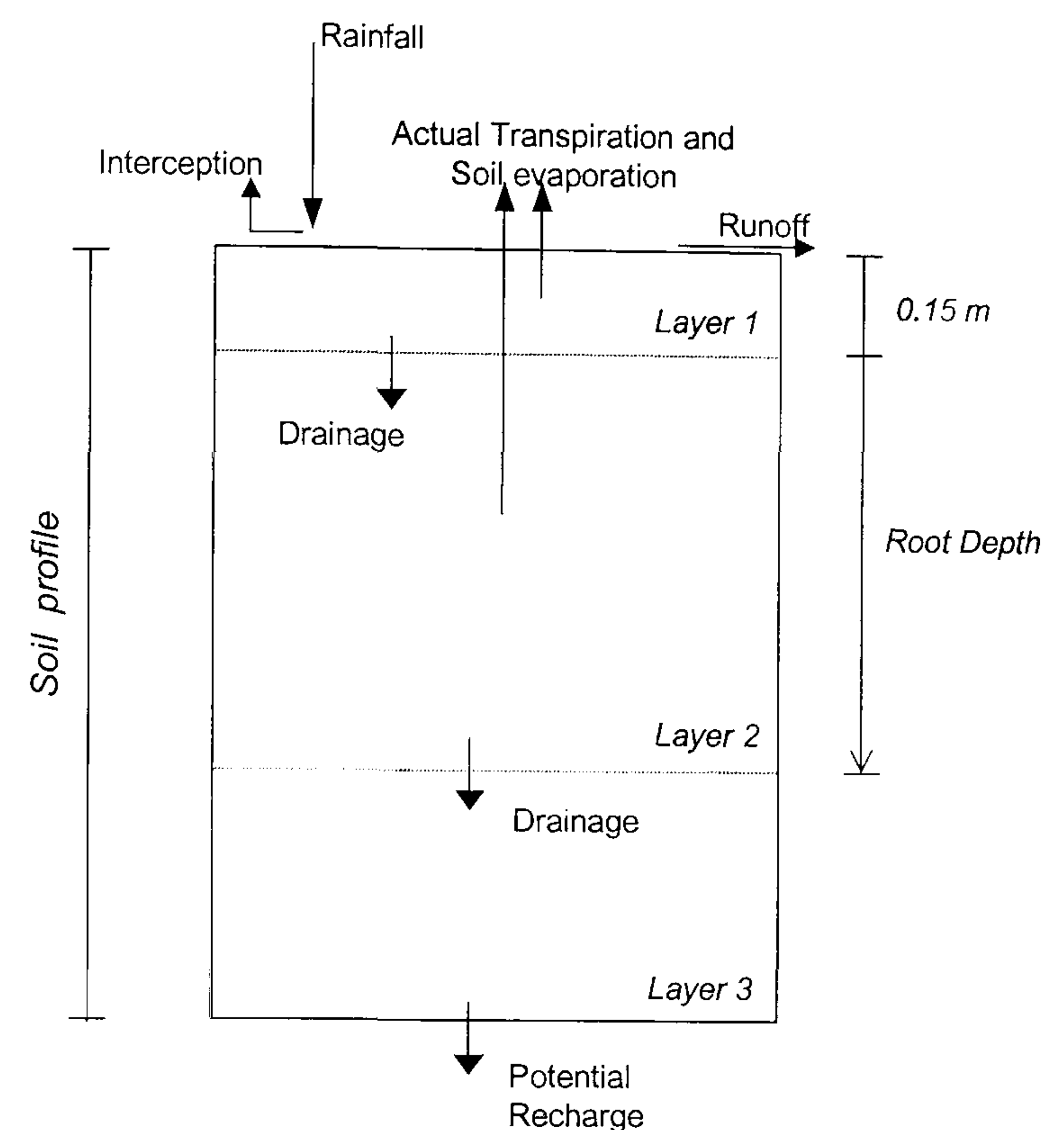


Figure 2. 13. Inputs and outputs to and from the soil water stores in the BALANCE model.

Then, the two components are summed according to the proportion of surface cover. Surface runoff is estimated using the curve-number technique developed by the USDA Soil and Conservation Service.

2.7.4. The Four Root Layers Model (FRLM)

The FRLM model is a soil moisture model developed at the Institute of Hydrology, UK, for the estimation of soil moisture deficits in sites under permanent grass cover (Ragab et al., 1997).

The FRLM model attempts to reflect the root density distribution dividing the soil profile into four layers, each layer representing 25% of the total root depth (figure 2.14). Each layer contributes differently to the total root water uptake depending on its root density. Therefore, the upper layer contributes 70% of the total water uptake and the others the remaining 30%. Water movement between layers occurs when the water content exceeds field capacity, recharge being the drainage from the bottom layer.

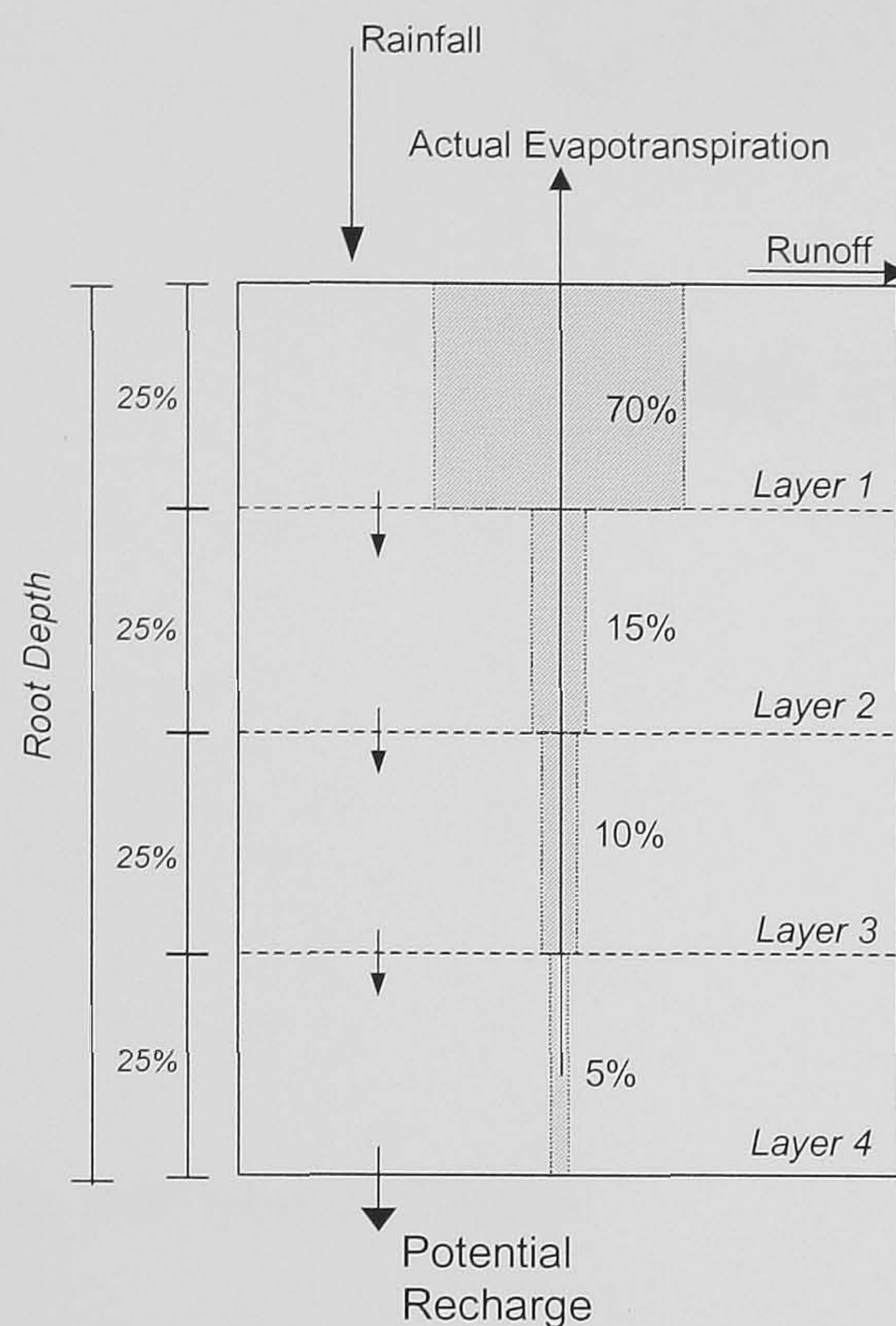


Figure 2. 14. Schematic representation of the FRLM model.

Finch (1998) adapted this approach in order to represent the seasonal variations in the root growth changing the contribution made from each layer according to the crop growing stage.

The FRLM method is based on the principle that the distribution of roots with depth is approximately triangular in shape, the most of the water uptake by plants being from the soil near surface. However, it seems to have no clear experimental basis as the process of root growth is dynamic and not partitioned into fixed layers (see section 2.2.6)

2.7.5. Discussion

The soil water balance models presented in Section 2.7 provide valuable insights into the hydrologic processes occurring in the soil-crop-water system. However, they were not designed to represent the principal hydrological processes which actually affect recharge in such environments (as mentioned in Section 2.4).

Chapter three presents the development of a new conceptual soil water balance model that includes alternative procedures in order to take into account key hydrologic processes, thus improving the physical robustness of the soil water balance technique for recharge estimation in semi-arid conditions.

A qualitative comparative analysis between the new conceptual model and the alternative models here described is carried out at the end of Chapter three.

CHAPTER THREE

THE SAMBA MODEL

Chapter three presents the fundamentals of the methodology developed for recharge estimation based on a soil water balance technique. New procedures are included in order to represent physical mechanisms affecting recharge in semi-arid conditions.

3.1. INTRODUCTION

Section 2.6 presents a series of arguments for the use of the soil water balance technique for recharge estimation in semi-arid areas. One of the principal issues is about the physical credibility of the conceptual model. Following the criticism from authors such as Allison (1988), Gee and Hillel (1988) and Lerner et al. (1990, the soil water balance approach has been considered too simplistic to represent the complex physical processes actually involved in the water movement within the soil. However, this thesis challenges this overall criticism and shows that by the association of an appropriate representation of the principal hydrological processes and their relationship with an adequate time step, the soil water balance technique is a useful and important tool for the understanding of the groundwater system and their implications for estimating groundwater recharge.

Chapter three presents a conceptual model based on the physical understanding of the processes acting in a typical semi-arid climate. The model follows a series of existing procedures in order to determine the soil water balance components. However, new aspects have been included and combined in order to adequately represent actual field conditions. They are:

- the development of a new procedure to estimate surface runoff based on climatic and soil moisture conditions,

- the inclusion of the bare soil period before and after the crop season and,
- the inclusion of key physical features regarding the complex water movement within the soil profile which reflect: (a) the condition of water stress during the crop period, (b) the root water uptake following rainfall and, (c) the retention of part of the rainfall near the surface, thus making it available for shallow roots in the following days.

These aspects incorporate the physical features present in situations where the conventional concepts of soil water balance are not totally adequate.

The following sections present the conceptual description of the soil water balance model and how the alternative procedures are incorporated into the overall daily soil water balance. The model has been named SAMBA (Semi-Arid Model using the soil water BAlance technique).

3.2. THE SAMBA MODEL

The SAMBA model is a single layer soil water balance model that incorporates the physical processes identified in section 2.4. It follows the procedures of the conventional soil water balance technique as described in section 2.5.8 (see figure 2.9 and 2.10) and section 2.7.1. In summary, they are the water accounting of the inputs and outputs within a representative soil volume where the soil water content is represented by the concept of soil moisture deficit *SMD* (as defined in section 2.5.8). A further discussion about *SMD* is set out in later sections of this chapter.

However, the conventional soil water balance technique is not adequate for application in semi-arid climates, as is demonstrated during this chapter. The new aspects cited in section 3.1 need to be included. The followings sections present the components of the model using diagrams and numerical examples.

3.3. A PRACTICAL APPROACH FOR THE INCLUSION OF SURFACE RUNOFF

Usually, soil water balance models do not attempt to estimate surface runoff (e.g. the models presented in section 2.17, 2.7.2 and 2.7.4). However, as discussed in section 2.2.3, surface runoff is an important component of the soil water balance in semi-arid areas and therefore needs to be considered. The SAMBA model estimates runoff using an alternative approach associating practicability with the problem of data limitation. It is a preliminary representation of runoff based on the empirical relationship of three key aspects affecting surface runoff at a plot scale. They are the surface characteristics (soil and vegetation), rainfall intensity and the antecedent soil moisture conditions.

However, it is not the main objective of this research to develop a procedure for estimating runoff. There are many methodologies and models available to estimate runoff, although they usually require a large number of variables at short time intervals - such as in the EUROSEM model (Morgan et al., 1998). The present study highlights the importance of the inclusion of runoff in recharge determination in semi-arid environments and presents a plausible procedure for its quantification. This procedure is flexible and provides a basis for further research to verify the relation between the variables at a plot scale.

Runoff is estimated correlating rainfall and the antecedent soil moisture conditions (soil moisture deficit at start of the day) through a matrix of coefficients (r_c), as illustrated by the table 3.1.

Table 3. 1. Example of coefficients for runoff determination for a sandy soil

Soil moisture deficit at start of day (mm)	Rainfall intensity (mm/day)		
	0 – 20	20 – 50	> 50
0 – 20	0.10	0.20	0.30
20 – 50	0.05	0.15	0.25
> 50	0.00	0.10	0.20

Runoff Ro is calculated according to the following equation, P being the precipitation (in mm/day) and r_c the runoff coefficient (from the matrix),

$$Ro = P \times r_c \quad 3.1$$

The coefficients relating rainfall and soil moisture deficit aim to represent the complex physical process involved in the relationship between rainfall, soil surface conditions and infiltration capacity limits. They are based on the assumption that runoff is significant when the soil is very wet and rainfall intensity is high. Such conditions are likely to occur in semi-arid areas due to the rainfall characteristics.

It is assumed that rainfall intensity is implicitly included in the daily rainfall values. High values of rainfall are assumed to have intensities enough to cause runoff, whereas low rainfall events have generally little or no runoff. In semi-arid regions, rainfall is highly intense. Lebel et al. (1997) observed in semi-arid Niger that half of the annual rainfall falls at intensities exceeding 35 mm.h^{-1} , with one-third being at intensities higher than 50 mm.h^{-1} . Moreover, half of the seasonal rainfall falls in events of less than 5 hours. These rates were fairly constant during the period of observation (1990-93).

The definition of the matrix coefficients depends on a series of factors and should be modified to represent different soil types and surface conditions, such as slope, tillage and crusting. The method allows for the parameterisation and validation of the coefficients as well as the inclusion of further intervals of rainfall intensity and soil moisture deficit.

The coefficients presented in table 3.1 are preliminary values based on limited observations of runoff over sandy soil in semi-arid areas. They attempt to realistically represent the gradual increase of runoff rates from a threshold rainfall value (20 mm.day^{-1}) associated with the initial soil moisture conditions. Some values from field experiments and observation in semi-arid West Africa are given below.

The minimum rainfall intensity necessary to begin overland flow over a millet plot in sandy soil observed by Peugeot et al. (1997) was 18 mm.h^{-1} in Niger. Owonubi et al. (1991) based on results from northeast Nigeria state that all rainstorms of more than 20 mm in 24 hours resulted in some runoff over different surface covers. In addition, Audu (1999) based on values of modelled runoff at 13 sites in northeast Nigeria suggests the rainfall daily value of 16 mm as the threshold value for runoff to occur in a dry soil condition.

However, despite the attempt to relate the values of table 3.1 to field information they have to be understood as starting point values and further investigation has to be made in order to better define them. Chapter Five presents and compares some results using the table 3.1 coefficients to overall results from the literature.

Figure 3.1 gives two numerical examples of a runoff calculation using the parameters from table 3.1.

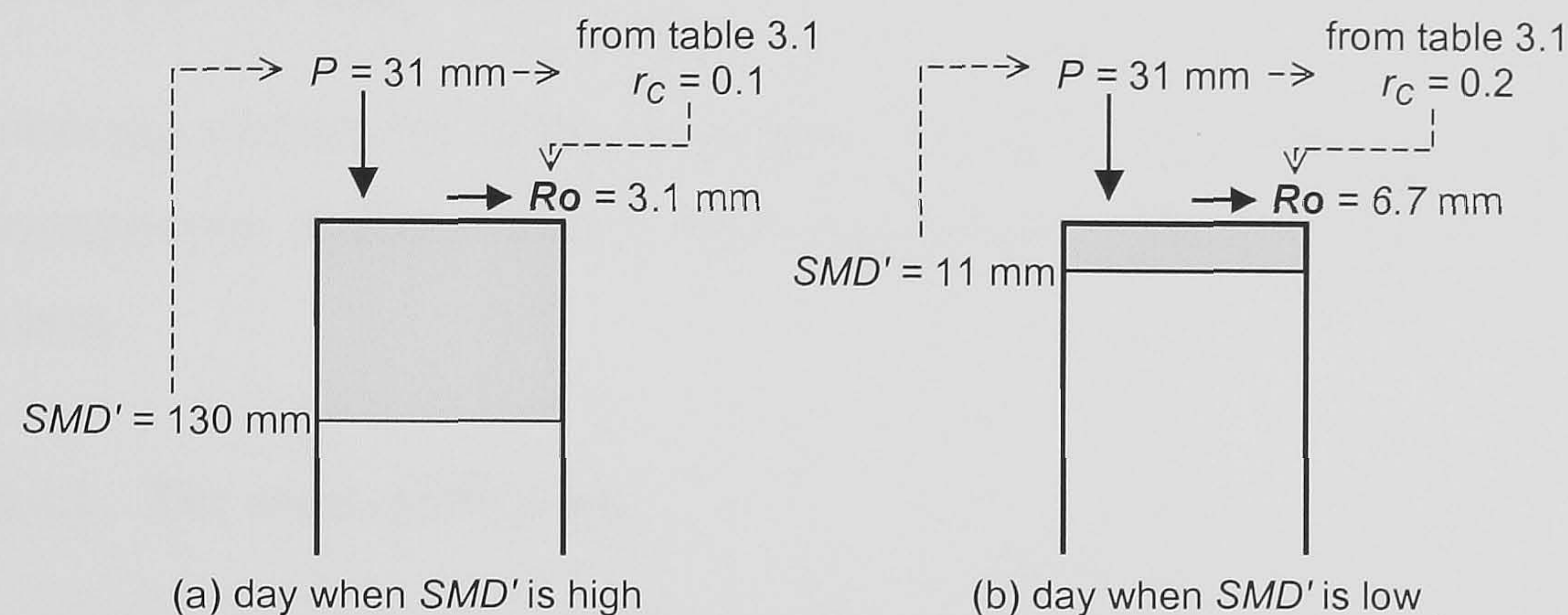


Figure 3. 1 Two examples of runoff calculation

Two days with significant rainfall are considered. Figure 3.1 (a) shows a day at the start of the rainy season when the soil is relatively dry and the moisture deficit at start of the day SMD' is high. A rainfall of 31 mm depth results in 3.1 mm of runoff (10%). However, when the soil is wetter, such as during the rainy season, the estimated runoff reflects the soil moisture conditions and runoff is double for the same rainfall value (figure 3.2.b).

3.4. THE POTENTIAL EVAPORATION FOR A CROPPED AREA INCLUDING THE PERIOD WHEN BARE SOIL IS PREDOMINANT

In semi-arid areas, the periods when the direct evaporation from the soil surface is the dominant process are key periods because they define the initial conditions of the soil water balance during the rainy season, consequently affecting the determination of recharge (see section 2.2.4). They are the periods before the beginning of the crop development and after harvest.

The SAMBA model includes those periods in the determination of the potential evapotranspiration from a cropped unit area. The procedures follow the FAO method described in section 2.7.2. However, as the FAO method has been designed for the assessment of crop water requirements during the crop season, no attempt is made to represent the periods before and after the crop growth. Therefore that concept has been modified in order to include the period when direct soil evaporation is the dominant evaporative process.

The following sections show the procedures for the determination of the potential evapotranspiration starting with a brief description of the FAO concept of crop coefficients.

3.4.1. The crop coefficients

The determination of the evapotranspiration from a cropped surface at optimum moisture conditions PE according to the FAO method is a two-step technique where the evapotranspiration from a reference surface ET_o (see section 2.7.2 for definitions) is related to a particular crop through the utilisation of crop coefficients

$$PE = K_c \cdot ET_o \quad 3.2$$

The crop coefficient K_c integrates the main characteristics that distinguish a typical crop from the reference surface (Allen et al., 1998). That is:

- the crop height which influences the aerodynamic resistance term in the Penman-Monteith equation;
- the albedo of the crop-soil surface which is affected by the fraction of ground covered by vegetation and by the soil surface wetness;
- the canopy resistance which is affected by the number of stomata and leaf age, among others features; and
- the direct evaporation from the exposed soil fraction.

The effect of changing crop characteristics over the growing season on crop coefficients is obtained dividing the growing period into distinct stages:

- Initial stage, representing the initial period from planting when the evapotranspiration is predominantly from direct soil evaporation;
- Crop development stage, from the end of the initial period to the beginning of flowering. As the crop develops, transpiration from vegetation gradually becomes the major process;
- Mid-season stage, from flowering to the start of maturity. At this stage the crop coefficient reaches its maximum value;
- Late season stage, from maturity to harvest. The late stage ends when the crop is harvested, dries out naturally, reaches full senescence or experiences leaf drop.

The determination of the stage lengths and crop coefficients K_c will depend on crop variety and growth conditions. For instance, $K_{c\ end}$ is high if the crop is harvested fresh before the end of the rainy season. On the other hand, if the crop is allowed to dry out before harvest, $K_{c\ end}$ is lower. Figure 3.2 illustrates the stages described above and their respective K_c coefficients. The modification included in the SAMBA model to represent the initial crop stage ($K_{c\ ini}$) is explained in the following subsection.

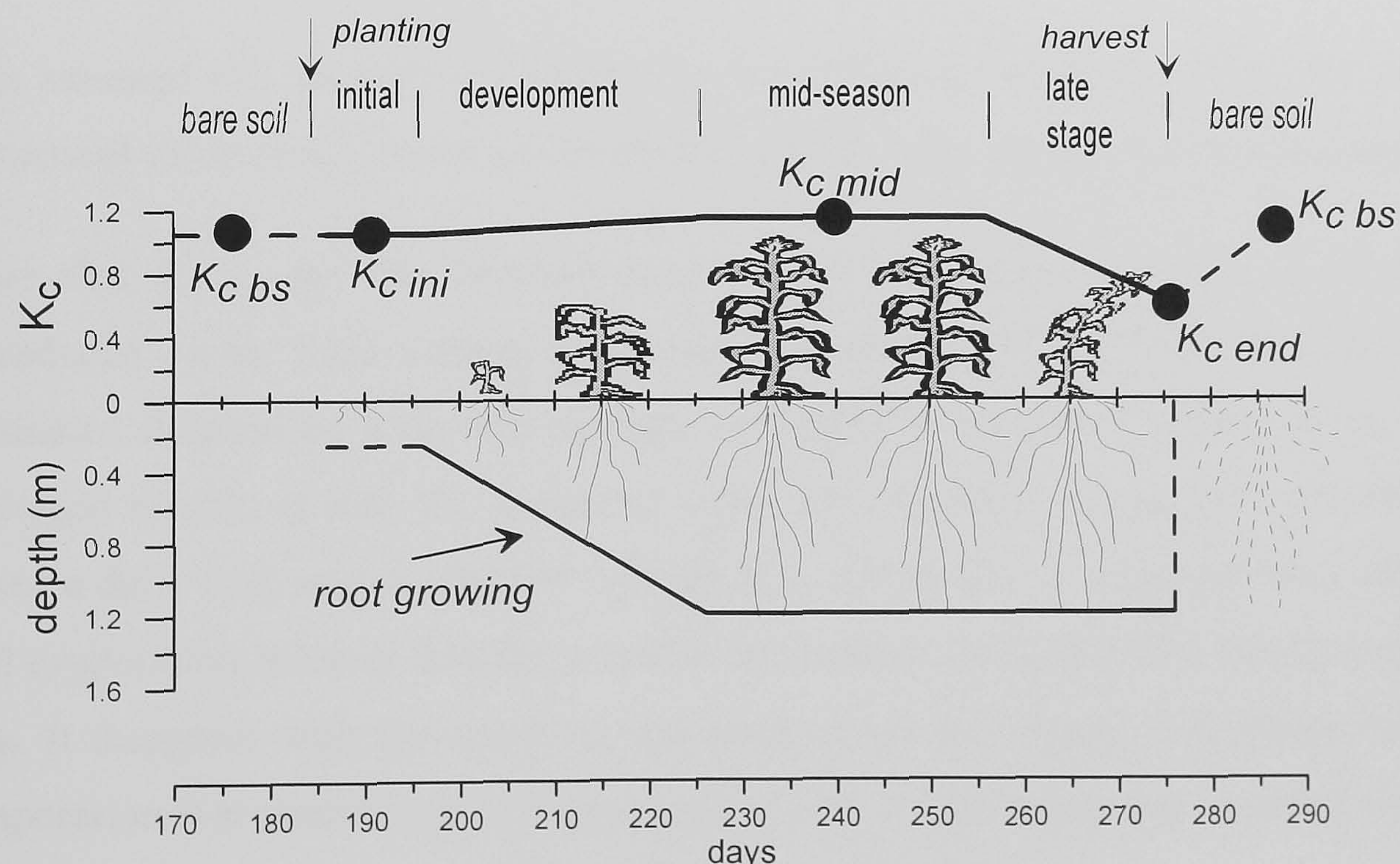


Figure 3. 2. Growing stages including the period before and after the crop when the surface is predominantly bare. The figure includes the respective K_c coefficients and the resulting interpolated crop coefficient curve. The schematic root development according to the crop stages is also included.

Doorenbos and Pruitt (1977) and Allen et al. (1998) list typical values for full grown and irrigated crops which can be used as reference for determining $K_{c\ mid}$ and $K_{c\ end}$. However, it is advised that the values of the coefficients and growth stage length should be adjusted for particular field conditions.

Alternative and more local specific values can be achieved from field experiments such as the studies cited by Abdulmin and Missari (1990) for the Sub-Saharan semi-arid region. However, caution should be exercised when using crop coefficients from literature. Sometimes the crop coefficient represents the direct relationship between actual and reference evapotranspiration. In this case, K_c is usually a lower value because it incorporates the periods of water stress. Note that in the SAMBA model K_c is the relationship between reference and potential evaporation from a cropped surface.

3.4.2. The K_c coefficient for periods when soil evaporation is predominant

The SAMBA model includes a coefficient $K_{c\ bs}$ in order to relate the reference evapotranspiration ET_o to the potential evaporation from bare soil before and after crop planting.

It is assumed that the crop coefficient for the initial crop stage, $K_{c\ ini}$, has the same numerical value as $K_{c\ bs}$ since soil evaporation is the dominant process at that stage.

Note that $K_{c\ ini}$ for the SAMBA model is in fact different to the coefficient introduced by the FAO method (Doorenbos and Pruitt, 1977). The FAO technique estimated it based on a decadal average of rainfall in the initial period. When the averaged rainfall is low, the suggested value for the period is also low. However, when a daily time step is utilised, a small $K_{c\ ini}$ can lead to a situation when actual soil evaporation is lower than the potential evaporative demand (ET_o) during a rainy day. It disagrees with the observed soil evaporation (see figure 2.4) where actual evaporation is at potential rates during a period of 1-3 days following a rainfall event.

In fact, there is no general agreement on the value of $K_{c\ ini}$ when using the FAO method. For instance, Agnew (1991) shows several values of $K_{c\ ini}$ for the early growth period ranging from 0.05 to 0.8. This variation is in part explained by the

ambiguity in the crop coefficient definitions, as pointed out at the end of the previous sub-section, as well as by the dependence of $K_{c\ ini}$ on the rainfall time step adopted Wallace et al. (1993).

Figure 3.3 shows an example of determination of daily potential evapotranspiration during a crop season. The reference evapotranspiration ET_o utilised is a long-term monthly average from semi-arid Nigeria (Hess, 1998). The growth stage lengths and the crop coefficient $K_{c\ mid}$ are determined based on the values listed in Allen et al. (1998) for a millet crop and adjusted for a semi-arid condition with moderate winds. $K_{c\ end}$ represents the situation when the crop is drying out but providing continuous cover to the soil.

The coefficient for the bare soil period $K_{c\ bs}$ and the coefficient for the initial stage $K_{c\ ini}$ have the same value determined from the example number 31 of the FAO publication 56 (Allen et al., 1998) for the potential evapotranspiration from a bare soil surface at maximum rate.

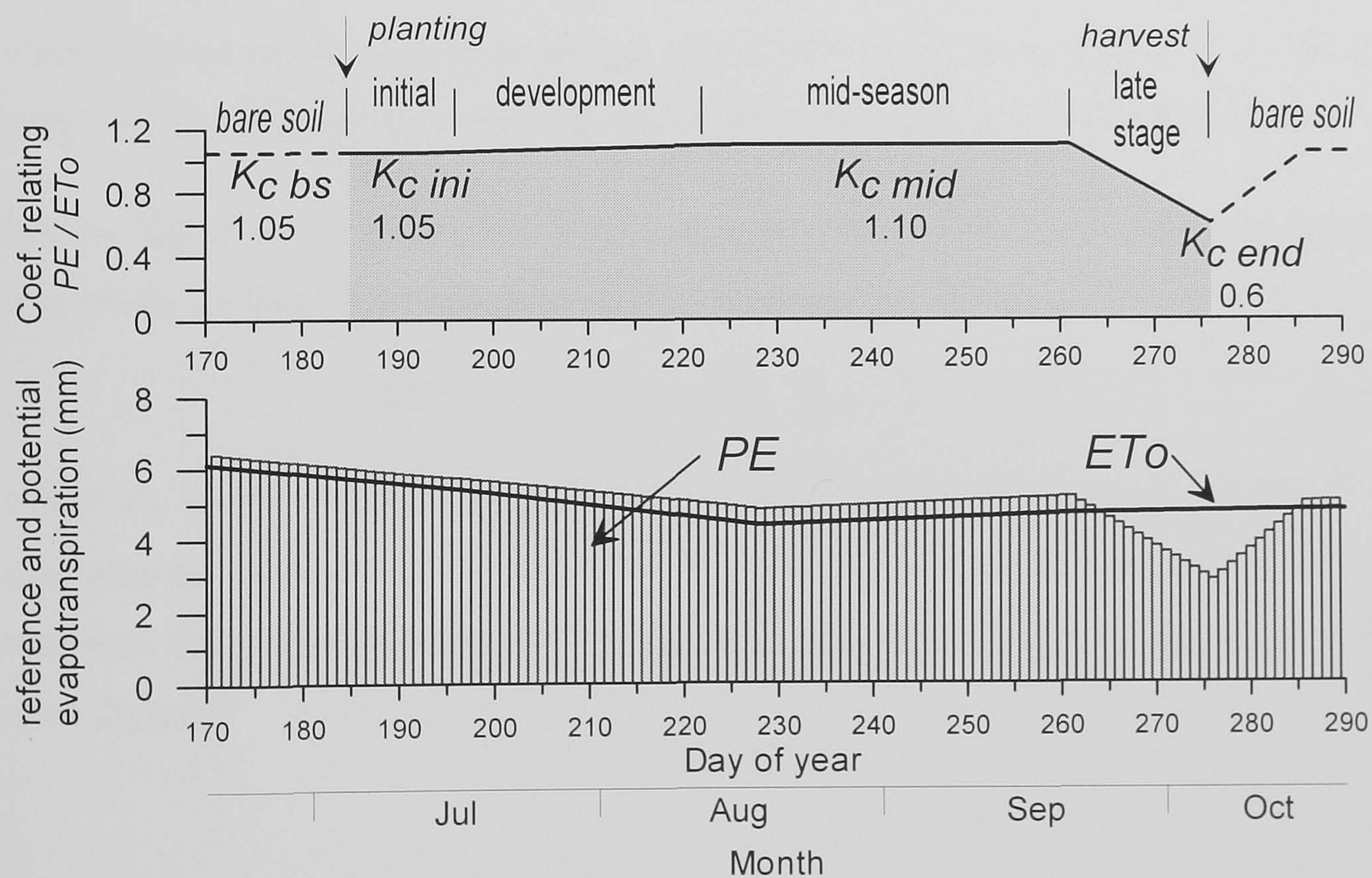


Figure 3. 3. Daily potential evapotranspiration PE calculation for a crop season

3.5. OVERALL PROCEDURES FOR ACTUAL EVAPOTRANSPIRATION DETERMINATION

The SAMBA model estimates actual evapotranspiration using a modified version of equation 2.4 (section 2.7.1). That is,

$$AE = In + K_s \cdot (PE - In) \quad 3.3$$

where K_s is the water stress coefficient representing the actual evapotranspiration decay as function of the soil moisture deficit. K_s is calculated according to the equations presented by figure 2.12 (section 2.7.2). Infiltration In represents the amount of water from rainfall which infiltrates after the runoff component has been deducted.

The root growth process that determines the depth of soil water which can be effectively utilised by the crop is visualised in figure 3.2. During the crop initial stages the root depth is taken as a shallow depth, normally 0.25 – 0.30 m, representing the soil depth from which the small seedlings effectively abstract their water. During the development period, root extension is characterised by a deeper growth until full development at the start of the mid-season when root growth ceases.

For the fallow period before the planting and after harvest, the total amount of water that can be depleted by evaporation TEW is estimated as,

$$TEW = 1000 \cdot (\theta_{FC} - 0.5 \cdot \theta_{WP}) \cdot Z_e \quad 3.4$$

where Z_e is the depth of the surface soil layer which is subject to drying by evaporation, adopted as 0.25 m. Note that a coefficient of 0.5 is introduced to represent the minimum water content as being midway between wilting point and oven dryness.

3.6. THE ACTUAL EVAPOTRANSPIRATION FOR WATER STRESS PERIODS

Actual evapotranspiration is calculated following the procedures cited in section 3.4. However, these approaches fail to fully represent crucial actual field conditions in a semi-arid area. In particular the following condition is of interest:

- When the profile water content is low (high soil moisture deficit) and a significant rainfall event occurs. Evapotranspiration should happen at potential rates during the rainfall event and on succeeding days since at least some of the excess rainfall is retained in the upper part of the soil profile. This amount of water is readily available for direct evaporation from bare soil during the periods when soil evaporation is the dominant process and readily available for the shallow roots during the crop season.

These conditions frequently occur in semi-arid areas and, therefore, there is a need to conceptually represent them.

The SAMBA model introduces an alternative procedure based on actual physical mechanisms of water redistribution within the soil profile and root water uptake. This approach considers that following a rainfall event, a proportion of the excess water is retained near the soil surface, so that it is available for extraction by the shallow roots (and/or by direct soil evaporation) instead of being immediately transferred downward to reduce the soil moisture deficit. The proportion of water retained depends on the soil hydraulic properties and varies for different types of soil, that is, less permeable soils are prone to hold a larger proportion of a rainfall event, hence more water is available for evapotranspiration in the following days.

The following diagrams illustrate schematically the adopted approach showing a situation during the crop season when the initial soil water content is low after a prolonged dry spell. Therefore, the soil moisture deficit at the start of day SMD' is high and the crop is under water stress. Immediately after a significant rainfall event (e.g. 40 mm.day^{-1}) the water content at near surface increases allowing the water uptake by the shallow roots. Thus actual evapotranspiration is at potential rates (5 mm.day^{-1} in figure 3.4)

Figure 3.4 illustrates the inputs and outputs and the actual soil water distribution within the soil profile, showing the soil water content curve at start and end of the day (based on the curves presented in figure 2.6 for a sandy soil). Note that runoff is not considered to simplify the calculations.

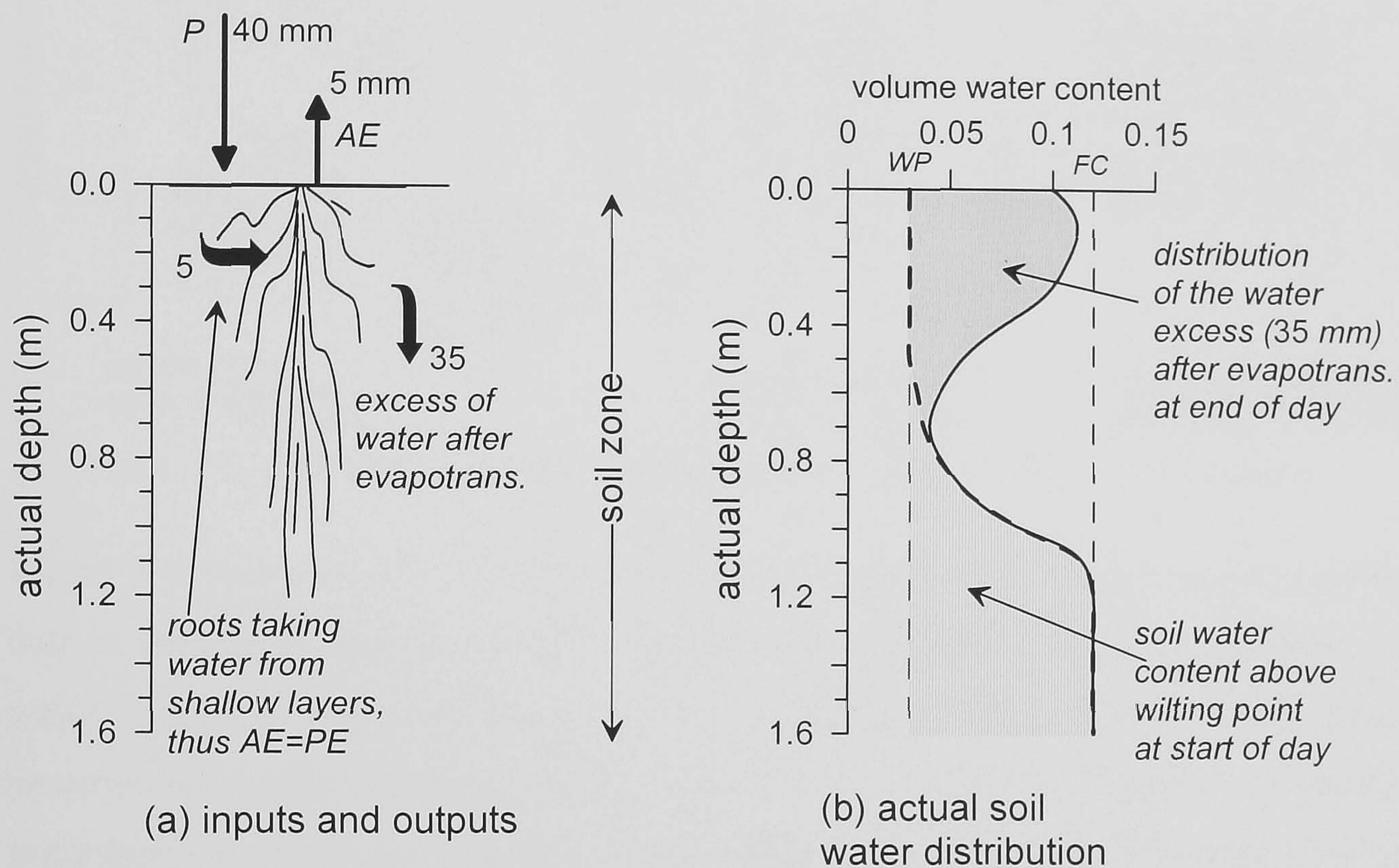


Figure 3. 4. Physical situation after a day with a significant rainfall (40 mm). Actual evapotranspiration equals potential (5 mm) due to the near surface storage of water. The vertical scale is a true depth.

The soil water distribution at the end of the day is not uniform due to the variation in the hydraulic conductivity and a proportion of the water excess (after AE) remains near surface.

The conventional soil water balance approach immediately transfers the surplus of water (35 mm after the roots had taken 5 mm during the rainy day) to the soil moisture deficit (see section 2.7.2).

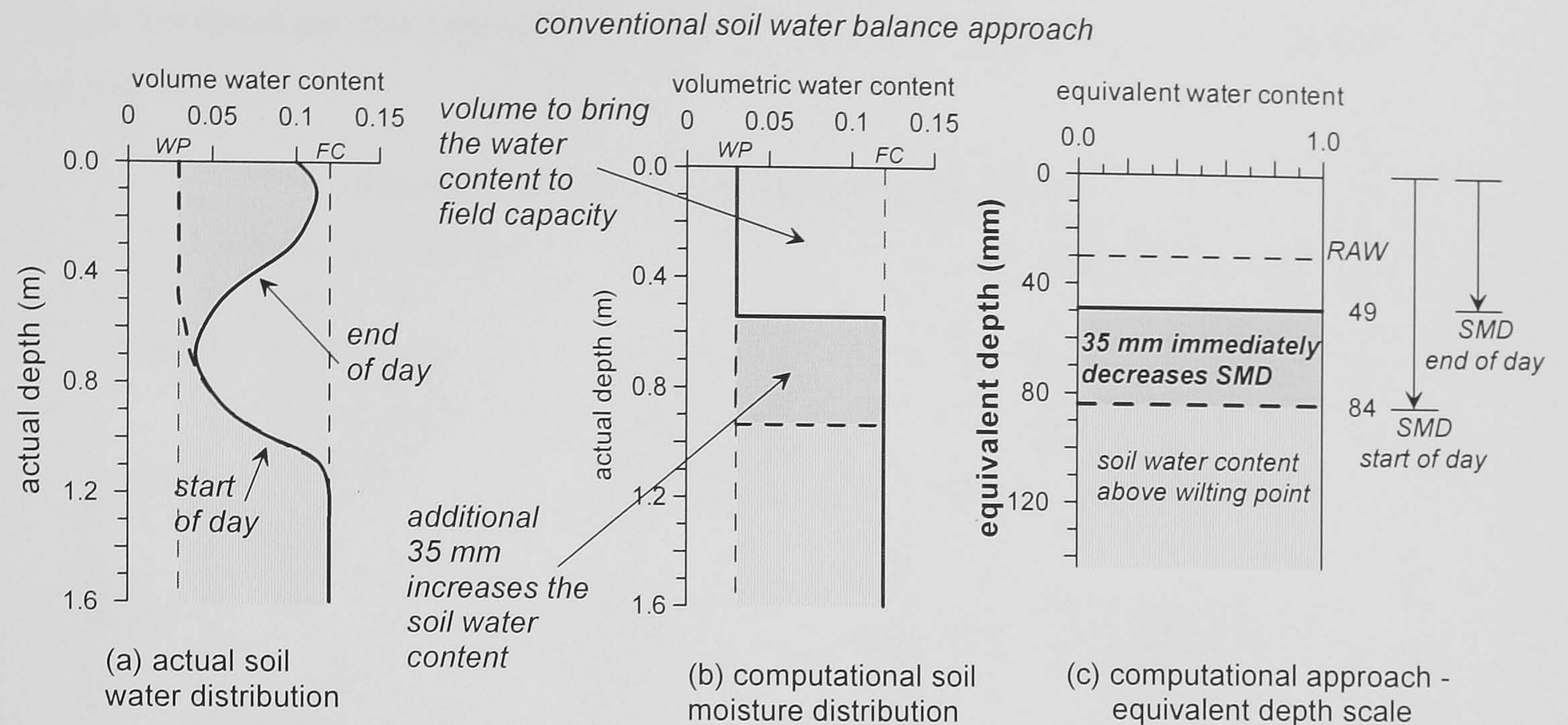


Figure 3. 5. The conventional soil water balance during a rainy day following a dry period.

Figure 3.5 shows the actual soil water profile (3.5.a) and the computational profile with a simplified step-like water distribution representing the overall volume of water related to the available water for evapotranspiration (3.5.b). Figure 3.5.c presents the computational approach in terms of an equivalent depth of water (in mm) showing the volume of water above wilting point and the soil moisture deficit at the start of day SMD' and end of day SMD . It is important to stress that SMD is an equivalent depth of water and does not relate to a real depth of soil.

The surplus of water (35 mm) brings the initial soil moisture deficit of 84 mm to 49 mm. However, the equivalent water content is still not enough to reach the amount of water readily available for evapotranspiration RAW at that particular day. RAW is equal to 30 mm assuming that the root depth is 0.55 m during the crop development stages (depletion factor p equal to 0.5). In the absence of another significant rainy day, actual evapotranspiration in the following day is at reduced rates. However, this conventional approach misrepresents the actual physical conditions as previously mentioned.

The SAMBA model utilises an alternative approach for the conditions cited at the beginning of this section. A proportion of the infiltrated water goes to near surface storage from where it is available for evapotranspiration in the following days, while the surplus reduces the soil moisture deficit.

Figure 3.6 illustrates the approach showing the same representative day of figures 3.4 and 3.5.

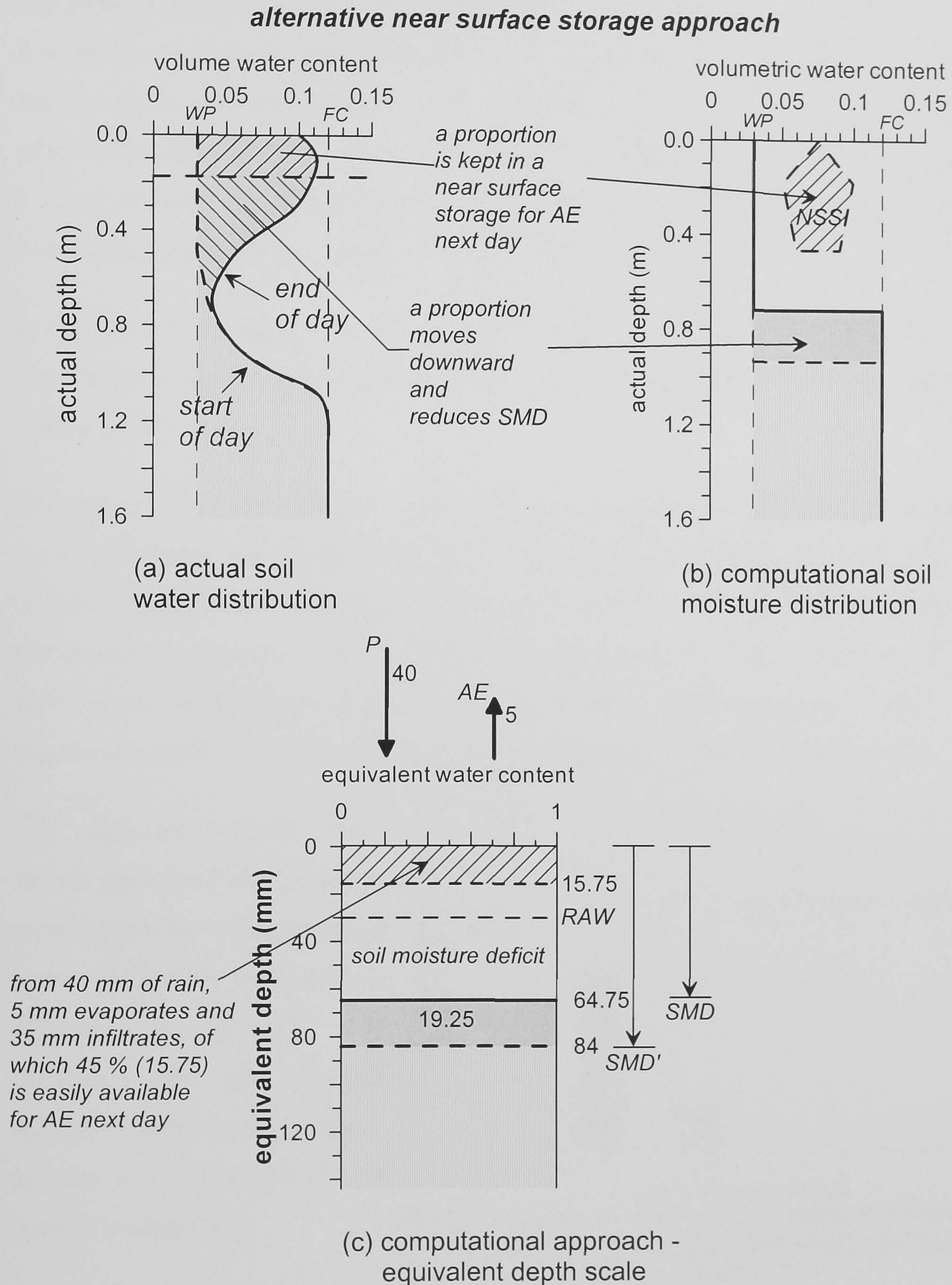


Figure 3. 6. The alternative near surface storage approach included in the *SAMBA* model in order to estimate *AE* following a heavy rainfall when *SMD'* is bigger than *RAW*.

After a day with significant rainfall (40 mm) and potential evapotranspiration of 5 mm, a proportion of the water surplus ($40 - 5 = 35$ mm) is assumed to be kept in near

surface storage. For instance, assuming a proportion of 45%, about 16 mm ($35 \text{ mm} \times 0.45 = 15.75 \text{ mm}$) goes to the near surface storage *NSS* while the remaining 19.25 mm reduces the soil moisture deficit, moving the initial deficit from 84 to 64.75 mm. Note that actual evapotranspiration during the rainy day is at potential rates (5 mm) because water is readily available near surface for the shallow roots during and just after infiltration. The near surface storage is represented as a pentagon in figure 3.6(b) in order to differentiate it from a pre-fixed sub-layer as usually utilised by soil water balance models (e.g. the BALANCE model, section 2.7.3).

In the absence of significant rain in the following day, the near surface storage is depleted by actual evapotranspiration and by the proportion of water moving down to reduce *SMD'*.

The proportion of the infiltrated water that goes to the near surface storage is based on a preliminary association with the soil texture type. That is, the value of 0.45 utilised in figure 3.6 and for several examples throughout this thesis was chosen to represent the water movement in a permeable sandy soil. The *NSS* proportion for fine textured soils with lower infiltration rates (see table 2.1) is progressively bigger, e.g. suggested values for loamy soils and clayey soils are 0.55 and 0.75 respectively.

This approach assures that a certain amount of water remains near surface for a longer period in less permeable soils such as illustrated by figure 3.7. For instance, the near surface storage is successively reduced to zero over a period of 1-2 days in a sandy soil.

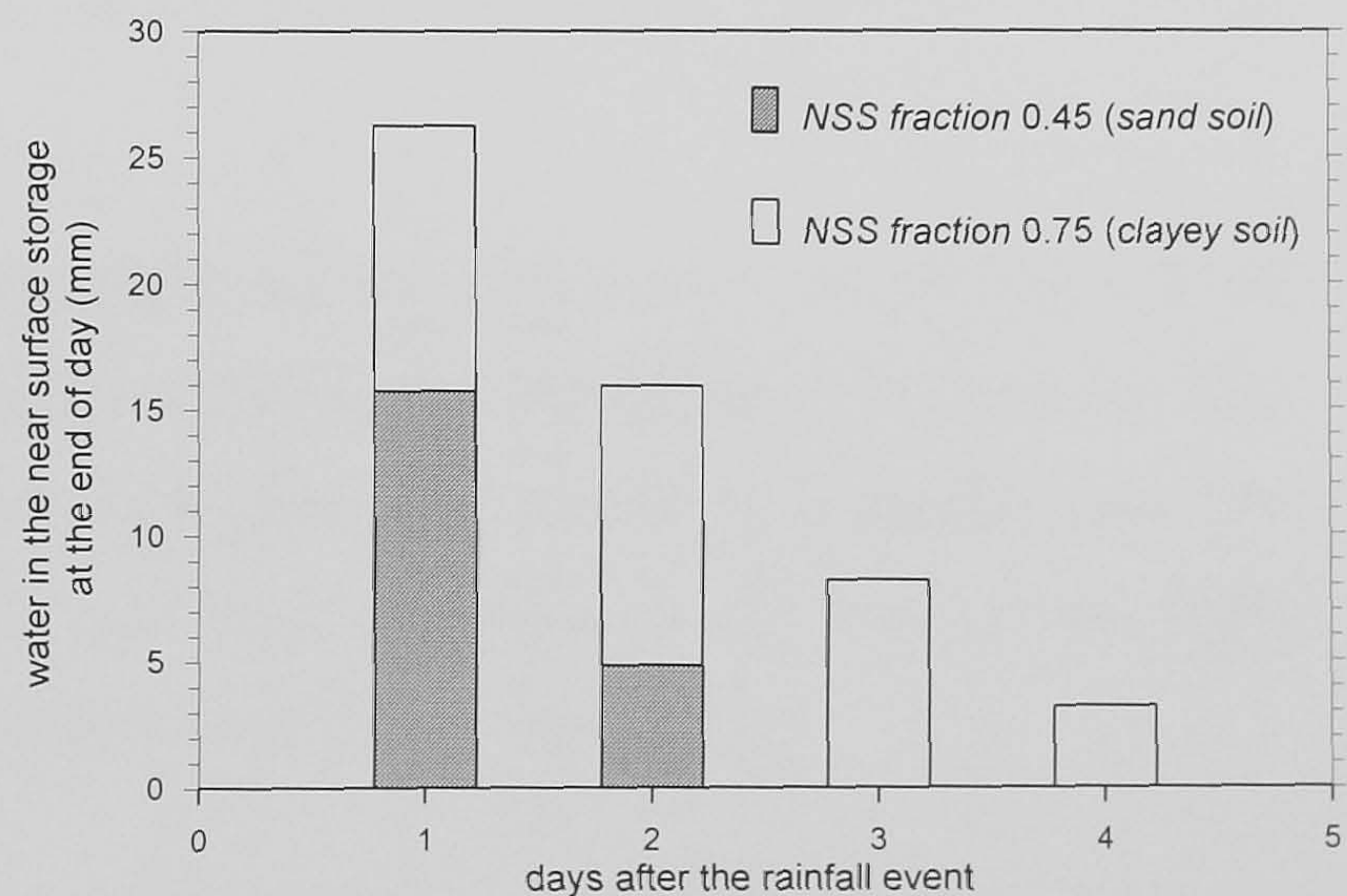


Figure 3. 7. *NSS* storage after a rainfall event of 40 mm and actual evapotranspiration 5 mm for two different values associated with soil type.

However, it is necessary to stress that the values adopted for the *NSS* fraction are preliminary and further investigation must be made (see section 7.5). In section 5.2.3 different values of *NSS* fraction are compared to field measurements from a semi-arid plot experiment. The effect of this alternative approach on the overall soil water

balance, and hence on recharge estimation, is presented in Chapter Four using real data from a semi-arid region.

3.7. SOIL WATER BALANCE FOR A SERIES OF CLIMATIC AND SOIL MOISTURE CONDITIONS

Actual evapotranspiration is estimated based on the current soil moisture conditions calculated by a daily soil water balance. The soil water balance input is the infiltration (depending on precipitation and runoff) and the main output is actual evapotranspiration. A series of different conditions can occur regarding the soil moisture deficit and the soil water availability for evapotranspiration. They are:

- i. When water is easily extracted by the roots from the soil storage (soil moisture deficit is less than the readily available water RAW);
- ii. When the crop is under water stress (soil moisture deficit greater than RAW);
- iii. A particular situation after harvest or at start of the rainy season when the soil moisture deficit is greater than the total available water for evapotranspiration (soil moisture deficit greater than readily evaporable water REW or total evaporable water TEW).

A series of soil water balances are illustrated in figures 3.7 – 3.9. In each of the figures, rainfall P (25 mm), runoff Ro (calculated according to the section 3.3), infiltration In ($P - Ro$) and potential evapotranspiration PE (5 mm) are the same. The difference is the magnitude of the soil moisture deficit at the start of day SMD' relative to the readily and total available water (RAW and TAW).

3.7.1. Soil moisture deficit less than RAW

Figure 3.8 shows the situation when the soil moisture deficit at start of the day SMD' is less than RAW (or REW in bare soil conditions), hence water is readily extracted to supply the evaporative demand of the atmosphere ($K_s = 1.0$, see figure 2.12). Consequently, actual evapotranspiration is always at potential rates (equation 3.3

with K_s equal 1.0). The variation in the soil moisture deficit depends on the occurrence and magnitude of rainfall.

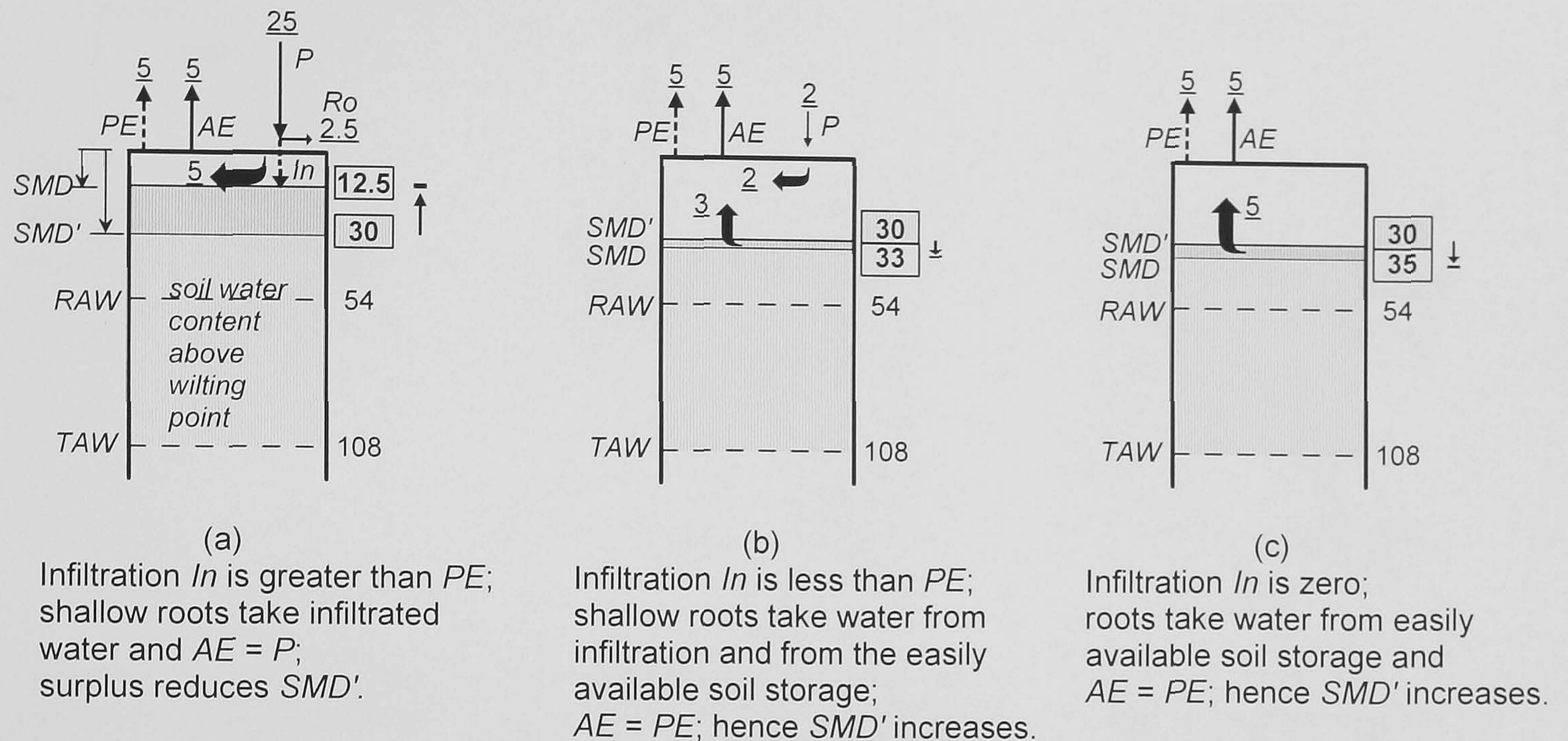


Figure 3. 8. Actual evapotranspiration when $SMD' < RAW$. All values in mm/day

3.7.2. Soil moisture deficit greater than RAW

Figure 3.9 illustrates the situations when the soil moisture deficit is greater than the readily available amount of water for evapotranspiration and the crop is under water stress. This condition often occurs in semi-arid areas due to by prolonged dry spells during the crop season.

Although the crop is under water stress, when infiltration is greater than potential evapotranspiration, the shallow roots can meet the evaporative demand (figure 3.9.a). In this case K_s is equal to 1.0 and from equation 3.3, $AE = PE$. Note that the computational concept of near surface storage is not fully presented, in order to simplify the numerical example. In fact, it is partially represented by the utilisation of the infiltrated water by the shallow roots.

When $PE > In$ (figure 3.9.b), infiltration is readily transpired by the shallow roots but the remaining evapotranspiration demand ($PE - In$) is only met at reduced rates. In this case K_s applies to the proportion of water withdrawn by roots from the stressed soil. When infiltration is zero (figure 3.9.c), AE is a K_s proportion of the remaining total available water.

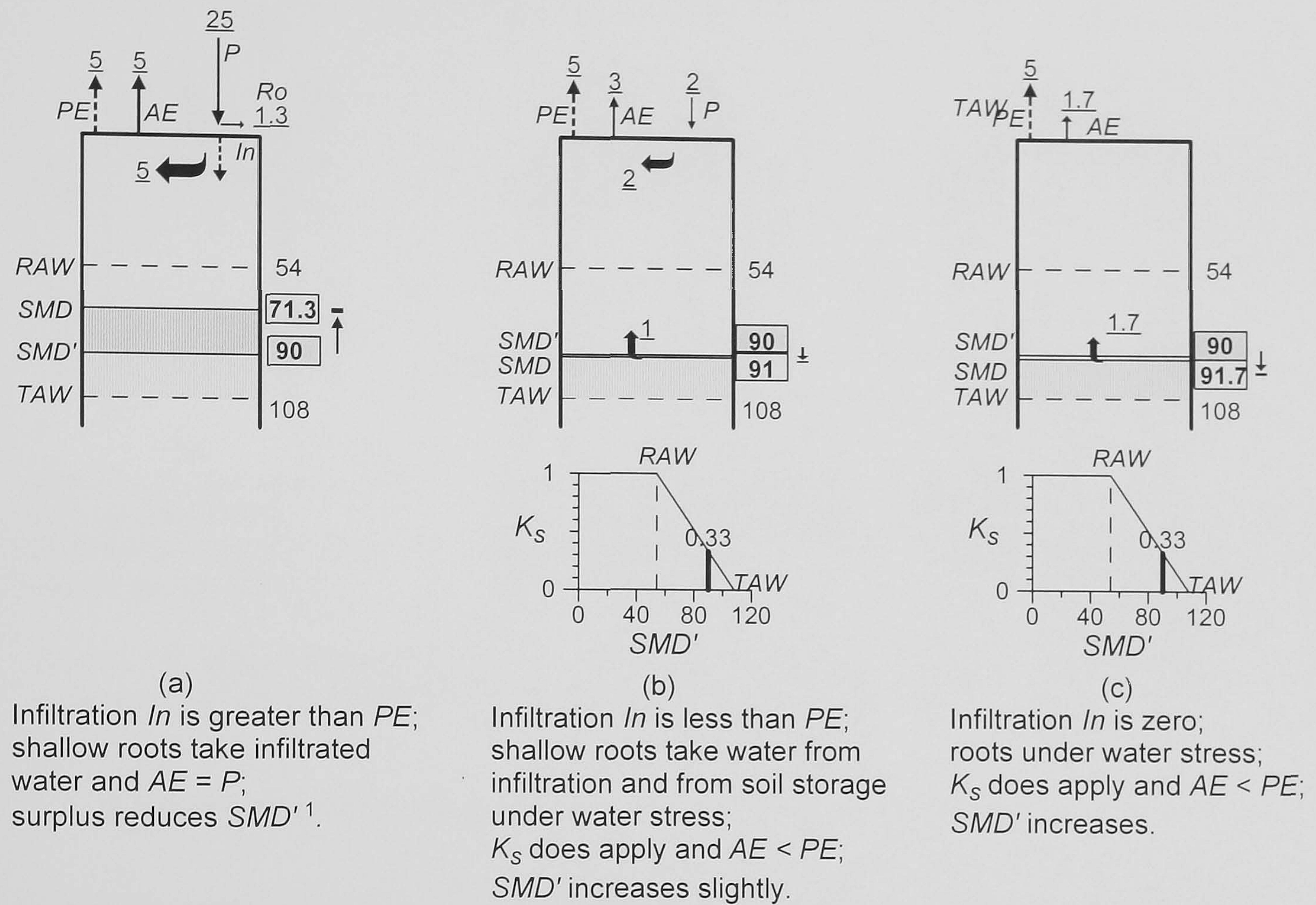


Figure 3. 9. Actual evapotranspiration when $SMD' > RAW$. All values in mm/day. ¹ The computational concept of near surface storage was not fully applied in order to simplify the example (see text).

3.7.3. Soil moisture deficit greater than TEW

Figure 3.10 shows a particular situation when the soil moisture deficit is significantly greater than the amount of water potentially available for evapotranspiration. It might occur: (i) during the dry season when the soil profile is very dry and (ii) after harvest, when SMD might remain high.

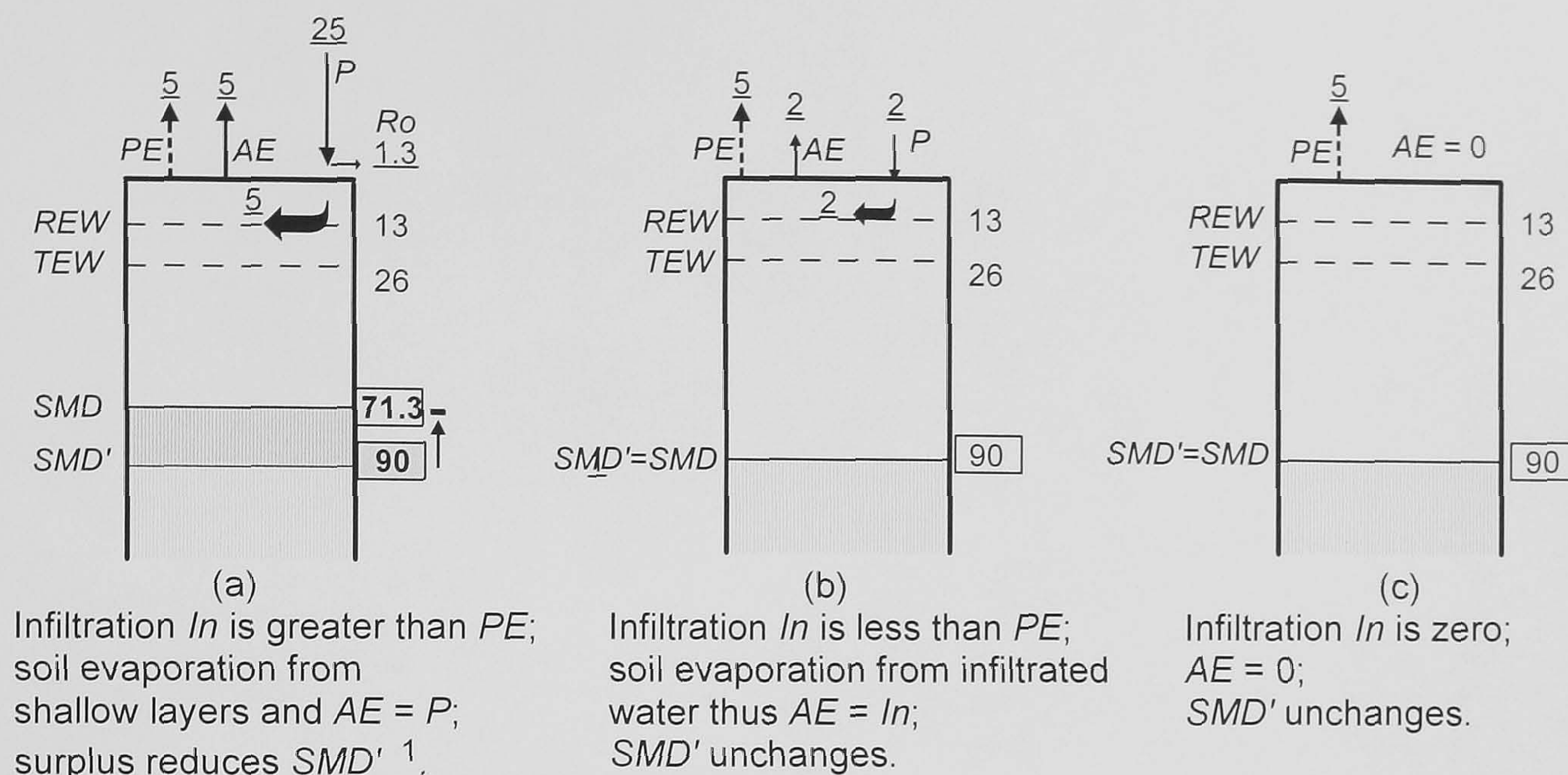


Figure 3.10. Actual evapotranspiration when $SMD' > TEW$. All values in mm/day. ¹ The concept of near surface storage was not fully applied in order to simplify the example (see section 3.7.4).

3.7.4. Near surface storage NSS

In figures 3.9(a) and 3.10(a) the computational concept of near surface storage was not fully applied in order to simplify the numerical examples. They did not show the resulting soil water balance for the following days. Figure 3.11 shows the resulting computational water balance for the situation of figure 3.9(a) showing three consecutive days following a significant rainfall of 25 mm. A similar water balance would occur in the situation of figure 3.10(a).

A rainfall fraction going into near surface storage equal to 45% was adopted as discussed in Section 3.6. Note that actual evapotranspiration is at potential rates in the following day, which would not be possible adopting directly a conventional soil water balance technique. For the following days, AE gradually decreases. On day 2, water readily available from the near surface storage allows AE to equal PE . On day 3, AE is equal to the remaining water in the near surface storage plus the fraction of water from the soil (under water stress). For the following days, in the absence of rainfall AE is reduced by the stress coefficient K_s .

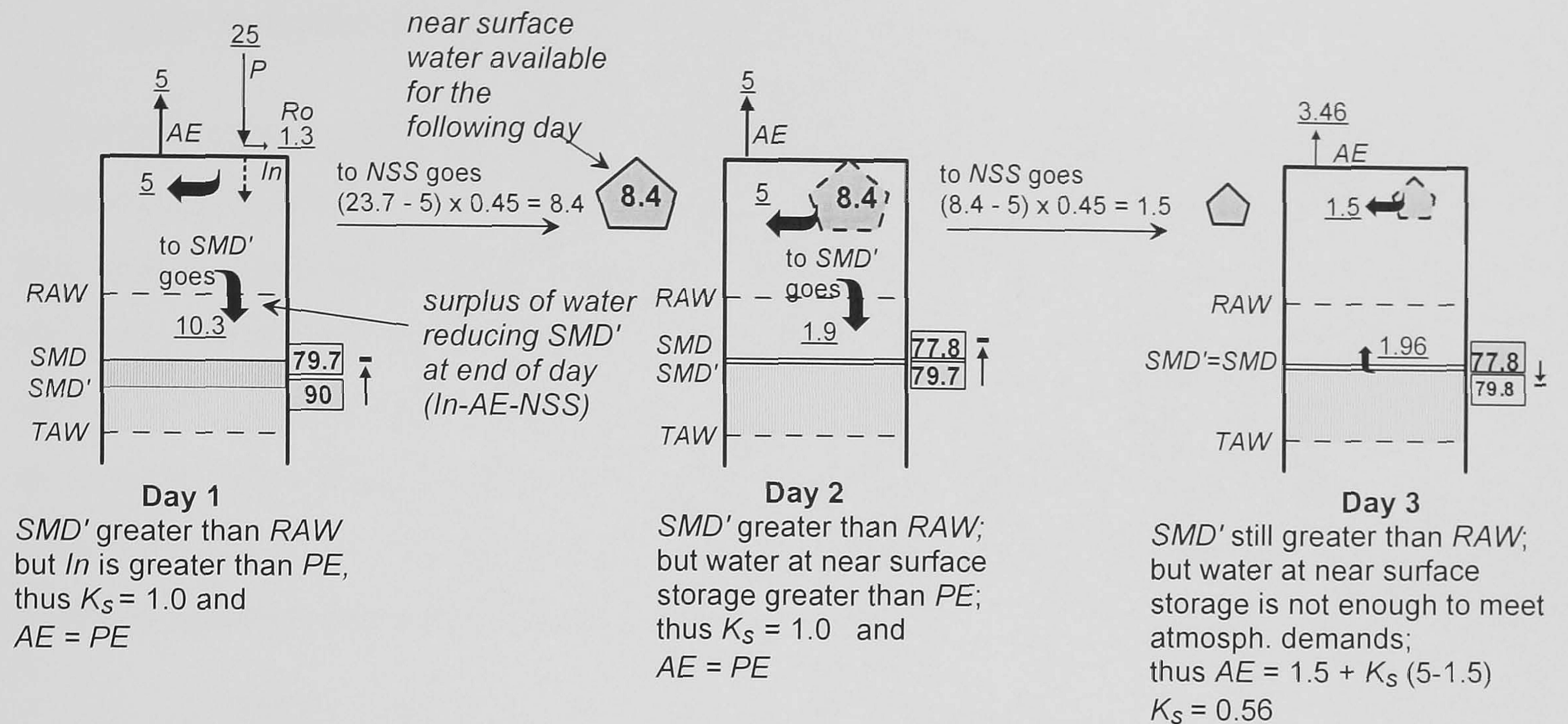


Figure 3. 11. Example of daily water balance including the near surface storage approach. All values in mm/day.

Table 3.2 shows the calculations using two alternative NSS fractions for the situation when $SMD > TEW$, thus AE is zero when rainfall or the near surface storage is also zero. A smaller fraction represents more permeable soils.

The variable AWE in the fourth column is the water available for evapotranspiration at the start of the day, that is, AWE is the sum of the infiltrated water at start of day plus the near surface storage at start of day NSS' . Thus,

$$AWE = In + NSS'.$$

		SMD'		fraction		fraction	
		90 mm		0.45		0.75	
day	In	PE	AWE	AE	NSS'	NSS	SMD
1	25.0	5.0	25.0	5.0	0.0	9.0	79.0
2	0.0	5.0	9.0	5.0	9.0	1.8	76.8
3	0.0	5.0	1.8	1.8	1.8	0.0	76.8
4	0.0	5.0	0.0	0.0	0.0	0.0	76.8
5	0.0	5.0	0.0	0.0	0.0	0.0	76.8
		Total		11.8		Total	
						16.9	

Table 3. 2. Near surface storage for actual evapotranspiration estimation (all values in mm/day)

Note that for a larger NSS fraction, more water is evaporated, over a longer period.

Chapter four investigates the effect of the inclusion of near surface storage on recharge determination using real data from a semi-arid area for selected periods.

3.7.5. Potential recharge

Potential recharge occurs when infiltration brings the soil moisture deficit to a negative value, that is, the soil water content becomes greater than field capacity. At this point, soil is assumed to be free draining hence all excess water drains through the bottom of the soil zone. Therefore, SMD at the end of the day is set to zero and the negative value becomes a positive recharge. Figure 3.12 illustrates the calculation of recharge when a rainfall of 25 mm brings SMD' at end of the day to a negative value of 5 mm. It is assumed that 5 mm will drain through the soil and the soil moisture deficit at start of the next day is zero.

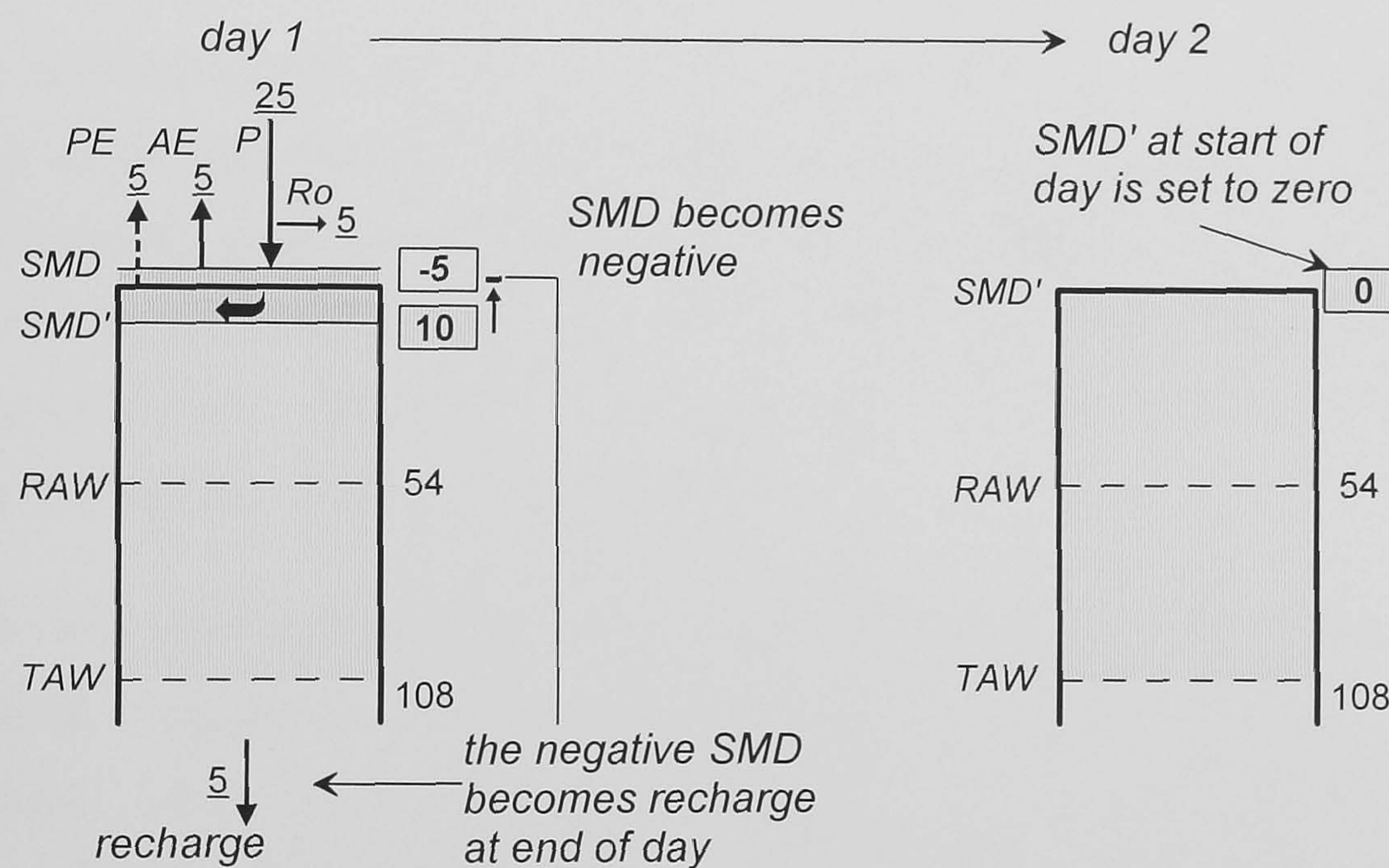


Figure 3. 12. Soil water balance and the resulting potential recharge.

When SMD reaches and then exceeds field capacity, computationally, the near surface storage is also considered as recharge and added to the negative SMD . In this case, NSS at the end of the day is set to zero to avoid the misconception of keeping water within the soil zone above field capacity.

3.8. THE SPREADSHEET SOIL WATER BALANCE MODEL

The computational algorithms for the soil water balance are presented in Appendix A. The implementation of these algorithms in a spreadsheet model is described in Appendix B.

3.9. COMPARISON BETWEEN THE METHODOLOGY PRESENTED IN THIS STUDY AND ALTERNATIVE SOIL WATER BALANCE TECHNIQUES

This section outlines the main features of the soil water balance method described in this study in comparison to the alternative models presented in the section 2.7. The objective is not to carry out a numerical comparison but to stress the similarities, and differences, between the main model components.

3.9.1. The conventional single layer model

The SAMBA model has several points of similarity with the conventional single layer model (the Penman-Grindley approach, section 2.7.1). For instance, Penman and Grindley represent the soil profile as a single reservoir from which the roots draw water to meet the evaporative demand. The concept of soil water stress is also adopted in both models and the algorithms and equation for estimating actual evapotranspiration (equation 3.3) are similar.

However, the SAMBA model has adopted a series of procedures that expand the physical representation of key physical processes. For instance, the root growth process, the concept of water availability for evapotranspiration and the determination of water stress period are more physically realistic than the highly empirical concept of root constant utilised by the conventional technique. Moreover, unlike the Samba model, the Penman-Grindley approach does not explicitly include the soil properties.

The major difference between the models is the inclusion of procedures for representing crucial processes in semi-arid areas. The concept of near surface storage provides the SAMBA model with a better representation of the real conditions in semi-arid areas. Moreover, the SAMBA model takes runoff into account while the conventional technique does not attempt to estimate it.

In fact, the conventional soil water balance was designed for different climate (temperate climate) where the crop is not expected to be under extreme water stress conditions such as in semi-arid regions. Therefore, there was a need to include new

aspects and procedures to expand the conventional technique to wider range of applications.

3.9.2. The CROPWAT model

Similar to the discussion above, there are several common points between the FAO-CROPWAT model and the SAMBA model. The approach for determination of potential evapotranspiration strongly follows the FAO method (section 3.4). The concepts of total and readily available water, root growth, crop growth stages and crop coefficients are similar for both models.

However, there are important differences between the models. First of all, the CROPWAT model was originally developed for estimating water requirements of irrigated crops during a crop season. Therefore, the expected soil moisture conditions are closer to the optimum conditions, with a well watered crop transpiring at potential rates during the majority of the crop period.

For that reason, the periods of water stress are not important in terms of the overall soil water balance and they are represented by the simplistic relationship between a stress coefficient (K_s) and potential evaporation. For instance, no attempt is made to represent the daily soil water balance for a heavy rainfall when soil moisture deficit is high. The SAMBA model is therefore more realistic as it adopts the algorithms and equations described in section 3.7, and the concept of near surface storage.

The concept of crop coefficients for the periods when soil evaporation is the dominant process is another contrasting issue. The FAO technique is based on the knowledge of the “time-averaged” magnitude and interval between wetting events occurring over a period of a week or longer. However, at a shorter time scale this procedure fails to represent the process of soil evaporation when sporadic and highly intense rainfall occurs. Moreover, crop coefficients are utilised by the SAMBA model for determining the maximum potential evaporative demand. The actual evapotranspiration is a function of the soil water availability. On the other hand, the FAO approach, including the most recent proposals by Allen et al. (1998), attempts to directly determine actual evapotranspiration using the crop coefficients, since the irrigated crop is not under significant stress.

The CROPWAT model does not consider the non-cropped period since the initial conditions are pre-fixed. However, this period is crucial for estimating recharge as is demonstrated in chapter four. Also, no attempt is made to partition the surface runoff and deep drainage components.

3.9.3. The BALANCE model

The BALANCE model (section 2.7.3) is a comprehensive model in the sense that it integrates climatic, soil and vegetation factors for estimating the daily soil water balance of cropped or fallow surfaces. The model considers the period between crops which allows for the model run of an inter-related sequence of years.

The main differences in relation to the SAMBA model are the estimation of soil evaporation and crop transpiration as separate processes in a crop plot and the subdivision of the soil profile into three layers. These approaches include more complexity and the consequent needs to determine the model parameters such as the constant of soil evaporation in Ritchie's equation.

The concepts of water distribution within the soil profile are based on two different approaches. The BALANCE model considers the water distribution as a step-like process, that is, water movement to lower layers occurs according to an exponential function after the water content reaches field capacity. It includes a depth fixed top sub-layer that needs be at field capacity before water moves to the root zone. The SAMBA model does not include a top layer since the concept of near surface storage represents the initial near surface wetting stage without the need to define a pre-fixed depth layer. This depth is difficult to determine in the field and can vary during the season due to rainfall intensity and crop stages.

3.9.4. The Four Root Layers Model (FRLM)

- The main characteristics of the FRLM model are its conceptual representation of root growth and root distribution in four fixed depth layers. It assumes that the contribution of each layer to actual evapotranspiration depends on the proportional root distribution within the soil profile. The version of the FRLM model presented by Ragab et al. (1997) does not consider the root growth process since it is designed for

permanent grass surface application. Finch (1998) has proposed the inclusion of root growth as a function of time for cropped surfaces, varying the proportions of the root distribution. However, the division of the root zone into four static layers and the empirical definition of how much each layer is contributing to the water uptake seem to not have a clear experimental basis. As cited in section 2.2.6, roots are able to draw water from the soil independently of their distribution. When the upper layers approach the wilting point, the deeper roots increase their uptake capacity in order to meet the evaporative demands.

The application of the FRLM model in semi-arid climates would clearly underestimate actual evapotranspiration because the approach reduces actual evapotranspiration each time the soil water content is below field capacity in any layer. Moreover, no attempt is made to estimate evaporation from a bare soil surface during a non-crop period.

CHAPTER FOUR

SELECTED MODEL OUTPUTS

Chapter four presents the main characteristics of the SAMBA model outputs, stressing the features which have been introduced to improve the physical credibility of the soil water balance technique for estimating recharge in semi-arid areas.

4.1. INTRODUCTION

The SAMBA model incorporates new alternative procedures which improve the representation of the real physical processes in a semi-arid area where the rainy and crop season are coincident. The purpose of Chapter Four is to investigate how these processes, and others key aspects presented in the model, perform in a daily soil water balance using real field data.

The analysis is based on selected model outputs. First, the individual components of the model are presented for a representative year. In the second part of this chapter the computational procedures for key aspects of the model are discussed with emphasis on how they represent actual field conditions. These aspects are: (i) the correct selection of crop planting dates, (ii) the estimation of surface runoff and (iii) the introduction of near surface storage. Finally, an investigation of the overall effect of these procedures on potential recharge is made using selected representative periods of days and years from a field data set.

4.1.1. Field data

The data utilised for this analysis are from a semi-arid region located in the Sub-Saharan area of Africa (figure 4.1). The climatic data is from the meteorological station of Nguru (12° 53'N, 10° 28'E, alt.343 m) situated in the Northeast region of Nigeria. The region lies in the Quaternary Chad Formation sediments of the Yobe river basin

system which consist of sands and clays, overlain by superficial aeolian and alluvial materials.

Rainfall in Northeast Nigeria has a unimodal distribution with the rainy season starting in May/June, peaking in August and finishing in September/October. The rest of the year is effectively dry. The regional mean annual rainfall is approximately 500 mm. The mean annual rainfall at Nguru for the period of 1962 to 1997 is 431 mm (see Appendix C for a detailed presentation of the climatic data set).

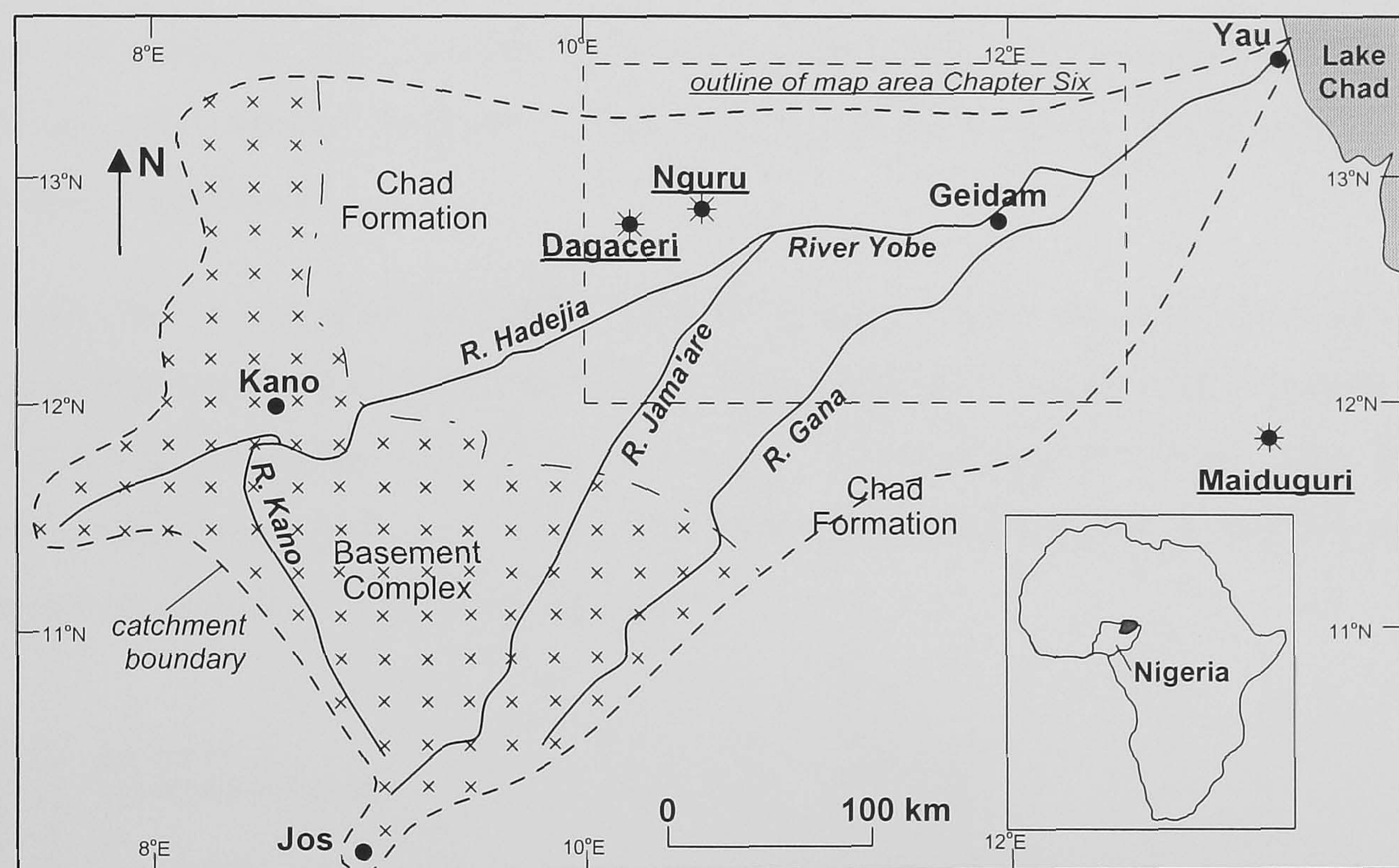


Figure 4. 1. Yobe river system. Northeast semi-arid zone of Nigeria. Data from the underlined localities are utilised in this research.

The values of daily reference evapotranspiration ET_o utilised are derived from a long-term monthly mean calculated by the Penman- Monteith equation for this region (Carter, pers comm.).

The soil and vegetation properties were estimated based on a typical Sahelian sandy soil and a millet crop respectively. Millet is grown extensively throughout the region and it constitutes one of the main crops in Northeast Nigeria (Hess and Grema, 1994). Moreover, a significant conversion from savannah vegetation to millet cultivation has occurred in the sub-Saharan region with substantial effects on the water balances (Bromley et al., 2002).

4.2. MODEL INPUTS AND OUTPUTS FOR A REPRESENTATIVE YEAR

The SAMBA model involves a series of inputs and outputs regarding the principal hydraulic components of the soil water balance. The following figures illustrate the components of the soil water balance for a whole representative year (1964) including the period before and after the crop season.

4.2.1. Rainfall

Figure 4.2 shows the daily rainfall collected at Nguru climatic station for the year of 1964 and the resulting estimated surface runoff calculated following the approach described in Section 3.3.

The year 1964 is representative of the overall long-term series in the sense that it shows the unimodal rainfall distribution starting in May, peaking in August and finishing in the middle of September (figure 4.2). There is no rain during the rest of the year. The year is a relative wet year with a total of 536 mm (above the long-term average of 431 mm) and six rainfall events of over 30 mm.

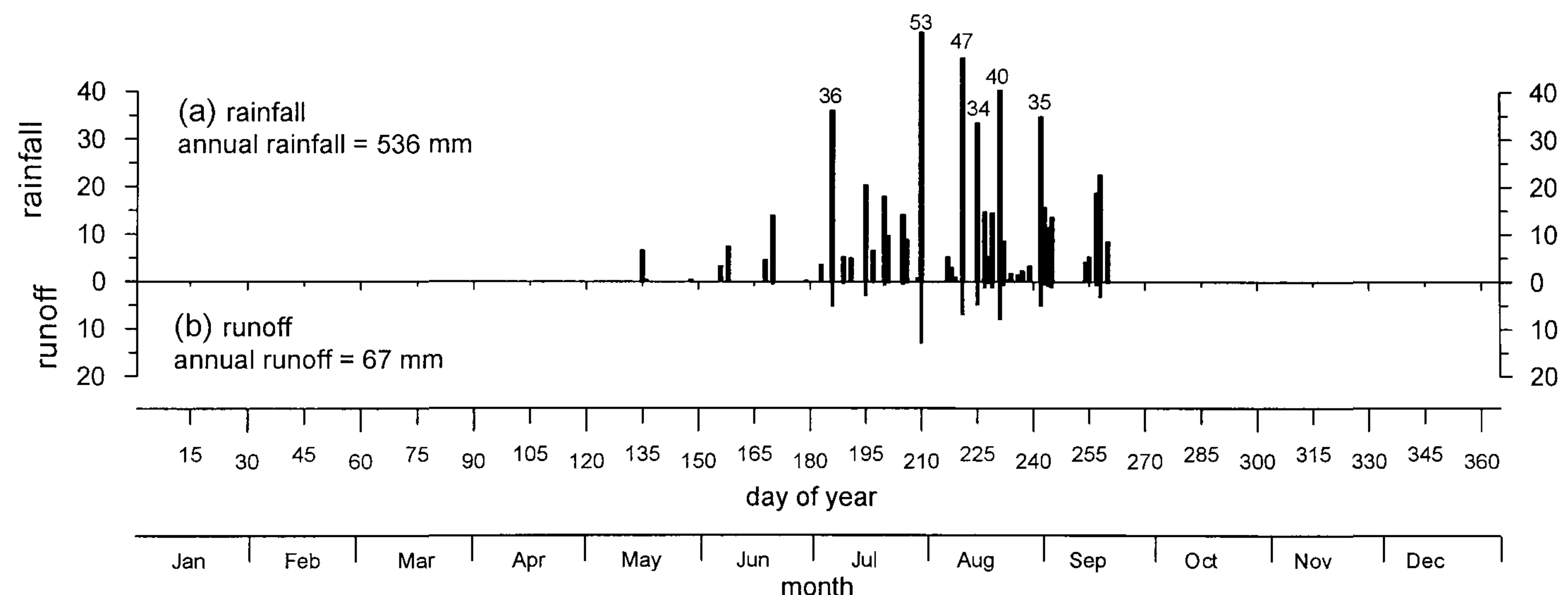


Figure 4. 2. Daily rainfall and estimated runoff from the SAMBA model for the representative year of 1964 at Nguru, Nigeria. Values in mm.

The effect of rainfall as the main input to the soil water balance calculated by the SAMBA model is illustrated in the following graphs.

4.2.2. Surface runoff

Runoff is presented in figure 4.2 immediately below rainfall. The calculated annual runoff figure for 1964 is 67 mm (12% of the annual rainfall). The diagram shows that

the estimated runoff during the early rainfall events is insignificant due to the high soil moisture deficits. However, as the soil becomes wetter and heavy rainfall occurs, runoff becomes more frequent. Runoff is calculated using the runoff coefficients from table 3.1.

4.2.3. Near surface storage

The water held near the soil surface at the end of the day and made available for evapotranspiration at start of the following day is shown in figure 4.3. The near surface storage is calculated using a fraction of 45% of the infiltrated water. Note that during the rainy season a significant amount of water is initially held in the upper part of the soil profile instead of being immediately redistributed. The effect of the near surface storage during different conditions is investigated later in this chapter.

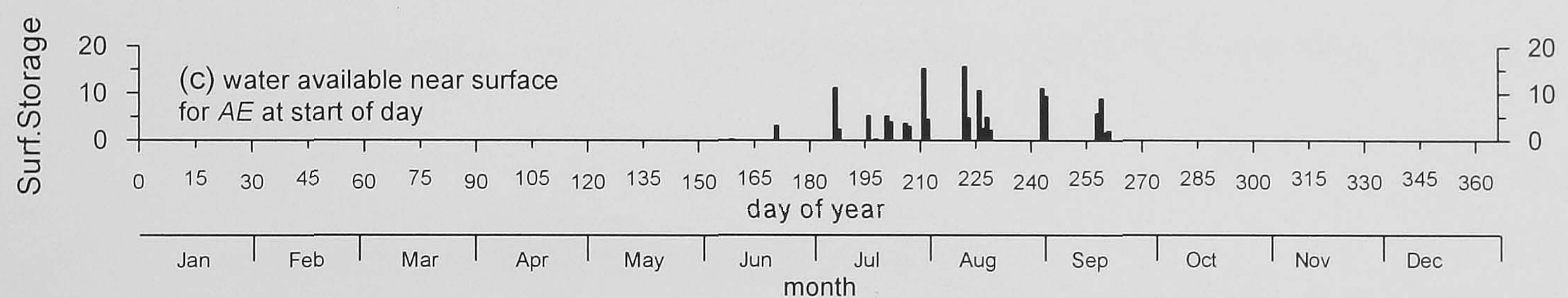


Figure 4. 3. Estimated water available near surface from the SAMBA model . 1964, Nguru, Nigeria. Values in mm.

4.2.4. Potential and actual evapotranspiration

Figure 4.4 shows the estimated potential and actual evapotranspiration (PE and AE respectively) calculated according to the procedure described in Sections 3.4, 3.5 and 3.6.

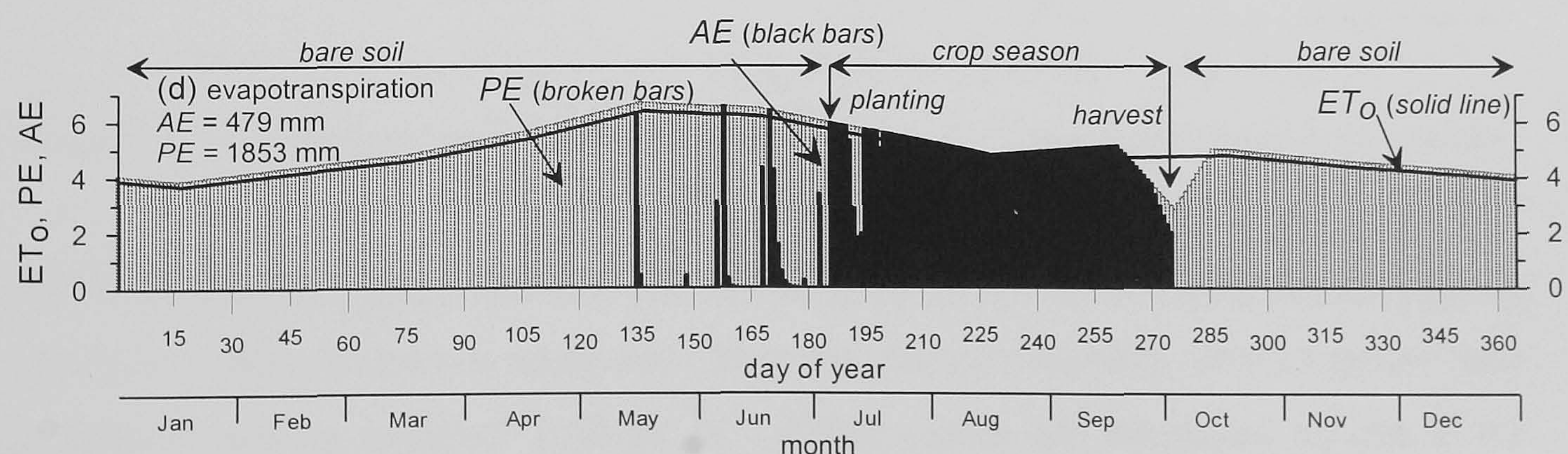


Figure 4. 4. Estimated potential and actual evapotranspiration from the SAMBA model. 1964, Nguru, Nigeria. Values in mm.

The atmospheric demand represented by the reference evapotranspiration ET_o varies between 4 and 6 mm each day, reaching a maximum in May, after the spring equinox. During the rainy season (summer) ET_o decreases slightly.

Potential evapotranspiration is estimated for the two principal soil surface conditions (crop season and bare soil) using the crop parameters based on the studies of Kowal and Kassam (1978), Grema and Hess (1994) and Abdulmumin and Misari (1990) for millet in Northeast Nigeria. It was assumed that the millet crop was planted early July with a growing season of 90 days. The crop coefficients utilised are: (i) $K_{c\ mid}$ equal 1.1; (ii) $K_{c\ end}$ equal to 0.6, regarding the relative soil wetness caused by late rains; and (iii) $K_{c\ ini}$ equal to the soil evaporation coefficient $K_{c\ bs}$ of 1.05 as discussed in Section 3.4.2. The value of 1.05 is from the FAO publication 56, example 31 (Allen et al., 1998).

Note that assuming a coefficient $K_{c\ ini}$ equal to $K_{c\ bs}$, the actual evapotranspiration during rainy days at the times when soil evaporation is dominant (May, June and early July) is at the potential rates.

4.2.5. Soil moisture deficit and available water

Figure 4.5 shows the depth of water available for evapotranspiration and the resulting soil moisture deficit.

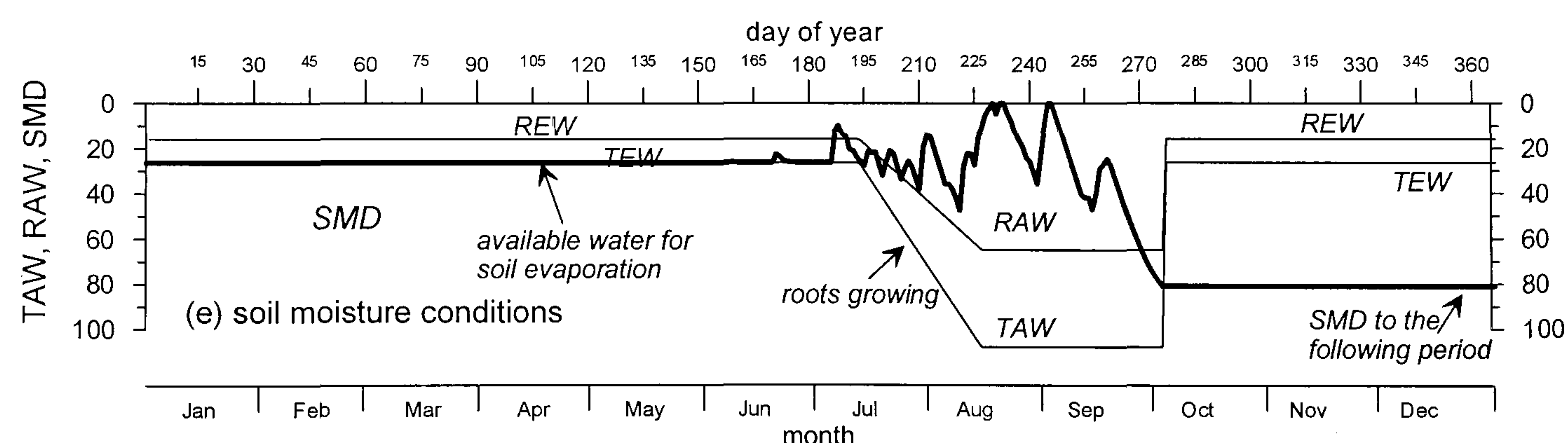


Figure 4. 5. Estimated soil moisture deficit and available water for evapotranspiration from the SAMBA model. 1964, Nguru, Nigeria. Values in mm.

The total available water TAW and the total evaporable water TEW are estimated based on the soil properties for a sandy soil. They are the field capacity of $0.12\text{ m}^3.\text{m}^{-3}$ and permanent wilting point of $0.03\text{ m}^3.\text{m}^{-3}$. These values are suggested by the FAO publication 56 (Allen et al., 1998) and they are in line with the values utilised by Rockström et al. (1998) and Grema and Hess (1994). The readily available fractions

RAW/REW were calculated using a depletion factor p equal to 0.6 (Rockström et al., 1998). The root growth is visualised by the linear increase of *TAW* from approximately 8 days after planting to a maximum root depth of 1.2 m at maturity (Rockström et al., 1998; Payne et al., 1991).

Figure 4.5 shows three aspects of the soil moisture deficit: (i) with onset of rains and subsequent crop development, the soil moisture deficit changes by the evapotranspiration demands and the surplus of water from rainfall. Occasionally, soil moisture content reaches field capacity; (ii) during the late crop period, the crop relies on the residual moisture to survive and *SMD* increases; (iii) after the crop season and in the absence of rains, *SMD* remains unaltered until the following rainy season. This assumption assumes that the soil is kept bare after the crop season and the further root extraction and the downward water movement below the soil zone is not significant during the dry season. It is a basic assumption which does not consider the effects of natural vegetation and/or late-cropping periods. Chapter six presents the overall recharge results and compares to the alternative assumption considering that the soil profile reaches the permanent wilting point during the dry season.

Considering that the remaining water at end of rain season carry over from one year to the next differences in the soil water deficit affects the water balance of the following crop season, that is, the greater the post-harvest soil moisture deficit, more rainfall is required to restore the soil water deficit at the beginning of the following growing season (Kowal and Kassam, 1978). Therefore, the assumption of an unaltered *SMD* during the dry season determines the initial soil moisture conditions for the following crop season.

Figure 4.5 illustrates the adopted assumption by the difference between *SMD* before and after the crop season. The initial *SMD* in 1964 reflects the rainfall distribution of the previous year (1963). The year 1963 was wet (annual rainfall equal to 650 mm) with the rainy season extending to the middle of October. Rainfall after the harvest decreased the *SMD*, which is then assumed to remain unaltered to the beginning of the next rainy season in 1964.

Appendix E presents the water balance of the complete rainfall data series where it is possible to observe the effect of later rainfall on the initial conditions of the subsequent

year. Also, figure 4.19 illustrates this aspect for the years of 1966 and 1967. Section 6.4.1 investigates the effect of an alternative initial soil moisture assumption on recharge estimation.

4.2.6. Overall view of the water balance and the resulting potential recharge

Figure 4.6 shows figures 4.2 to 4.5 on a single graph to provide a view of potential recharge as result of the overall soil water balance. For example, a heavy rainfall of 53 mm at day 209 (Julian days) results in no recharge because the surplus water is utilised to restore the soil moisture deficit. On the other hand, lower rainfall events associated with wetter soil conditions (e.g. in the days following day 225) are enough to generate potential recharge. For the year of 1964, the estimated annual potential recharge is approximately 8% of annual rainfall.

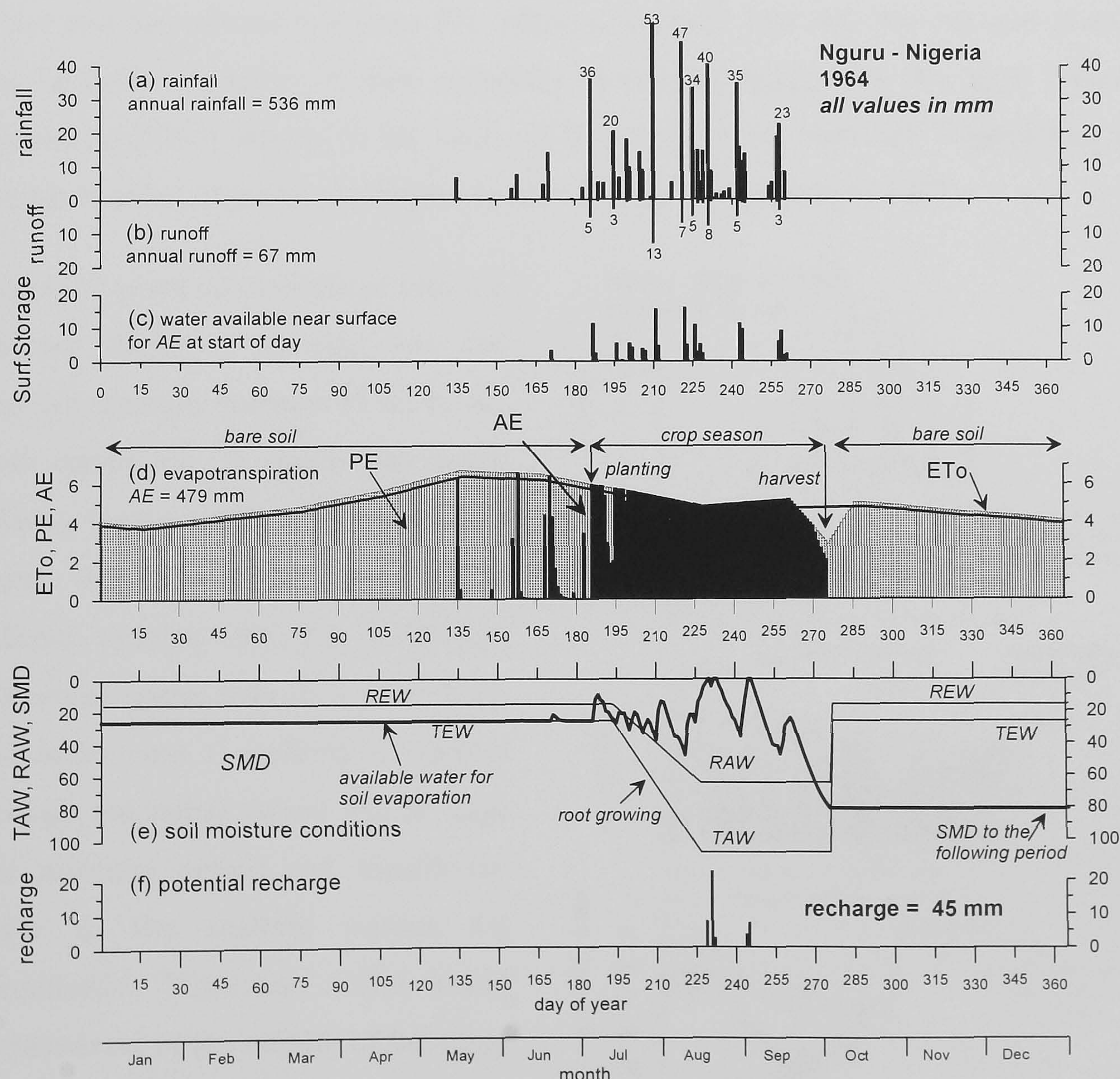


Figure 4. 6. Water balance components and the resulting potential recharge from the SAMBA model. 1964, Nguru, Nigeria. Values in mm.

After the overall view of the water balance components above, the following sections demonstrate particular characteristics of the SAMBA model using representative periods of time from the data set cited in Section 4.2.

4.3. THREE IMPORTANT ASPECTS OF THE SAMBA MODEL

4.3.1. The planting date and the onset of rains

This section points out the importance of choosing the correct planting date when investigating the soil water balance of a cropped area in semi-arid conditions.

In semi-arid areas, the establishment of successful rainfed crops depends on the perfect adjustment of the growth cycle to the seasonal distribution of rainfall (Mortimore, 1989). In fact, the time for planting can vary considerably depending on the region, labour and agricultural practices. For instance, a farmer can take the risk and plant at first rainfalls. However, it may necessary to re-plant again if a dry spell follows. Mortimore (1989) reports in his study of five villages in Northeast Nigeria that the average number of millet plantings was three during the dry year of 1975.

Figure 4.7 gives an example of how the arbitrary choice of planting date can lead to a misrepresentation of the actual field conditions. It shows the water balance for the same year presented in figure 4.6 (1964) but adopting a different planting date, that is, the first rain event larger than ET_o (day 135). The millet crop is unlikely to survive through the initial stages due to high soil moisture deficit and insufficient water in the shallow surface for germination. Moreover, millet would be harvested in the middle of the rainy month of August, which would impede the ripening process.

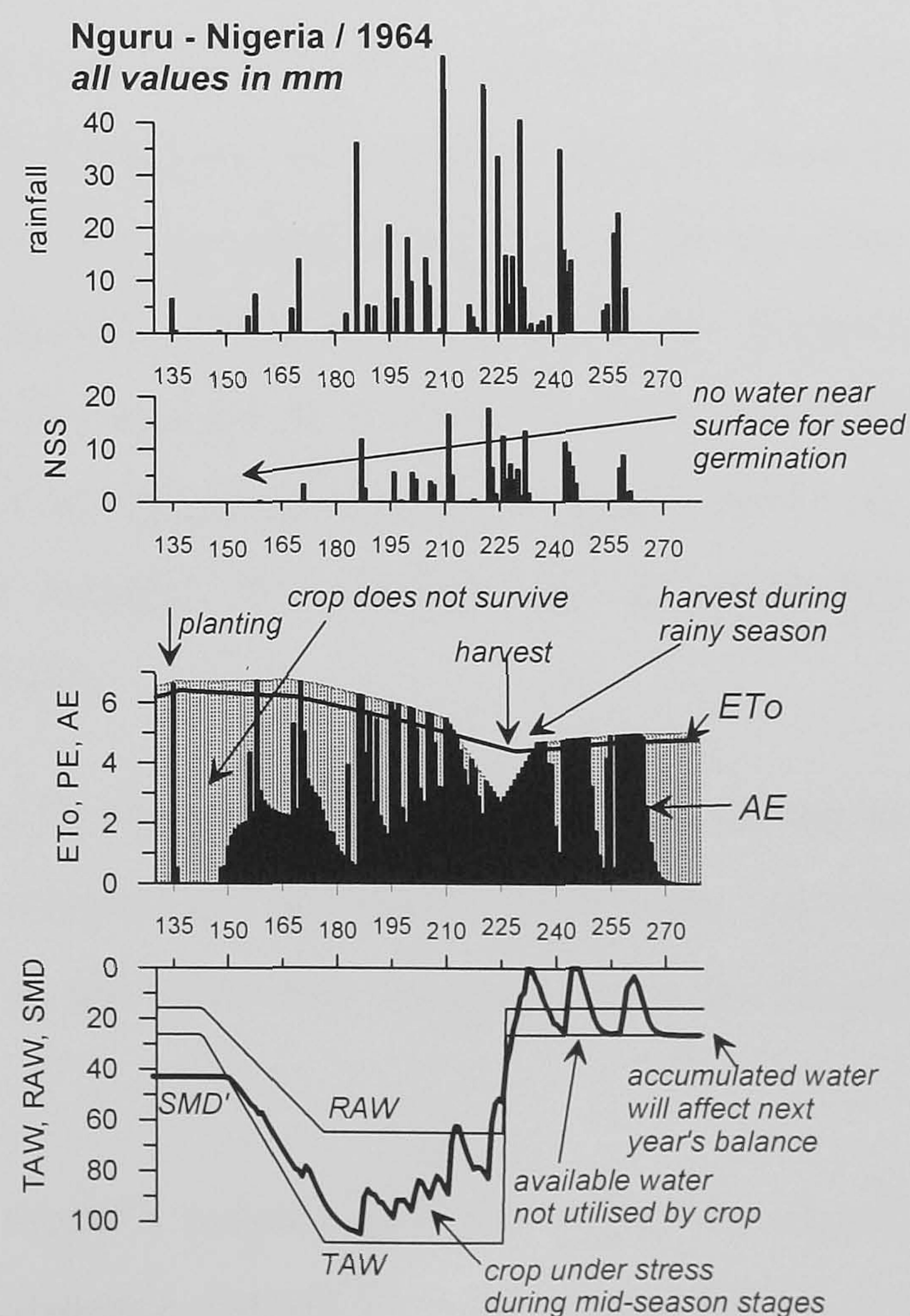


Figure 4. 7. Selecting arbitrary planting date for a crop during the rainy season.

In addition, the effect of an incorrect planting date choice can affect the water balance of the following period. That is, as the crop season finishes earlier, water that should be effectively utilised by the crop is erroneously added to the soil moisture deficit, thus affecting the initial conditions of the following year.

Usually, the date of planting is determined based on the concept of rainfall onset and false start. Several authors have defined different criteria for the onset of rains and false start for crops in West Africa. Table 4.1 lists some examples of these definitions.

Table 4. 1. Definitions for onset of rains in the region of semi-arid West Africa

Author	Criteria for onset of rainst
Kowal and Kassam (1978)	Rainfall at least 25 mm in 10 days with a subsequent 10 day period in which the amount of rainfall is at least equal to 0.5 ETo
Agnew (1991)	Rainfall at least 10 mm in 5 days with a subsequent 15 days in which the amount of rainfall is at least 10 mm
Sivakumar (1992)	Rainfall at least 20 mm in 3 days with a subsequent 30 days in which no dry spell exceeds 7 days.
Benoit (1977)	Rainfall at least 0.5 ETo over any period with no more than 5 dry days immediately following.
Stern et al. (1982)	Rainfall at least 20 mm in 2 days with a subsequent 30 days in which no dry spell exceeds 10 days.

Each of these methods above has been tested for the period of 36 years (1962-97) from Nguru. The results show a considerable variation of the rain onset date for each method (see Appendix D). The method of Sivakumar (1992) and Stern et al. (1982) are found to be too strict for the data set, resulting in years with no onset of rains at all. On the other hand, the methods of Agnew (1991) and Benoit (1977) do not identify the frequent presence of dry spells following planting. The method of Kowal and Kassam (1978) was devised for decadal rainfall and may not be appropriate for daily rainfall. However, in 92% of the 36 years tested with this method, the planting date is reasonably determined between the months of June and July.

In the following numerical examples in this thesis, planting date is determined first by one of the criteria above, then individually analysed and adjusted in order to be coherent with observed agricultural practices. That is, the latest date for planting is the second week of July (day 196).

Another aspect provided by the SAMBA model's outputs in figure 4.7 is the overall understanding of the modelled system. For instance, information about crop failure can be drawn out from a visual interpretation of figure 4.7, or assuming that actual

evapotranspiration can not be zero in the days following planting. This type of alternative approach is better explored in Chapter Five when alternative outputs are compared to lateral information from field observations.

4.3.2. Surface runoff

The method for estimating surface runoff, as described in Section 3.3, is based on the relationship between rainfall and the soil moisture deficit for a particular soil type. This association is made through a matrix of coefficients, such as table 3.1. This procedure gives insight into the physical processes resulting in surface runoff since the soil surface conditions and the climatic factors are implicitly included in the method.

This section explores the determination of surface runoff using selected periods from the Nguru long-term series. Figure 4.8 shows the calculated runoff component for the year 1990, using the same assumptions utilised in the Section 4.2.

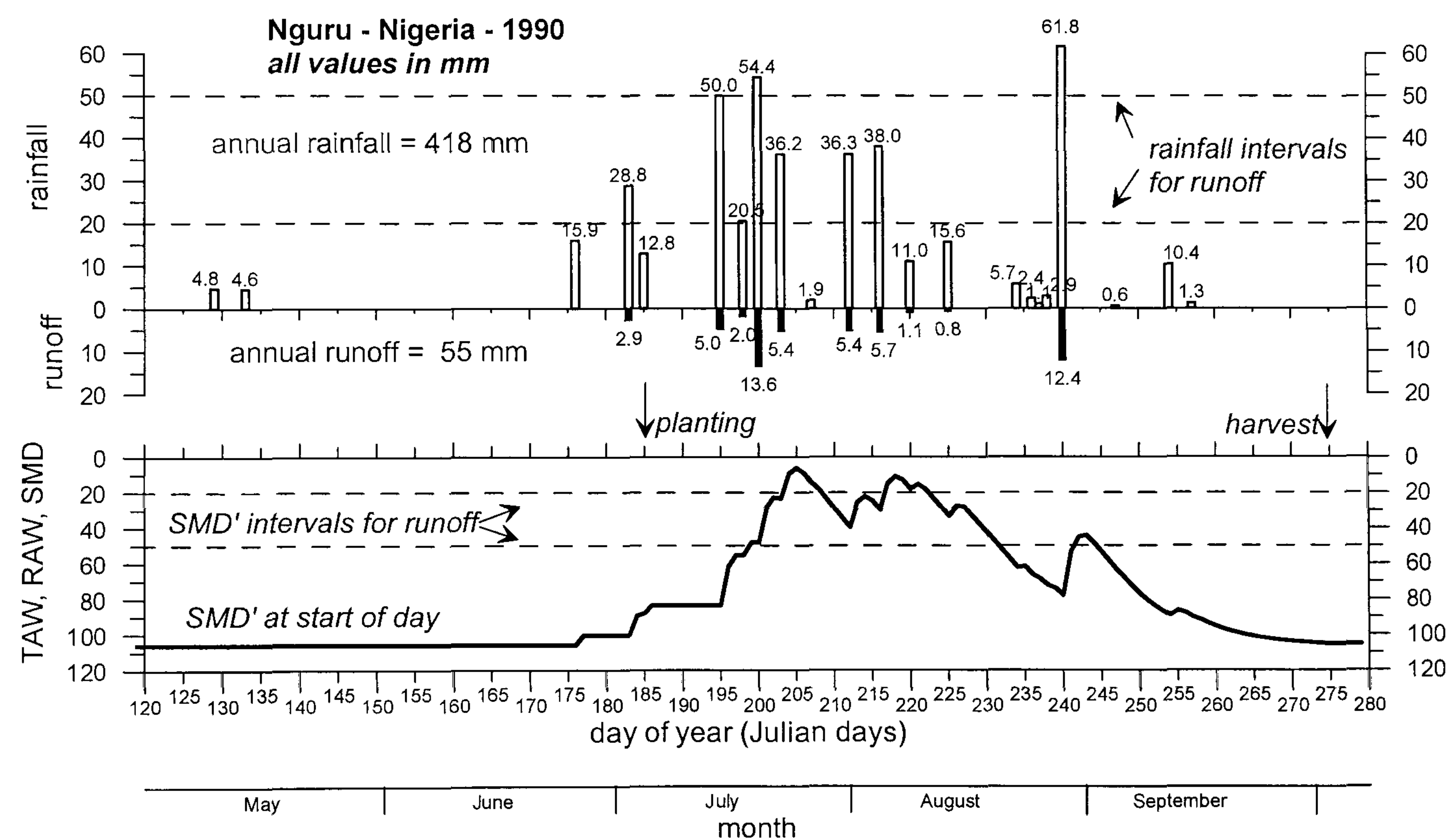


Figure 4. 8. Surface runoff determination for a climatic average year in Nguru using the SAMBA model.

In 1990, the annual rainfall of 418 mm was slightly below the long-term average (431 mm) and fairly distributed between late June and middle September (the two rainfall events in May were not significant). The soil moisture deficit before the start of the

rainy season is high, reflecting the moisture conditions of the previous year (annual rainfall in 1989 was below average).

Surface runoff is not significant during the early period because of the high *SMD'*. However, this situation changes with the heavy rains after the middle of July.

The variability in estimated daily runoff resulting from the estimation approach is observed in two selected days. At day 195, a rainfall of 50 mm results in 5 mm of runoff. Five days later, due to the wetter soil, a similar rainfall of 54 mm results in a runoff approximately three fold greater (13.6 mm). At the end of the crop season, the soil becomes drier due to the crop water uptake, consequently a larger rainfall event of approximately 62 mm generates a lower runoff value of 12 mm.

This variability is characteristic of the hydraulic processes present in a rainy season. Therefore, it is expected that runoff also varies between years with similar annual rainfall since the soil moisture conditions are unlikely to be the same. This can be observed in the following analysis using all the 36 years from the Nguru data series (figure 4.9). The model's parameters are the same as those utilised in Section 4.2. All the model parameters were kept constant in each year with the exception of planting date which varies according to the climatic criteria defined previously, and the initial soil moisture deficit which varies for each year as discussed in Section 4.2.5.

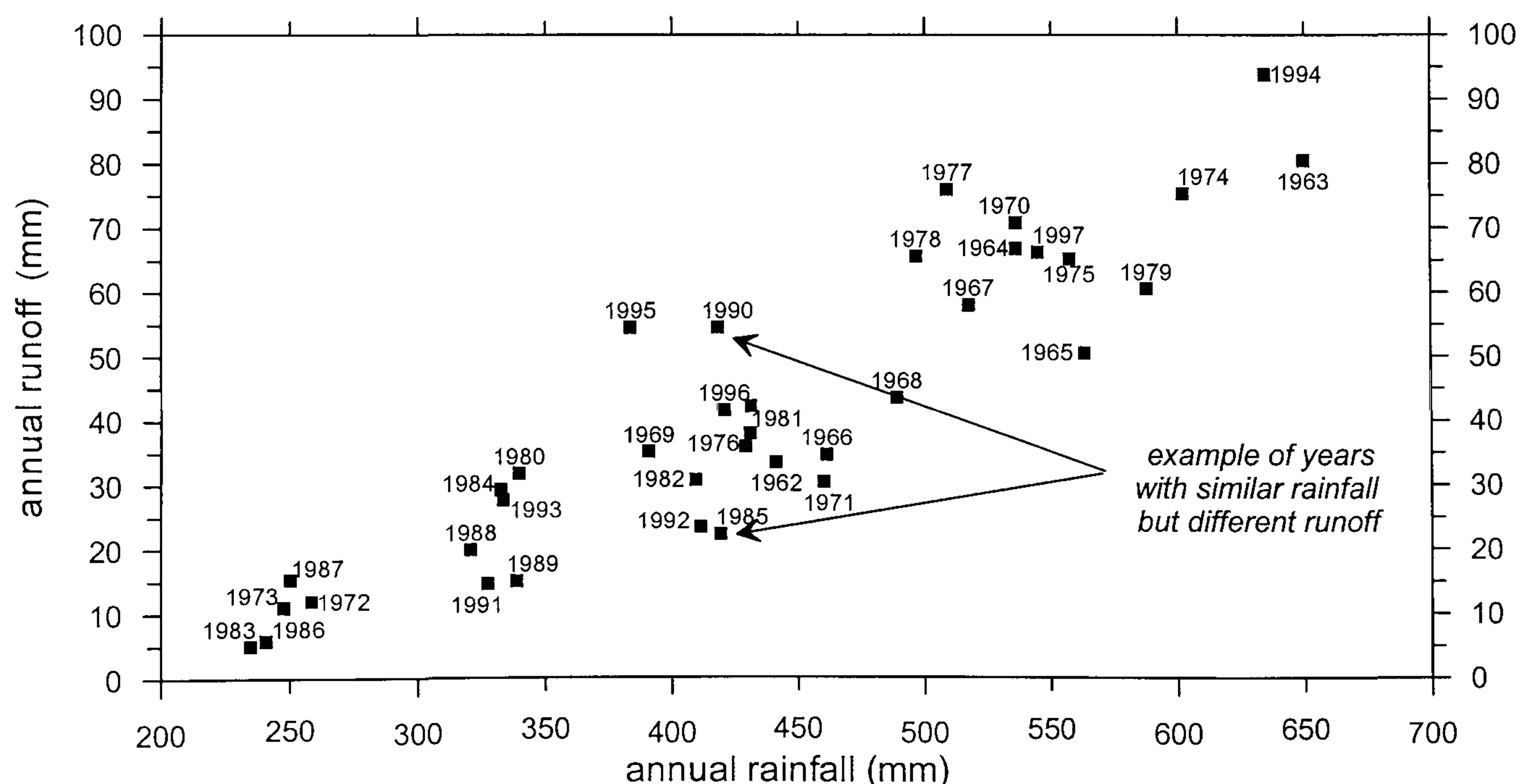


Figure 4. 9. Relationship between annual rainfall and estimated annual runoff for 36 years in Nguru.

It is possible to observe a linear trend between annual rainfall and runoff ($R^2=0.82$), but the inter-annual variability is high. Runoff varies from 2 to 15% of the annual rainfall.

The inter-annual variability can be visualised comparing the years of 1990 and 1985. In the year 1990 (rainfall 418 mm, figure 4.8), annual runoff was 55 mm (13% of annual rainfall). However, in 1985 the annual rainfall of 419 mm has generated a runoff value 60% lower than in 1990 (22 mm, 5% of annual rainfall).

This can be explained by the occurrence of high rainfall events and the variations in soil moisture conditions. In 1990, eight rainy days were above 20 mm including three storms of over 50 mm (responsible for 56% of the total runoff). In addition, the soil moisture deficit was low during the peak of the rainy season in July and August. On the other hand, in 1985 (figure 4.10) no rainfall was larger than 50 mm and the soil moisture deficit remained high during the entire rainy season due to frequent dry spells.

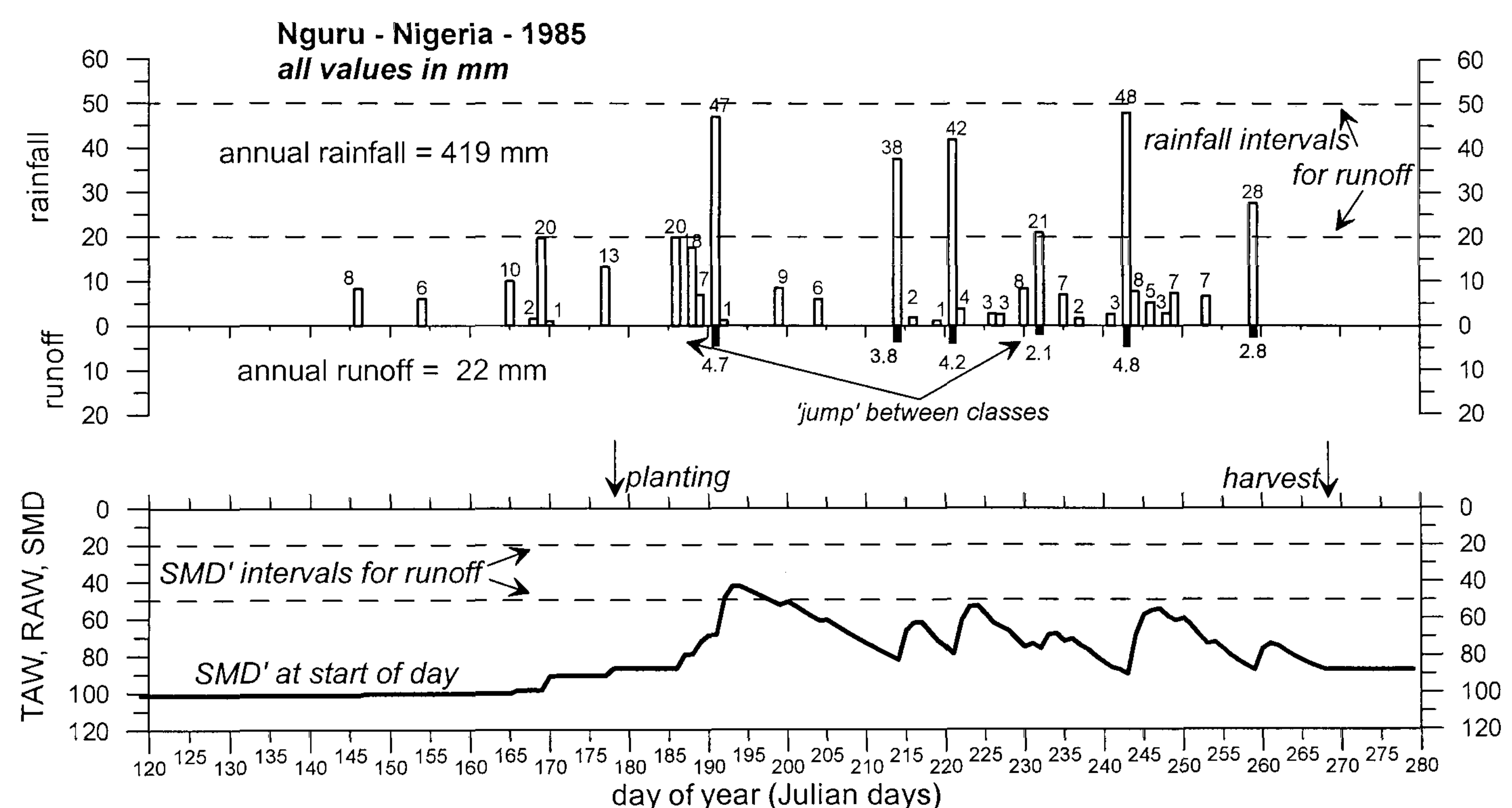


Figure 4. 10. Year with similar annual rainfall but different estimated annual runoff than 1990.

In an overall analysis, the numerical examples above show that runoff estimated by the procedure adopted by the SAMBA model is reasonable in terms of timing and total amount. Including the effect of antecedent soil moisture alongside rainfall characteristics, the determination of surface runoff implicitly takes into account key physical factors. The present procedure is flexible and allows for an easy adjustment of the coefficients, as well as of the number of intervals in the matrix of coefficients.

However, the values of the matrix of coefficients (e.g. table 3.1) are only a rough estimate of the actual local conditions, as discussed in Section 3.2.1. More detailed

investigation could be carried out in order to achieve a better representation of the local values. For instance, figure 4.10 shows a particular example when two similar rainfall days (20 and 21 mm respectively) with similar soil moisture deficits (*SMD* above 50 mm) are responsible for different runoff values (zero and 2.1mm respectively). This is caused by a discontinuity in the matrix coefficients (zero and 0.1 respectively). The inclusion of more intervals might avoid this problem.

The effect of the surface runoff component on potential recharge is investigated later in Section 4.4.

4.3.3. Near surface storage

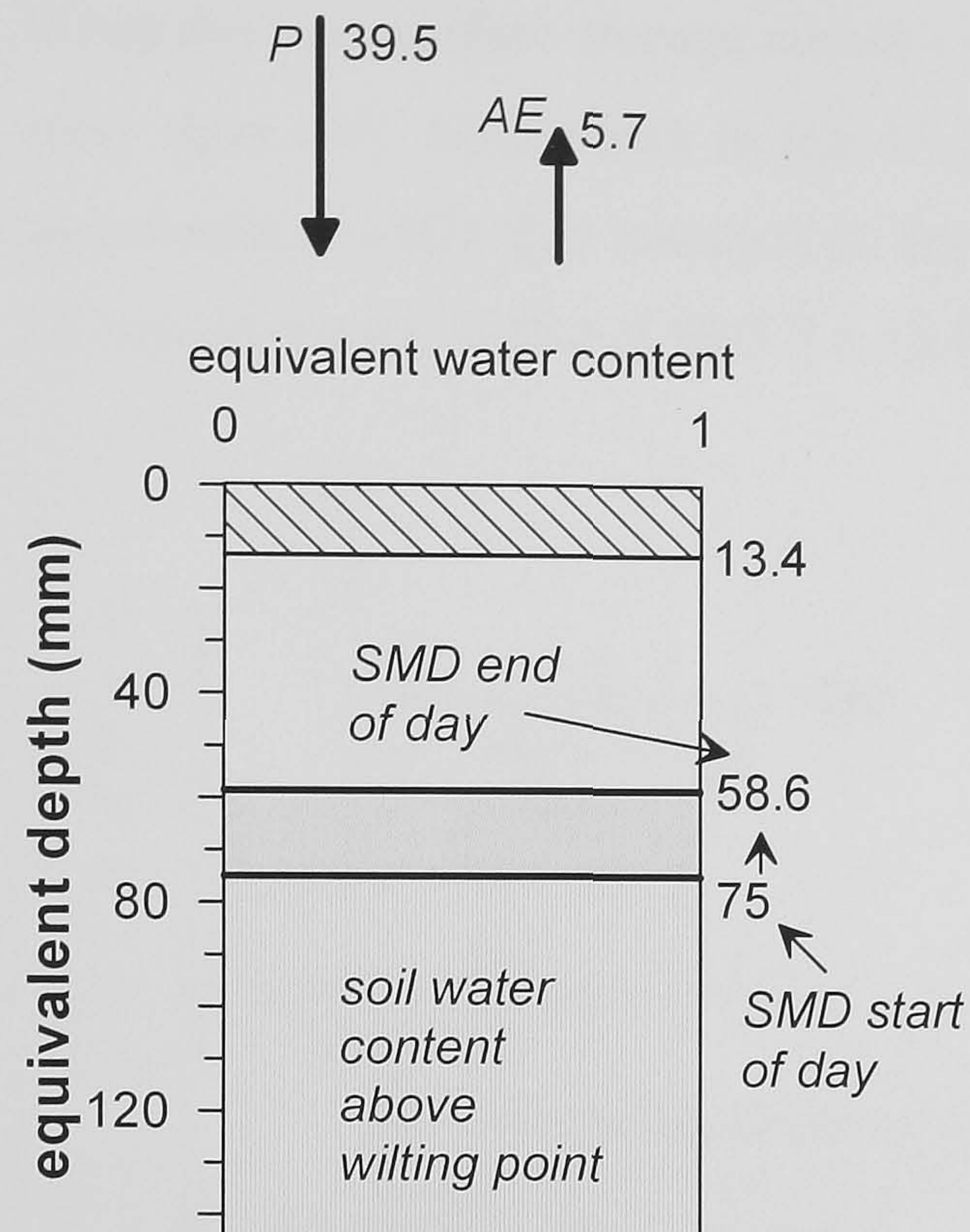
As described in Section 3.6, the SAMBA model keeps a proportion of the infiltrated rainfall in a near surface store so that it is available for evapotranspiration during the following days. This section aims to demonstrate the application of this procedure using real field data, for two situations cited in Section 3.6 where the concept of near surface storage is a significant improvement in terms of soil water balance technique in semi-arid areas.

Situations when SMD is greater than TAW or TEW

Figure 4.11 shows the soil water balance for a single day when the direct evaporation from the soil surface is the dominant process and the soil moisture deficit at the start of the day (75 mm) is larger than the total evaporable water (26 mm). This situation might occur at the beginning of the rainy season when the soil profile is very dry following the long dry season.

Rainfall P during that day is 39.5 mm (this value is real data for 11th May 1992 in Nguru). The resulting runoff Ro is 3.95 mm (10%) and actual evapotranspiration AE is equal to PE (5.7 mm).

The surplus of water ($In - AE$) is 29.85 mm. Assuming a fraction of storage equal to 45%, 13.43 mm of water is kept near surface (figure 4.11.a). The remaining 16.42 mm is added to the soil moisture content in the deeper layers resulting in a reduction of the soil moisture deficit (from 75 to 58.58 mm).



(a) computational approach-equivalent depth scale

These quantities are shown in figure 4.11(b) to actual depth scale, keeping the proportions among the areas.

Figure 4.11(c) shows the actual change in volumetric water content during the day. The change in moisture content is divided into two parts; the lower part corresponds to the decrease in soil moisture deficit; the upper part represents the increase in near surface storage.

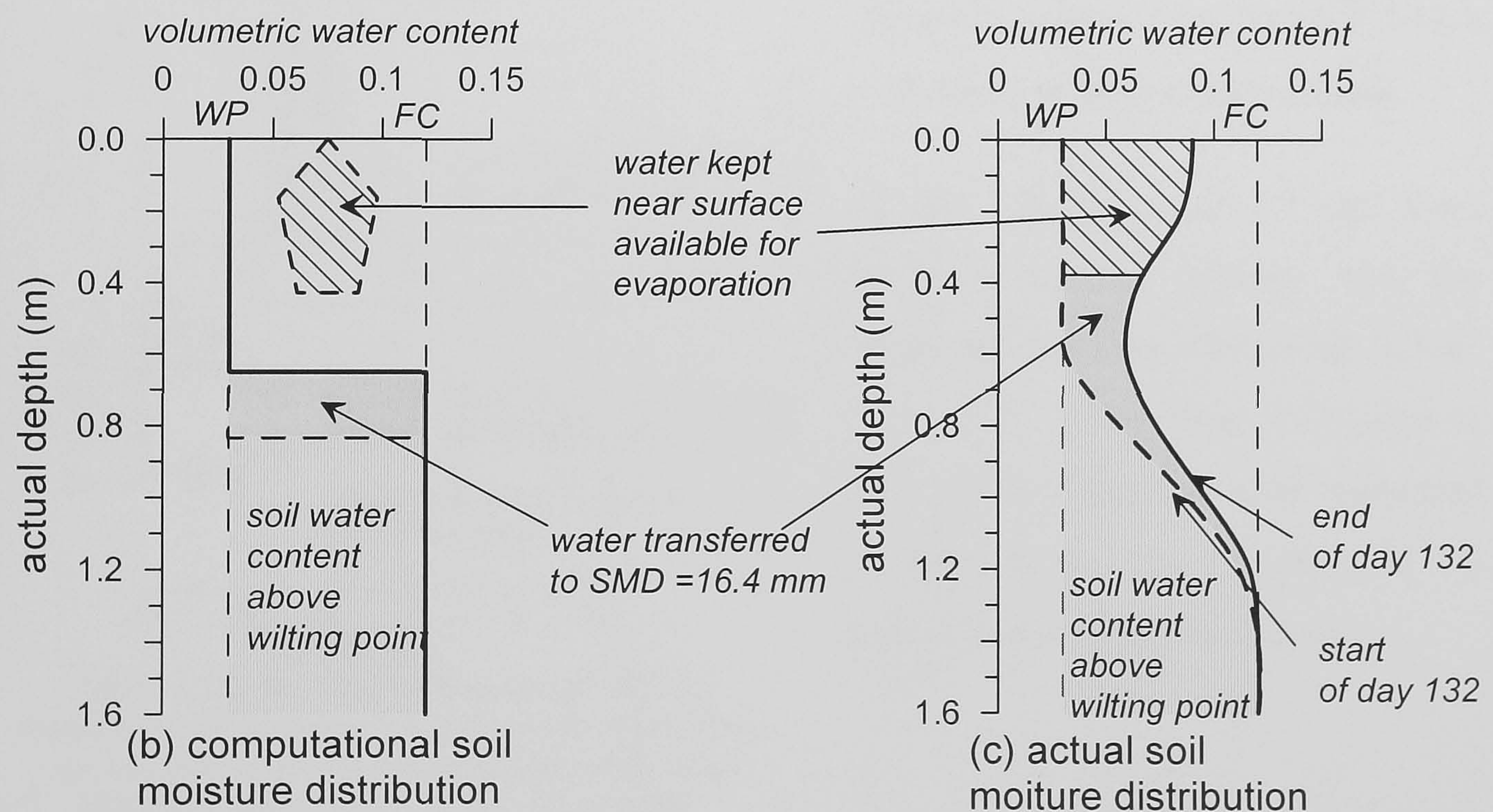


Figure 4. 11. Soil water balance from the SAMBA model for one particular day in Nguru (11th May1992) showing the estimated actual soil water distribution after a heavy rain.

The effect of this procedure in the estimation of AE for a period of 10 days including the day presented in figure 4.11 is presented in the figure 4.12. Two situations are considered with the purpose of illustrating the concept of near surface storage: (i) assuming a fraction of storage equal zero and, (ii) a fraction of storage equal to 0.45.

When the near surface storage concept is not applied, actual evaporation occurs during rainy days only (white bars in the *AE* graph), being the excess of water immediately transferred to *SMD*. For instance, at the end of day 132, *SMD* decreases approximately 30 mm ($P-Ro-AE = 39.5-3.95-5.7 = 29.85$ mm), changing from 75 mm to 45 mm.

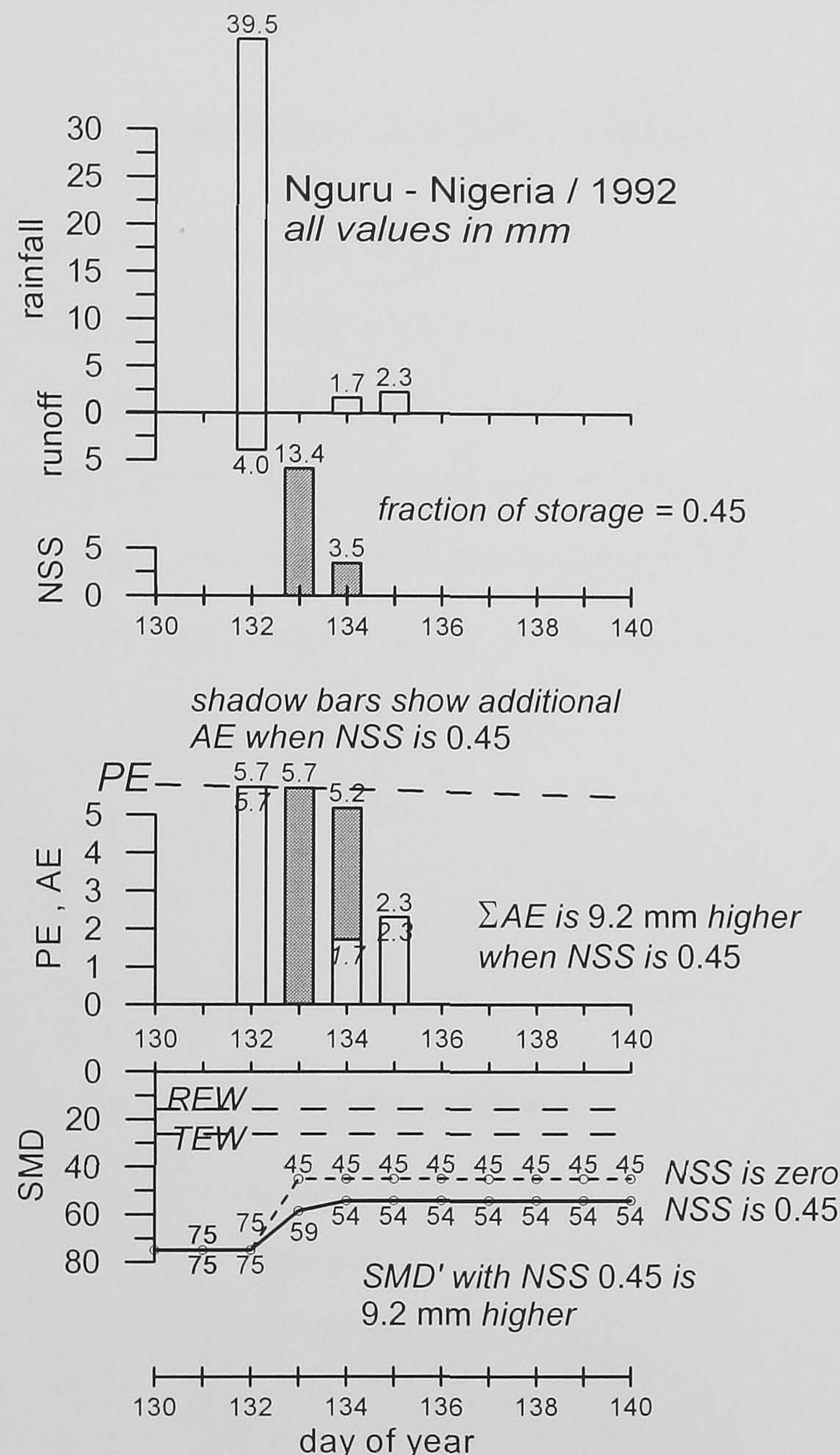


Figure 4. 12. The near surface storage and soil evaporation for a sequence of ten days in Nguru when the soil surface is bare and soil evaporation is the principal process (*SMD* is larger than *TEW*).

Comparing the two situations in figure 4.12, the results show that the estimated actual evapotranspiration is higher when the concept of near surface storage is applied. Consequently, the resulting soil moisture deficit is also higher. In figure 4.12, *SMD'* at the end of the analysed period is approximately 9 mm higher with *NSS*.

The actual evapotranspiration estimated with the near surface storage approach agrees reasonably well with the observed soil evaporation process (Section 2.2.4). Figure 2.4

When a fraction of storage of 0.45 is applied, the resulting actual evapotranspiration (soil evaporation as the dominant process) is higher (additional *AE* is shown as shaded bars).

For example, at the end of day 132, 13.4 mm is kept near surface ($P-Ro-AE) \times 0.45 = 13.4$ mm and the remaining water ($39.5-3.95-5.7-13.4 = 16.45$ m) is transferred to *SMD*.

At day 133, *AE* takes 5.7 mm from the near surface storage and the remainder is again partitioned, $(13.4-5.7) \times 0.45 = 3.47$ mm. This value is added to the small rainfall registered on day 134 (1.7 mm) resulting in 5.2 mm of actual soil evaporation.

shows that soil evaporation occurs close to the potential rates during the first 2-3 days after rainfall (first stage). After that, evaporation decreases rapidly to residual values (second stage), which persist for longer (third stage). The computational concept adopted by the SAMBA model represents the first and the second stage of the soil evaporation drying process.

Situations when SMD' is larger than RAW or REW

Figure 4.13 shows a period of 15 days during the crop period of 1972 in Nguru. The year of 1972 was a very dry year (annual rainfall of 248 mm) with rainfall sparsely distributed. Consequently, the soil moisture deficit was high and the millet crop was under water stress. In order to compare the effect of near surface storage in a sequence of days, two conditions are analysed: (i) the NSS concept is not applied (figure 4.13.a) and (ii) applying a fraction of storage equal to 0.45 (figure 4.13.b)

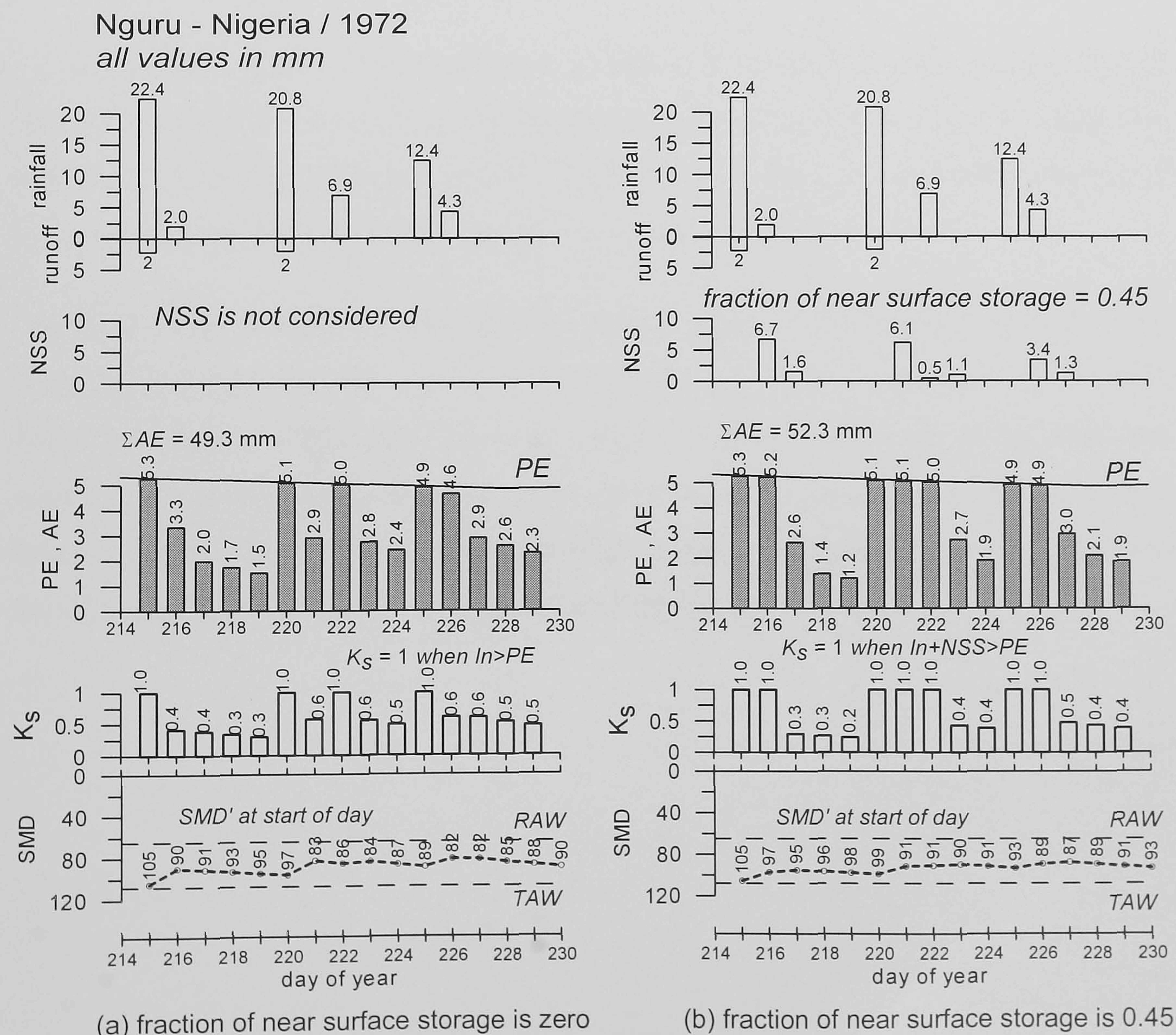


Figure 4. 13. The soil water balance components during 15 days when SMD is larger than RAW .

In figure 4.13 (a) at day 215, rainfall is 22.4 mm and actual evapotranspiration is at the potential rate (5.3 mm). On the following day (day 216), as rainfall is low and the resulting infiltration is less than PE , the stress factor K_s applies. AE is calculated using the equation 3.3, that is, $AE = In + K_s \cdot (PE - In) = 2 + 0.4 (5.2 - 2) = 3.3$ mm. One day later (day 217), rainfall is equal to zero and AE is a fraction K_s of PE . Note that the stress factor K_s is equal to 1.0 when infiltration is larger than PE .

When the storage factor is equal to 0.45 (figure 4.13.b), on day 216 the water available for evapotranspiration (i.e. rainfall during the day plus the water stored from the previous day, $2 + 6.7 = 8.7$ mm) is greater than PE (5.2 mm), therefore $AE = PE$. One day later (day 217), rainfall is zero but there is still water remaining near surface (1.6 mm), therefore, AE is calculated using the equation 3.3, that is, $AE = NSS + K_s \cdot (PE - NSS) = 1.6 + 0.3 (5.1 - 1.6) = 2.6$ mm. On the following day (218) rainfall is zero and no water is available from the previous days, so AE is a fraction K_s of PE .

Comparing the figures for the period of 15 days, AE estimated with the near surface storage approach is larger than the situation when no surface storage is taken into account (52.3 mm and 49.3 respectively). Consequently, the resulting SMD at the end of the period is slightly higher when the NSS approach is applied.

The figures above illustrate that the near surface approach better represents the crop evapotranspiration for the periods when plants are under water stress. On days following heavy rainfall, plant roots are likely to extract water from near surface and transpire at potential rates. Moreover, the increase in AE suggests that this procedure better represents the semi-arid conditions where evapotranspiration rates are high due to the high atmospheric demand on plants and on the bare soil fraction.

4.4. EFFECT OF RUNOFF AND NEAR SURFACE STORAGE ON POTENTIAL RECHARGE FOR A SEQUENCE OF YEARS

This section investigates the effect of surface runoff and the near surface storage on the estimation of potential recharge. The analysis is developed using the 36 years from the Nguru rainfall data set. Subsequently, years with representative characteristics are explored in more detail.

4.4.1. Effect of surface runoff on recharge estimation

Figure 4.14 compares the estimated annual potential recharge assuming that runoff is zero to recharge estimated with runoff (using the runoff coefficients from table 3.1). All others parameters were kept equal except the planting date and the soil moisture deficit initial conditions that changed for each year.

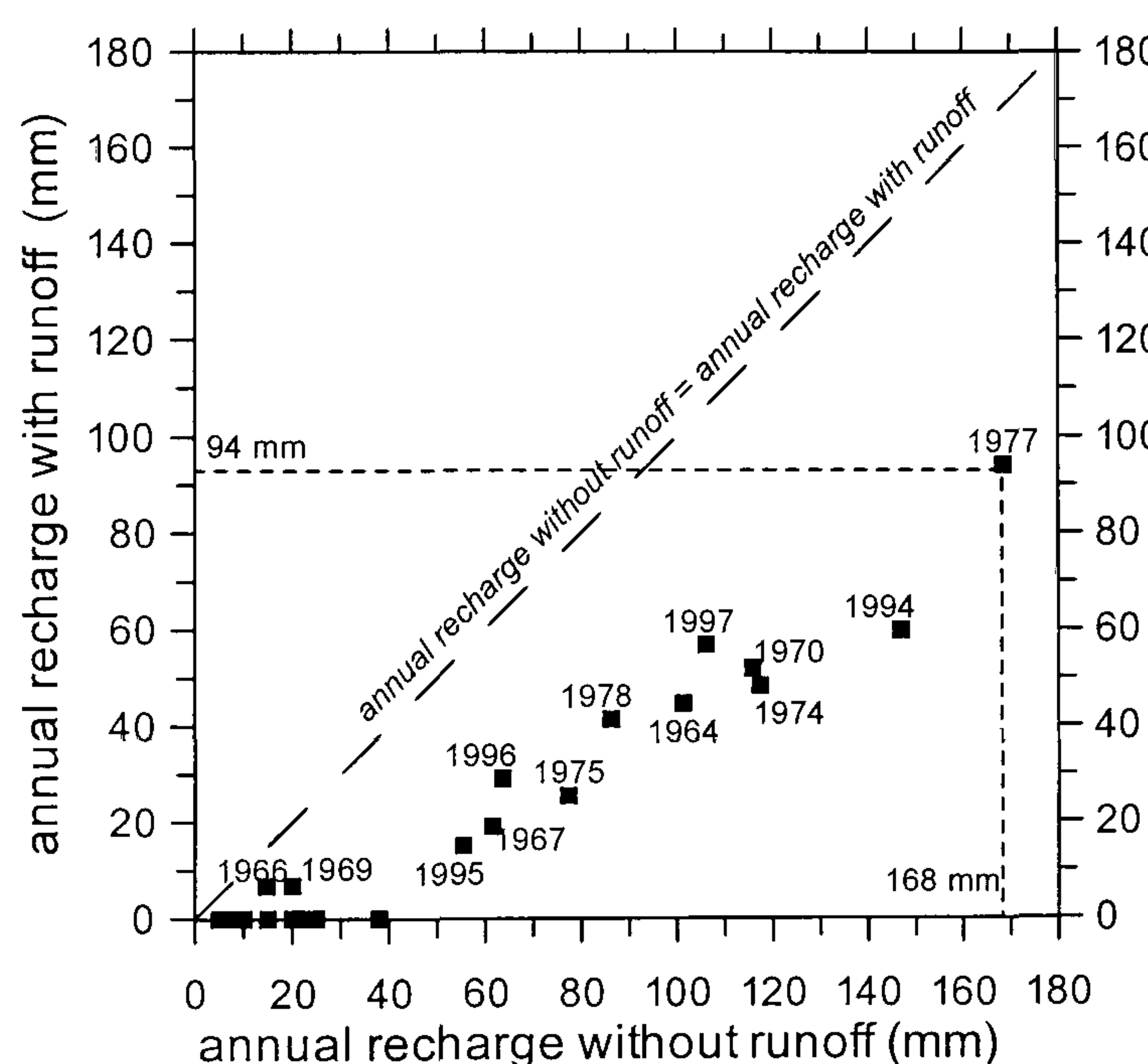


Figure 4. 14. Recharge calculated with and without the runoff component

The figures show a significant reduction of potential recharge ($p < 0.01$) when including runoff. The overall decrease is around 60%. For instance, in 1977 the estimated recharge not considering runoff is 168 mm. However, when runoff is included, the resulting recharge is 91 mm (a decrease of 46%). The difference in 1995 is more significant. Recharge decreases from 55 to 15 mm (a reduction of 73%)

Figure 4.15 illustrates a period of 20 days during the 1995 rainy season. If runoff is assumed zero, a heavy rainfall at day 197 (57 mm) brings the *SMD* from 30 mm to a negative value of 21.3 mm ($30 - 5.7 + 57 = -21.3$ mm). As defined in Section 3.7.5, a negative *SMD* represents the water above field capacity which drains as potential recharge. Note that *NSS* does not apply when the amount of water stored is larger than the soil moisture deficit at the end of the day.

On the other hand, with a surface runoff of 14.3 mm, the resulting potential recharge decreases from 23.1 to 7.1 mm ($57 - 14.3 - 5.7 - 30 = 7.1$ mm).

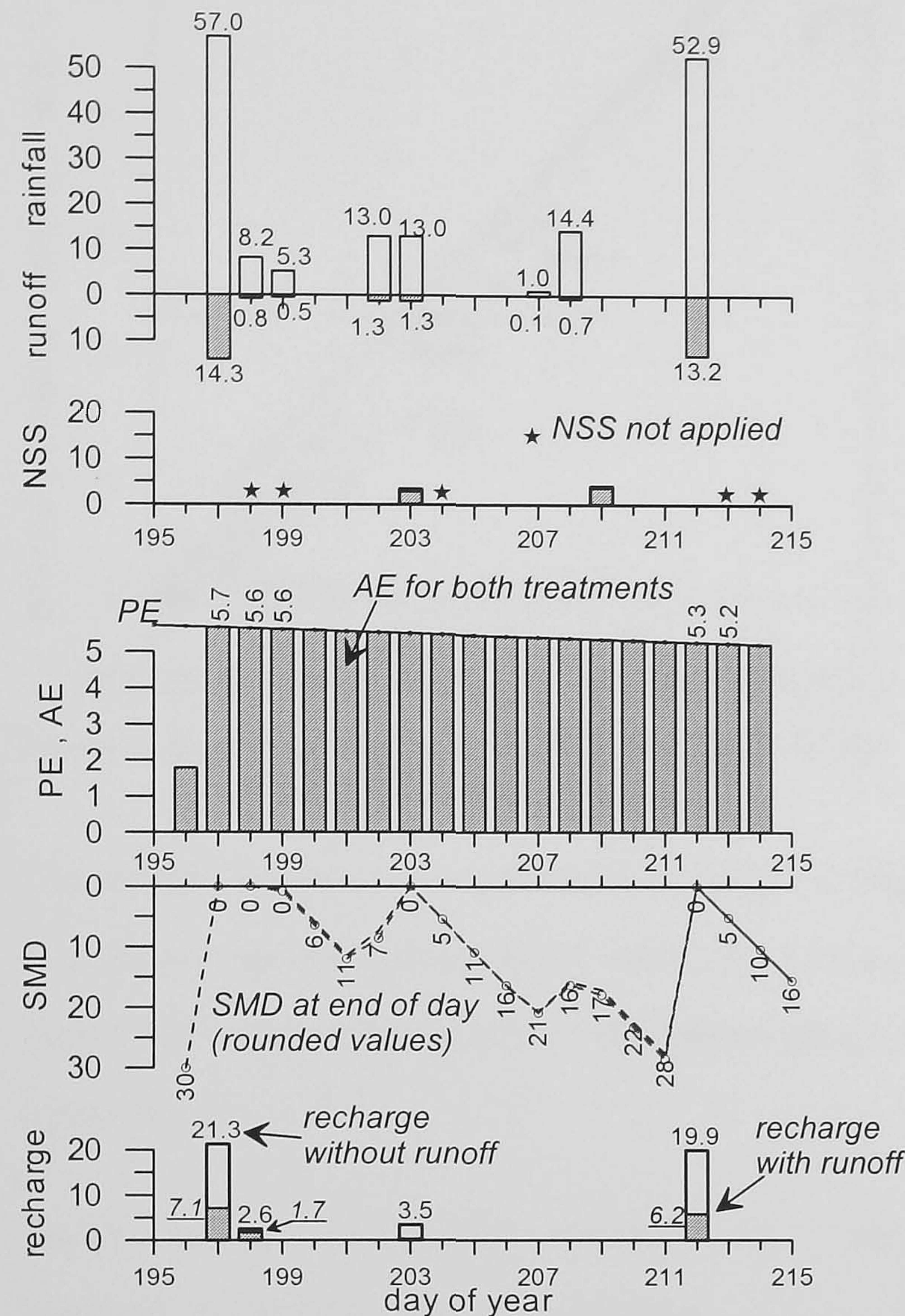


Figure 4. 15. Effect of surface runoff on potential recharge estimation. Nguru, 1995.

A similar pattern occurs at day 212, when a rainfall of 52.9 mm results in an estimated potential recharge of 19.9 mm if runoff is zero ($52.9 - 5.3 - 27.7 = 19.9$ mm). When runoff is included, recharge is 69% lower (6.2 mm).

These examples show that the soil water balance is sensitive to surface runoff in a semi-arid area. Runoff losses are likely to be significant due to the occurrence of heavy daily rainfall and, consequently, omission or inclusion of runoff will affect the resulting estimated potential recharge. Further investigation of the effect of runoff is carried out in Chapter Six.

4.4.2. Effect of the near surface storage approach on recharge estimation

Applying the *NSS* concept to the water balance seeks to improve the representation of potential recharge. The following figures investigate the overall effect of *NSS* on annual recharge. In addition, two particular years are considered in more detail in order to investigate key aspects of the *NSS* approach.

Figure 4.16 shows the potential recharge estimated for 36 years for two situations: (i) assuming that the fraction of storage near surface is nil; and (ii) the fraction of near surface storage is set to 0.45.

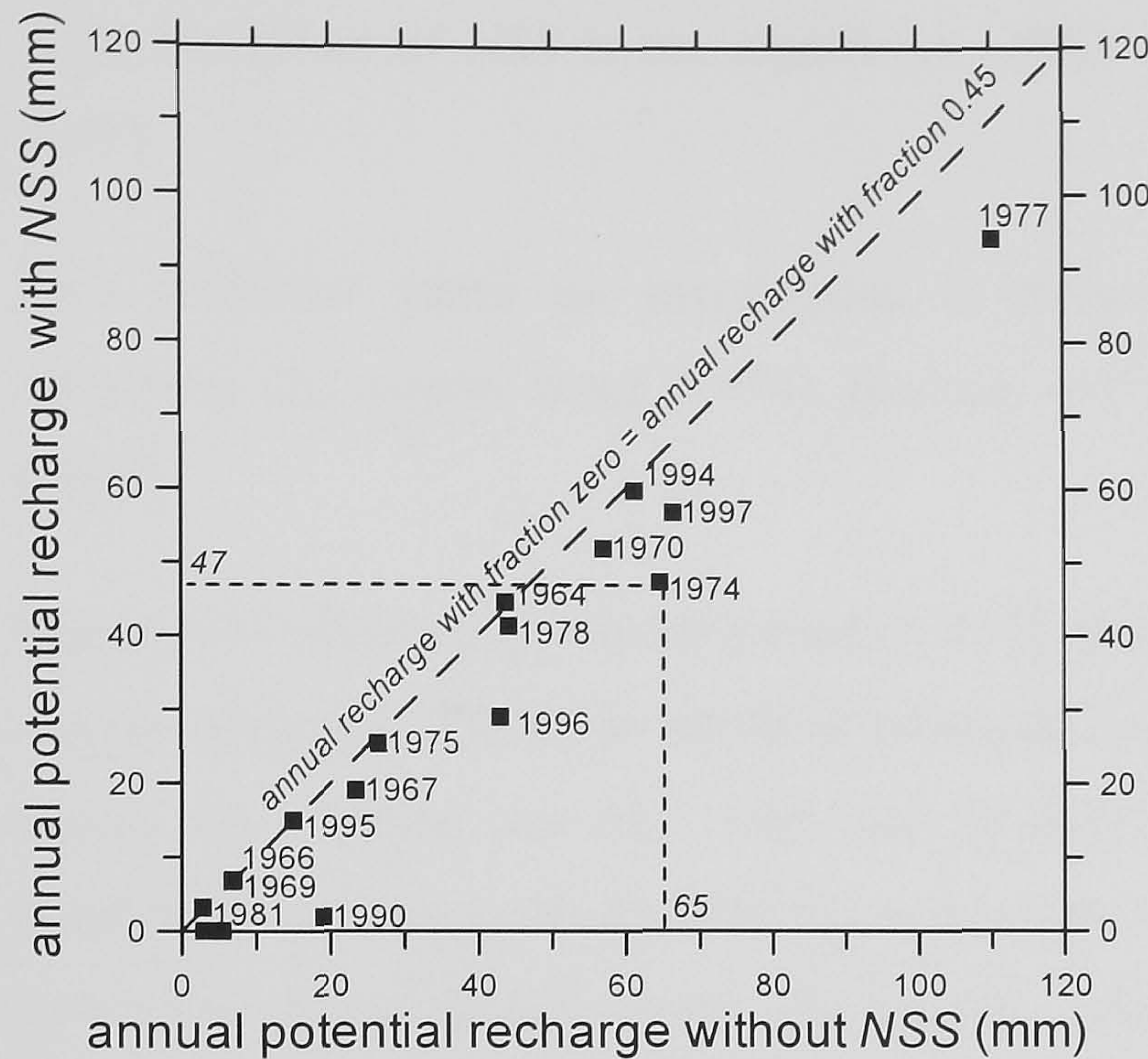


Figure 4. 16. Comparing the effect of the inclusion of NSS on annual potential recharge

The figures show a significant ($p < 0.01$) reduction of approximately 15%. For example, recharge for the year of 1974 changes from 65 mm to 47 mm (17% lower) when a near surface storage of 45% is considered. The year of 1990 presents the highest variation, with estimated recharge changing from 19 to 3 mm (a decrease of 90%).

The overall reduction in potential recharge is in part explained by the expected increase of actual evapotranspiration AE when the near surface approach is applied. For instance, figure 4.17 shows the actual evapotranspiration AE calculated for the same period as in figure 4.16.

There is a significant ($p < 0.01$) increase of actual evapotranspiration (approximately 15%, the same as the recharge decrease rate). For example, in 1990 AE increases 20 mm (from 361 to 341mm) and the resulting recharge is 16 mm lower (variation in runoff account for the other 4 mm). However several years such as 1964, 1994, 1977 and 1967, break this pattern.

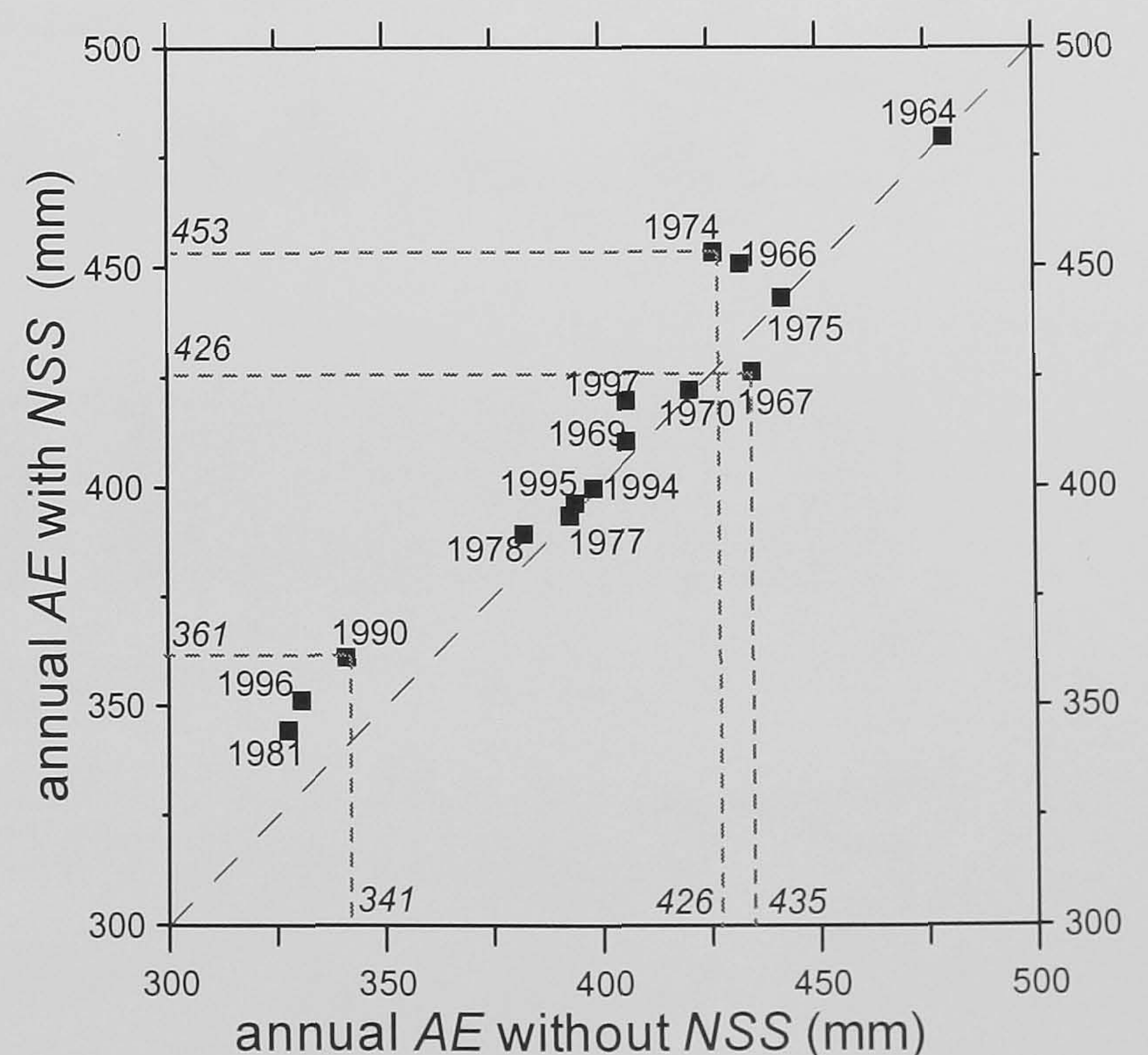


Figure 4. 17. Comparing the effect of the inclusion of NSS on annual actual evapotranspiration AE .

For some years such as 1964 and 1994 the near surface storage appears to have no effect on annual AE . The fact that these years have a high annual rainfall suggests that for wet

years the effect of *NSS* is not significant (*SMD* on the majority of days is less than *RAW*).

Two particular years are investigated in more detail. First, a typical year which represents the overall trend (1974). Second, one year that presents an opposite trend (1967).

Figure 4.18 shows a particular period of 1974 when the application of the near surface storage approach affects the resulting estimated recharge. The period comprises 20 days before the planting day (day 190) and 30 days during the initial crop stages. Two conditions are simulated: (i) *NSS* fraction equals zero and (ii) *NSS* fraction equals 45%. The figure shows some key aspects, which illustrate the effect of the *NSS* approach on the water balance.

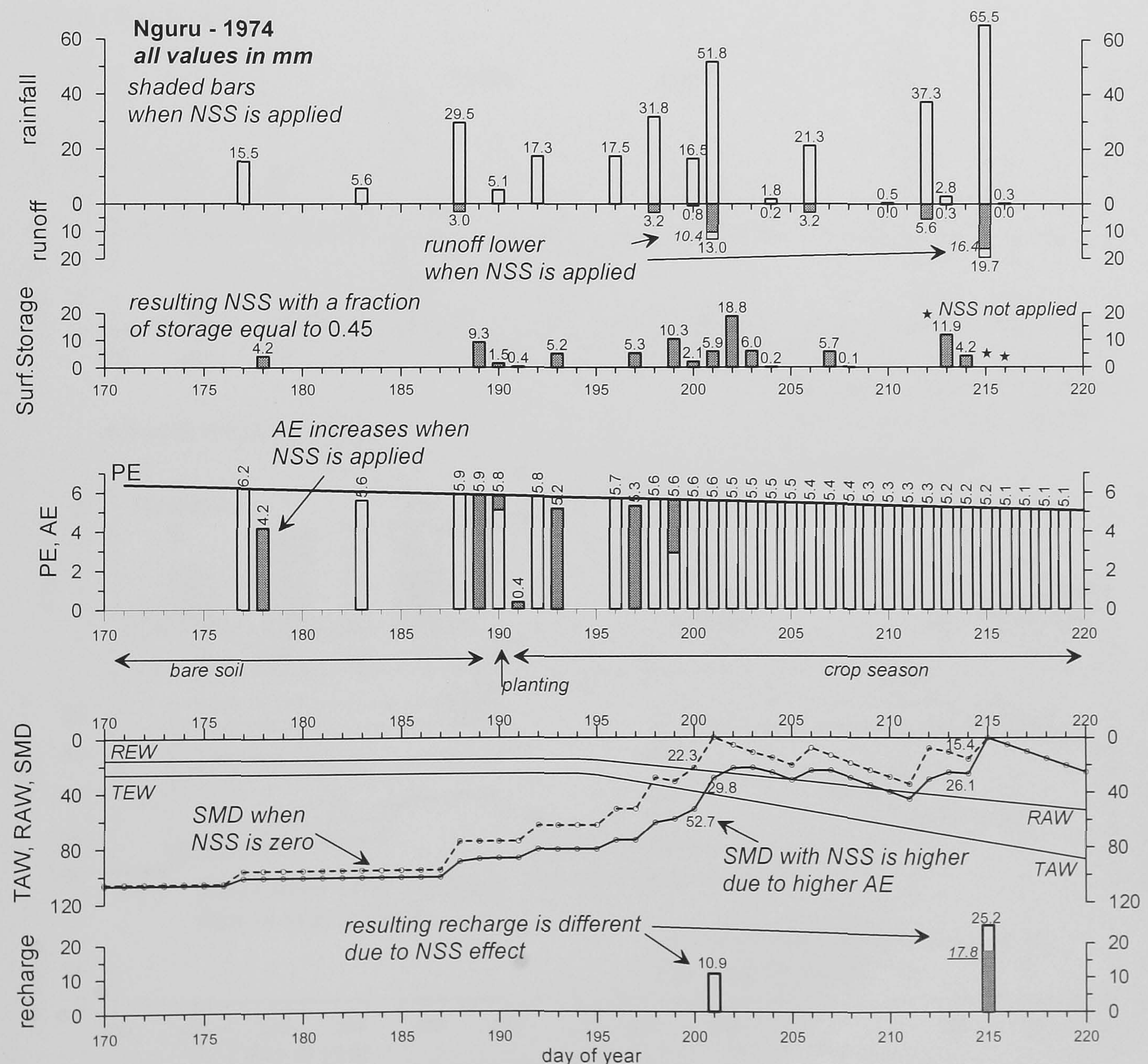


Figure 4. 18. The effect of *NSS* on recharge and actual evapotranspiration for a particular period of 1974.

Actual evapotranspiration AE increases when NSS is considered (shown as shaded bars). As a result of the increase in AE , the soil moisture deficit SMD is higher during the period. This affects the estimation of potential recharge in the following way. When a heavy rainfall at day 201 occurs (51.8 mm), the SMD' at start of day for the first condition is 22.3 mm. The estimated recharge is the excess of water equal to $P-R_o-AE-SMD'$ (51.8 - 13.0 - 5.6 - 22.3 = 10.9 mm). As the SMD for the second condition is higher (52.7 mm), the rainfall event is not enough to bring the soil profile to field capacity and estimated recharge is nil. At the day 215 a rainfall of 65.5 mm results in different values of recharge (25.3 and 17.8 mm).

Therefore, the effect of the application of NSS during the early period affects recharge estimation later on during the peak of the rainy season. However, there are other factors involved in the water balance which have to be considered in the analysis of a particular output (figure 4.19)

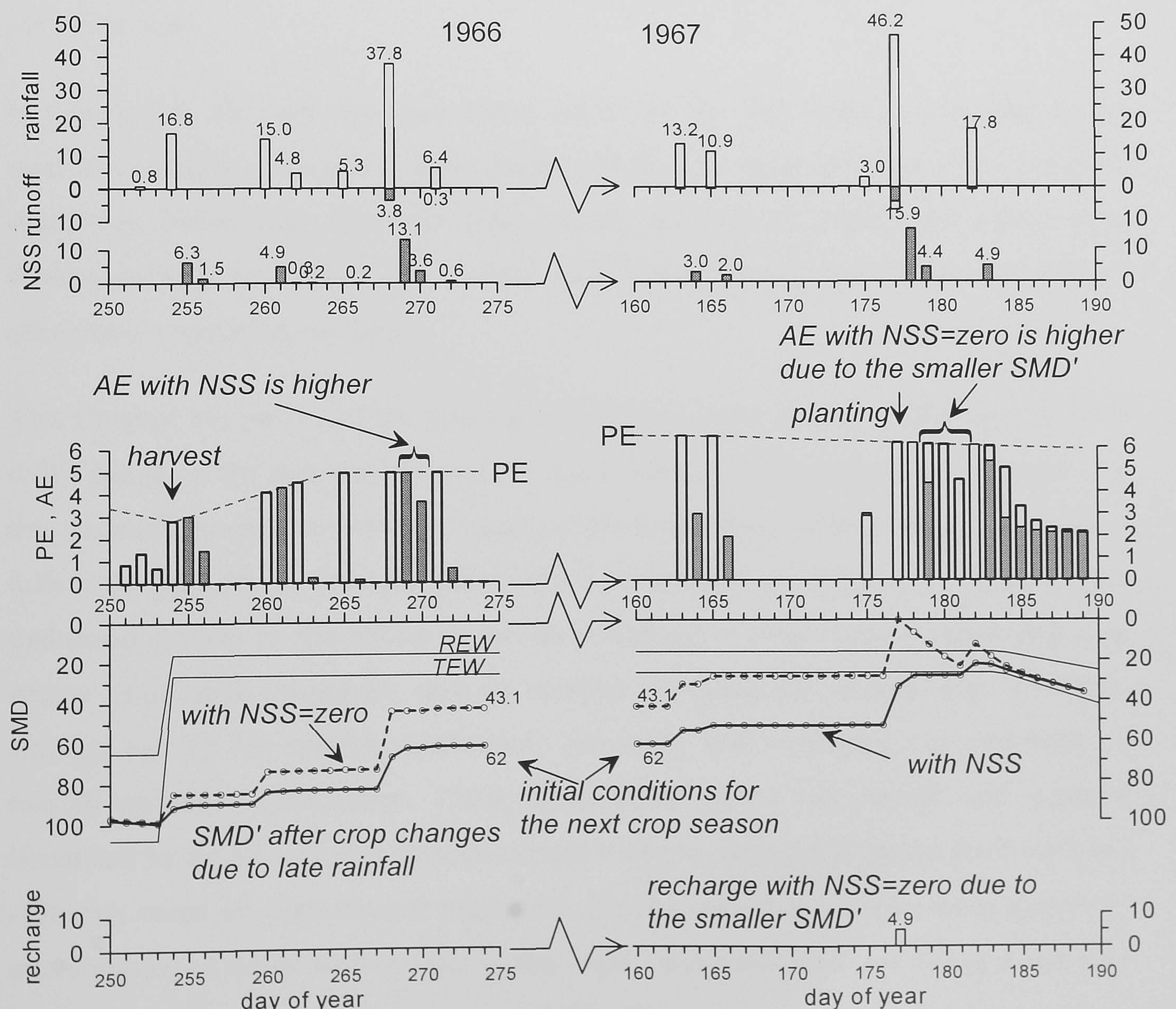


Figure 4. 19. Example of how the conditions from the previous rainy season affects the AE and recharge estimation in the following growing season (1967). Values in mm.

For example, the results for the year of 1967 present an unexpected pattern, that is, AE decreases despite the fact that more water is available for evapotranspiration. The reasons are the initial conditions associated with the water regime of the previous year, as shown by figure 4.19.

In 1966, late rainfall after harvest changes the soil moisture deficit for the two NSS conditions considered. When the factor of storage near surface is 0.45, AE is larger and consequently SMD is also higher, as expected (additional AE represented as shaded bars). Assuming that the water content does not change significantly during the dry season, at the beginning of the next rainy season (1967) the initial soil moisture conditions are the conditions determined at the end of the previous year. Therefore, for the case with no near surface storage, less water is needed to bring the soil water content to values closer to RAW . Consequently, actual evapotranspiration at early crop stages is slightly greater when NSS is zero, which explains the unexpected result for this particular year.

In conclusion, the two examples above point out the importance of the initial soil moisture conditions for a soil water balance study in a semi-arid area. The hydraulic conditions before and after the crop season, as well as during the initial crop development period affect the overall soil water balance and hence the correct estimation of potential recharge.

This Chapter has presented the results from the soil water balance assuming sole crop millet based on the characteristics of the short season (90 days) pearl millet since it is the predominant crop in northeast Nigeria (Grema and Hess, 1994; Philips, 1977) and it makes the best of the unpredictable rainfall distribution in the region. Actually, the traditional system in the region is the intercropping of short season crops and long season crops such as guinea corn or cowpea providing the farmer with a reliable strategy against the rainfall distribution variability and increasing the efficiency of rainfall utilisation (Mortimore, 1989). Thus, years with a late rainfall end, such as illustrated by figure 4.19, can provide enough water for the growth and maturing of late-maturing crops or even natural vegetation. Grema and Hess (1994) investigated the intercropping of millet with cowpea in this region and concluded that during a wet year

(1992) residual soil moisture deficit after millet harvest was enough to maintain a long duration cowpea crop.

Therefore, further investigation should be made associating the annual rainfall distribution with the crop length, based on the intercrop of short and long season crops. However, as stated by Kowal and Kassam (1978) the duration of the growing season is highly site-specific since it depends on a number of local variables such as climatic, soil, crop characteristics and external factors such as pest and insect damage.

CHAPTER FIVE

PLAUSIBILITY ANALYSIS OF THE MODEL RESULTS

Chapter five investigates the credibility of the results generated by the SAMBA model using the concept of ‘plausibility analysis’ developed in this research. A wide range of quantitative and qualitative observations ranging from local field experiment to overall studies in semi-arid areas, were gathered in order to judge the model results, as well as the model structure and its capacity to provide valuable insights into the physical system.

5.1. INTRODUCTION

One fundamental stage of groundwater modelling is to assess the credibility and robustness of results generated by a model. This process, conventionally named validation, usually requires the use of extensive datasets in an attempt to match the modelled output to the corresponding measured values.

However, when modelling with limited data, as in this research, this kind of validation is usually not possible and a different approach, based on more lateral thinking, has to be utilised.

This research uses the concept of “plausibility” analysis as an alternative approach for situations when modelling with limited data (Carter et al., 2002). The term “plausibility” is utilised in the sense of verification that the model adequately reflects those features of the groundwater system which really matter, based on as wide as possible a range of quantitative and qualitative observations from the hydrological system which can corroborate the insights provided by the model outputs.

The concept of ‘plausibility’ includes judgments about the structure of the model as well as tests of the model results. It assumes that if the model is reasonable in

representing the complex hydrological system through a series of processes, then it will be able to compute a credible water balance. Therefore, this procedure is not limited to an analysis of the recharge component alone but also to a series of other additional outputs which can help to assess the credibility of the representation of the water balance as a whole.

This approach is particularly relevant in developing countries where there is limited information about soil and aquifer properties and where monitoring and record keeping may be poor.

The following analyses make use of a range of quantitative and qualitative information from studies in semi-arid regions in order to compare with selected model outputs. They are: (i) observations from an independent field trial carried out in semi-arid Nigeria during one crop season (Section 5.2); (ii) a qualitative descriptive analysis of ‘hungry’ and ‘plenty’ years taken from a regional study developed for over a decade in a village located in semi-arid West Africa (Section 5.3) and; (iii) overall determination of groundwater recharge in semi-arid regions worldwide using representative studies (Section 5.4).

5.2. AN EXPERIMENTAL FIELD TRIAL IN SEMI-ARID NIGERIA.

The objective of this analysis is to compare the outputs of the model (water storage and recharge) to soil water content measured in a cropped plot in a semi-arid region. The data utilised is from an experimental trial conducted at the University of Maiduguri (11°54’N, 13° 05’E), Northeast region of Nigeria (see figure 4.1) as reported by Grema and Hess (1994), Grema (1994) and Hess (1999). The objective of this experiment was to examine the water use of sole and intercrop crops of millet and cowpea in a semi-arid environment.

Various data sets are available, such as climatic variables, crops and soil description, and measurements of soil water content obtained by neutron probe access tubes. A summary of these data is given below.

5.2.1. Field trial data

The long-term average rainfall in Maiduguri is 553 mm (1961-1990) with a unimodal distribution starting in June and lasting until the end of September. The rainfall data set available is a series of daily observations from 24th June to 31st October of 1992. The seasonal rainfall of 519 mm is 34 mm below the average. However, 1992 was significantly wetter than usual during the middle of the season. Approximately 50% of the rainfall (256 mm) was concentrated in the month of August, when four days with heavy rainfall above 30 mm were observed.

Daily reference evapotranspiration ET_o was calculated by the Penman-Monteith equation using meteorological data collected at the site and from a meteorological station situated nearby (Maiduguri Airport).

A representative soil water profile

The soil moisture profiles available reflect typical conditions under a semi-arid rainfed crop where the processes of water infiltration and redistribution within the soil profile happen simultaneously with the process of evapotranspiration.

Measurements of soil moisture content from 0.10 to 1.90 m were made on a weekly basis at intervals of 0.2 m using neutron probe access tubes installed in each crop plot. The soil at the experimental plots has a sandy loam texture and is weakly aggregated (Grema, 1994).

Figure 5.1 shows two sets of representative soil water profiles measured in one access tube under a millet crop. Two phases are identified, that is: from the onset of rains to the start of the month of September (wetting phase), and a second period when rainfall declines and the soil water content decreases due to root water uptake and due to drainage from the soil profile (dry phase).

The millet crop in this particular experimental plot was planted at the end of June (25th of June) and harvested at the end of September (21st of September). At the beginning of the rainy season, the soil water content of the profile averaged $0.03 \text{ m}^3 \cdot \text{m}^{-3}$ (curve of 24th of June) as a result of the prolonged dry season. The value of $0.03 \text{ m}^3 \cdot \text{m}^{-3}$ is

assumed to be the minimum (wilting point) value at which no water is available for plants.

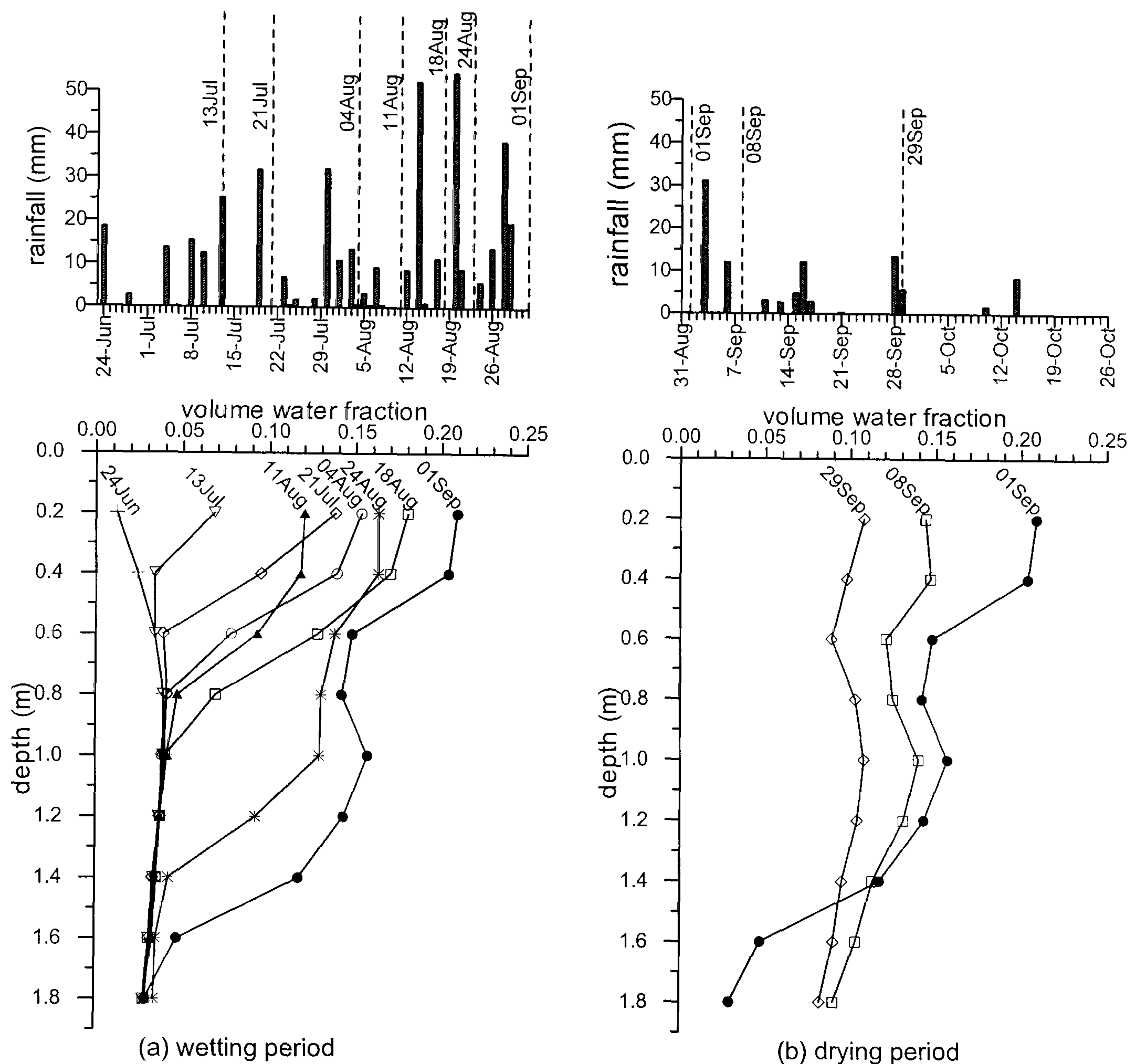


Figure 5. 1. Representative soil moisture profile under a millet crop and the respective seasonal daily rainfall. (From Hess, pers. comm.)

The curve of 13th of July was measured a few days after the onset of rains. The wet upper layers provide water for germination and the initial development of the crop. As rainfall continues, the wetted topsoil allows for the development of the roots. The infiltrated water remains preferentially in the upper part of the soil profile until the water content reached values above $0.10 \text{ m}^3 \cdot \text{m}^{-3}$, then the hydraulic conductivity becomes sufficiently high to permit the transmission of water to deeper layers. After 24th of August, heavy rainfalls bring the entire profile water content close to field capacity and water moves quickly downwards. Note that from 8th and 29th September water drains through the bottom of the measured profile. After 1st September soil water

content in the upper profile decreases due to the plant water uptake (roots are at their maximum depth) and drainage occurs from the bottom of the measured profile.

The soil moisture curves in figure 5.1 are taken to show a representative pattern of redistribution in a typical semi-arid soil profile where during the early crop stages water stays in the upper soil layers for a long time. This characteristic allows for the development of crops even in a dry soil profile with a high soil moisture deficit at the beginning of the season.

The approach utilised in this research attempts to reflect this behaviour keeping a proportion of water near surface and making it available for the shallow roots. Therefore, a preliminary qualitative analysis of actual soil moisture profiles goes some way towards the test of the physical plausibility of the model.

5.2.2. The SAMBA model output

The SAMBA model parameters were set based on the information provided by Grema and Hess (1994), Hess (1999). They are:

- The soil water retention properties (field capacity and permanent wilting point) are estimated based on the measured soil water curves (figure 5.1). The permanent wilting point adopted is $0.03 \text{ m}^3.\text{m}^{-3}$, assuming that at this point no water is available for plants. Field capacity is estimated as the average of the measurements taken on 1st and 8th September, resulting in a value of $0.13 \text{ m}^3.\text{m}^{-3}$. The values adopted are close to values suggested for sandy soils (e.g. FAO publication 56), although the soil has been classified as a sandy loam. Hess (1999) also pointed out this aspect when he estimated field capacity as approximately $0.09 \text{ m}^3.\text{m}^{-3}$. Therefore, despite the silt and clay content, it is assumed that this soil is highly permeable and having properties close to those of a sand. Based on a permeable soil, a factor of near surface storage equal to 0.45 is utilised.
- The initial conditions were based on the assumption adopted in Section 4.2 that the soil moisture remains unaltered between the rainy seasons. However, as daily rainfall from the previous years was not available, the model was run using the same year (1992) repeatedly for a sequence of 5 years.

- The surface runoff parameters utilised are the same matrix values presented in table 3.1 for a sandy soil. Runoff was taken into account despite the assumption of zero runoff adopted by Grema and Hess (1994). However, as the authors concluded, this assumption was inappropriate since runoff was visually observed as a result of rainfall intensities greater than 60 mm.h^{-1} .
- The crop growth stage durations, planting and harvest dates and maximum root depth (1.3 m) are set according to the values suggested by Grema and Hess (1994). Crop coefficients are the same values as utilised in Section 4.2 for a millet crop.
- Finally, the daily reference evapotranspiration is estimated based on the ET_0 calculated by Grema and Hess (1994). However, due to the high variability of the daily values, a monthly averaged series with linear interpolation was utilised. An analysis of the effect of the use of daily values derived from long-term average on the water balance is presented in the Chapter Six.

Figure 5.2 shows the resulting outputs from the SAMBA model.

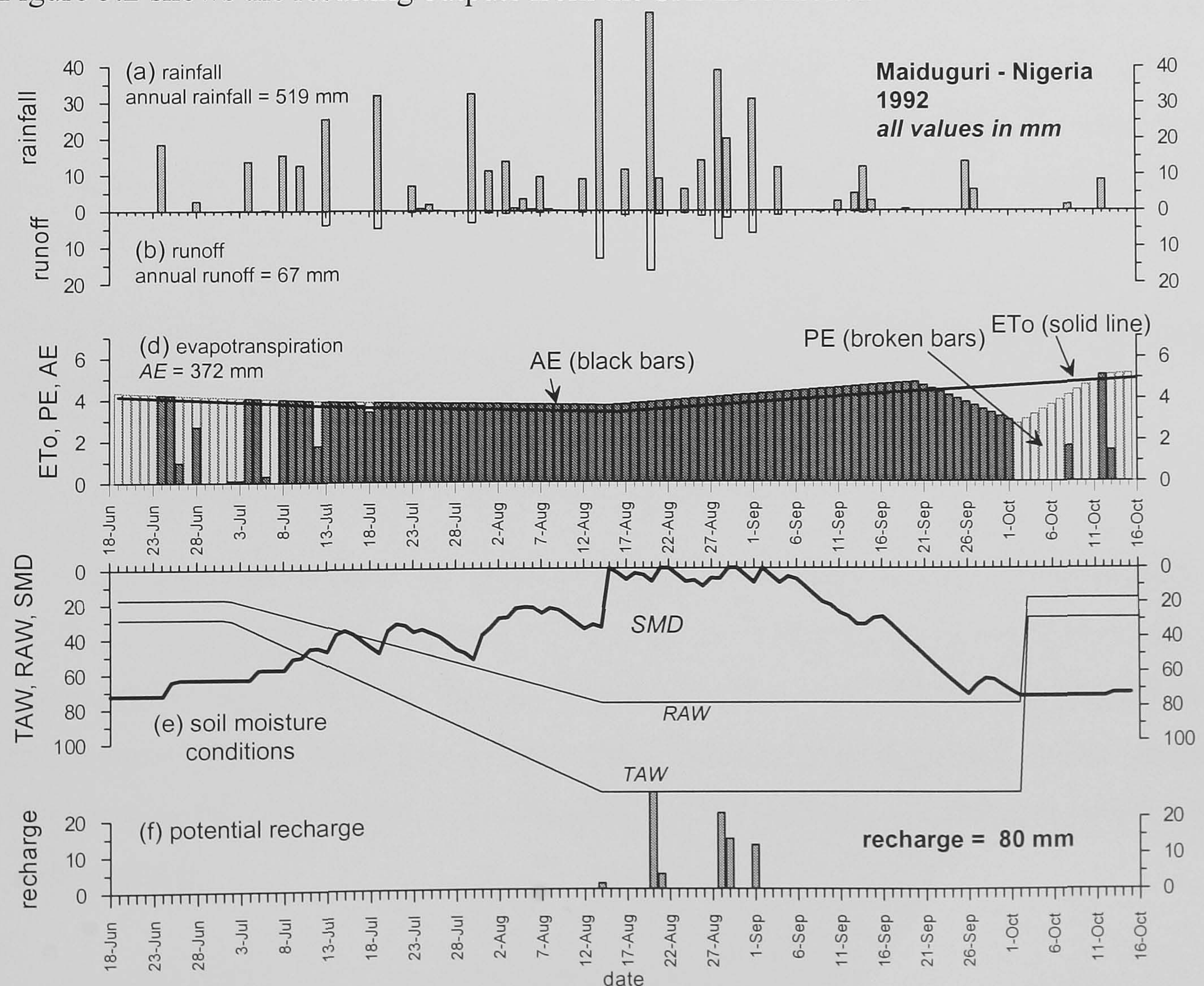


Figure 5. 2. Predicted values from the SAMBA model using the information from the field trial at Maduguri. Crop season of 1992.

This preliminary analysis shows that recharge below the maximum root depth (1.3 m) is estimated as 80 mm in the period by the SAMBA model, using the assumptions and parameters described above.

5.2.3. Comparative analysis between modelled and measured soil water content

This section compares the observed soil water content from the Grema and Hess measured profile (1.9 m) and the modelled soil water content from the SAMBA model to the same depth. The objective of this analysis is to verify if the modelled water balance follows the patterns observed in the field trial.

Soil water storage estimated from the measured water content

The total storage of water in the measured soil profile was calculated based on the soil water content curves shown in figure 5.1. The soil water storage was calculated by integrating the measured water content at each sub-interval to the maximum measured depth of 1.9 m. The first 10 cm of soil is not considered in this calculation due to neutron probe measurement limitations in top layers. The total soil moisture deficit was calculated based on the amount of water necessary to bring the whole profile to field capacity, that is, $1.8 \text{ m} \times 0.13 \text{ m}^3 \cdot \text{m}^{-3} = 234 \text{ mm}$.

The total soil storage for each soil profile measurement is presented in the following figure 5.3.

Soil water storage modelled by the SAMBA model

The modelled water content to a depth of 1.9 m was estimated by the SAMBA model using the same parameters of figure 5.2 except the initial conditions which was set to represent the initial soil water content measured by Grema and Hess at the start of the crop season. That is, soil water content close to wilting point as shown by the profile measured on 24th June (figure 5.1). The solid line in the figure 5.3 shows the predicted water content.

Comparative analysis between modelled and measured soil water content

Figure 5.3 shows the measured water storage (observed values) and that predicted by the model. The water content and the resulting soil moisture deficits are consistent with the observed values. Water content gradually increases with rainfall until it peaks at the end of August. The model is sensitive to dry spells as can be observed in the fall of the curve around 4th and 11th August. The end of the wetting period is well predicted by the model and it reasonably represents the period when field capacity is exceeded.

Although the objective of this analysis is not to fit statistically the modelled curve to the observed values, as in a conventional calibration procedure, some insights can be drawn from the graph. For instance, the model appears to overestimate the water content until it reaches field capacity. This suggests that actual evapotranspiration might be underestimated.

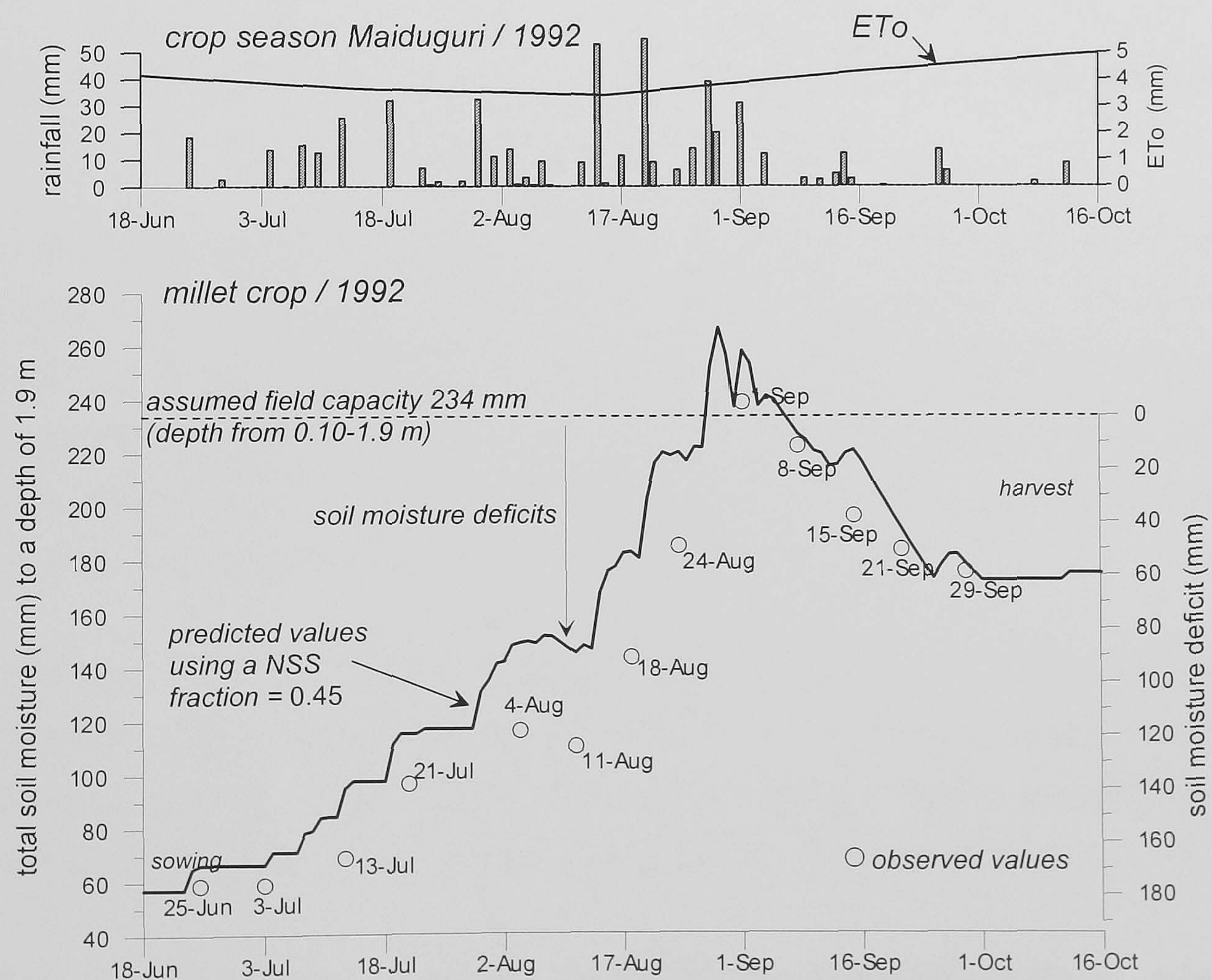


Figure 5. 3. Observed and predicted water contents in a sandy soil under a millet crop field experiment in semi-arid Nigeria.

In order to investigate this issue, a larger factor of near surface storage was utilised, thus attempting to increase the amount of water available for evapotranspiration. Figure 5.4 shows the resulting modelled water content using a fraction equal to 0.75.

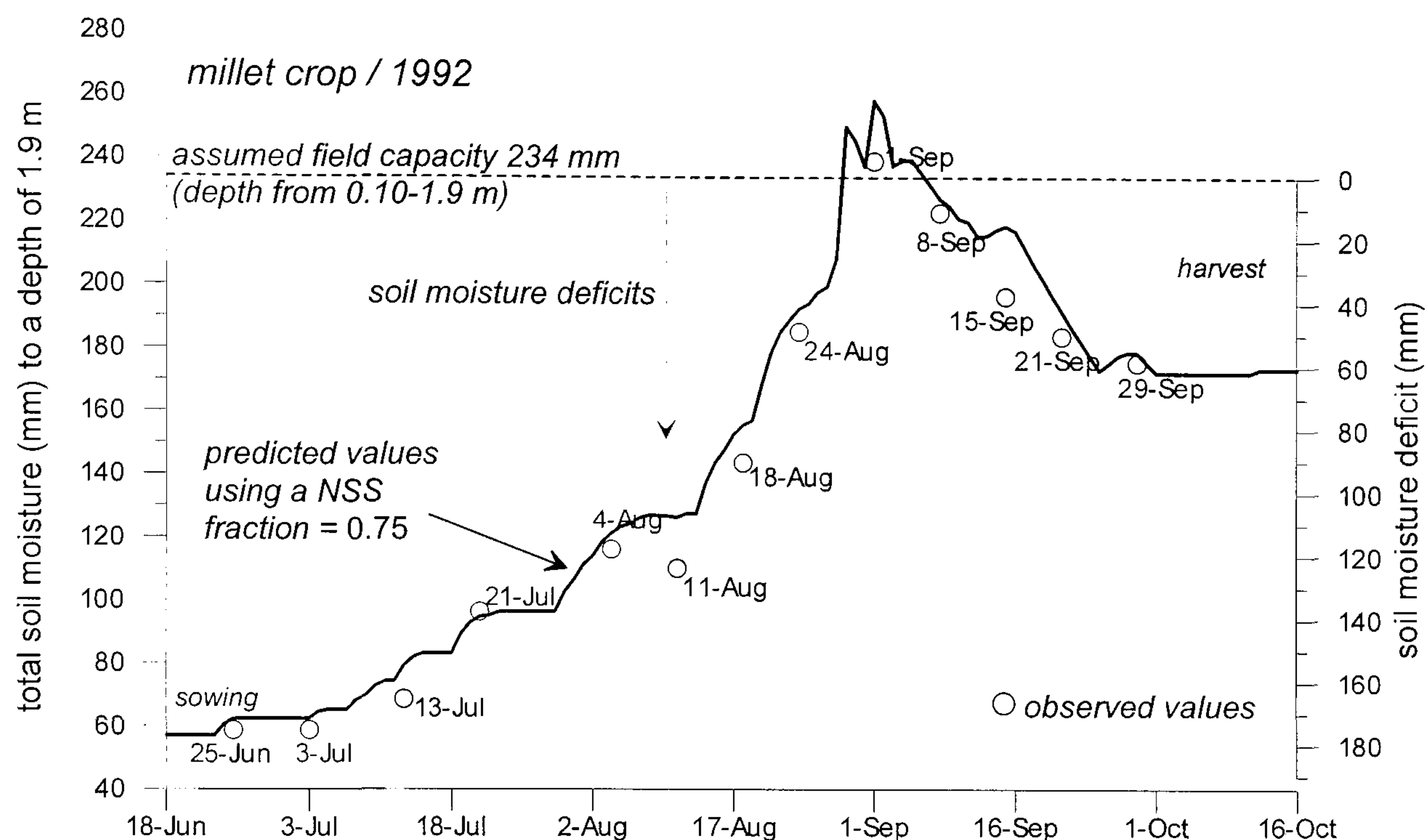


Figure 5. 4. Predicted soil water content from the *SAMBA* model using a factor of near surface storage 0.75.

By increasing the factor of storage, the predicted water content appears to better represent the field values in the early and mid-season. Although the soil shows a high permeability, a larger factor of storage can be supported by the observation of the soil water profiles in figure 5.1. During the early stages, the downward movement of the wetting front is very slow due to the dryness of the lower soil layers. Consequently, more water is available near surface for the shallow roots.

In addition, a larger fraction of storage may account for the soil evaporation losses from the shallow surface. The year 1992 was significantly wetter than the average during the mid-season and with a high frequency of rainfall during the entire growing season. Therefore, the soil surface was kept wet leading to high soil evaporation rates.

5.2.4. Comparative analysis of modelled and estimated drainage below a maximum root depth (potential recharge).

The previous analysis illustrated by figures 5.3 and 5.4 shows the soil water content measured in the soil interval of 0.10 - 1.9 m. However, the effective millet roots are

unlikely to reach the latter depth (Payne et al., 1991; Rockström et al., 1998). Therefore, the estimated drainage calculated by Grema and Hess (1994) is compared to recharge modelled by the SAMBA model as shown by figure 5.2.

Grema and Hess (1994) assume that drainage beyond a depth of 1.3 m is the sum of the change in water content between 1.3 m and the bottom of the measured profile (1.9 m), during the period when the soil water content at 1.9 m remains unchanged (corresponding to 1st September in figure 5.1). When the soil water content at 1.9 m ceases to be negligible, drainage is considered as the water flux below 1.9 m, estimated using the method of Klaij and Vachaud (1992). This method assumes conditions of unit hydraulic gradient at the bottom of the profile, and hence drainage equal to the unsaturated hydraulic conductivity corresponding to the water content of the bottom of the measured profile. However, as mentioned in section 2.5.6, this method is strongly affected by the uncertainties associated with the sensitivity of the function $K(\theta)$ to θ , the soil water content. For instance, Grema and Hess (1994) have determined a function $K(\theta)$ for this particular experiment resulting in a hydraulic conductivity range of 0.1 to 10 mm.day⁻¹ over the soil water content interval of 0.05–0.2 m³.m⁻³. Therefore, these values must be viewed with caution as the assumptions made by the method and the sensitivity of the relationship between K and θ can lead to a miscalculation of drainage.

Figure 5.5 shows the cumulative drainage below the maximum root depth estimated by Grema and Hess for the millet crop (circles). The dashed curve shows the cumulative drainage below 1.3 m estimated by the SAMBA model using the parameters described previously (factor of storage near surface equal to 0.45).

Although the numerical values of drainage from the two methods are different, the model is able to predict the moment when water starts to drain through the root zone (between 18 and 24 August). From 24 August to 8 September, the modelled drainage increases rapidly as a result of the heavy rainfall in the period. The model also predicts these events, however, not in a gradual manner but as instantaneous output of water typical of the soil water balance method based on the concept of field capacity.

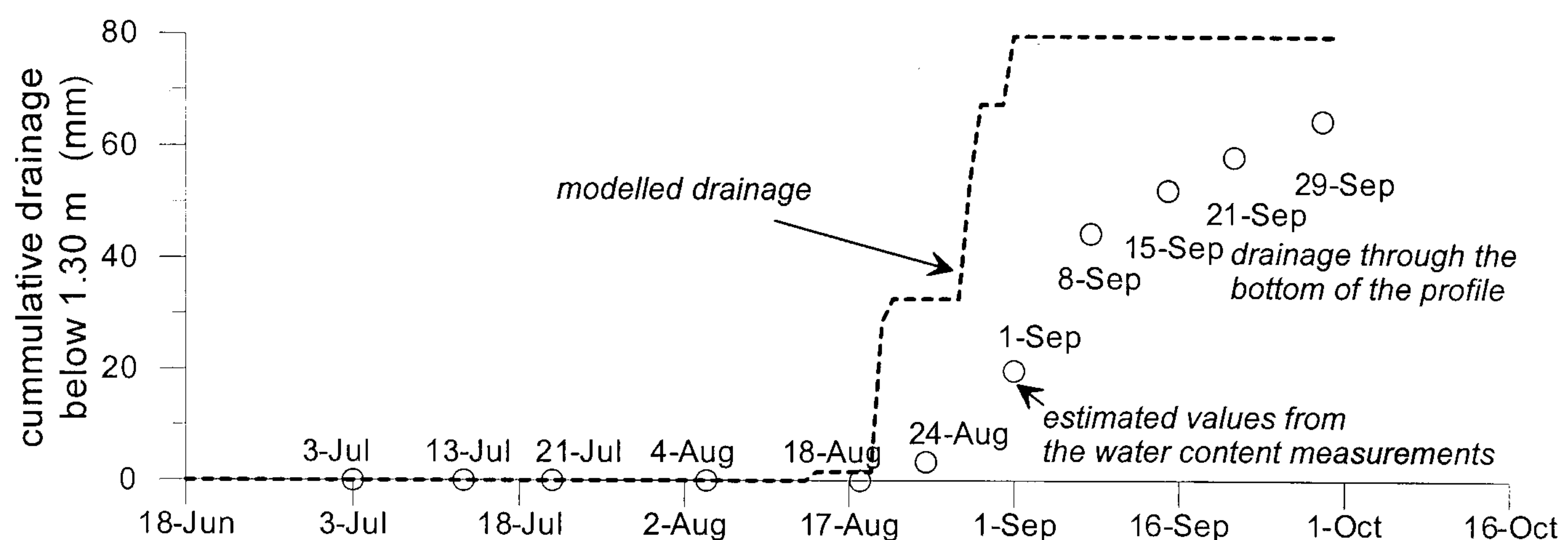


Figure 5. 5. Cumulative drainage below the root zone (1.3 m) in a millet crop plot estimated by the model and by the soil water content measurements.

The overall recharge estimated by the model (79.7 mm) is 23% greater than the drainage calculated from the soil water profiles (64.7 mm). However, in terms of proportion of the seasonal rainfall they are similar (15% and 13% respectively)

5.3. QUALITATIVE ANALYSIS OF ALTERNATIVE INFORMATION FROM HISTORICAL RECORDS

The analysis above showed that modelled potential recharge and water storage are plausible when compared to actual conditions. However, as the concept of plausibility implies, a more lateral approach can be taken in order to verify the credibility of the conceptual model using as large a range of outputs as possible.

The following analysis illustrates how existing information, taken for other purposes than modelling, can be utilised to investigate the credibility of the conceptual model. Alternative information can be inferred, such as crop survival and overall crop yields. These outputs may be corroborated by field observations which can give some kind of lateral information about the real situation at a particular site during a particular period of time.

An example of field observations which contain valuable information about the physical system is the work of Mortimore (1989) in the semi-arid region of Northeast Nigeria. In his book, Mortimore presents a chronicle of several years in the life of the small village of Dagaceri (see figure 4.1) from where a range of information was gathered, such as data about rainfall and crop yield for each crop season. The author gives a brief

narrative for each year in which he describes the overall crop yield and its relation to the rainfall patterns.

These observations from a typical semi-arid village supply a major source of information which can be utilised to corroborate the SAMBA model.

The present analysis compares some selected outputs from the SAMBA model with Mortimore's field observations. The information about rainfall is given in a descriptive manner in Mortimore's work. However, as the model requires daily data, the long-term data from Nguru (see Section 4.2) are utilised based on the fact that Dagaceri is geographically not far from Nguru (approximately 30 Km Southwest). Moreover, the rainfall pattern in Dagaceri is similar to the pattern observed in Nguru, that is, a unimodal distribution with the rainy season starting in May-June and finishing in September or October.

The crop yields are directly related to the rainfall patterns, that is, a major rainfall deficiency in June means late planting and delayed growth; in July and August, wilting and loss of yields in the early millet; and in September, partial or complete loss of guinea corn or cowpeas.

Table 5.1 summarises the information about crop yield (millet) and rainfall patterns in Dagaceri using the qualitative information extracted from Mortimore's report. In addition, a brief summary of the rainfall distribution in Nguru is presented.

An overall view of table 5.1 shows that the rainfall pattern was quite similar in Dagaceri and Nguru, thus allowing the information about the crop yields to be related to the model outputs. Moreover, the table shows the direct relationship between rainfall and crop yield in Dagaceri. For example, in 1974 a good rainy season provides enough water for satisfactory millet development. On the other hand, in the following year a late start provides good conditions for disease and insect attacks, thus damaging the final yield.

Table 5. 1. Summary of crop yields and rainfall pattern in Dagaceri (from Mortimore, 1989) and Nguru (from the Nguru data series)

Year	Dagaceri Crop aspects	Dagaceri Rainfall pattern	Nguru Rainfall pattern
1972	Very low millet yield.	Drought.	Low annual rainfall (248 mm).
1973	Disastrous millet yield.	Drought.	Low annual rainfall (258 mm).
1974	Satisfactory millet yield; late start.	Good rainy season; low in June.	Above average; low in June; heavy storms in July and August.
1975	Poor millet yield; damage to young millet due to diseases and insect attacks; late start.	Low in June; good rains in September.	Reduced and sparse early rains; good in August and September.
1976	Good grain harvest.	Good rains in June, July and September; low in August.	Low in June and August.
1977	Poor harvest of grains (and millet); late start.	Low in June and July; peak in August; premature end.	Good in July and August; strong peak in July; premature end.
1978	Good millet harvest; timely start.	Start in May and June; heavy falls in July; premature end.	Start in June; sparse early rains; good in July; low in August; premature end.
1979	Good grain yield.	Good rains; late end.	Above average; good distribution; late end.
1980	Millet suffered grasshopper damage in the ripening; poor yield.	Late start; deficient August.	Low and sparse early rains;
1981	Poor millet yield; late start.	Erratic short-lived rains; low in June.	Erratic distribution; low in August.
1982	Satisfactory millet yield.	Late start; good in August and September; late end.	Late start; regular distribution; heavy rainfall in September.
1983	Very poor millet yield.	Late start followed by a dry spell of 25 days; light rains.	Low and sparse rains; lowest annual rainfall of the period (235mm).
1984	Poor millet yield	Dry August	Low and sparse rains.
1985	Poor millet yield; little better than 1984.	Below long-term mean; well distributed.	Late start; below long-term mean; fairly distributed.
1986	Millet yield close to nil due to attack of pests (rodents and grasshoppers).	May be enough for millet; premature end.	Low annual rainfall (241 mm); low in July and August.

In an attempt to incorporate this type of information into the study of plausibility, two alternative outputs were generated. First, the actual evapotranspiration is related to potential evapotranspiration during the crop season. It is expected that a healthy crop, with no water supply limitation, will transpire at or near to potential rates. The second output shows the number of days when the soil moisture deficit is bigger than the amount of water readily available for evapotranspiration (*RAW*) during the stage of crop development (35 days). This output assumes that a crop under water stress during its development stage will not give an optimum yield.

The information about crop yield is split into two categories in order to carry out a comparative analysis with the model outputs. Thus, yield is divided into the years with a

satisfactory crop yield and the years when crop failure occurs or crop yield is significantly low (years of “hungry”). The relationship between the outputs from the model and the categories of crop yield from table 5.1 is shown in figure 5.6.

The SAMBA model was set up using the parameters described in Section 4.2. Runoff was calculated using the coefficients in table 3.1. A near surface storage fraction of 45% was utilised. The daily rainfall data is from Nguru.

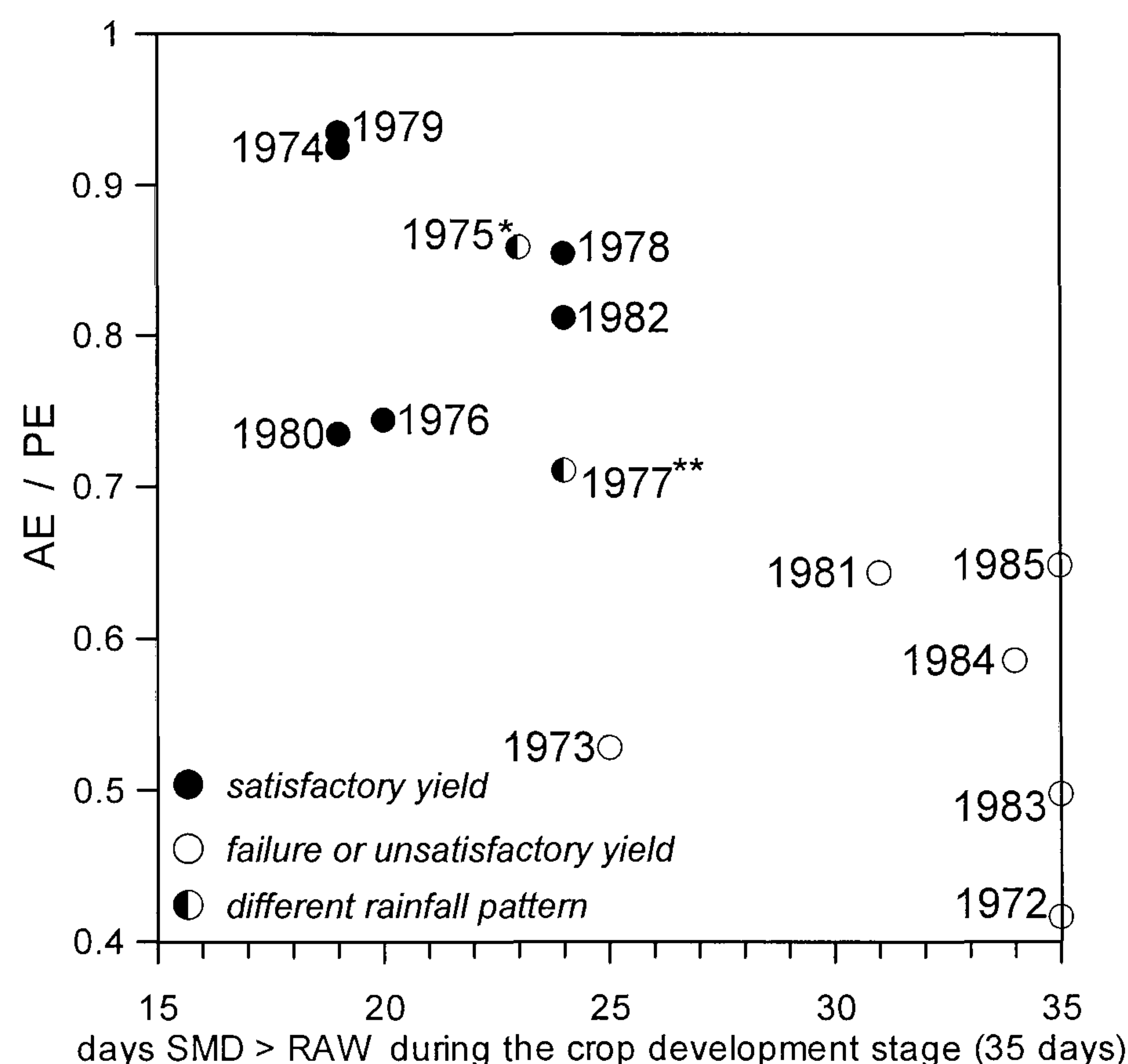


Figure 5. 6. Relationship between two alternative outputs from the SAMBA model and field observations from Mortimore (1989).

* diseases and insect attack damage the yield.

** significant different rainfall pattern.

Figure 5.6 shows that for the years with a satisfactory millet crop, the modelled actual evapotranspiration during the crop season is close to the potential rates. On the other hand, in years with a reported failure or unsatisfactory crop yield, the relationship between the modelled *AE* and *PE* is significantly smaller (less than 70%), thus reflecting the water stress conditions.

The other alternative output shown in the axis X of figure 5.6 is the number of days when the soil moisture deficit is bigger than *RAW* during the development stage. The number of days in which the crop is under water stress is greater for the years with a restricted crop yield.

The exceptions are the years of 1975 and 1977. The comparative analysis in these years is prejudiced by other factors. In 1975, insect attacks and diseases damaged the crop yield. The model cannot predict this type of scenario. Moreover, in 1977 the rainfall patterns were different in Dagaceri and Nguru. In Nguru, rainfall peaked in July, thus supplying water for the young millet. In Dagaceri, rainfalls in June and July were low and sparse, affecting crop development.

Therefore, the qualitative analysis above suggests that the model can reasonably represent the physical mechanisms which affect the millet crop yield. In addition, a closer look at two particular crop seasons shows how the model can provide insights into the soil-water system (figure 5.7).

In 1983, the crop season was considered one of the worst seasons in Dagaceri, as a result of the low rainfall rates. The millet yield was less than 10% of the overall expectations and in several plots it completely failed. On the other hand, in 1974 the millet crop yield was satisfactory (table 5.1).

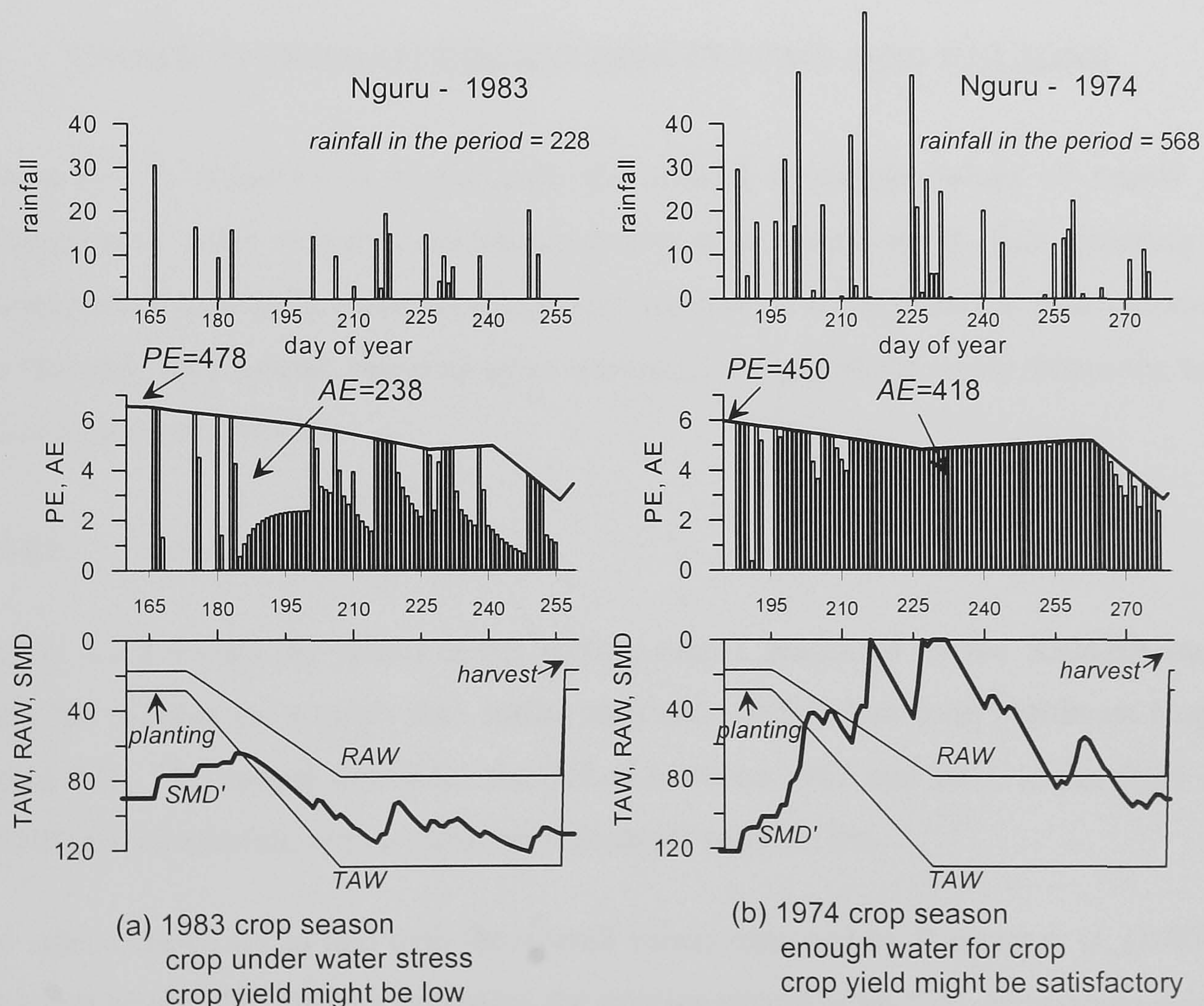


Figure 5. 7. Modelled crop season in Nguru for two years. The model outputs agree with the observed crop yield in Dagaceri. All values in mm.

Figure 5.7.(a) shows the results from the modelled crop season in 1983, using the rainfall series from Nguru. The modelled actual evapotranspiration and the soil moisture deficit suggest that the millet crop would not develop adequately. The crop is under water stress during the whole season (actual evapotranspiration is 50% of the potential during the crop season). During ripening and late stages, the soil water content is close to wilting point, certainly damaging the overall crop yield. Note that rainfall in 1983 is one of the lowest of the whole series (seasonal rainfall of 228 mm).

In 1974 (figure 5.7.b), a good and well-distributed rainfall supplies the soil with enough water for the development of the millet crop. The modelled outputs show a low soil moisture deficit during the majority of the crop stages. During crop development and maturing, the soil water content is readily available for the roots ($SMD < RAW$).

Therefore, insights into the system as illustrated above, which can be drawn out from alternative model outputs, are corroborated by lateral field observations.

5.4. INSIGHTS FROM OTHER STUDIES IN SEMI-ARID REGIONS

This section's objective is to compare the overall estimated values of runoff and recharge with other recharge estimation studies in semi-arid areas, first, pointing out representative studies in areas which present equivalent settings as the data utilised in this research; and second, showing other semi-arid regions of the world where the same processes may be present.

5.4.1. Surface runoff

Section 4.3.2 shows the values of the surface runoff generated by the SAMBA model using the settings for a sandy soil, millet crop and rainfall data from Northeast Nigeria (figure 4.9). The annual runoff for the period between 1962 and 1997 varies from 2 to 15% of annual rainfall, with a mean annual runoff equal to 9%.

The values above are in line with the overall values obtained by Peugeot et al. (1997) in semi-arid Niger. The authors estimated the average runoff to be 17% of seasonal rainfall for a millet crop with no tillage and 5% for a millet crop plot with traditional tillage,

both located on sandy hillslopes with a topographic gradient of a few percent. The climatic, soil and crop conditions present in this experiment are similar to the conditions utilised in the present research, that is, rainfall with a unimodal distribution (annual rainfall for the two years studied were respectively 430 and 470 mm) and a sandy soil.

Rockström and Valentin (1997) also measured runoff for experimental millet plots during two seasons in Sub-Sahel Niger. Millet was cultivated in deep, sandy hillslope plots equipped with a collector system which was observed after each storm. The results show an average runoff varying from 6 to 13% of total rainfall, depending on the gradient. The measurements were taken in wet years (596 and 517 mm respectively) with 35% of the events having an intensity exceeding 50 mm h^{-1} .

Lal (1991) reports some values of runoff from several studies of different crop systems. They vary from 1.5% to 20% of annual rainfall. However, these values have to be observed with caution because most of the measurements were made either on small plots or under simulated rains.

Therefore, the overall values of surface runoff predicted by the SAMBA model are plausible when compared with the runoff determined by other investigations in semi-arid areas.

5.4.2. Potential recharge

The results of the recharge estimation by the SAMBA model are compared with other studies using different methods such as chemical methods, water level fluctuations and numerical modelling. This comparative analysis begins with some selected studies related to field situations close to those so far utilised in this research (northeast Nigeria). The results from this region provide an opportunity to compare recharge estimates obtained from different techniques.

Figure 5.8 shows the relationship between annual rainfall and modelled annual potential recharge using the parameters from Section 4.2 and daily rainfall from Nguru (1962-97).

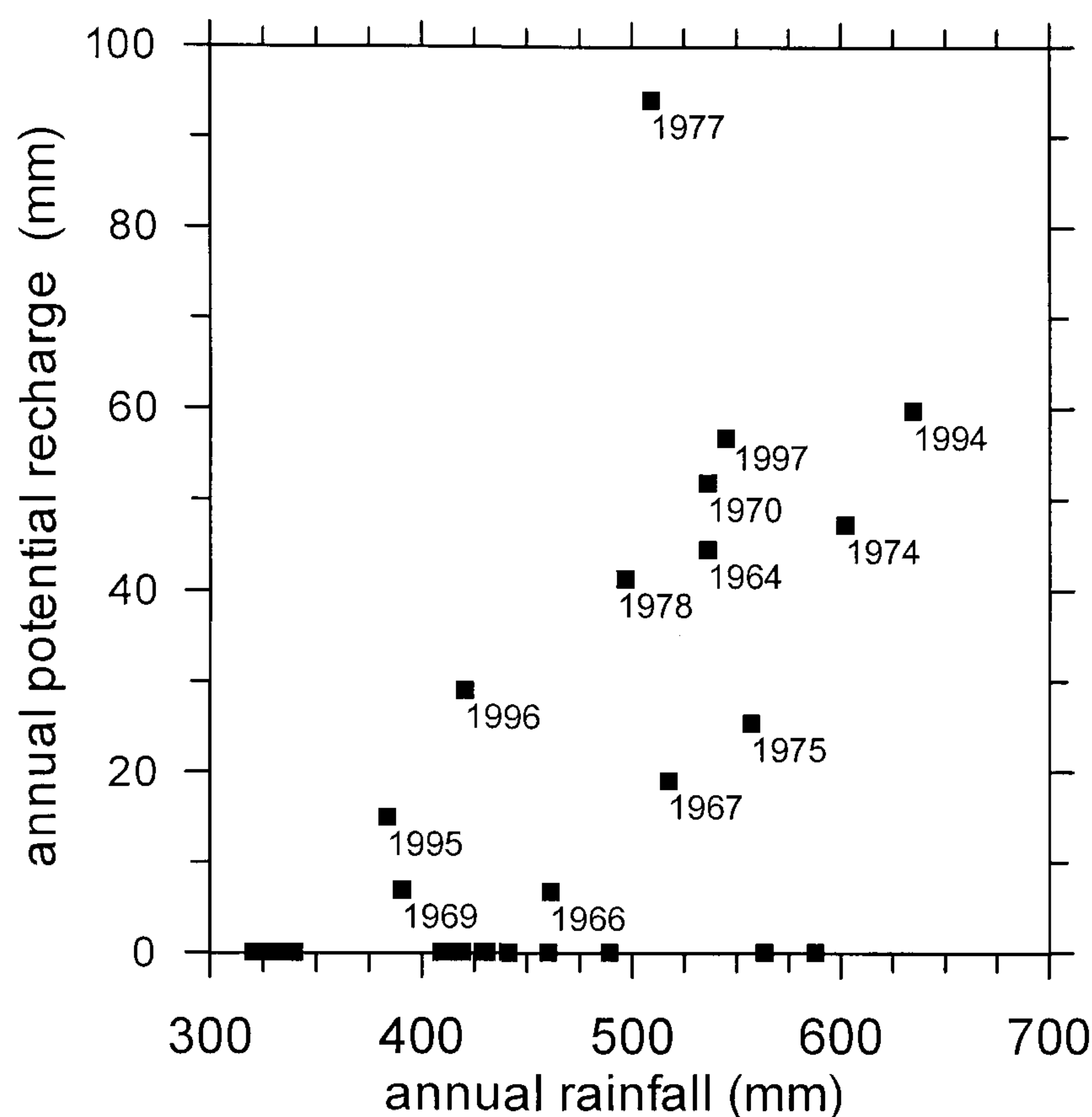


Figure 5. 8. Rainfall and modelled potential recharge for Nguru, Northeast Nigeria under a millet crop on a sandy soil using the *SAMBA* model.

Recharge varies significantly from year to year depending upon distribution, intensity and frequency of rainfall events.

For the 36 years (1962-97), recharge varies from zero to 95 mm/year (22% of mean annual rainfall in Nguru, 431 mm). The mean recharge for the total period is 14 mm (3% of mean annual). Note that without the extreme 1977 value, recharge varies from zero to 60 mm/year.

Rockström et al. (1998) estimated recharge in an experimental field cultivated with millet in semi-arid Niger using field measurements (climatic data, runoff collection and neutron probe measurements) and modelling. The model utilised for the water balance was a numerical model based on numerical solutions to the Richard's equation in a layered soil profile (14 layers). The resulting deep drainage (below a depth of 160 cm) for three years (1996-1998) ranged from 160 to 290 mm (33-48% of annual rainfall) when runoff was considered negligible. When runoff was included, deep drainage varied from 100 to 198 mm (20 to 33% of annual rainfall). These high values may in part be explained by the: (i) annual rainfall above average for the whole period (488, 517 and 596 mm for 1996, 1997 and 1998 respectively), (ii) setting of parameters during the parameterisation of the model, such as the definition of a reduced fraction of total available water for evapotranspiration ($0.6 \text{ m}^3 \cdot \text{m}^{-3}$) based on the measured soil hydraulic properties for a very sandy soil.

Other studies using cultivated millet plots in the Sahel region are the work of Klaij and Vachaud (1992) and Grema and Hess (1994), respectively in Niger and Nigeria. Both studies used neutron probe measurements of soil water content during one crop season. Recharge ranged from 12% (Grema and Hess) to 48% (Klaij and Vachaud) of annual

rainfall (440 and 470 mm/year respectively). However, surface runoff was considered zero in both works, despite it being visually observed in the field. Also, hydraulic assumptions had to be applied in order to deal with the drainage from the bottom of the measured profile caused by the high infiltration rates of the very sandy soil.

Regional studies such as the work of Carter (1994) and Edmunds et al. (1999; 2002) present values more consistent with the values estimated by the SAMBA model. Carter (1994) based on modelling and on a regional water balance of the Manga Grasslands in northeast Nigeria, estimates the annual recharge to the shallow aquifers within the range 30-60 mm (10-20% of mean annual rainfall). This value is corroborated by the study of Edmunds et al. (1999), who estimated direct recharge in the Manga Grassland region as 44 mm/year using the chloride mass balance technique. Edmunds et al. (2002) using the same chemical technique and based on data collected in 360 shallow wells, estimated direct regional recharge in northern Nigeria as 43 mm/year.

A similar value was obtained by Leduc et al. (1997) using hydrodynamic observations from water-table records in southwest Niger. The authors conclude that a recharge of 50-60 mm/year (10% of annual rainfall) would explain the observed mean inter-annual rise of the regional water-table. Further investigation carried out by Leduc et al. (2001) estimated the mean regional recharge as 20 mm/year.

Finally, numerical modelling was utilised by Milville (1991) to estimate recharge in another Sahelian region (Burkina Faso). The results show average recharge rates ranging from 47 mm/year (8% of average rainfall 551 mm: 1985-1988) to 107 mm/year (15% of average rainfall 720 mm: 1953-1988).

Extending the comparative analysis to worldwide areas with semi-arid climates, the work of Rangarajan and Athavale (2000) in several basins and watersheds of India shows values in line with the values obtained here. The authors estimated recharge through an extensive application of the tritium injection method over 25 years in 17 major river basins of India. The overall recharge ranged from 24 to 198 mm /year (4.1 to 20% of the average seasonal rainfall). However, this range includes humid areas with seasonal rainfall greater than 1000 mm. Taking into account the measurements in semi-arid areas only the regional recharge is from 4 to 15% of annual rainfall.

Kennett-Smith et al. (1994) present the result of several recharge estimation studies carried out in southwestern Australia where mean annual rainfall varies from less than 255 to 580 mm/year. The estimation of potential recharge was carried out using the chloride method and water balance techniques for several sites with different soil and vegetation conditions. The results varied from 2 to 40 mm/year (1 to 12% of mean annual rainfall).

Kitching et al. (1980) estimated recharge using several lysimeters constructed in the semi-arid region of Cyprus (mean annual rainfall 450 mm/year). The final figures show an annual recharge of 5 mm/year.

In conclusion, the estimates of recharge by the SAMBA model are consistent with the results obtained by other studies in semi-arid areas, including different conditions and methods of estimation.

CHAPTER SIX

VARIABILITY OF POTENTIAL RECHARGE

Chapter six explores some implications of the SAMBA model outputs showing how the model's parameters reflect the inherent complexity of the system. Variability of recharge as a result of climatic, vegetation and soil properties are investigated using field observations from a semi-arid region.

6.1. INTRODUCTION

The insights from the different studies mentioned in Section 5.4 illustrate the variation of potential recharge as a result of the complexity and heterogeneity of the soil-water system. Variability in soil type, vegetation characteristics, and climate result in variations in recharge. This issue is well documented and discussed in literature such as Simmers (1997), Lerner et al. (1990), Gee et al. (1994), Kennett-Smith et al. (1994), Gregory (1991) and Zhang et al. (1999a).

This research has so far considered the temporal variability of recharge through the utilisation of a long-term series of daily rainfall. However, further analyses can be carried out to investigate other aspects which add to the variability of recharge.

Chapter six presents a series of preliminary investigations with the purpose of exploring the sensitivity of the model's outputs to the field variability. Some aspects of model utilisation are explored illustrating some of ways in which the model can be applied to different field conditions which are common in semi-arid regions.

This exploratory analysis begins by assessing the effect of time averaging the reference evapotranspiration ET_o on the resulting soil water balance by comparing the modelled soil moisture deficit calculated using daily ET_o with an alternative simulation using time-averaged ET_o values.

A second analysis illustrates the variability of modelled recharge for different situations involving surface runoff. Three alternative conditions are considered: (i) when the net runoff is positive and water flows out the plot, (ii) when the net runoff is zero, and (iii) when the plot receives an inflow of water (runon) from an adjacent area (runoff is assumed to be absent).

The sensitivity of the model output to field variability is explored through an examination into how the modelled recharge varies as a function of the variation of selected parameters.

Finally, preliminary insights into the spatial distribution of recharge are developed using daily rainfall from 18 rainfall stations situated in the northeast arid zone of Nigeria (see figure 4.1, Chapter Four).

6.2. EFFECT OF TIME-AVERAGING OF REFERENCE EVAPOTRANSPIRATION ON THE SOIL WATER BALANCE

The weather and surface observations necessary to estimate the reference evapotranspiration ET_o are often limited in space and time. Consequently, daily values derived from the disaggregation of long-term means are often utilised.

This section investigates the effect of the use of averaged ET_o on the model output, comparing the resulting soil water deficits for two alternative situations:

- (a) using daily reference evapotranspiration from the Maiduguri meteorological station, Nigeria (see map figure 4.1). ET_o was calculated by the FAO modified Penman-Monteith equation (Smith et al., 1992) from 24th June to 31st October 1992 (Grema and Hess, 1994).
- (b) using daily values derived from the monthly mean of the values calculated in (a). The daily values are estimated by linear interpolation between the monthly means. Note that the mean for June is calculated from the 7 days available only.

The daily ET_o and the monthly means are presented in figure 6.1. Daily ET_o varies from 2.2 mm/day on 17th August (the height of the rainy season) to 5.4 mm/day at 16th October (end of rainy season).

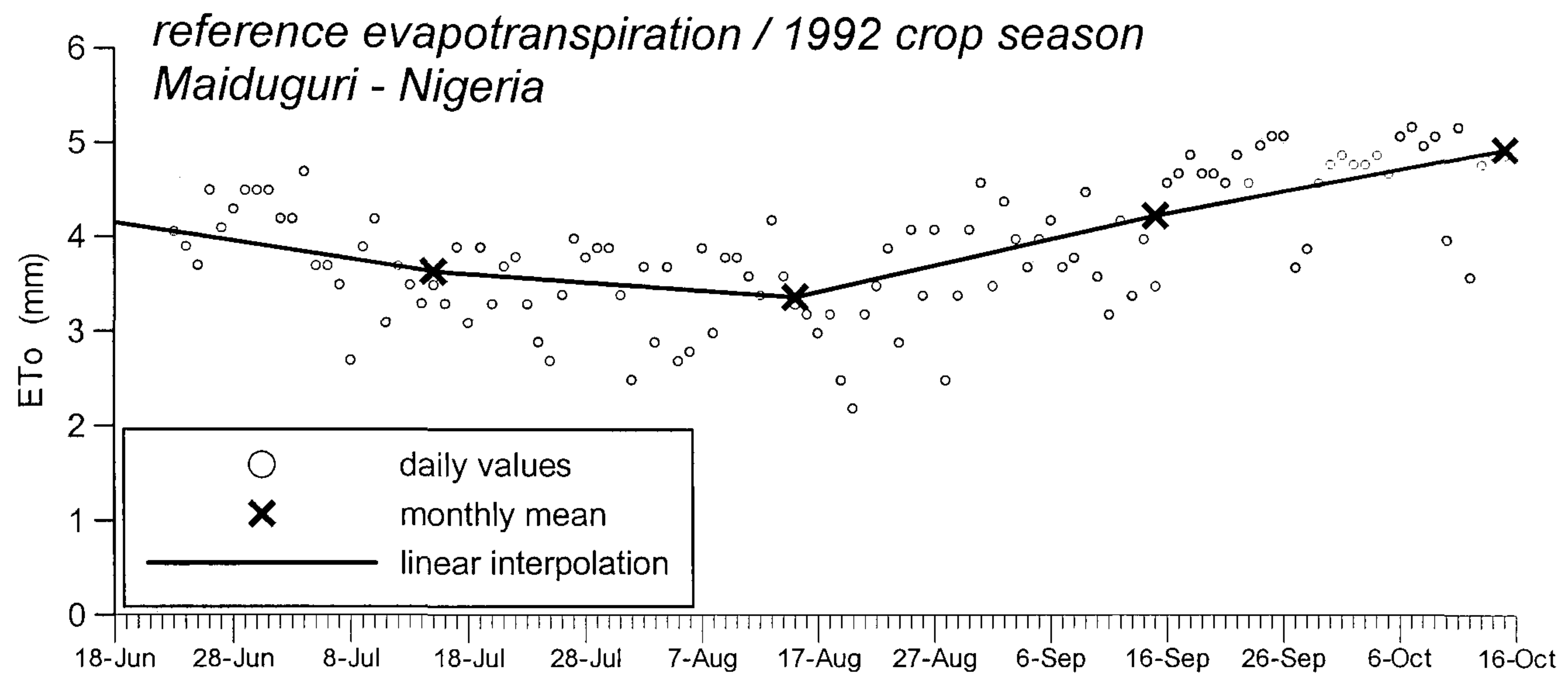


Figure 6. 1. Reference evapotranspiration estimated for the 1992 season in Maiduguri, Nigeria (from Grema and Hess, 1994)

Figure 6.2 shows the modelled soil moisture deficit for the two situations. The soil and crop conditions utilised by the model are the same as in Section 5.2.2.

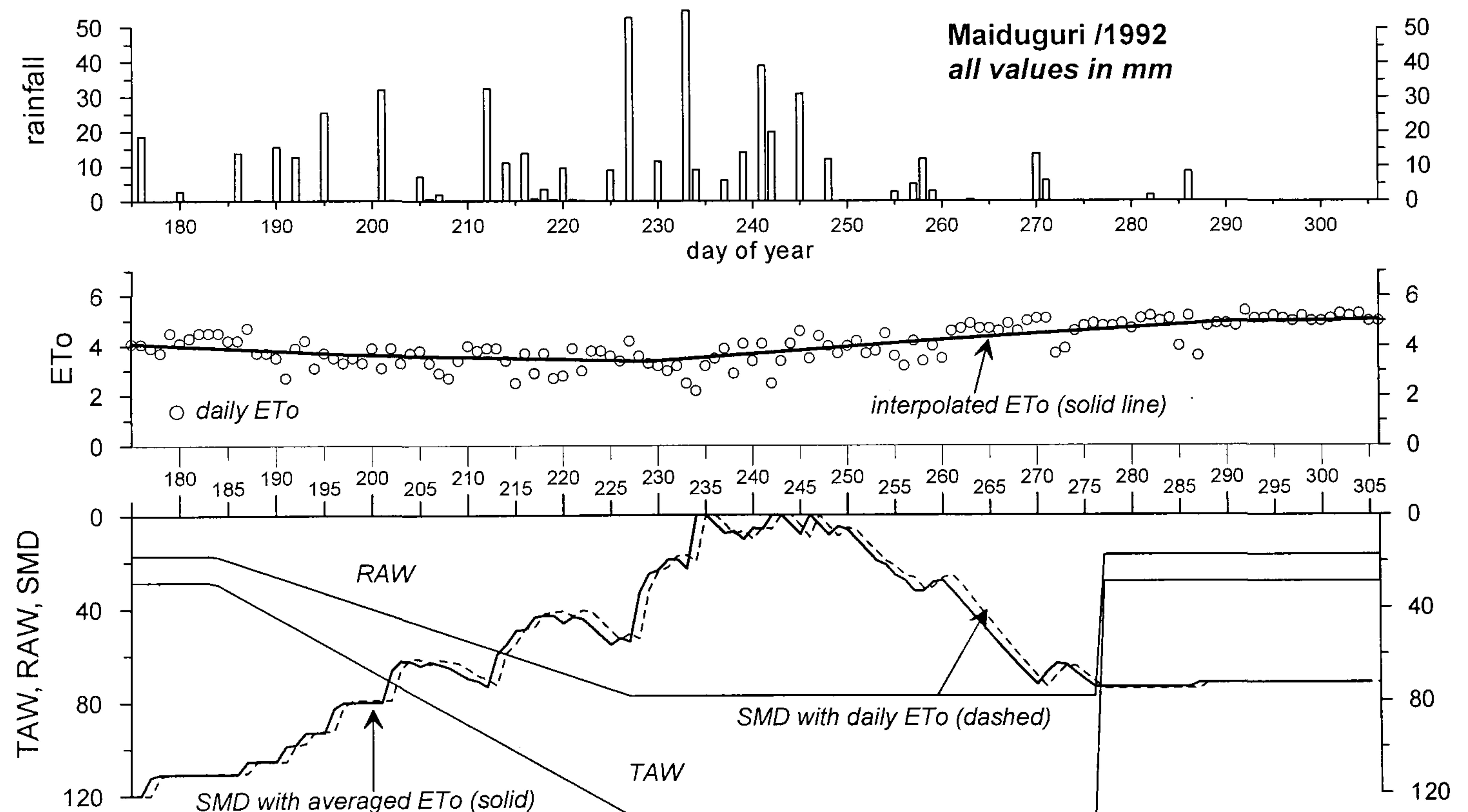


Figure 6. 2. Soil water balance for a millet crop and the resulting soil moisture deficit curves from the SAMBA model using daily estimated ET_o and monthly mean ET_o varying linearly. Daily rainfall input is common to both model runs.

Note that the resulting curves are very similar and the difference between the daily values is not significant ($p < 0.01$). The maximum difference of daily *SMD* is 3 mm only.

The estimated recharge using the daily observed *ET_o* and using the interpolated values are 69 mm and 66 mm respectively. This result suggests that the soil water balance is relatively insensitive to the averaging of the potential evapotranspiration.

Fowler (2002) came to a similar conclusion when comparing daily potential evapotranspiration calculated from the weather records to alternative methods of estimating daily values from averaged values. By running a single-layer soil water balance model for 13 years, Fowler concluded that the use of a simple disaggregation function, such as dividing the monthly *ET_o* by the number of days, is a reasonable alternative to daily observations.

6.3. WATER BALANCE FOR THREE SURFACE RUNOFF SITUATIONS

Surface runoff is the product of the net balance between the outflow of water leaving a plot (called simply runoff) and the inflow from an adjacent area (runon). For further information about surface runoff see Sections 3.3, 4.3.2 and 4.4.1.

One of the assumptions of this thesis, based on field observations in semi-arid areas, is that surface runoff (net positive runoff) is an important physical process which has to be included in the soil water balance.

However, depending on local conditions the contribution of water from an upper adjacent area (runon) may be significant, such as occurs in localised ponding. Consequently the net surface runoff may be negative and runon has to be considered. For example, Rockström et al. (1998) observed a significant inflow of water (8 to 18% of annual rainfall) entering a millet crop plot in a field trial in semi-arid Niger, resulting from the upstream runoff produced on the adjacent non-agricultural land.

The occurrence of runon to and runoff from an area is associated with the high spatial variability of infiltration (Gaze et al., 1997). Leduc et al. (2001) have associated the overall increase in regional recharge in southwest Niger with the decreases in

infiltration capacity caused by changes in land-use (intense land clearing). The authors suggest that the increasing runoff from areas with lower infiltration rates led to increasing runoff to localised ponds, from which water percolated to the regional aquifer.

The studies above are examples of runoff as an important component of the soil water balance, resulting in an additional input of water.

The purpose of this section is to illustrate how increases in water availability caused by variations in the net balance of surface runoff affect the resulting recharge. Three scenarios are simulated:

- (a) A net runoff producing surface. Runoff is calculated by the approach described in the Section 3.3, using the matrix coefficients presented in table 3.1.
- (b) A situation when the net runoff equals zero. This condition assumes that the balance between runoff to and runoff from the crop plot is zero. Therefore, the amount of water that was assumed to leave the plot in situation (a) is no longer subtracted from the soil water balance. Additional water is thus available to the system.
- (c) A surface receiving an inflow of water from an adjacent area (negative net runoff). The amount of water added to the daily soil water balance is assumed to be equal to the daily runoff coming from a neighbouring plot with the same soil-vegetation conditions. Therefore, the daily incoming runoff has the same value as the calculated runoff but it is added to the soil water accounting instead of being deducted. Positive net runoff is assumed to be zero.

For the three situations above the model was run using the same soil crop parameters utilised in Section 4.2 (assuming a millet crop and a sandy soil). The surface with a positive net runoff (scenario a) is considered as the reference surface since it represents the conditions presented so far in this research.

The daily rainfall data utilised are from a series of 5 years (1993-97) at Nguru (Section 4.1.1). The mean annual rainfall for the period is 436 mm with a range of 333-634 mm. This period has alternating years with annual rainfall above and below the long-term

mean of 431 mm and it is taken to be typical of the temporal variability observed at the station. Table 6.1 shows the resulting water partitioning for each year of the period.

Table 6. 1. Water partitioning for the three surface conditions. Note runoff is given negative values and runoff positive. All values in mm/year.

Rainfall in Nguru		Scenario (a) (reference surface)				Scenario (b) (runoff = zero)			Scenario (c)			
Year	P	Roff	AE	ΔS	Rech	AE	ΔS	Rech	Ron	AE	ΔS	Rech
1993	333	-28	315	-9	0	333	-11	10	28	337	-14	38
1994	634	-94	400	81	60	406	81	147	94	413	81	234
1995	384	-53	393	-78	15	405	-77	55	53	407	-77	107
1996	421	-42	351	-1	29	359	-2	64	42	367	-2	97
1997	545	-64	412	12	57	424	15	106	64	434	20	155

The observed variation in recharge between the treatments is significant ($p < 0.01$). In the wettest year (1994), 92% of the additional 94 mm when shifting from the reference surface to the scenario of zero net runoff appears as recharge (87 mm). Note that actual evapotranspiration AE does not change significantly, suggesting that evapotranspiration is already at or closer to potential rates and nearly all additional water is directly transferred to recharge.

In a drier year this proportion decreases. In 1993, 28 mm of water is available when changing from a positive to zero net runoff. From this value, 10 mm (36% of the additional 28 mm) becomes recharge. The other 18 mm is utilised by the crop. However, when an additional 28 mm is added as runoff, all water goes to recharge. The small increase in AE shows that the crop does not need the extra water.

The overall figures show that the greatest proportion of the additional water available when shifting from a net runoff to net runoff goes to recharge.

Figure 6.3 summarises the figures of table 6.1 showing the difference between the mean values for the 5-year period. Figure 6.3.a shows the reference surface with a mean net runoff of 56 mm/year (12% of the mean rainfall P) and actual evapotranspiration equal to 463 mm/year (81% of P and 92% of the infiltrated water). Mean recharge of 32 mm/year is 7% of mean rainfall and 8% of the infiltrated water.

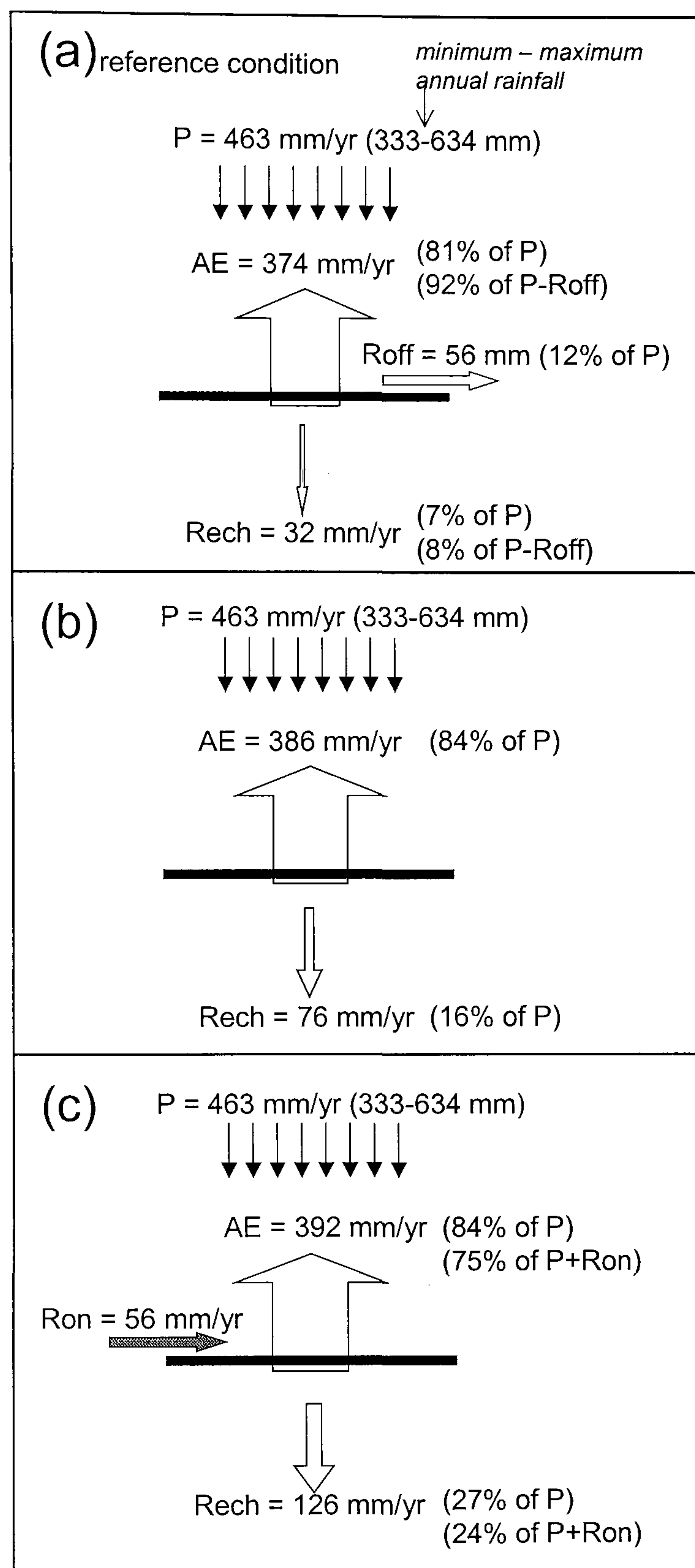


Figure 6. 3. Water balance for three net surface runoff scenarios. (a) net runoff, (b) zero net runoff, (c) net runoff. Mean values for 1993–97. Values are in mm/year.

When net runoff is zero (figure 6.3.b), there is an additional 56 mm/year of water and AE increases (12 mm). The remaining 44 mm/year becomes recharge resulting in a mean recharge of 76 mm/year (120% greater than in the situation with runoff).

When shifting to net runoff (figure 6.3.c), the additional 56 mm contributes mainly to recharge. Mean recharge is 126 mm/year, approximately four times the mean recharge with runoff. Recharge increases 94 mm/year (126–32 mm/year), or 84% of the additional 112 mm/year. Note that AE hardly changes as a result of the additional runoff.

In summary, when shifting from a net runoff to a net runoff, the largest proportion of the additional water becomes recharge.

The overall results are in line with the study of Rockström et al. (1998) who also concluded that the greatest fraction of the water additionally supplied by runoff is deep drainage and the crop did not benefit from the additional inflow of water.

The analysis presented here is a preliminary investigation of the effect of surface runoff variability on recharge. The results suggest that the corresponding variability of recharge is large and it should not be neglected. Depending on local conditions, the net runoff can vary significantly, thus affecting recharge. Therefore, any investigation of recharge needs to first define the local conceptual model of surface runoff in order to realistically represent field conditions.

6.4. ANALYSIS OF SPATIAL VARIABILITY

The analyses presented so far in this thesis have considered the temporal variability of recharge through the use of an extensive daily weather data set. However, as illustrated in Section 5.4, recharge varies significantly due to the inherent spatial variability of the complex soil-vegetation system. Spatial variations of rainfall and ET_o , as well as in the physical properties of soil and vegetation will produce changes in recharge.

The purpose of this section is to explore how the variability of the model parameters, based on knowledge of actual field variation, affects the modelled recharge. Three model variables were selected as representative of the field variability. They are the crop coefficient for the mid-season $K_{c_{mid}}$, the maximum root depth and the soil water holding properties (field capacity and permanent wilting point). In addition, the effects on modelled recharge of surface runoff variability and initial soil water content are investigated.

The first two variables ($K_{c_{mid}}$ and maximum root depth) represent some of the variability of the crop properties. The crop coefficient can vary due to a series of factors such as effect of climate and/or agricultural practice (Allen et al., 1998). Maximum root depth can also vary in the field as discussed in Section 2.2.6.

The variability of the soil water properties is represented by variation in the total available water ($TAW = \text{field capacity } FC \text{ minus permanent wilting point } PWP$). The values of field capacity and permanent wilting point for a particular soil type can vary, as shown by Allen et al. (1998), table 19.

The variability of net runoff depends on a series of field factors as discussed in Section 2.2.3. For this analysis, runoff variability is simulated by the increasing and decreasing ($\pm 20\%$) the runoff coefficients presented in table 3.1.

Finally, the variability of the initial soil moisture conditions is explored for a condition when the water content at the beginning of each rainy season is assumed to be at permanent wilting point. This condition is based on field observations made by Hess (1999) in Nigeria. The author suggests that water in the soil profile may be lost before the next rainy season by continued slow drainage and/or soil evaporation.

The variability of modelled recharge is assessed by running the model for the maximum and minimum values of the interval of variation of each single model parameter. The results are presented showing the variation from a reference condition in order to compare the overall variability.

The reference surface is a sandy soil planted with a millet crop as described in Section 4.2.

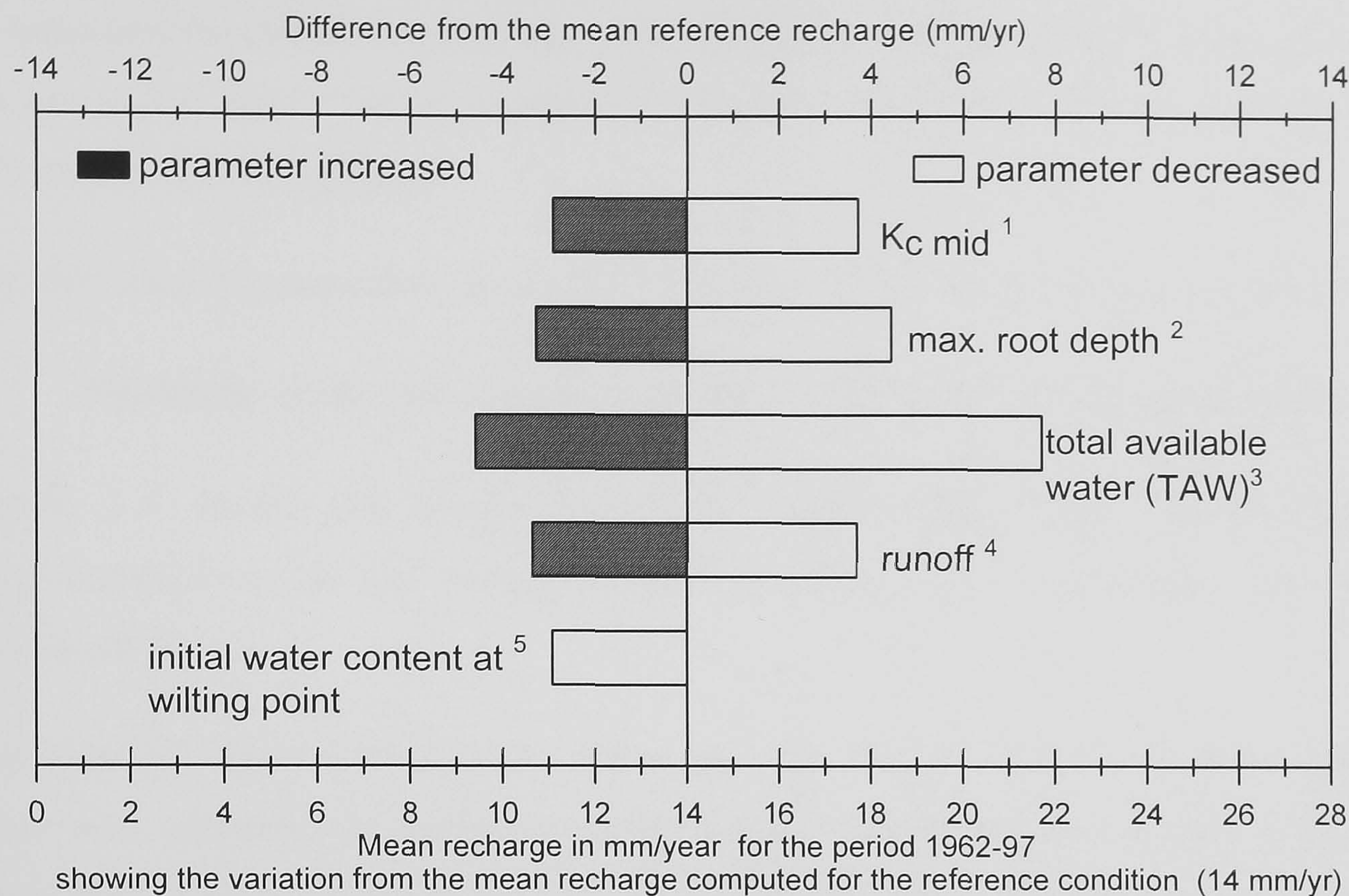
Variability is investigated using the entire 36 year rainfall dataset from Nguru, northeast Nigeria (see Section 4.1). In a further analysis, a sequence of wet years from the same data series is selected.

It is important to point out that this analysis does not consider any interaction between the parameters and it has to be understood as a preliminary sensitivity analysis.

6.4.1. Analysis using the long-term rainfall series (1962-97)

The annual rainfall at Nguru for the period 1962-1997 varies from 235 to 650 mm and the mean annual rainfall is 431 mm/year. The estimated mean annual recharge for the reference surface using the long-term rainfall data series is 14 mm/year (Section 5.4.2).

The variation of modelled recharge from the mean reference value can be observed in figure 6.4. The box below the figure presents the range of variation and the sources of information on which the interval is based. The range of variation of the parameters is based on overall values from literature, field observation and insights from reality.



1. K_c mid (1.10 ± 0.10) based on Table 12 and Figure 32, FAO publication n.56 (Allen et al., 1998).
2. maximum root depth (1.2 ± 0.25 m) within range observed by Payne et al. (1991).
3. available water ($0.09 \pm 0.02 \text{ m}^3 \cdot \text{m}^{-3}$) within range of Table 19, FAO publication n.56 (Allen et al., 1998) for a sandy soil.
4. runoff coefficients from table $3.1 \pm 20\%$
5. initial soil water content at permanent wilting point each 1st January.

Figure 6. 4. Variations of the mean recharge for the period 1962-97 from a mean reference value (14 mm/year) as function of the single variations of selected parameters.

The overall figures show that recharge varies from 9.5 to 21.8 mm/year, a difference of -4.6 and $+7.7$ mm/year from the reference value of 14.1 mm/year (-33 to 55%). Modelled recharge is most sensitive to variation in the total available water.

The overall interval of the mean recharge resulting from the variation of the crop parameters (crop coefficient and maximum root depth) was approximately ± 4 mm/year ($\pm 30\%$ from the reference). The figures for the variability of runoff are also very similar.

The assumption of the soil water content at wilting point before the start of the rainy season appears not to affect the mean recharge significantly. The reduction of the mean recharge was only 3 mm.

In summary, the overall recharge figures for the period of 36 years with a millet crop on a sandy soil in Nguru can be presented by the mean of 14 mm/year varying within the interval of 10 to 22 mm/year.

The following diagrams show the overall results for the soil water balance components

Variability of the soil water balance due to variations in the $K_{c\ mid}$ parameter

Figure 6.5 shows the mean values of annual runoff R_{off} , annual actual evapotranspiration AE and annual potential recharge $Rech$ when varying the crop coefficient $K_{c\ mid}$.

The mean AE varies approximately $\pm 1\%$ from the reference value (-5 to +4 mm/year). This result suggests that variations of the crop coefficient for the mid-season do not affect the estimation of AE significantly. At first sight this is curious since changes in $K_{c\ mid}$ directly affect the potential atmospheric demand, and so a more significant variation in AE would be expected.

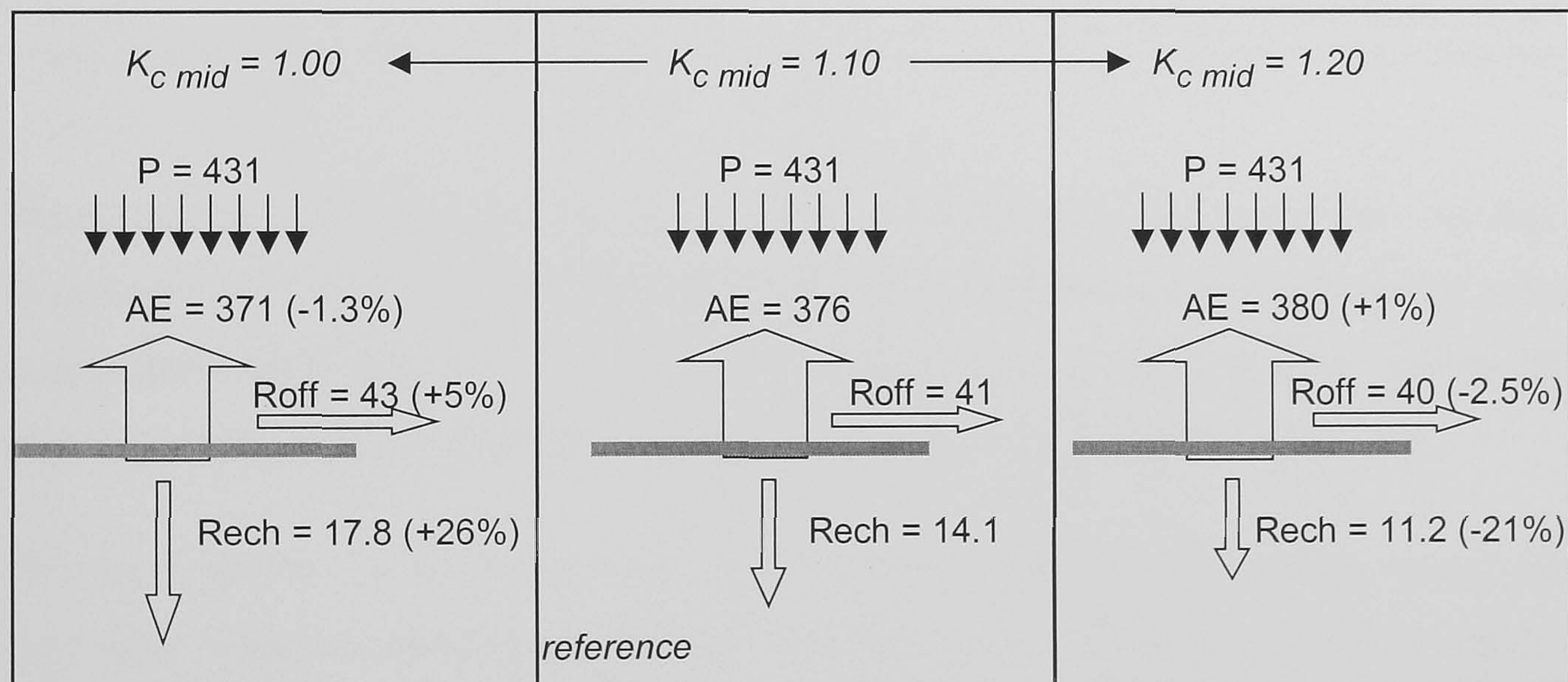


Figure 6. 5. Water balance partitioning resulting from the variation of the crop coefficient $K_{c\ mid}$. Mean values for the period 1962-97 are shown. Percentages indicate the relative variation from the reference surface.

The explanation for this outcome may be in the averaging of dry and wet years. During dry years, the crop is likely to be under water stress and AE is a fraction of the potential rate. In wet years PE is higher, and so is AE . Over the long term these variations in AE balance out.

The runoff component does not vary significantly also. The changes in mean runoff are approximately 2 mm/year.

Variability of the soil water balance due to variations in the maximum root depth

The effect of the variation of the maximum root depth on the soil water balance is illustrated by figure 6.6.

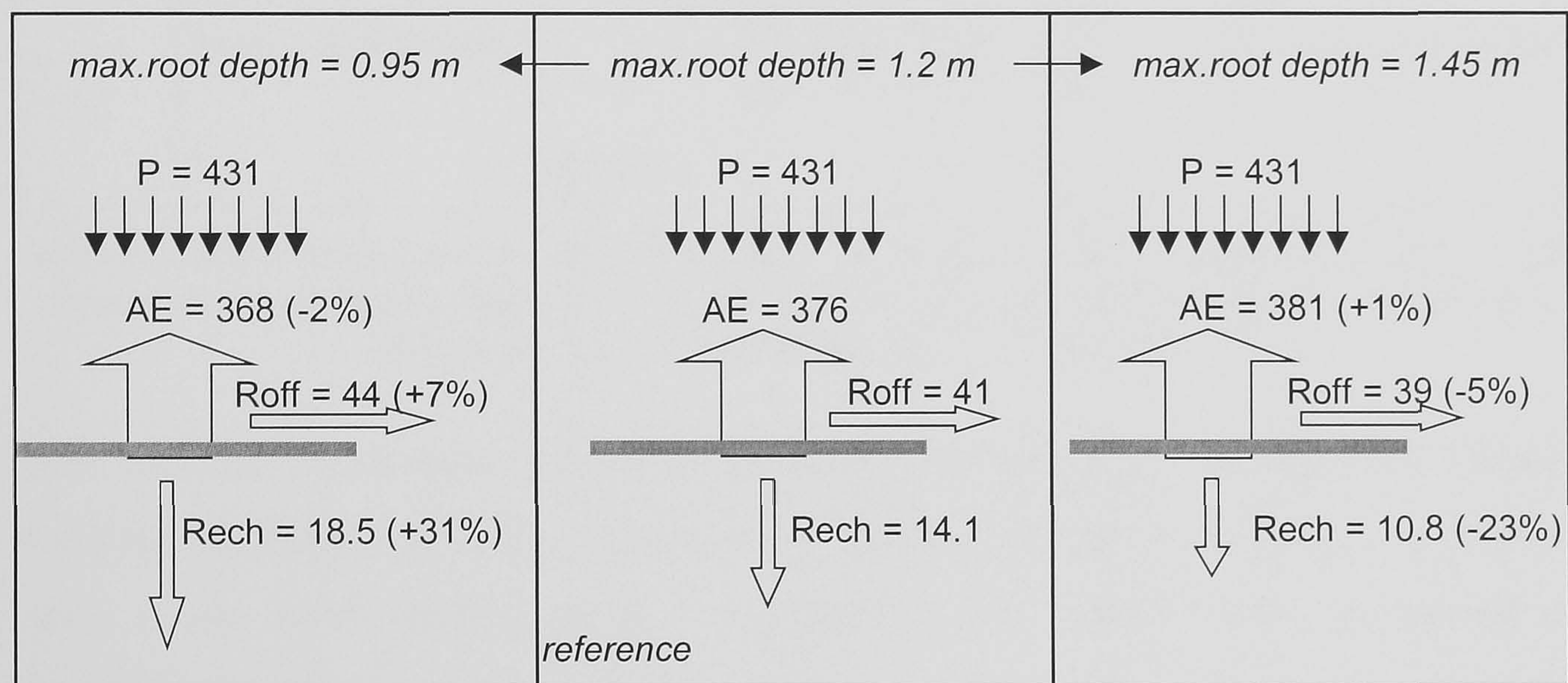


Figure 6. 6. Water balance partitioning resulting from the variation of the maximum root depth. Mean values for the period 1962-97 are shown. Percentages indicate the relative variation from the reference surface.

Mean recharge decreases 7.7 mm (18.5 - 10.8 mm/year) when the maximum root depth is altered from 0.95 to 1.45 m. This decrease may be explained by the increase of actual evapotranspiration (13 mm). Mean actual evapotranspiration varies from 368 to 381 mm/year as the roots make their way deeper to extract more water from the soil.

Moreover, as the soil becomes drier by the action of the root deepening, runoff also decreases. Mean runoff changes from 44 to 39 mm/year, a difference of 5 mm which is made available for evapotranspiration.

Variability of the soil water balance due to variations in the available water for evapotranspiration

Figure 6.7 shows the resulting soil water balance when the total available water *TAW* (defined as the difference between field capacity and permanent wilting point) is altered within a symmetric range.

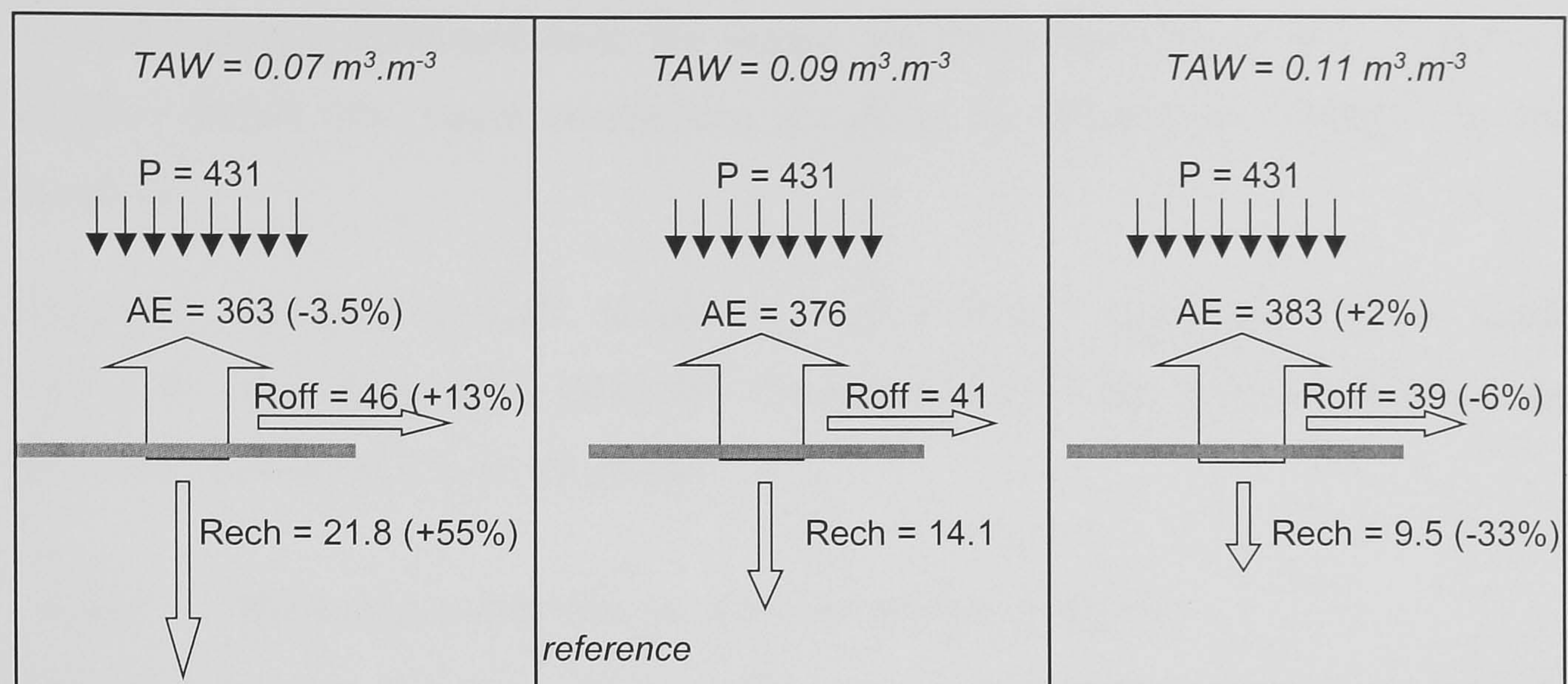


Figure 6. 7. Water balance partitioning resulting from the variation of total available water (TAW = field capacity minus permanent wilting point). Mean values for the period 1962-97 are shown. Percentages indicate the relative variation from the reference surface.

Mean recharge varies from 9.5 to 21.8 mm/year. This introduces the biggest difference in terms of recharge variability. A small TAW means a low soil water retention capacity, hence a low field capacity point. Consequently, more water drains downward as potential recharge.

In addition, if TAW is lower, AE decreases. Mean AE varies from 383 to 363 mm/year (a difference of 20 mm) when changing the soil water holding properties from a situation with a higher field capacity (and wilting point) to a situation where the field capacity is low.

The runoff component decreases 7 mm when the total available water increases from 0.07 to 0.11 $m^3.m^{-3}$. This result appears to be contradictory, since with a lower field capacity point it would be expected that water would move vertically more rapidly, resulting in a drier soil near surface and a higher infiltration capacity.

This result can be explained by the simplified assumptions adopted in the present analysis. The matrix of runoff coefficients (section 3.3) is kept the same, with no changes in the interval of soil moisture deficit for the three conditions presented in figure 6.7. However in reality, variations in field capacity directly affect the concept of soil moisture deficit. That is, in the condition when field capacity is low (lower total available water) the soil moisture deficit is also low during the rainy season (note that less water is needed to bring the soil water content to field capacity). Consequently, the

selected runoff coefficients from the matrix are from the interval with a low soil moisture deficit (the larger coefficients according to the approach adopted in this research).

Therefore, variation of TAW should be linked to the adjustment of the runoff coefficients also. This aspect illustrates the interrelation of the variables which is not taken into account in this investigation.

6.4.2. Recharge variability for a wetter period (1974-79)

The period from 1962 to 1997 includes several years with low annual rainfall (see Appendix C). This can affect the overall results by averaging the effect of the variability of the model parameters. For instance, in dry years the effect of $K_{c\ mid}$ on AE may not be significant since the crop is under water stress and evapotranspiration is not at potential rates.

In order to investigate the variability during a wetter period, a period of six years is selected (1974 - 1979). In this period, annual rainfall varies from 431 to 602 mm with a mean rainfall of 531 mm/year.

Figure 6.8 presents the variability of recharge due to the variation of the selected parameters. Note that the intervals of variation of the parameters are the same as presented in the box under figure 6.4. The mean potential recharge estimated for the reference surface is 35 mm/year.

Comparing figure 6.8 to figure 6.4, it is possible to observe that the pattern of variability is similar, but the range of variation in terms of mean recharge is greater. For example, recharge varies from 26.2 to 48.5 mm/year (a difference of 22.3 mm) due to the variability in total available water. The range is almost twice as large as that estimated using the data series including dry years (9.5 to 21.8 mm/year, a range of 12.3 mm).

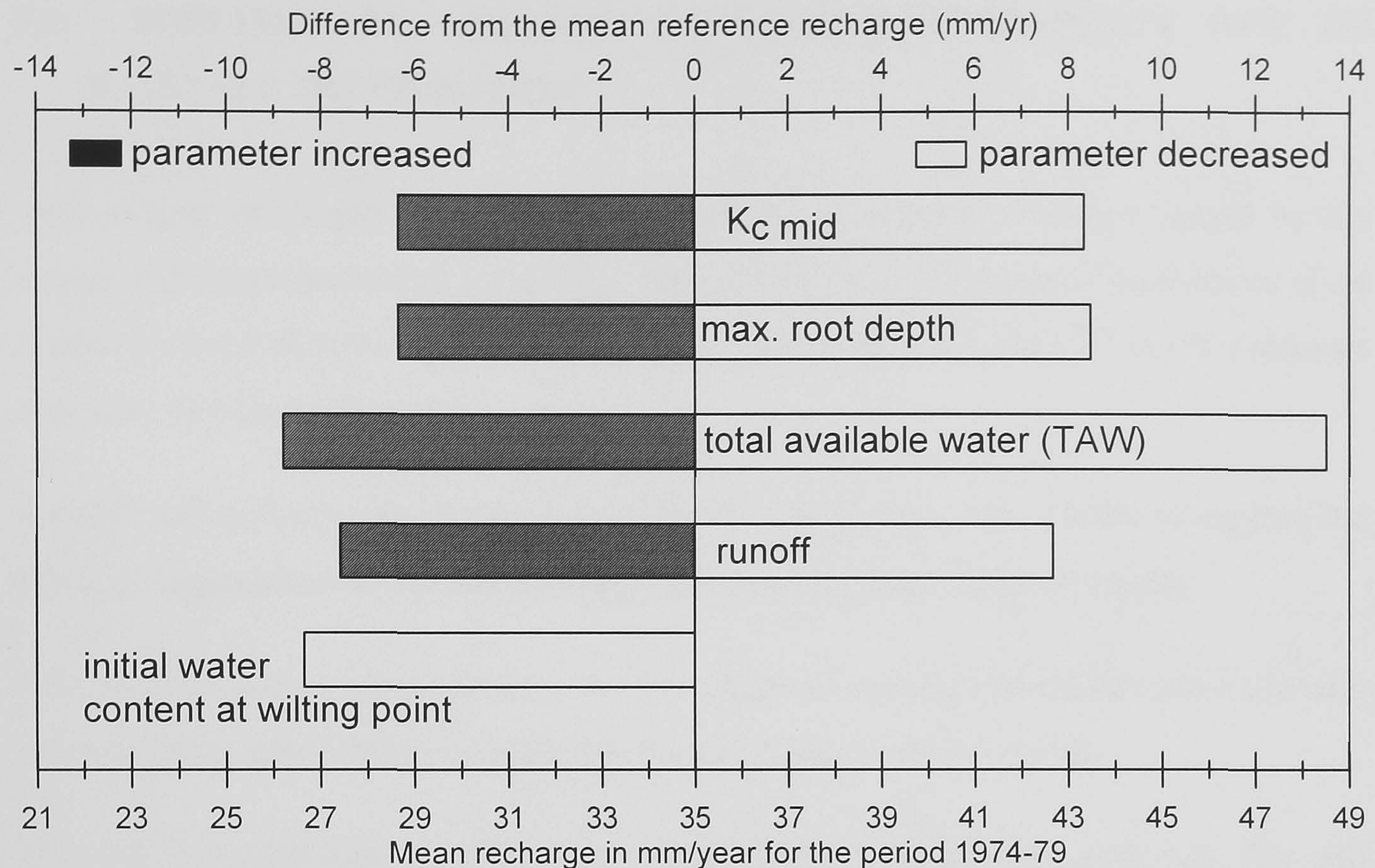


Figure 6. 8. Variation of the mean recharge from a mean reference value for the wet period of 1974-79. The bars show the variation from the mean recharge for the reference condition (35 mm/yr). The intervals of variation of the parameters are as in figure 6.4.

Although greater in absolute terms, the variability is lower in terms of proportional variation from the reference condition. When changing the available water parameter, the variation from the mean recharge of 35 mm/year is -25 and +40%, instead of -33 and +55% from the mean recharge of 14.1 mm/year. It suggests that the proportional variability is more accentuated during droughts.

Regarding the initial water content variability, the mean recharge difference between the two treatments for a wetter period is greater (8 mm instead of 3 mm estimated using the long term period). It suggests that the soil water balance is more sensitive to the moisture initial conditions during wet years.

The reason for this is that during a sequence of wetter years, the soil water content at beginning of the rainy season is likely to be greater than the permanent wilting point. Therefore, studies which assume that the soil profile is completely dry at the start of the rainy season may be underestimating recharge.

6.5. PRELIMINARY ANALYSIS OF SPATIAL VARIABILITY DUE TO RAINFALL DISTRIBUTION

This section investigates the pattern of variability of regional recharge caused by the spatial and temporal rainfall variability. This analysis is based on rainfall observed at 18 rainfall stations distributed over an area of approximately 45,000 km² in the northeast arid zone of Nigeria (figure 4.1).

Rainfall and recharge are presented for different temporal scales in order to explore the effect of aggregation of the results when presenting regional recharge results.

This analysis is only a preliminary survey of regional recharge variability since the only variable taken into account as leading to recharge variability is rainfall.

The soil and vegetation conditions are assumed to be constant for each site. The soil water balance is carried out for a reference surface based on the conditions presented in Section 4.2.

The rainfall data available from the 18 rainfall stations are daily-recorded rainfall for a period of four years (1992–1995). The reference evapotranspiration is not available; therefore, *ET_o* is derived from the long-term monthly *ET_o* as described in the Section 4.2.

Rainfall has a unimodal annual distribution with the rainy season beginning in April-May and ending in September-October. The wettest month is August with a mean rainfall of 200 mm/month, followed by July with 129 mm/month. The wettest and driest years of the period are respectively 1994 (regional mean of 677 mm) and 1993 (regional mean of 353 mm). For more detailed information about the rainfall characteristics of each site see the Appendix F.

Figure 6.9 shows the mean annual rainfall for the period 1992-95 at each site and the isohyets drawn at 50 mm unit intervals constructed by interpolation between the stations. The regional mean annual rainfall for 1992-95 is 445 mm/year, varying from 561 mm/year at Dagona in the west part of the area to 220 mm/year at Kanamma in the northeast. It is in line with the regional trend as described by Kowal and Kassam (1978), that is, the mean annual rainfall declines from southwest to northeast.

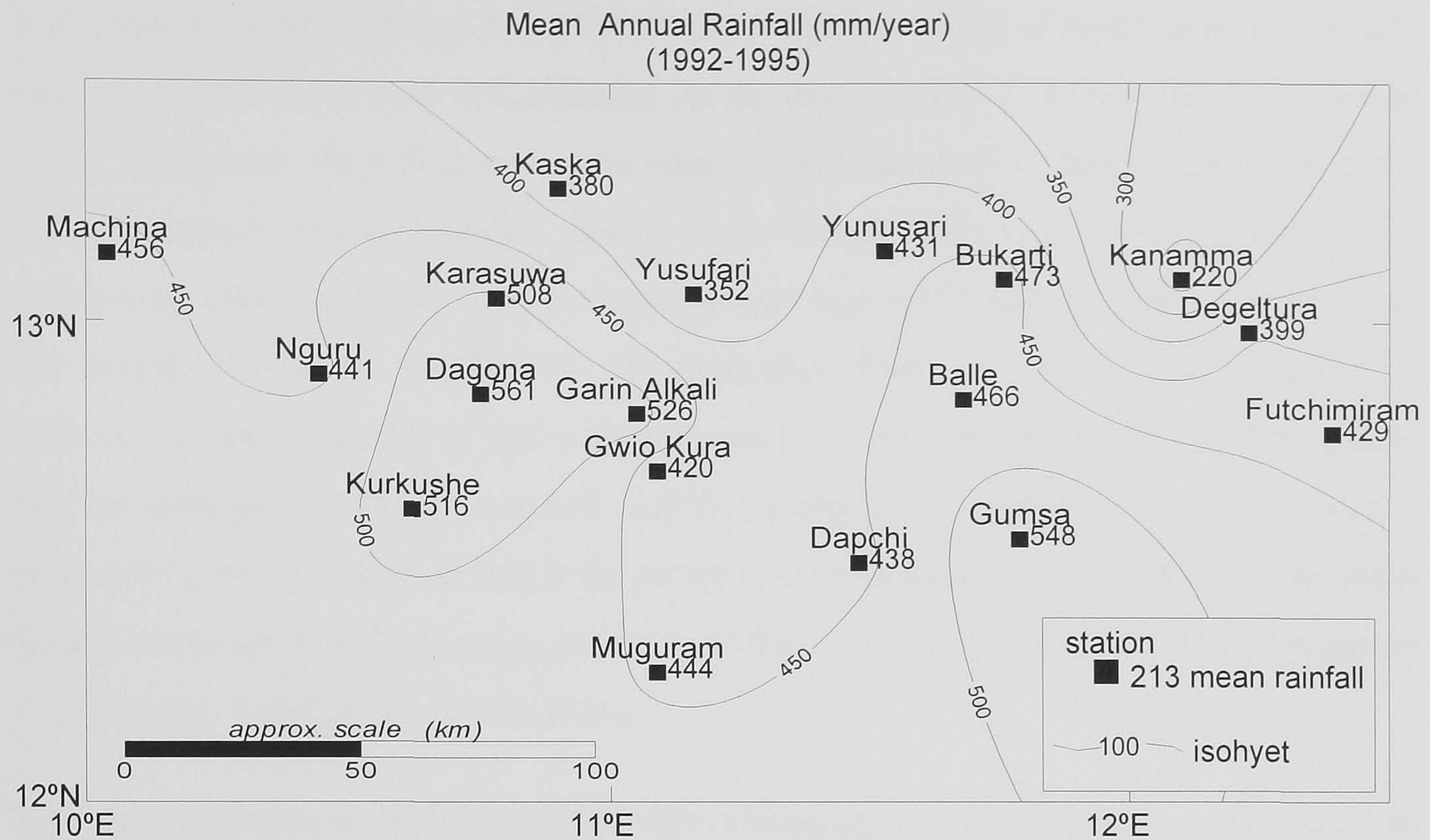


Figure 6. 9. Spatial distribution of the mean annual rainfall for 1992-1995. Values in mm/year.

Figure 6.10 shows the modelled mean annual potential recharge resulting from the daily soil water balance of the reference surface at each site. The figure shows the point values estimated at each station. These point values are represented in the format of proportional circles in order to better visualise the spatial variability.

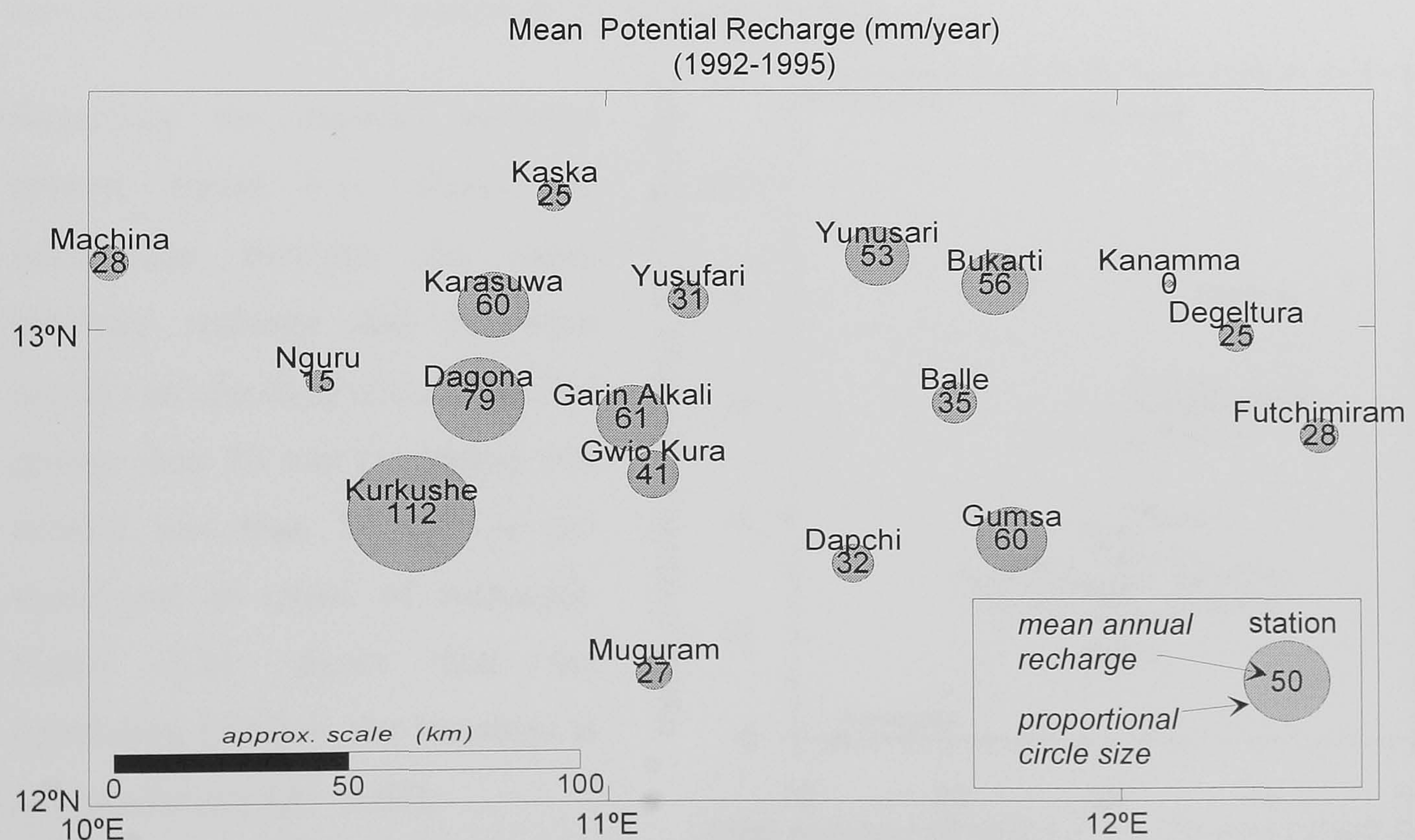


Figure 6. 10. Mean annual potential recharge (1992-95) estimated for a reference surface using the daily rainfall data from 18 rainfall stations. The mean recharge for each site is shown. The circles are proportional to the values of mean recharge. All values in mm/year.

The mean regional recharge for 1992-95 is 43 mm/year (9% of mean annual rainfall). Recharge varies from zero at Kanamma (in the northeast) to 112 mm/year in Kurkushe in the southwest. At a first view, the spatial distribution of recharge agrees with the overall tendency of mean rainfall, that is, higher values in the southwest area and lowers in the northeast. However, a closer look reveals that mean rainfall and mean recharge are poorly correlated (coefficient of correlation linear $R^2=0.56$). For instance, at Kurkushe a mean rainfall of 516 mm/year results in a mean recharge of 112 mm/year. For the same period, at Garin Alkali, approximately 50 km northeast, mean recharge is 61 mm/year (45% lower) although the mean rainfall is higher (526 mm/year). The same aspect can be observed comparing Balle with Yunusari, Kaska with Yusufari, Muguram with Dapchi, and Nguru with Machina.

This poor correlation is explained by the averaging of rainfall, such as discussed in Section 2.2.1. Mean annual rainfall is an unreliable guide to the year-to-year variability of recharge. Therefore other rainfall descriptors associated with regional recharge are investigated.

Hess et al. (1995) concluded that the number of rain-days has an important role in the regional hydrology. Their decline is associated with the overall decline in annual rainfall over a long-term period in the northeast Nigeria.

Regarding the regional recharge pattern, figure 6.11 shows the relationship between the mean potential recharge and the total number of rain-days when rainfall is greater than 20 mm (assuming that rainfall less than 20 mm is not significant in terms of recharge). Figure 6.11 shows that the correlation between the variables is not satisfactory ($R^2=0.47$).

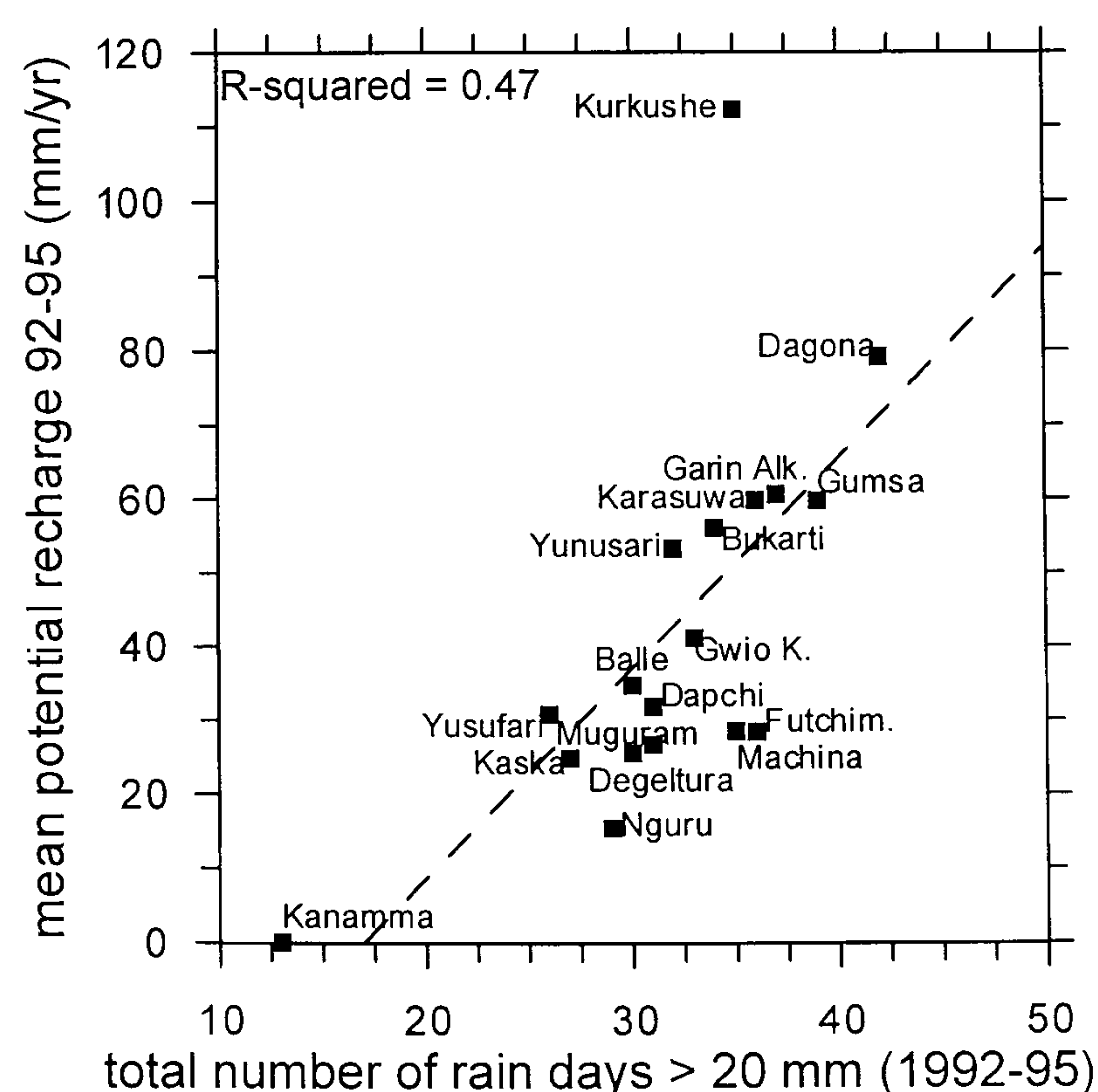


Figure 6. 11. Correlation between the total rainy days with rainfall bigger than 20 mm and the mean annual recharge for 1992-95 for the study area.

For instance, the variable does not explain why the mean recharge in Kurkushe is higher than in Dagona (located 30 km northwards) despite the smaller number of days with a significant rainfall at Kurkushe.

The results above suggest that the number of days above a significant rainfall threshold is not a good indicator of the occurrence of recharge. It is necessary to know about the magnitude of the daily rainfall values. The following analyses consider the total rainfall for a selected shorter period.

Taking the Kurkushe site as an example, approximately 80% of the estimated recharge for the four-years period is in 1994. The resulting annual recharge for 1994 was 354 mm, against 50, 45 and zero mm for 1992, 1993 and 1995 respectively. The same characteristic is observed in the other stations with the exception of Kanamma where estimated recharge is nil for the whole period due to the low rainfall rates, and Garin Alkali which is the only station where annual rainfall in 1995 is greater than 1994.

In 1994, annual rainfall varies from 222 mm in Kanamma to 1042 mm in Dagona. The regional mean rainfall for 1994 is 667 mm. The regional recharge estimated for 1994 is 125 mm (18% of annual rainfall) varying from zero to 354 mm. The correlation between the annual recharge and annual rainfall for 1994 is shown in figure 6.12.

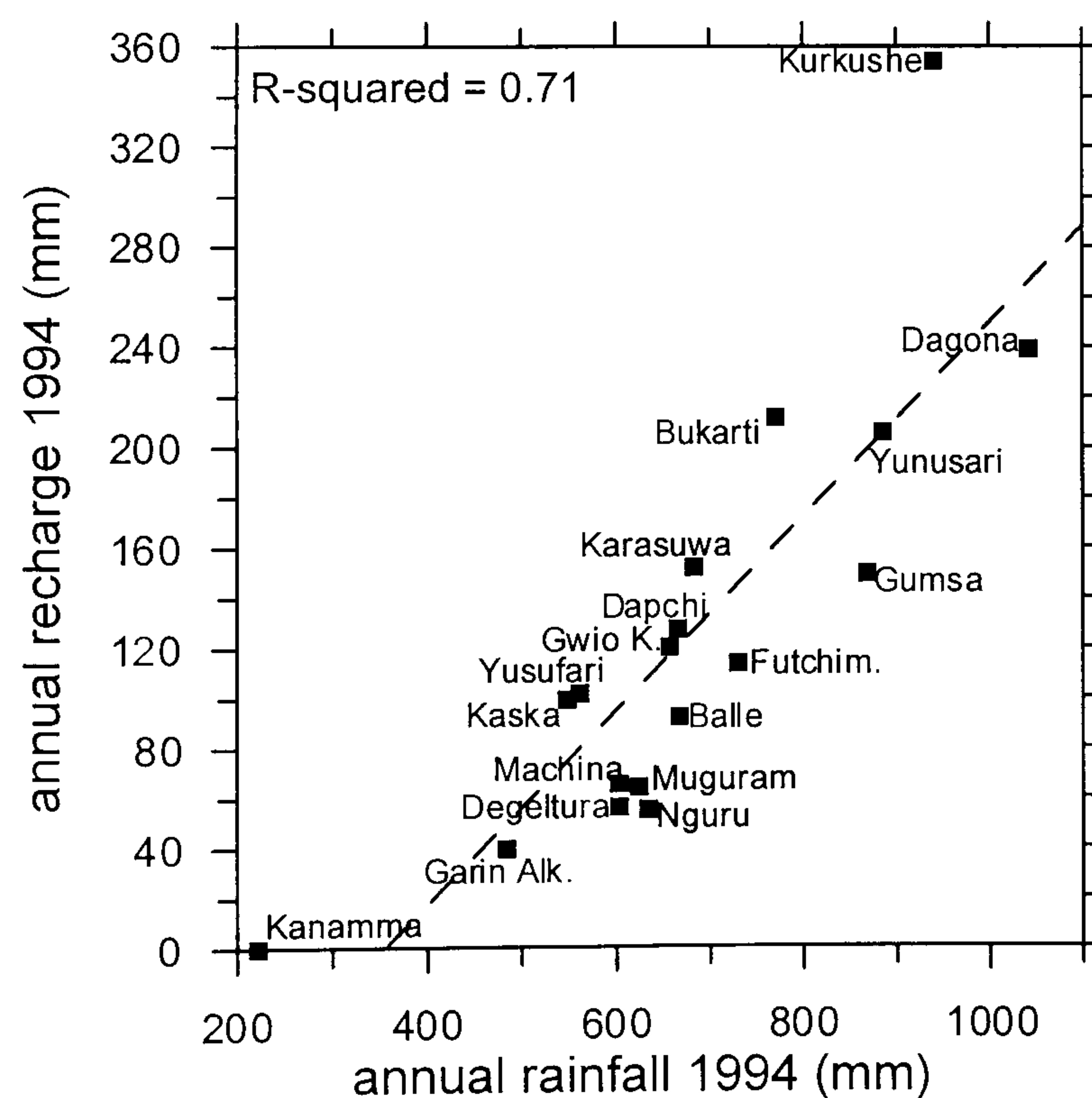
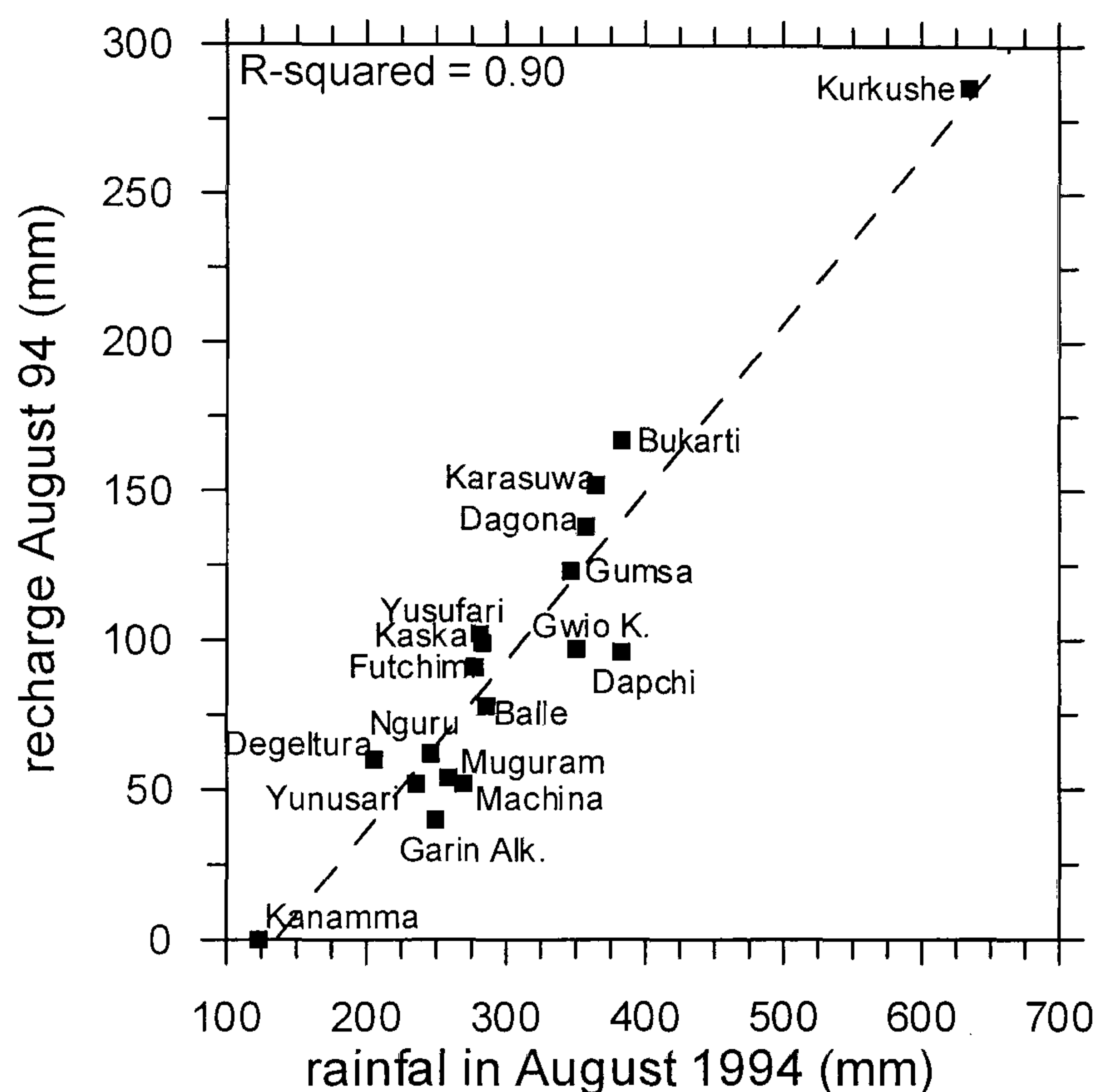


Figure 6. 12. Correlation between the annual rainfall and annual recharge for 1994 for the study area.

The correlation is better than the previous matches ($R^2 = 0.71$). It suggests that for wet years such as 1994, recharge and annual rainfall are well correlated. However, the correlation does not explain the variability between Gumsa and Yunusari for instance. Rainfall in 1994 was similar at both sites (869 and 886 mm respectively) but recharge at Yunusari is 38% greater.

Recharge in semi-arid areas is not a simple function of the annual rainfall. Recharge is caused by sporadic and highly intense rainstorms occurring during a short rainy season. For example, 81% of the estimated recharge in 1994 at Kurkushe occurs during the month of August (the wettest month). This pattern is observed in all the stations, with the exception of Yunusari where July is the wettest month.

The correlation between rainfall for the month of August 1994 and the corresponding estimated monthly recharge is presented in figure 6.13.



There is a good correlation between monthly rainfall and monthly recharge ($R^2=0.9$). High rainfall rates during the peak of the rainy season result in a consequent high potential recharge.

However, the variability of the recharge values such as between Dapchi and Bukarti can only be explained by the daily distribution of rainfall.

Figure 6. 13.. Correlation between rainfall during the month of August 1994 and estimated annual recharge for the study area.

For example, at Dapchi, total rainfall in August is 382 mm and the estimated recharge is 96 mm (25% of monthly rainfall). Meanwhile, at Bukarti an equivalent rainfall (383 mm) results in recharge equal to 167 mm (43% of monthly rainfall). This variability is explained by the occurrence of high daily events during the month and the moisture station differences in conditions at start of the peak of the rains.

At both sites the number of rain-days with rainfall bigger than 20 mm is the same (7 days), however, at Bukarti 75% of the monthly rainfall is concentrated in the first two weeks with two storms of more than 40 mm/day within a week (one storm of 96 mm in

a day). In addition, rainfalls in June and July were good enough to increase the soil water content to a point closer to field capacity. On the other hand, in Dapchi the high rainfalls during the month of August are more spread out and rainfall during the months preceding August was not significant.

The distribution of monthly rainfall and estimated recharge is shown in figures 6.14 and 6.15 respectively.

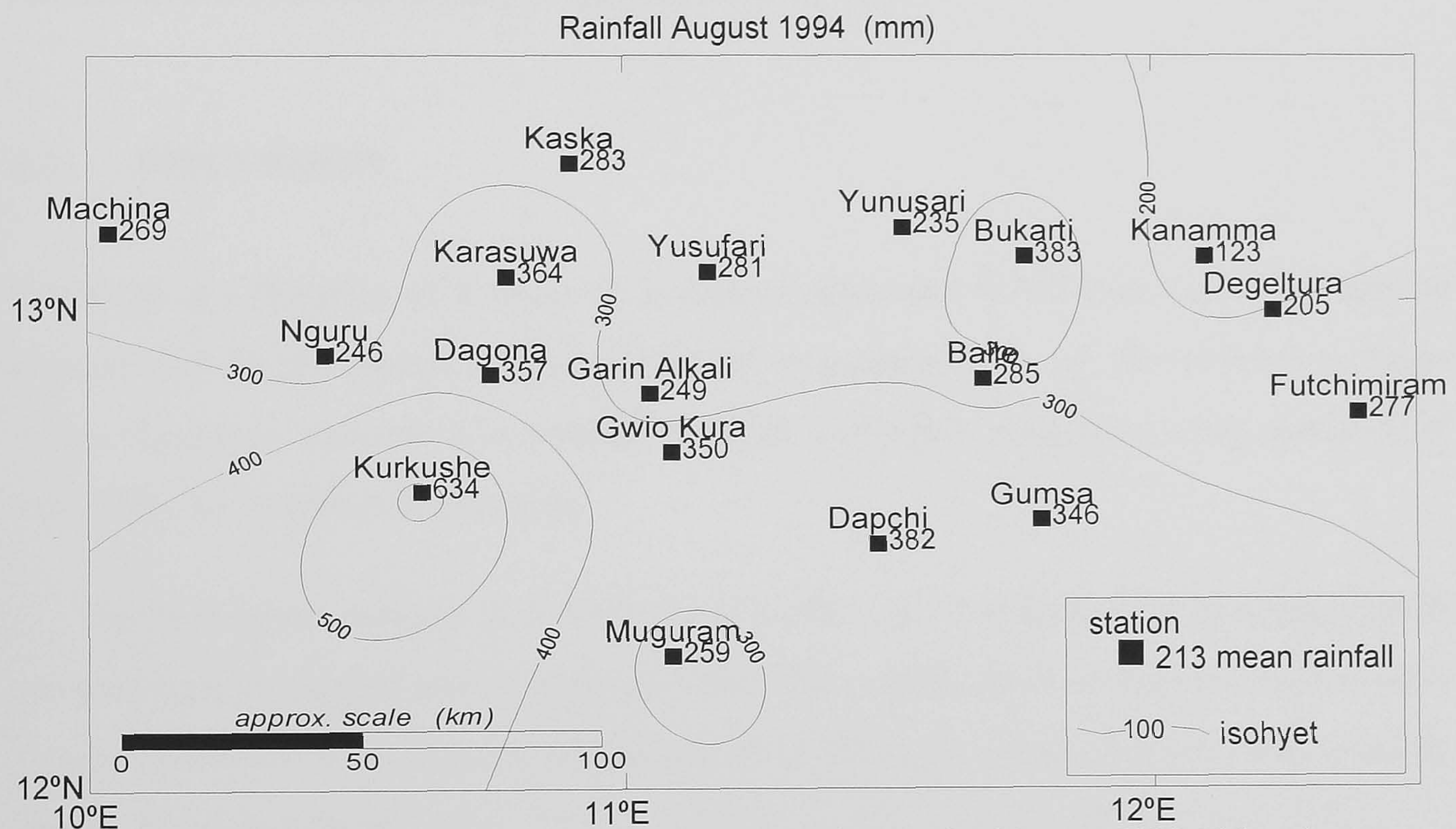


Figure 6. 14. Spatial distribution of the total rainfall for the month of August 1994. Values in mm.

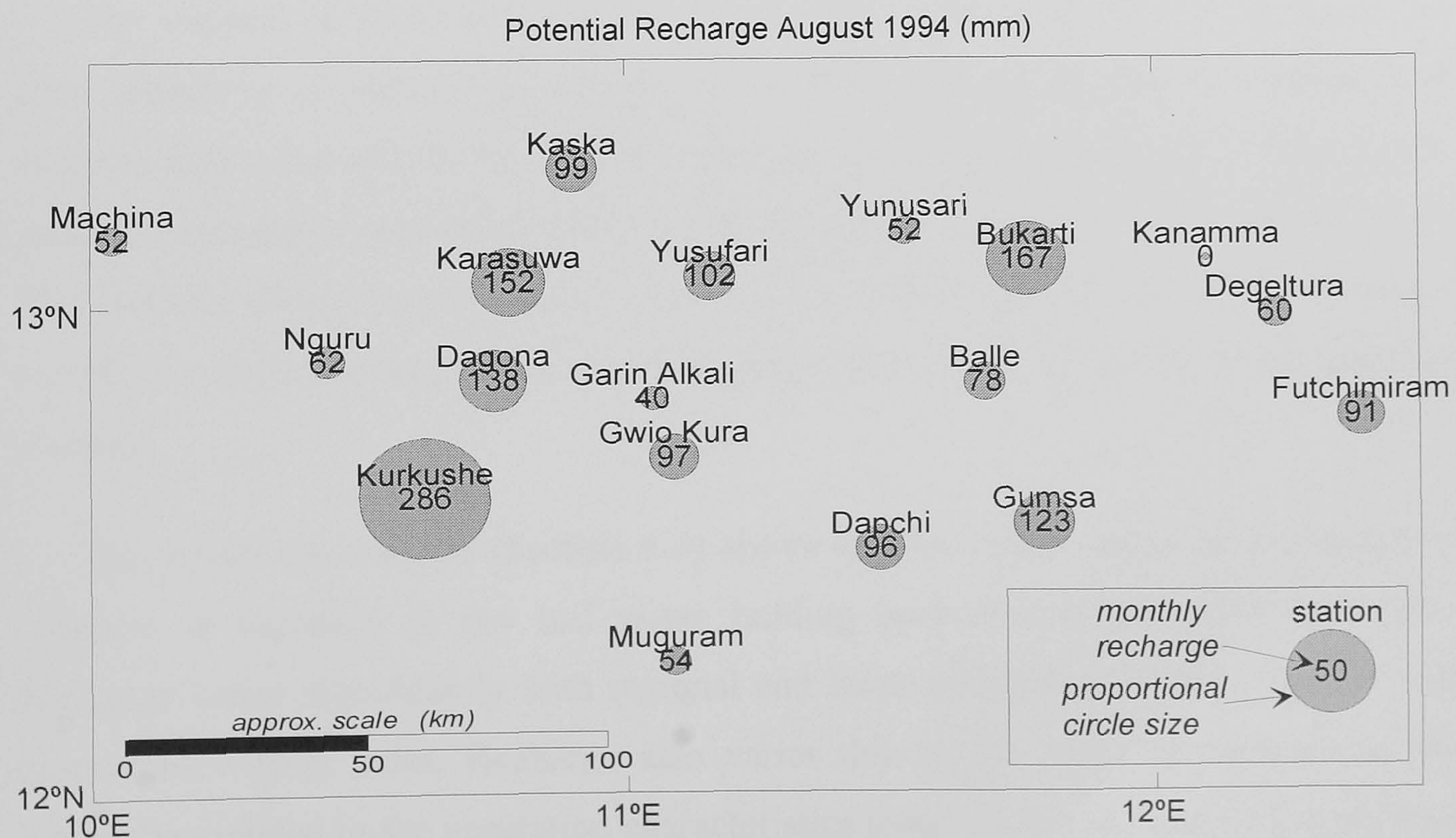


Figure 6. 15. Monthly potential recharge for August 1994 estimated for a reference surface using the rainfall daily data from 18 rain gauge stations. All values in mm.

Rainfall varies from 634 mm/month in the southwest to 123 mm/month in the northeast. The resulting monthly recharge varies from zero (Kanamamma) to 286 mm/month (Kurkushe). The mean recharge for the area is 97 mm/month. Because of the good correlation between the two components, the spatial distribution of monthly recharge follows a similar pattern to rainfall. For the sites with monthly rainfall less than 300 mm, recharge is less than 100 mm. On the other hand, for the stations with monthly rainfall above 300 mm recharge is greater than 100 mm.

6.6. DISCUSSION

Recharge is a function of a range of factors as discussed in Chapter two. They can be summarised by the factors related to climate, vegetation and soil characteristics. From this preliminary analysis it is possible to draw out some insights relating actual field variability to its effect on recharge.

- The soil water balance is not sensitive to the use of reference evapotranspiration derived from averaged values (Section 6.2). The overall difference between the daily values calculated from climatic observation and the daily values derived from monthly means is not significant.
- The analysis of three different surface runoff conditions (Section 6.3) shows the great sensitivity of recharge to changes in the net runoff component. The addition of water by runoff has resulted in a fourfold increase in recharge during a wet year. In dry years the increase is smaller because part of the additional water is utilised by the crop. This analysis shows the importance of defining a correct conceptual model of surface runoff. Net runoff can be an important component at localised areas (e.g. where ponding occurs).
- The sensitivity analysis (Section 6.4) shows that the main source of variability in recharge is variation in the soil water holding properties (total available water). Recharge varies significantly with regional and local differences in field capacity and permanent wilting point. Recharge also varies due to the regional variation in the parameters related to the vegetation characteristics (crop coefficient and maximum root

depth). The variations in the crop coefficient may represent variations in climate as well as in agricultural practices, such as crop row spacing or intercropping.

- The spatial variability of recharge, as investigated in Section 6.5 gives an opportunity to compare the recharge estimated by the soil water balance model with the results obtained by Edmunds et al. (2002) using the chloride method. Regional recharge estimated by the soil water balance model at 18 sites (figure 4.1) and considering a period of four years (1992-95) is 43 mm/year, varying from zero to 112 mm/year. This variation is caused exclusively by the rainfall variability since soil and vegetation properties were kept the same.

Edmunds et al. (2002) estimated the regional recharge using the chloride method based on data from 360 shallow wells located in the same region (northeast Nigeria) as also 43 mm/year. The authors did not provide details of the regional variability of recharge, but insights can be inferred from the sampled variation of Cl. The distribution of Cl ranges from 0.8 to 96 mg/l Cl with a median value of 6.35 mg/l Cl. This range reflects the heterogeneity of different soil types, vegetation and topography.

Therefore, despite the different conditions and assumptions utilised by the chloride method (see Section 2.5.4) the overall values for regional recharge are very similar.

CHAPTER SEVEN

CONCLUSIONS AND RECOMMENDATIONS FOR FURTHER INVESTIGATIONS

7.1. THE SOIL WATER BALANCE TECHNIQUE FOR RECHARGE ESTIMATION IN SEMI-ARID REGIONS

The first objective of this thesis was “*to develop a model which is physically credible and adequately reflects the principal physical processes which affect recharge*”. This has been achieved through the development of a soil water balance conceptual model (Chapter Three). This conceptual model introduces new aspects for the representation of semi-arid field conditions. They are: (i) the estimation of surface runoff based on a pragmatic and realistic physically-based procedure; (ii) the bare soil period, before and after the crop season, which affects the initial soil water content conditions; (iii) the estimation of actual evapotranspiration based on the physical knowledge of water uptake by roots and on the water movement into soil profile; (iv) addition of the near surface storage concept to account for evapotranspiration following rainfall on dry soil.

The inclusion of these procedures has improved the physical robustness of the soil water balance technique, linking practicability with data availability, and so fulfilling the second objective of this research, “*to develop a model which makes use of a small number of key parameters to describe the principal physical processes*”.

The parameters utilised by the soil water balance model (table 7.1) are in general readily available from standard agro-hydrological databases and/or from the literature. The exceptions are the factor of near surface storage and the surface runoff coefficients, which need further research as discussed later in this Chapter.

Table 7. 1. Input parameters of the soil water balance model.

<i>climatic input data</i>	<ul style="list-style-type: none"> • daily rainfall (mm.day^{-1}) • daily reference evapotranspiration (disaggregated from monthly means) (mm.day^{-1})
<i>crop characteristics</i>	<ul style="list-style-type: none"> • planting date • length of crop development stages (days) • crop coefficients • maximum root depth (m) • factor of depletion
<i>soil characteristics</i>	<ul style="list-style-type: none"> • field capacity point ($\text{m}^3.\text{m}^{-3}$) • permanent wilting point ($\text{m}^3.\text{m}^{-3}$) • factor of near surface storage
<i>surface runoff</i>	<ul style="list-style-type: none"> • surface runoff coefficient matrix

7.2. THE SOIL WATER BALANCE MODEL AND THE RESULTS USING ACTUAL FIELD DATA

From the conceptual diagrams and algorithms described in Chapter three a spreadsheet model (SAMBA model) was developed and applied to actual semi-arid field conditions.

The use of a spreadsheet computational environment has allowed for the easy understanding of the model structure and of the dynamic internal logic, thus facilitating the learning process. Moreover, its computational language is widely understood allowing for the easy utilisation and adaptation of new components if necessary.

By focusing on conditions where the crop growth and recharge occur at the same time, the approach has been tested under the most complex conditions, taking into account the real variability present in semi-arid regions. It is in line with the thesis's objective "*to apply the model to real conditions and present the results in diagrammatic form to help understand the interactions of the physical processes*".

Chapter four presents the results of the analyses in a graphical format that allows the reader to observe the complex interaction between the model inputs and outputs and easily visualise the temporal variability. The analysis approach makes use of selected periods in order to illustrate the effects of key model characteristics on recharge, beginning with the presentation of a representative year followed by the use of selected periods of days and years. The results show that the alternative procedures adequately represent semi-arid field conditions. That is:

- The near surface storage is a practical computational device for estimating actual evapotranspiration after rainfall during water stress periods. Figure 4.12 illustrates how the near surface storage adequately represents the soil evaporation process during the bare soil period and early crop stages, as well as providing means for determining actual evapotranspiration in the middle of the crop season when the crop is under water stress.

The effect of the near surface storage on potential recharge is directly associated with its effect on actual evapotranspiration. Overall potential recharge decreases when the near surface storage concept is applied.

- The surface runoff coefficient matrix appears consistent with runoff observed in field experiments (figures 4.8 and 4.10). Runoff becomes gradually more significant through the rainy season as the soil becomes wetter. The procedure is sensitive to rainfall distribution characteristics such as illustrated by the inter-annual variability in figure 4.9. The effect of surface runoff on potential recharge (Section 4.4.1) suggests that studies which neglect the surface runoff component are in danger of overestimating recharge rates.
- The importance of establishing a correct representation of the initial conditions is pointed out by the determination of a correct date of planting (Section 4.3.1). Planting date is based on the climatic criteria of onset of rains, taking account of false starts, thus realistically representing the actual agricultural practices utilised in the modelled area.

7.3. CREDIBILITY OF THE APPROACH

The thesis objective “*to demonstrate the credibility and plausibility of the model results when applied to typical semi-arid conditions*” is achieved by the assessment of the credibility of the soil water balance model described in Chapter five. Credibility is tested by a new approach entitled “plausibility analysis”. This approach allows for the assessment of qualitative implications of the model outputs in situations where conventional validation processes are not possible or not justified.

The concept of analysis of plausibility incorporates a range of insights from the physical system which corroborate the whole modelling process. It is a functional approach that

deals with the common problem of modelling with limited data. However, its application is more extensive and should be adopted in overall groundwater recharge modelling.

The analysis carried out in this thesis makes use of three different sources of information in order to give an overview of the plausibility of the model structure and results. They are (i) the standard analysis using soil water content measurements during a crop season, (ii) the utilisation of crop yield information from a sequence of years in a representative site and, (iii) overall results from several studies of recharge estimation in areas with equivalent climate.

The final results show that the outputs obtained from the soil water balance model are in accordance with the information gathered from the sources listed above, thus leading to the conclusion that the model is hydrologically credible.

7.4. INVESTIGATING RECHARGE USING THE SOIL WATER BALANCE MODEL

Chapter six presents a series of analyses the aim of which is “*to investigate the variability and sensitivity of recharge to a series of selected factors*”. The following conclusions emerge:

- Recharge rates vary widely depending on particular site conditions. Section 6.3 shows that recharge increases significantly when localised runoff is included into the soil water balance. Therefore it is crucial to define the correct conceptual model in relation to runoff and runoff.
- From the analysis in Section 6.4 it is shown that recharge is sensitive to a range of field parameters. The greatest variation results from the geographical variation in soil water holding properties (field capacity and permanent wilting point). These results are not unexpected since recharge is a function of several factors that are inherently variable due to the field heterogeneity (soil and vegetation characteristics).
- The spatial variability of recharge (Section 6.5) has provided a unique opportunity to compare the regional recharge estimated by two different methods (as discussed in

Section 6.6). The results obtained for regional recharge in the same area of northeast Nigeria are very similar, despite the differences of assumptions and techniques adopted in each study. The overall results show the high variability of recharge caused by rainfall distribution at each site. Moreover, the analysis of climatic factors such as rain-days shows that recharge is better explained by the magnitude of rainfall during the wettest period of the year.

- In a real field situation, many of the sources of variability are strongly spatially and temporally correlated. Therefore, it is necessary to carry out further analyses in order to establish the variability of recharge due to the complex interdependent climate-soil-vegetation system.

7.5. FURTHER INVESTIGATIONS

The procedures included in the SAMBA model for surface runoff and actual evapotranspiration associate practicality with physical credibility. However, further research is necessary in order to define some of its parameters. Two parameters in particular require further work:

1. The procedure for runoff determination allows for the direct parameterisation of the coefficients based on field observations. It is recommended that field measurements of runoff should be carried out in order to represent the spatial and temporal variability of the local hydraulic properties. Further investigation based on field experiment is also necessary in order to define typical coefficient values for different soil types.
2. In addition, the proportion of water kept near surface for evapotranspiration needs to be determined based on field observation of the water movement within the soil profile. Field measurements, which represent the soil water content of the topsoil following significant rainfall events, are important for determining the fraction of storage for different types of soil. Moreover, the fraction of storage seems to be not only a hydraulic soil property but also depends on crop development stages and/or farmer practice, as the variation of the factor of storage in figure 5.3 and 5.4 suggests. That is, during the early crop stages when soil profile is dry, the fraction of storage might be greater than during the peak of the rainy season, because of the different soil water

content (and hydraulic conductivity) conditions. Farmer practices can affect the fraction of storage through changing the infiltration capacity of the soil by tillage for example.

The model conceptualisation developed in this research provided insights into the processes affecting recharge in semi-arid environments, making use of the available information about the principal hydrologic processes. However, further knowledge and consequent improvement of the concepts and assumptions adopted can arise from experimental field investigation. For instance, measurements of soil water content at short intervals and including the top layers (0-30 cm), usually not measured by soil neutron probes, are necessary to investigate the storage of water near surface and its temporal variability. Moreover, long-term measurement of soil water content during the dry period can provide further information about the initial soil moisture conditions.

Although the SAMBA model was developed based on semi-arid field conditions, it has the potential for wider application. For instance, similar conditions with high soil moisture deficits can be found in regions such as East England, caused by the occurrence of long dry spells and low rainfall during summer. In these conditions the alternative procedure of near surface storage certainly represents an improvement from conventional soil water balance techniques in order to describe the actual crop evapotranspiration process and consequently the overall soil water balance. Therefore, the SAMBA model should be tested not only in semi-arid situations but also in different climates.

REFERENCES

- Abd El Samie, S.G. and Sadek, M.A. (2001) Groundwater recharge and flow in the Lower Cretaceous Nubian Sandstone aquifer in the Sinai Peninsula, using isotopic techniques and hydrochemistry. *Hydrogeology Journal*, 9, 378-389.
- Abdulmumin, S. and Misari, S.M. 1990. Crop coefficients of some major crops of the Nigerian semi-arid tropics. *Agricultural Water Management*, 18, 159-171.
- Abdulrazzak, M.J., Sorman, A.U. and Alhames, A.S. (1989) Water balance approach under extreme arid conditions – A case study of Tabalah Basin, Saudi Arabia. *Hydrological Processes*, 3, 107-122.
- Agnew, C. (1991). Evaluation of a soil water balance model for the analysis of agricultural drought in the Sahel. In: M.V.K. Sivakumar, J.S. Wallace, C. Renard and G. Girous, eds. *Soil Water Balance in the Sudano-Sahelian Zone: proceedings of the Niamey Workshop, February 1991*. IAHS, 199, 583-592.
- Allen, R., Pereira, L.S., Raes, D. and Smith, M. (1998) Crop evapotranspiration: guidelines for computing crop water requirements. *Irrigation and Drainage Paper* 56. FAO, Rome, Italy.
- Allen, R.G., Smith, M., Pruitt, O. and Pereira, S. (1996) Modifications to the FAO Crop Coefficient Approach. In: C.R. Camp, E.J. Sadler and R.E. Yoder, eds. *Evapotranspiration and Irrigation Scheduling: proceedings of the International Conference, San Antonio, Texas, 3-6 November 1996*. ASAE, 124-132.
- Allen, S.J. (1990) Measurements and estimation of evaporation from soil under sparse barley crops in Northern Syria. *Agricultural and Forest Meteorology*, 49, 291-309.
- Allison, G.B., Gee, G.W. and Tyler, S.W. (1994). Vadose-zone techniques for estimating groundwater recharge in arid and semiarid regions. *Soil Sci Soc Am Journal*, 58, 6-14.
- Allison, G.B. (1988) A review of some of the physical, chemical, and isotopic techniques available for estimating groundwater recharge. In: I. Simmers (ed), *Estimation of natural groundwater recharge*. D. Reidel Publishing Company, Dordrecht, The Netherlands: 49-72.
- Arora, V.K. and Gajri, P.R. (1996) Performance of simplified water balance models under maize in a semiarid subtropical environment. *Agricultural Water Management*, 31, 51-64.
- Audu, I. (1999) *Development and application of runoff model for water harvesting in North-East Nigeria*. PhD Thesis. Cranfield University, Silsoe.

- Barnes, C.J., Jacobson, G. and Smith, G.D. (1994). The distributed recharge mechanism in the Australian arid zone. *Soil Science Society American Journal*. 58, 31-40.
- Benoit, P. (1977) The start of the growing season in Northern Nigeria. *Agricultural Meteorology*, 18, 91-99.
- Borg, H. and Grimes, D.W. (1986) Depth development of roots with time: An empirical description. *Transactions of the American Society of Agricultural Engineers*, 29 (1), 194-197.
- Bromley, J., Taylor, C.M. and Gash, J.H.C. (2002) A comment on 'Long-term rise in a Sahelian water-table: the Continental Terminal in South-West Niger' by Leduc, C., Favreau, G., Schoroeter, P., 2001. *Journal of Hydrology* 243, 43-54. *Journal of Hydrology*, 255, 260-262.
- Campbell, G.S. (1985) *Soil Physics with Basic: Transport Models for Soil-Plant Systems*. Amsterdam. Elsevier, 150 p.
- Canadell, J., Jackson, R.B., Ehleringer, J.R., Mooney, H.A., Sala, O.E. and Schulze, E.D. (1996) Maximum rooting depth of vegetation types at the global scale. *Oecologia*, 108, 583-595.
- Carter, R.C, Rushton, K.R., Eilers, V.H.M. and Hassan, M. (2002) Modelling with limited data: "Plausibility" as a measure of model reliability. In: *Proceedings of 4th International Conference on Calibration and Reliability in Groundwater Modelling (ModelCARE'2002), Prague, June 2002*, 328-330.
- Carter, R.C. and Alkali, A.G. (1996). Shallow groundwater in the northeast arid zone of Nigeria. *Quarterly Journal of Engineering Geology*, 29, 341-355.
- Carter, R.C. (1994) The groundwater hydrology of the Manga Grasslands, northeast Nigeria: importance to agricultural development strategy for the area. *Quarterly Journal of Engineering Geology*, 27, 573-583.
- Cuenca, R.H., Brouwer, J., Chanzy, A., Droogers, P., Galle, S., Gaze, S.R., Sicot, M., Stricker, H., Angulo-Jaramillo, R., Boyle, S.A., Bromley, J., Chebhouni, A.G., Cooper, J.D., Dixon, A.J., Fies, J.C., Gandah, M., Gaudu, J.C., Laguerre, L., Lecocq, J., Soet, M., Steward, H.J., Vandervaere, J.P. and Vauclin, M. (1997) Soil measurements during HAPEX-Sahel intensive observation period. *Journal of Hydrology*, 188-189, 224-266.
- Doorenbos, J., Kassam, A.H., Bemtvelde, C. and Uittenboogaard, G. (1978) Yield response to water. *FAO Irrigation and Drainage Paper* 33. FAO, Rome.
- Doorenbos, J. and Pruitt, W.O. (1977) Guidelines for predicting crop water requirements. *FAO Irrigation and Drainage Paper* 24. FAO, Rome.
- Edmunds, W.M., Fellman, E., Goni, I.B. and Prudhomme, C. (2002) Spatial and temporal distribution of groundwater recharge in northern Nigeria. *Hydrogeology Journal*, 10, 205-215.

- Edmunds, W.M. and Tyler, S.M. (2002) Unsaturated zones as archives of past climates: toward a new proxy for continental regions. *Hydrogeology Journal*, 10, 216-228.
- Edmunds, W.M., Fellman, E. and Goni, I.B. (1999) Lakes, groundwater and palaeohydrology in the Sahel of NE Nigeria: evidence from hydrogeochemistry. *Journal of the Geological Society, London*, 156, 345-355.
- Finch, J.W. (1998) Estimating direct groundwater recharge using a simple water balance model – sensitivity to land surface parameters. *Journal of Hydrology*, 211, 112-125.
- Fowler, A. (2002) Assessment of the validity of using mean potential evaporation in computations of the long-term soil water balance. *Journal of Hydrology*, 256, 248-263.
- Gaze, S.R., Simmonds, L.P., Brouwer, J. and Bouma, J. (1997) Measurements of surface redistribution of rainfall and modelling its effect on water balance calculations for a millet field on sandy soil in Niger. *Journal of Hydrology*, 188-189, 267-284.
- Gee, G.W., Wierenga, P.J., Andraski, B.J. Young, M.H., Fayer, M.J. and Rockhold, M.L. (1994) Variations in water balance and recharge potential at three western sites. *Soil Science Society of America Journal*, 58, 63-72.
- Gee, G.W. and Hillel, D. (1988). Groundwater recharge in arid regions: review and critique of estimation methods. *Hydrological Processes*, 2, 255-266.
- Grindley, J. (1967) The estimation of soil moisture deficits. *The Meteorological Magazine*, vol.96, No. 1137, Meteorological Office, UK.
- Gregory, P.J. (1991) Soil and plant factors affecting the estimation of water extraction by crops. In: M.V.K. Sivakumar, J.S. Wallace, C. Renard and G. Girous, eds. *Soil Water Balance in the Sudano-Sahelian Zone: proceedings of the Niamey Workshop, February 1991*. IAHS, 199, 261-273.
- Grema, A.K. and Hess, T.M. (1994) Water balance and water use of pearl millet-cowpea intercrops in northeast Nigeria. *Agricultural Water Management*, 26, 169-185.
- Grema, A.K. (1994) *Productivity of Pearl Millet-Cowpea intercrops in North-East Nigeria*. PhD Thesis. Cranfield University, Silsoe.
- Hendrickx, J.M.H. and Walker, G.R. (1997) Recharge from Precipitation. In: I. Simmers, ed. *Recharge of aquifers in (semi-) arid areas*. International Contributions to Hydrogeology, 19. International Association of Hydrogeologists, Balkema, Rotterdam, 19-99.
- Hanks, R.J. (1992) *Applied Soil Physics: Soil Water and Temperature Applications*. Springer-Verlag New York, 2nd edition, 176p.

- Hess, T.M. (1999) *The impact of climatic variability over the period 1961-1990 on the soil water balance of upland soils in the North East Arid Zone of Nigeria*. PhD Thesis, Cranfield University, Silsoe.
- Hess, T.M. (1998) Trends in reference evapotranspiration in the North East Arid Zone of Nigeria, 1961-91. *Journal of Arid Environments*, 38, 99-115.
- Hess, T.M., Stephens, W. and Maryah, U.M. (1995) Rainfall trends in the North East arid zone of Nigeria (1961-1990). *Agricultural and Forest Meteorology*, 74, 87-97.
- Hess, T.M. and Grema, A.K. (1994) The impact of changing rainfall patterns on the efficiency of rainfall utilisation in pearl millet-cowpea intercropping systems in North-East Nigeria. *Aspects of Applied Biology*, 38, 231-238.
- Hess, T.M. (1994) *BALANCE: Soil Water Balance Program*. Water Management Department, Silsoe College, Cranfield University, UK.
- Hillel, D. (1998) *Environmental Soil Physics*. Academic Press, USA, 771p.
- Hillel, D. (1980) *Applications of Soil Physics*. Academic Press, USA, 385p.
- Hillel, D. (1977) *Computer Simulation of Soil Water Dynamics: A compendium of recent work*. International Development Research Centre, Ottawa, Canada.
- Hillel, D. (1971) *Soil and Water. Physical Principles and Processes*. Academic Press, USA, 288p.
- Howard, K.W.F. and Lloyd, J.W. (1979) The sensitivity of parameters in the Penman evaporation equations and direct recharge balance. *Journal of Hydrology*, 41, 329-344.
- Kennett-Smith, A., Cook, P.G., Walker, G.R. (1994) Factors affecting groundwater recharge following clearing in the south western Murray Basin. *Journal of Hydrology*, 1994, 154, 85-105.
- Kirbi, C., Newson, M.D. and Gilman, K. (eds) (1991) *Plynlimon research: The first two decades*. Report No. 109. Institute of Hydrology, Wallingford, UK.
- Kitching, R., and Shearer, T.R. (1982) Construction and operation of a large undisturbed lysimeter to measure recharge to a chalk aquifer, England. *Journal of Hydrology*, 58, 267-277.
- Kitching, R., Edmunds, W.M., Shearer, T.R. and Walton, N.R.G. (1980) Assessment of recharge to aquifers. *Hydrological Sciences Bulletin* 25 (3), 217-235.
- Kitching, R., Shearer, T.R. and Shedlock, S.D. (1977) Recharge to Bunter Sandstone determined from lysimeters. *Journal of Hydrology*, 33, 217-232.
- Kitching, R. and Bridge, L.R. (1974) Lysimeter installations in sandstone at Styrrup, Nottinghamshire. *Journal of Hydrology*, 23, 219-232.

- Klaij, M.C. and Vachaud, G. (1992) Seasonal water balance of a sandy soil in Niger cropped with pearl millet, based on profile moisture measurements. *Agricultural Water Management*, 21, 313-330.
- Kowal, J.M. and Kassam, A.H. (1978) *Agricultural Ecology of Savanna: A study of West Africa*. Clarendon Press, Oxford, UK.
- Kruseman, G.P. (1997) Recharge from intermittent flow. In: I. Simmers, ed. *Recharge of Phreatic Aquifers in (Semi-) Arid Areas*. International Contributions to Hydrogeology, 19. International Association of Hydrogeologists, Balkema, Rotterdam, 145-184.
- Kutilek, M. and Nielsen, D.R. (1994) *Soil Hydrology*. Catena Verlag, Cremlingen, Germany.
- Lal, R. (1991) Current research on crop water balance and implications for the future. In: M.V.K. Sivakumar, J.S. Wallace, C. Renard and G. Groux, eds. *Soil Water Balance in the Sudano-Sahelian Zone: proceedings of the Niamey Workshop, February 1991*. IAHS, 199, 31-44.
- Landon, J.R. (1991) (ed.) *Booker Tropical Soil Manual: A handbook for soil survey and agricultural land evaluation in the tropics and subtropics*. Longman Scientific and Technical, Harlow.
- Lebel, T., Taupin, J.D. and Amato, N.D. (1997) Rainfall monitoring during HAPEX-Sahel. 1. General rainfall conditions and climatology. *Journal of Hydrology*, 188-189, 74-96.
- Leduc, C., Favreau, G. and Schroeter, P. (2001) Long-term rise in a Sahelian water-table: the Continental Terminal in South-West Niger. *Journal of Hydrology*, 243, 43-54.
- Leduc, C., Bromley, J., Schroeter, P. (1997) Water table fluctuation and recharge in semi-arid climate: some results of the HAPEX-Sahel hydrodynamic survey (Niger). *Journal of Hydrology*, 188-189, 123-138.
- Lerner, D.N., Issar, A.S. and Simmers, I. (1990) *Groundwater recharge: a guide to understanding and estimating natural recharge*. International Contributions to Hydrogeology, Vol. 8. International Association of Hydrogeologists, Verlag Heinz Heise.
- Li, K.Y., De Jong, R. and Boisvert, J.B. (2001) An exponential root-water-uptake model with water stress compensation. *Journal of Hydrology*, 252, 189-204.
- Lloyd, J.W. (1986) A review of aridity and groundwater. *Hydrological Processes*, 1, 63-78.
- Lloyd, J.W., Drennan, D.S.H. and Bennell, B.M.U. (1967) A ground-water recharge study in north eastern Jordan. *Proceedings of Institution of Civil Engineers*, vol. 35, No. 6962, 615-631.

- Milville (1991) Etude hydrodynamique et quantification de la recharge des aquifères en climat soudano-sahélien: application à un bassin expérimental au Burkina Faso. In: M.V.K. Sivakumar, J.S. Wallace, C. Renard and G. Girous, eds. *Soil Water Balance in the Sudano-Sahelian Zone: proceedings of the Niamey Workshop, February 1991*. IAHS, 199, 311-328.
- Morgan, R.P.C., Quinton, J.N., Smith, R.E., Govers, G., Poesen, J.W.A. Auerswald, K., Chisci, G., Torri, D., Styczen, M.E. and Folly, A.J.V. (1998) The European soil erosion model (EUROSEM): documentation and user guide. Cranfield University, Silsoe.
- Mortimore, M. (1989) Adapting to Drought. Farmers, Famines and Desertification in West Africa. Cambridge University Press 1989.
- Odigie, D.I. and Anyaeche, O.L. (1991) The estimation of the direct recharge to the upper zone aquifer of the Lake Chad Basin. In: *Arid zone Hydrology and Water Resources: proceedings of the International Conference, Maiduguri, September 1985, Lagos, Nigeria*. University of Maiduguri Press, 1991. 349-365.
- Owonubi, J.J., Abdulmumin, S., Malgwi, W.B. and Muazu, S. (1991) Review of soil water balance studies in the Sudano-Sahelian zone of Nigeria. In: M.V.K. Sivakumar, J.S. Wallace, C. Renard and G. Girous, eds. *Soil Water Balance in the Sudano-Sahelian Zone: proceedings of the Niamey Workshop, February, 1991*. IAHS, 199, 329-338.
- Payne, W.A., Lascano, R.J. and Wendt, C.W. (1991) Annual soil water balance of cropped and fallowed millet fields in Niger. In: M.V.K. Sivakumar, J.S. Wallace, C. Renard and G. Girous, eds. *Soil Water Balance in the Sudano-Sahelian Zone: proceedings of the Niamey Workshop, February 1991*. IAHS 199, 401-411.
- Pereira, L.S., Perrier, A., Ait Kadi, M. and Kabat, P. (eds) (1992) *Crop Water Models*. Special issue of ICID Bulletin 41 (2), New Delhi.
- Penman, H.L. (1950) The water balance of the Stour catchment area. *Journal of Institute Water Engineering*, 4, 457-469.
- Penman, H.L. (1949) The dependence of transpiration on weather and soil conditions. *Journal of Soil Science*, 1, 74-90.
- Peugeot, C., Esteves, M. Galle, S., Rajot, J.L. and Vandervaere, J.P. (1997) Runoff generation processes: results and analysis of field data collected at the East Central Supersite of the HAPEX-Sahel experiment. *Journal of Hydrology*, 188-189, 179-202.
- Phillips, T.A. (1977). An Agricultural Notebook. Longman Ltd, London.
- Ragab, R., Finch, J. and Harding, R. (1997). Estimation of groundwater recharge to chalk and sandstone aquifers using simple soil models. *Journal of Hydrology*, 190, 19-41.

- Rangarajan, R. and Athavale, R.N. (2000) Annual replenishable ground water potential of India – an estimate based on injected tritium studies. *Journal of Hydrology*, 234, 38-53.
- Ritchie, J.T. (1972) Model for predicting evaporation from a row crop with incomplete cover. *Water Resources Research*, 8 (5), 1204-1213.
- Rockström, J., Jansson, P.E. and Barron, J. (1998) Seasonal rainfall partitioning under runoff and runoff conditions on sandy soil in Niger. On-farm measurements and water balance modelling. *Journal of Hydrology*, 210, 68-92.
- Rockström, J. and Valentin, C. (1997) Hillslope dynamics of on-farm generation of surface water flows: The case of rain-fed cultivation of pearl millet on sandy soil in the Sahel. *Agricultural Water Management*, 33, 183-210.
- Rosenzweig, C. (1998) Representing the land surface in global climate models. In: D. Hillel. *Environmental Soil Physics*. Academic Press, 691-706.
- Rushton, K.R. (1988) Numerical and conceptual models for recharge estimation in arid and semi-arid zones. In: I. Simmers, ed. *Estimation of natural groundwater recharge*. NATO ASI Series C, Reidel Publishing Co. Vol 222, 223-238.
- Rushton, K.R. and Ward, C. (1979) The estimation of groundwater recharge. *Journal of Hydrology*, 41, 345-361.
- Scanlon, R.B., Healy, R.W. and Cook, P.G. (2002) Choosing appropriate techniques for quantifying groundwater recharge. *Hydrogeology Journal*, 10, 18-39.
- Seymour, K.J., Wyness, A.J. and Rushton, K.R. (1998) The Fylde aquifer – a case study in assessing the sustainable use of groundwater resources. In: *Hydrology in a Changing Environment*, vol II. British Hydrological Society.
- Simmers I. (1998) Groundwater recharge: an overview of estimation problems and recent developments. In: N.S. Robins, ed. *Groundwater Pollution, Aquifer Recharge and Vulnerability*. Geological Society, London, Special Publications, 130, 107-115.
- Simmers, I. (1997) Groundwater recharge principles, problems and developments. In: I. Simmers, ed. *Recharge of Phreatic Aquifers in (Semi-) Arid Areas*. International Contributions to Hydrogeology, 19. International Association of Hydrogeologists, Balkema, Rotterdam, 1-18.
- Sivakumar, M.V.K. (1992). Empirical analysis of dry spells for agricultural applications in West Africa. *Journal of Climate*, vol 5, 532-539.
- Smith, M., Pereira, L.S., Berengena, J., Itier, B., Goussard, J., Ragab, R., Tollefson, L., and Van Hofwegen, P. (eds) (1996) *Irrigation Scheduling: From Theory to Practice*. Proceedings ICID/FAO Workshop, Sept. 1995, Rome. Water Report No. 8, FAO, Rome.

- Smith, M. (1992) CROPWAT. A computer program for irrigation planning and management. . *Irrigation and Drainage Paper No. 46*. FAO, Rome, Italy.
- Smith, M., Allen, R.G., Monteith, J.L., Perrier, A., Pereira, L., and Segeren, A. (1992) Report of the expert consultation on procedures for the revision of FAO guidelines for prediction of crop water requirements. UN-FAO, Rome, Italy, 54 p.
- Sophocleous, M.A. (1991) Combining the soilwater balance and water-level fluctuation methods to estimate natural groundwater recharge: practical aspects. *Journal of Hydrology*, 1991, 124, 229-241.
- Stern, R.D., Dennett, M.D. and Dale, I.C. (1982) Analysing daily rainfall measurements to give agronomically useful results. I. Direct methods. *Experimental Agriculture*, 18, 223-236.
- Walker, G.R., Zhang, L., Ellis, T.W., Hatton, T.J. and Petherram, C. (2002) Estimating impacts of changed land use on recharge: review of modelling and other approaches appropriate for management of dryland salinity. *Hydrogeology Journal*, 10, 68-90.
- Walker, H.B. and Langridge, J.L. (1996) Modelling plant and soil water dynamics in semi-arid ecosystems with limited site data. *Ecological Modelling*, 87, 153-167.
- Wallace, J.S. and Holwill, C.J. (1997) Soil evaporation from tiger-bush in south-west Niger. *Journal of Hydrology*, 188-189, 426-442.
- Wallace, J.S., Lloyd, C.R. and Sivakumar, M.V.K. (1993) Measurements of soil, plant and total evaporation from millet in Niger. *Agricultural and Forest Meteorology*, 63, 149-169.
- Wallace, J.S. (1991) The measurement and modelling of evaporation from semiarid land. In: M.V.K. Sivakumar, J.S. Wallace, C. Renard and G. Girous, eds. *Soil Water Balance in the Sudano-Sahelian Zone: proceedings of the Niamey Workshop, February 1991*. IAHS 199, 131-148.
- Wellings, S.R. and Bell, J.P. (1980) Physical controls of water movement in the unsaturated zone. *Q.J.Eng.Geol. London*, 15, 235-241.
- Zhang, L., Hume, I.H., O'Connell, M.G., Mitchell, D.C., Milthorpe, P.L., Yee, M., Dawes, W.R. and Hatton, T.J. (1999a) Estimating episodic recharge under different crop/pasture rotations in the Malle region. Part 1. Experiments and model calibration. *Agricultural Water Management*, 42, 219-235.
- Zhang, L., Dawes, W.R., Hatton, T.J., Hume, I.H., O'Connell, M.G., Mitchell, D.C., Milthorpe, P.L. and Yee, M. (1999b) Estimating episodic recharge under different crop/pasture rotations in the Malle region. Part 2. Recharge control by agronomic practices. *Agricultural Water Management*, 42, 237-249.

APPENDIX A

ALGORITHMS FOR THE SOIL MOISTURE BALANCE

There are several stages in carrying out a daily water balance to estimate recharge. The main components of the balance are calculated according to the equations described in the chapter three. The determination of the components related to the determination of the daily soil moisture deficit SMD can be represented in terms of algorithms as follow.

The computational soil water balance algorithms assume that all rainfall events occur early in the morning. Also, all drainage from the root zone (potential recharge)

Infiltration

The amount of water entering the soil system at beginning of the day In is given by the difference between the precipitation P and surface runoff Ro (calculated as in Section 3.3). Note that the total daily precipitation is assumed to be distributed over the day.

1. $In = P - Ro$

Water available for evapotranspiration and redistribution

The water available for evapotranspiration at start of the day AWE is the sum of the infiltrating water plus the near surface storage at start of day NSS' .

2. $AWE = In + NSS'$

If $SMD' < 0$, $AWE = In$ this ensure that available water at start of day is not greater than field capacity.

Near surface storage NSS

The proportion of water kept in the near surface storage at end of the day is

$$3. \quad NSS = (AWE - AE) \cdot FS$$

where FS is the storage fraction and AE is the actual evapotranspiration.

If $AWE - AE > SMD'$, $NSS = 0$ this ensure that no water is stored at near surface above field capacity

The depth of water available to reduce SMD at the end of the day is given by

$$4. \quad SMD = SMD' - (AWE - AE - NSS)$$

Actual evapotranspiration

Potential evapotranspiration PE

$$5. \quad PE = K_c \cdot ETo \quad \text{see Section 3.4 for definition of } K_c \text{ at different crop stages}$$

Total and readily available water TAW, RAW and stress coefficient K_s (Section 2.7.2)

$$6. \quad TAW = 1000 \cdot (\theta_{FC} - \theta_{WP}) \cdot Z_r \quad Z_r = \text{root depth calculated as in figure 3.2}$$

$$7. \quad RAW = p \cdot TAW \quad p = \text{depletion factor}$$

$$8. \quad AE = K_s \cdot PE$$

$$9. \quad K_s = \frac{TAW - SMD}{TAW - RAW} \quad \begin{array}{l} K_s = 1 \text{ when } SMD < RAW \\ K_s = 0 \text{ when } SMD > TAW \end{array}$$

The following algorithms represent the all possible conditions cited in the Section 3.7.

For a bare soil, TEW and REW are used instead of TAW and RAW respectively.

$$10. \quad \text{for } AWE \geq PE, \quad \text{then } AE = PE$$

$$11. \quad \text{for } SMD < RAW, \quad \text{then } AE = PE$$

$$12. \quad \text{for } TAW \geq SMD \geq RAW, \quad \text{then } AE = AWE + K_s(PE - AWE)$$

note that when AWE is zero $AE = K_s \cdot PE$

13. for $SMD > TAW$ and $AWE < PE$, then $AE = AWE$

Potential recharge

14. If $SMD < 0.0$, $Re\ ch = SMD + NSS$ and $SMD = 0; NSS = 0$

***SAMBA – a Semi Arid Model using the soil water
BALance technique***

	A	B	C	D	E	F	G	H	I	J	K	
1	SAMBA model											
2	<i>(Input data in shaded boxes. Non shaded values are calculated automatically)</i>											
3												
4	Crop:		Millet									
5												
6												
7			Julian		Max. Root Depth			Depletion coeff.				
8	Planting:		Date	Day	Zr max	1.2	m	p	0.5			
9	Crop stages		20-Jun	171	Length of growth							
10	development		5-Jul	186	stages (days)			Crop coefficient				
11	maturity		30-Jul	211	Lini	15		Kcini	1.05			
12	late		8-Sep	251	Ldev	25		Kcmid	1.00			
13	harvest		3-Oct	276	Lmid	40		KCend	0.60			
14					Llate	25		Kcbs	1.00			
15					Total	105						
16	Runoff Coefficients											
17	Rainfall		SMD			Soil properties			NSS			
18	intensity		0 - 20	20 - 50	> 50	FC	0.12	fraction				
19	mm/day		20	50		WP	0.03	0.4				
20	0 - 20	20	0.1	0.05	0							
21	20 - 50	50	0.2	0.15	0.1	Soil evap layer			SMD initial			
22	> 50		0.3	0.25	0.2	Ze			0.25	m	10	mm
23												

:	A	B	C	D	E	F	G	H	I	J	K
24											
25	day	month	year	J	Rain	SMD'	RoC	Ro	ETo	Kc	PE
26											
27	5	8	1994	217	30.0	10.0	0.2	6.0	5.00	1.00	5.00
28	6	8	1994	218	20.0	0.0	0.1	2.0	5.00	1.00	5.00
29	7	8	1994	219	0.0	0.0	0.1	0.0	5.00	1.00	5.00
30	8	8	1994	220	20.0	5.0	0.1	2.0	5.00	1.00	5.00
31	9	8	1994	221	0.0	0.0	0.1	0.0	5.00	1.00	5.00
32	10	8	1994	222	0.0	5.0	0.1	0.0	5.00	1.00	5.00
33	11	8	1994	223	0.0	10.0	0.1	0.0	5.00	1.00	5.00
34	12	8	1994	224	50.0	15.0	0.2	10.0	5.00	1.00	5.00
35	13	8	1994	225	0.0	0.0	0.1	0.0	5.00	1.00	5.00
36	14	8	1994	226	0.0	5.0	0.1	0.0	5.00	1.00	5.00

:	L	M	N	O	P	Q	R	S	T	U	V
24											
25	Zr	TAW	RAW	Ks	In	AWE	AE	NSS	to SMD	SMD	Rech
26											
27	1.20	108	54	1.0	24.0	24.0	5.00	7.6	11.4	-1.4	9.0
28	1.20	108	54	1.0	18.0	18.0	5.00	5.2	7.8	-7.8	13.0
29	1.20	108	54	1.0	0.0	0.0	5.00	0.0	-5.0	5.0	0.0
30	1.20	108	54	1.0	18.0	18.0	5.00	5.2	7.8	-2.8	8.0
31	1.20	108	54	1.0	0.0	0.0	5.00	0.0	-5.0	5.0	0.0
32	1.20	108	54	1.0	0.0	0.0	5.00	0.0	-5.0	10.0	0.0
33	1.20	108	54	1.0	0.0	0.0	5.00	0.0	-5.0	15.0	0.0
34	1.20	108	54	1.0	40.0	40.0	5.00	14.0	21.0	-6.0	20.0
35	1.20	108	54	1.0	0.0	0.0	5.00	0.0	-5.0	5.0	0.0
36	1.20	108	54	1.0	0.0	0.0	5.00	0.0	-5.0	10.0	0.0

Spreadsheet Formulas

J number of the day in the year	
	$J27 = \text{TRUNC}(275 * B27 / 9 - 30 + A27) - 2 + \text{IF}(B27 < 3, 2, 0) + \text{IF}(\text{AND}(\text{MOD}(C27, 4) = 0, B27 > 2), 1, 0)$
SMD' soil moisture deficit at the start of the day (mm)	
	$F27 = J22 \quad F28 = \text{IF}(U27 < 0, 0, U27 + V27)$
RoC Runoff constant	
	$G27 = \text{IF}(E27 > \$B\$21, \text{IF}(F27 > \$D\$19, \$E\$22, \text{IF}(F27 > \$C\$19, \$D\$22, \$C\$22)),$ $\text{IF}(E27 > \$B\$20, \text{IF}(F27 > \$D\$19, \$E\$21, \text{IF}(F27 > \$C\$19, \$D\$21, \$C\$21)),$ $\text{IF}(F27 > \$D\$19, \$E\$20, \text{IF}(F27 > \$C\$19, \$D\$20, \$C\$20))))$
Ro Runoff	$H27 = E27 * G27$
Kc coefficient	
	$J27 = \text{IF}(\text{AND}(D27 \geq \$D\$8, D27 < \$D\$10), \$J\$10,$ $\text{IF}(\text{AND}(D27 \geq \$D\$10, D27 < \$D\$11), \$J\$10 + (D27 - \$D\$10) / \$G\$11 * (\$J\$11 - \$J\$10),$ $\text{IF}(\text{AND}(D27 \geq \$D\$11, D27 < \$D\$12), \$J\$11,$ $\text{IF}(\text{AND}(D27 \geq \$D\$12, D27 \leq \$D\$13), \$J\$11 + (D27 - \$D\$12) / \$G\$13 * (\$J\$12 - \$J\$11),$ $\text{IF}(\text{AND}(D27 \geq \$D\$13, D27 < \$D\$13 + 10), \$J\$12 + (D27 - \$D\$13) / 10 * (\$J\$14 - \$J\$12), \$J\$14))))$
PE potential evapotranspiration (mm)	$K27 = J27 * I27$
Zr root depth (m) ; Ze soil evaporative surface	
	$L27 = \text{IF}(D27 < \$D\$8, 0, \text{IF}(D27 < \$D\$10, \$H\$22,$ $\text{IF}(D27 < \$D\$11, \$H\$22 + ((D27 - \$D\$10) / (\$G\$11) * (\$G\$6 - \$H\$22)), \text{IF}(D27 < \$D\$13, \$G\$6, 0))))$
TAW total available water ; TEW total evaporable water (mm)	
	$M27 = \text{MÁXIMO}((\$H\$18 - \$H\$19) * 1000 * L27, (\$H\$18 - 0.5 * \$H\$19) * 1000 * \$H\$22)$
RAW readily available water ; REW readily evaporable water (mm)	
	$N27 = M27 * \$J\6
Ks stress coefficient	
	$O27 = \text{IF}(G27 \geq M27, 0, \text{IF}(G27 > N27, (M27 - G27) / (M27 - N27), 1))$
In infiltration (mm)	$P27 = +E27 - H27$
AWE water available for evapotranspiration (mm)	
	$Q27 = \text{IF}(U26 > 0, P27 + S26, P27)$
NSS near surface water available for the following day (mm)	
	$S27 = \text{MAX}((Q27 - R27) * \$J\$18, 0)$
AE actual evapotranspiration (mm)	
	$R27 = \text{IF}(F27 < N27, K27, \text{IF}(Q27 \geq K27, K27, \text{IF}(F27 \geq M27, Q27, Q27 + O27 * (K27 - Q27))))$
toSMD water transferred to SMD (mm)	
	$T27 = Q27 - R27 - S27$
SMD soil moisture deficit at the end of the day (mm)	
	$U27 = F27 - T27$
Rech potential recharge (mm)	$V27 = \text{IF}(U27 < 0, U27 * -1 + S27, 0)$

APPENDIX C

THE NGURU CLIMATIC DATA SET

Appendix C summarises the climatic data from Nguru (1962-97). The daily rainfall is saved in the attached floppy disks.

Table C. 1. Monthly and annual rainfall. All values in mm.

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	TOTAL
1962	0	0	0	0	20	45	106	180	65	17	9	0	441
1963	0	0	0	0	140	56	110	166	152	26	0	0	650
1964	0	0	0	0	8	30	181	245	73	0	0	0	536
1965	0	0	0	0	0	207	134	191	32	0	0	0	563
1966	0	0	0	17	40	102	80	134	89	0	0	0	461
1967	0	0	0	0	12	75	92	265	74	0	0	0	518
1968	0	0	0	42	21	109	125	183	10	0	0	0	489
1969	0	0	0	0	19	18	122	147	70	15	0	0	391
1970	0	0	0	0	4	12	231	218	71	0	0	0	536
1971	0	0	0	0	61	19	139	154	88	0	0	0	460
1972	0	0	0	0	19	53	23	148	5	0	0	0	248
1973	0	0	0	0	6	16	86	117	34	0	0	0	258
1974	0	0	0	1	9	23	236	225	90	18	0	0	602
1975	0	0	0	0	42	22	126	258	108	2	0	0	557
1976	0	0	0	0	8	15	159	64	99	86	0	0	431
1977	0	0	0	0	0	32	186	258	34	0	0	0	509
1978	0	0	0	16	5	98	291	58	19	10	0	0	497
1979	0	0	0	0	11	131	152	196	71	27	0	0	588
1980	0	0	0	0	23	23	112	141	40	2	0	0	340
1981	0	0	0	0	5	110	149	74	91	0	0	0	429
1982	0	0	0	0	2	56	77	175	84	15	0	0	409
1983	0	0	0	0	0	56	42	99	38	0	0	0	235
1984	0	0	0	38	5	24	116	90	58	1	0	0	332
1985	0	0	14	0	8	52	107	180	58	0	0	0	419
1986	0	0	0	0	0	28	40	123	50	0	0	0	241
1987	0	0	0	0	3	19	74	136	18	0	0	0	250
1988	0	0	0	4	2	15	107	187	5	2	0	0	321
1989	0	0	0	0	1	22	75	161	76	4	0	0	339
1990	0	0	0	0	10	16	242	139	12	0	0	0	418
1991	0	0	0	0	23	44	130	88	42	0	0	0	327
1992	0	0	0	0	49	16	118	146	81	0	0	0	412
1993	0	0	0	0	8	3	156	149	18	0	0	0	333
1994	0	0	0	0	14	62	157	246	126	30	0	0	634
1995	0	0	0	15	1	4	213	119	31	1	0	0	384
1996	0	0	0	0	3	60	124	157	72	6	0	0	421
1997	0	0	0	6	18	22	270	130	86	13	0	0	545
Mean	0	0	0	4	17	47	136	160	60	8	0	0	431
Max	0	0	14	42	140	207	291	265	152	86	9	0	650
Min	0	0	0	0	0	3	23	58	5	0	0	0	235

Table C. 2. Number of days with rainfall greater than zero mm. All values in mm.

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	TOTAL
1962	0	0	0	0	3	9	12	13	8	3	1	0	49
1963	0	0	0	0	5	7	10	11	9	3	0	0	45
1964	0	0	0	0	3	5	12	18	6	0	0	0	44
1965	0	0	0	0	0	12	10	17	5	0	0	0	44
1966	0	0	0	2	3	9	11	8	9	0	0	0	42
1967	0	0	0	0	1	6	10	17	8	0	0	0	42
1968	0	0	0	2	4	10	12	14	5	0	0	0	47
1969	0	0	0	0	3	3	8	14	6	2	0	0	36
1970	0	0	0	0	2	3	13	13	9	0	0	0	40
1971	0	0	0	0	3	2	10	13	5	0	0	0	33
1972	0	0	0	1	4	7	7	12	4	0	0	0	35
1973	0	0	0	0	3	5	9	9	3	0	0	0	29
1974	0	0	0	1	2	3	12	12	9	3	0	0	42
1975	0	0	0	0	3	3	12	11	7	1	0	0	37
1976	0	0	0	0	1	3	8	10	8	5	0	0	35
1977	0	0	0	0	0	5	13	18	4	0	0	0	40
1978	0	0	0	1	1	3	17	8	6	2	0	0	38
1979	0	0	0	0	1	7	12	12	8	1	0	0	41
1980	0	0	0	0	4	4	8	11	3	1	0	0	31
1981	0	0	0	0	3	10	7	7	5	0	0	0	32
1982	0	0	0	0	1	6	9	13	3	1	0	0	33
1983	0	0	0	0	0	3	4	10	3	0	0	0	20
1984	0	0	0	1	1	3	10	5	2	1	0	0	23
1985	0	0	1	0	1	7	7	13	6	0	0	0	35
1986	0	0	0	0	0	5	5	8	6	0	0	0	24
1987	0	0	0	0	2	5	4	10	2	0	0	0	23
1988	0	0	0	2	2	6	7	14	2	1	0	0	34
1989	0	0	0	0	2	3	10	14	6	3	0	0	38
1990	0	0	0	0	4	1	9	8	3	0	0	0	25
1991	0	0	0	0	7	7	12	7	2	0	0	0	35
1992	0	0	0	0	7	2	13	15	8	0	0	0	45
1993	0	0	0	0	1	3	7	14	2	0	0	0	27
1994	0	0	0	0	1	3	6	10	5	4	0	0	29
1995	0	0	0	2	1	1	10	10	10	2	0	0	36
1996	0	0	0	0	2	5	10	11	6	3	0	0	37
1997	0	0	0	3	2	5	7	9	12	2	0	0	40
Mean	0	0	0	0	2	5	10	12	6	1	0	0	36
Max	0	0	1	3	7	12	17	18	12	5	1	0	49
Min	0	0	0	0	0	1	4	5	2	0	0	0	20

Table C. 3. Number of days with rainfall greater than ETo. All values in mm.

	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC	TOTAL
1962	0	0	0	0	1	3	6	8	4	1	1	0	24
1963	0	0	0	0	3	4	6	8	8	2	0	0	31
1964	0	0	0	0	1	2	8	11	5	0	0	0	27
1965	0	0	0	0	0	8	6	10	3	0	0	0	27
1966	0	0	0	1	1	4	5	6	6	0	0	0	23
1967	0	0	0	0	1	3	5	10	5	0	0	0	24
1968	0	0	0	2	1	6	7	8	0	0	0	0	24
1969	0	0	0	0	1	1	5	8	5	1	0	0	21
1970	0	0	0	0	0	1	9	9	5	0	0	0	24
1971	0	0	0	0	3	1	7	9	5	0	0	0	25
1972	0	0	0	0	1	3	1	9	0	0	0	0	14
1973	0	0	0	0	0	1	5	4	2	0	0	0	12
1974	0	0	0	0	1	2	9	8	6	2	0	0	28
1975	0	0	0	0	2	1	8	9	6	0	0	0	26
1976	0	0	0	0	1	1	5	4	5	4	0	0	20
1977	0	0	0	0	0	3	4	12	3	0	0	0	22
1978	0	0	0	1	0	2	10	4	2	1	0	0	20
1979	0	0	0	0	1	4	6	7	3	1	0	0	22
1980	0	0	0	0	1	2	4	6	3	0	0	0	16
1981	0	0	0	0	0	4	6	4	4	0	0	0	18
1982	0	0	0	0	0	2	5	9	2	1	0	0	19
1983	0	0	0	0	0	3	3	7	3	0	0	0	16
1984	0	0	0	1	0	1	7	4	1	0	0	0	14
1985	0	0	1	0	1	3	6	6	5	0	0	0	22
1986	0	0	0	0	0	2	3	6	4	0	0	0	15
1987	0	0	0	0	0	1	4	6	2	0	0	0	13
1988	0	0	0	0	0	1	4	8	0	0	0	0	13
1989	0	0	0	0	0	1	5	8	4	0	0	0	18
1990	0	0	0	0	0	1	7	5	1	0	0	0	14
1991	0	0	0	0	1	2	8	5	2	0	0	0	18
1992	0	0	0	0	1	1	6	10	4	0	0	0	22
1993	0	0	0	0	1	0	7	9	2	0	0	0	19
1994	0	0	0	0	1	2	5	7	3	2	0	0	20
1995	0	0	0	1	0	0	8	7	2	0	0	0	18
1996	0	0	0	0	0	2	5	9	4	0	0	0	20
1997	0	0	0	0	1	1	6	6	6	1	0	0	21
Mean	0	0	0	0	1	2	6	7	3	0	0	0	20
Max	0	0	1	2	3	8	10	12	8	4	1	0	31
Min	0	0	0	0	0	0	1	4	0	0	0	0	12

Table C. 4. Number of days with rainfall greater than 20 mm. All values in mm.

[illegible]

Table C. 5. Rainy season characteristics.

year	1 st rain-day (rain > zero mm)			Last rain-day (rain > zero mm)			Rainy season length number of days
	Day	Month	Julian day	Day	Month	Julian day	
1962	2	5	122	3	11	307	104
1963	10	5	130	18	10	291	186
1964	15	5	135	17	9	260	162
1965	6	6	157	30	9	273	126
1966	26	4	116	28	9	271	117
1967	21	5	141	27	9	270	156
1968	21	4	111	23	9	266	130
1969	10	5	130	15	10	288	156
1970	21	5	141	18	9	261	159
1971	9	5	129	18	9	261	121
1972	5	4	134	22	9	265	133
1973	23	5	143	22	9	265	132
1974	25	4	115	9	10	282	123
1975	13	5	133	7	10	280	168
1976	18	5	138	24	10	297	148
1977	6	6	157	25	9	268	160
1978	21	4	111	27	10	300	112
1979	20	5	140	12	10	285	190
1980	20	5	140	18	10	291	146
1981	18	5	138	23	9	266	129
1982	31	5	151	9	10	282	132
1983	15	6	166	25	9	268	103
1984	24	4	114	11	10	284	171
1985	29	3	88	16	9	259	172
1986	4	6	155	21	9	264	110
1987	21	5	141	4	9	247	107
1988	19	4	109	2	10	275	167
1989	6	5	126	6	10	279	154
1990	8	5	128	14	9	257	130
1991	10	5	130	5	9	248	119
1992	12	5	132	29	9	272	141
1993	25	5	145	17	9	260	116
1994	27	5	147	14	10	287	141
1995	16	4	106	24	10	297	192
1996	10	5	130	12	10	285	156
1997	21	4	111	3	10	276	166
earliest	29	3	88	4	9	247	
latest	15	6	166	3	11	307	

Figure C. 1. Long term monthly rainfall and ETo means in Nguru.

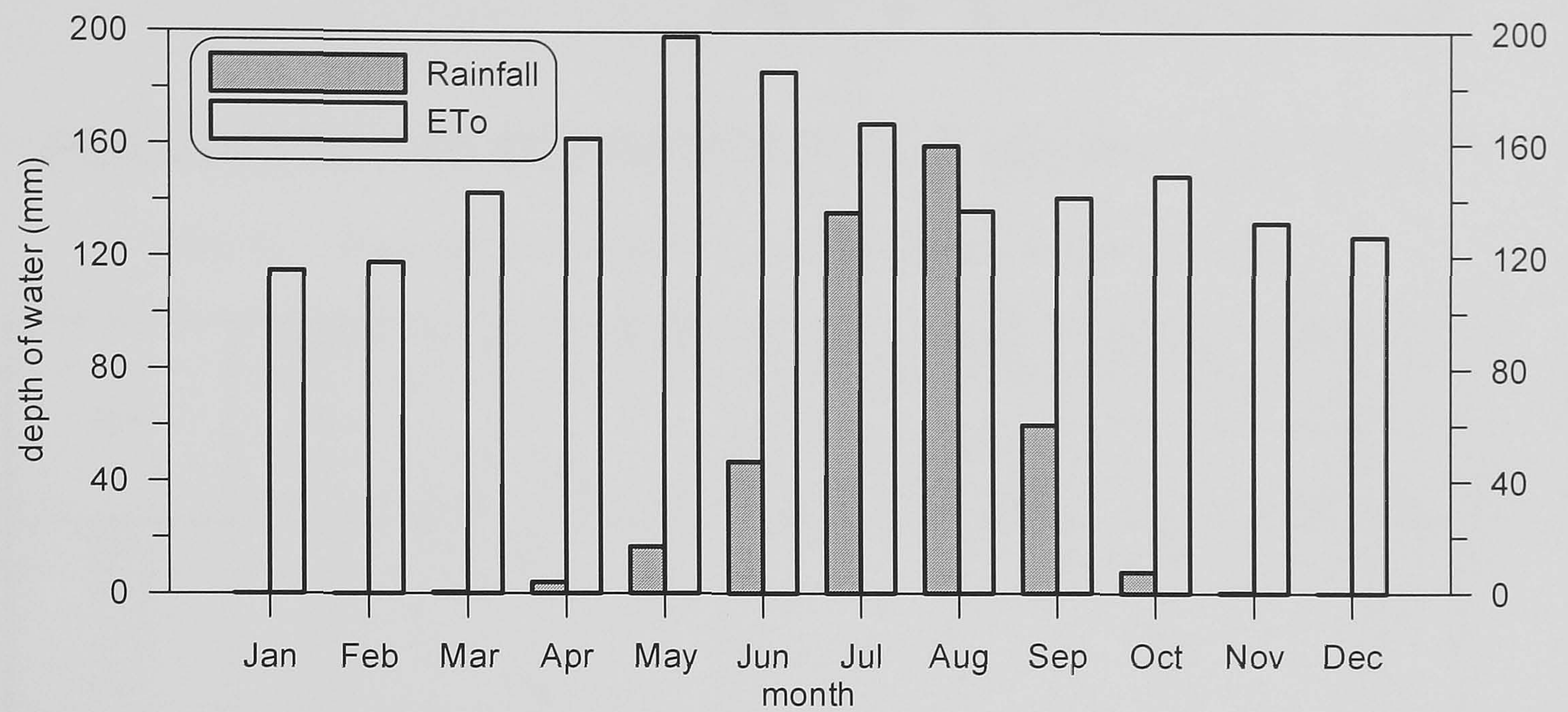
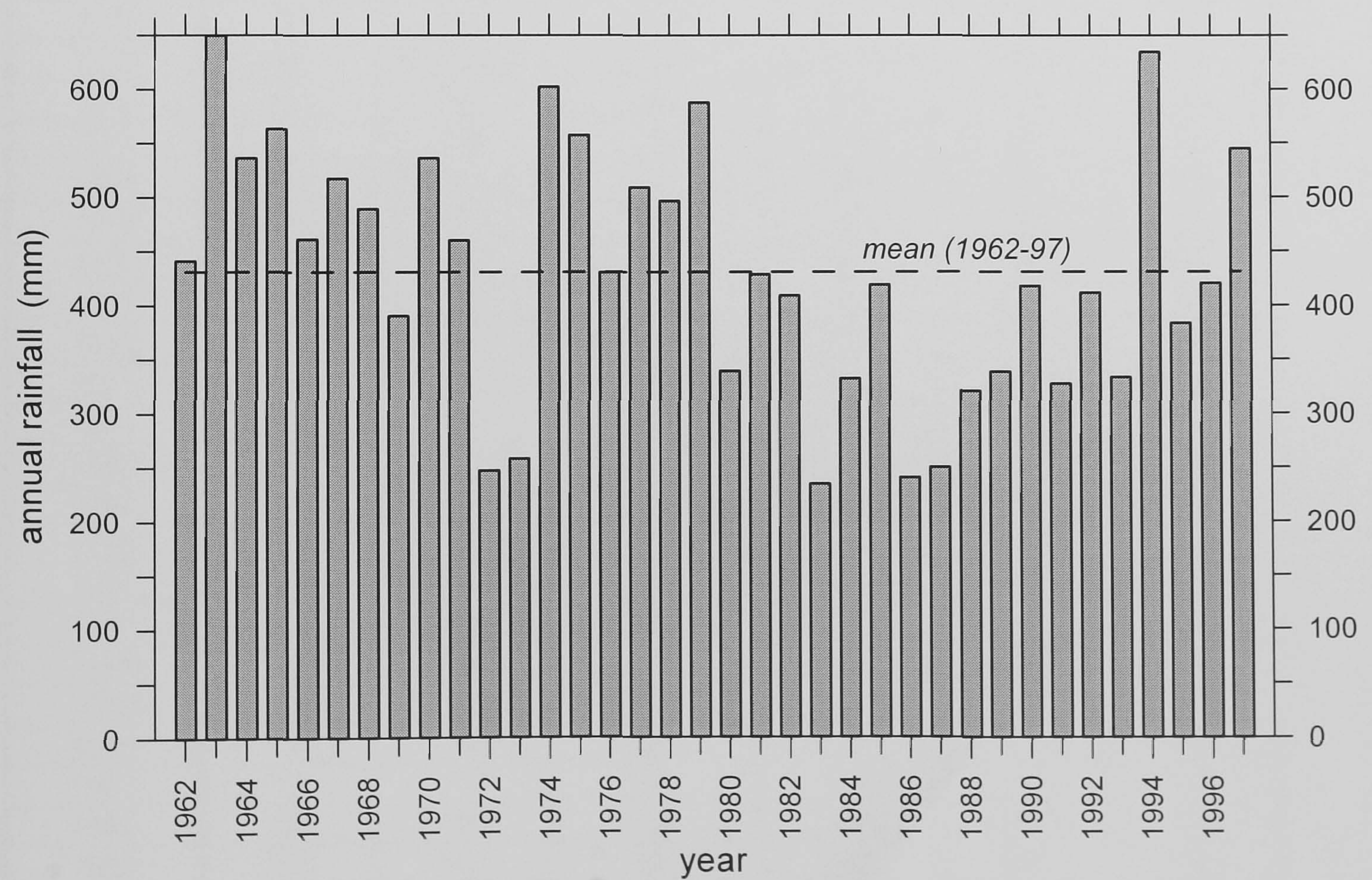


Figure C. 2. Annual rainfall 1962-97, Nguru.

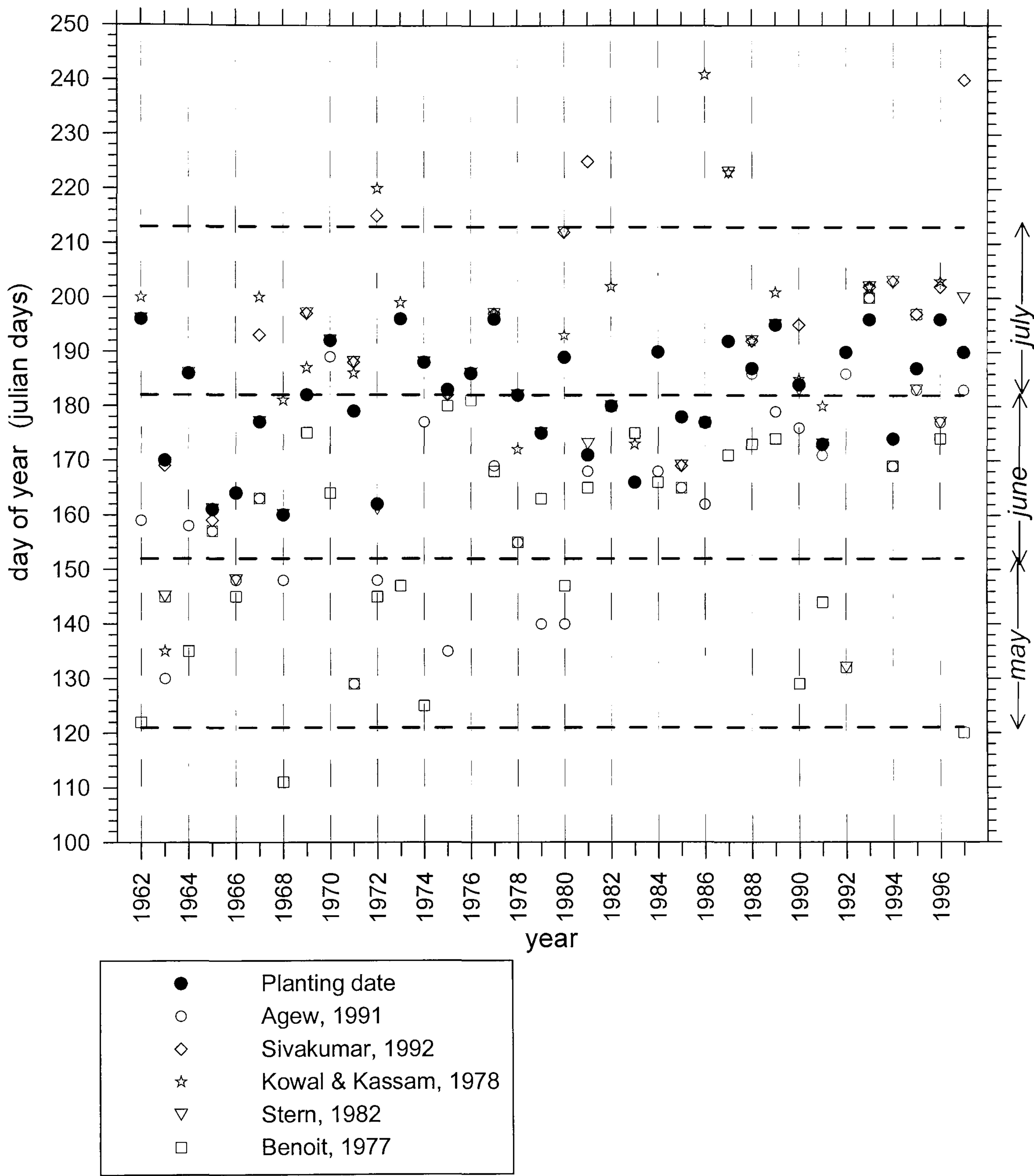


APPENDIX D

PLANTING DATE DEFINITIONS FOR NGURU (1962-97)

Table D. 1. Planting date based on criteria of onset of rains (Julian days)

Year	Agnew 10 mm in 5 days. 10 mm in next 15 days	Sivakumar 20 mm in 3 days. 7 dry days next 30 days	Kowal-Kassam 25 mm in 10 days. > than ETo next 10 days.	Stern 20 mm in 2 days. 10 dry days next 30 days	Benoit Rain > ETo followed by 5 dry days.	Planting date utilised in this research.
1962	159	196	200	196	122	196
1963	130	170	135	145	145	170
1964	158	186	186	186	135	186
1965	157	159	161	161	157	161
1966	148	No onset	164	148	145	164
1967	163	193	200	177	163	177
1968	148	160	181	160	111	160
1969	182	197	187	197	175	182
1970	189	192	192	192	164	192
1971	129	188	186	188	129	179
1972	148	215	220	162	145	162
1973	196	No onset	199	No onset	147	196
1974	177	188	188	188	125	188
1975	135	182	183	182	180	183
1976	186	No onset	186	186	181	186
1977	169	197	197	197	168	196
1978	155	182	172	182	155	182
1979	140	175	175	175	163	175
1980	140	212	193	212	147	189
1981	168	225	171	173	165	171
1982	180	180	202	180	180	180
1983	166	No onset	173	No onset	175	166
1984	168	No onset	190	No onset	166	190
1985	165	169	178	169	165	178
1986	162	No onset	241	No onset	177	177
1987	192	No onset	223	223	171	192
1988	187	192	192	192	173	187
1989	179	195	201	195	174	195
1990	176	195	184	183	129	184
1991	171	173	180	173	144	173
1992	186	190	190	132	132	190
1993	200	202	202	202	200	196
1994	169	203	174	203	169	174
1995	183	197	187	183	197	187
1996	177	202	203	177	174	196
1997	183	240	190	200	120	190

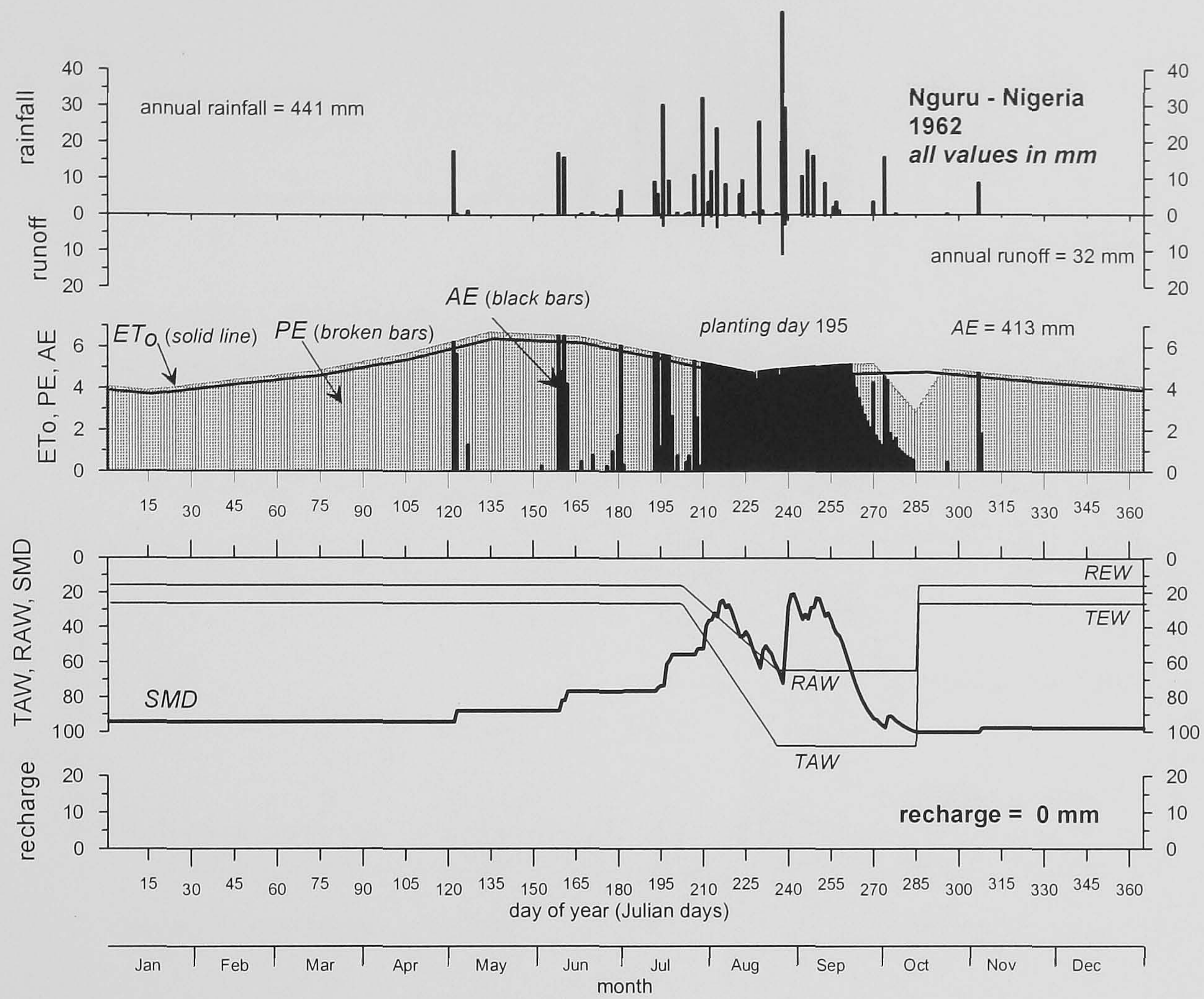
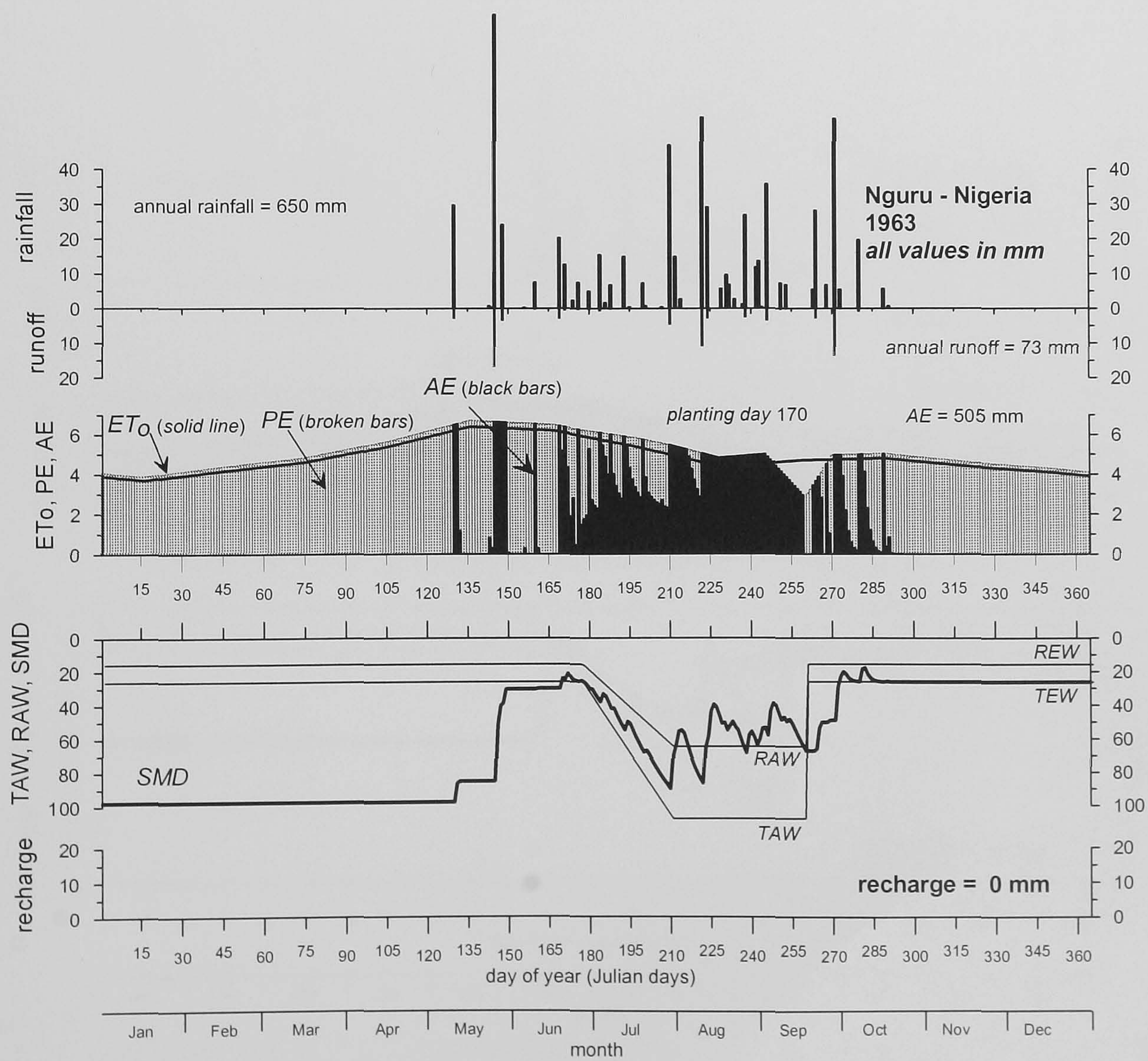


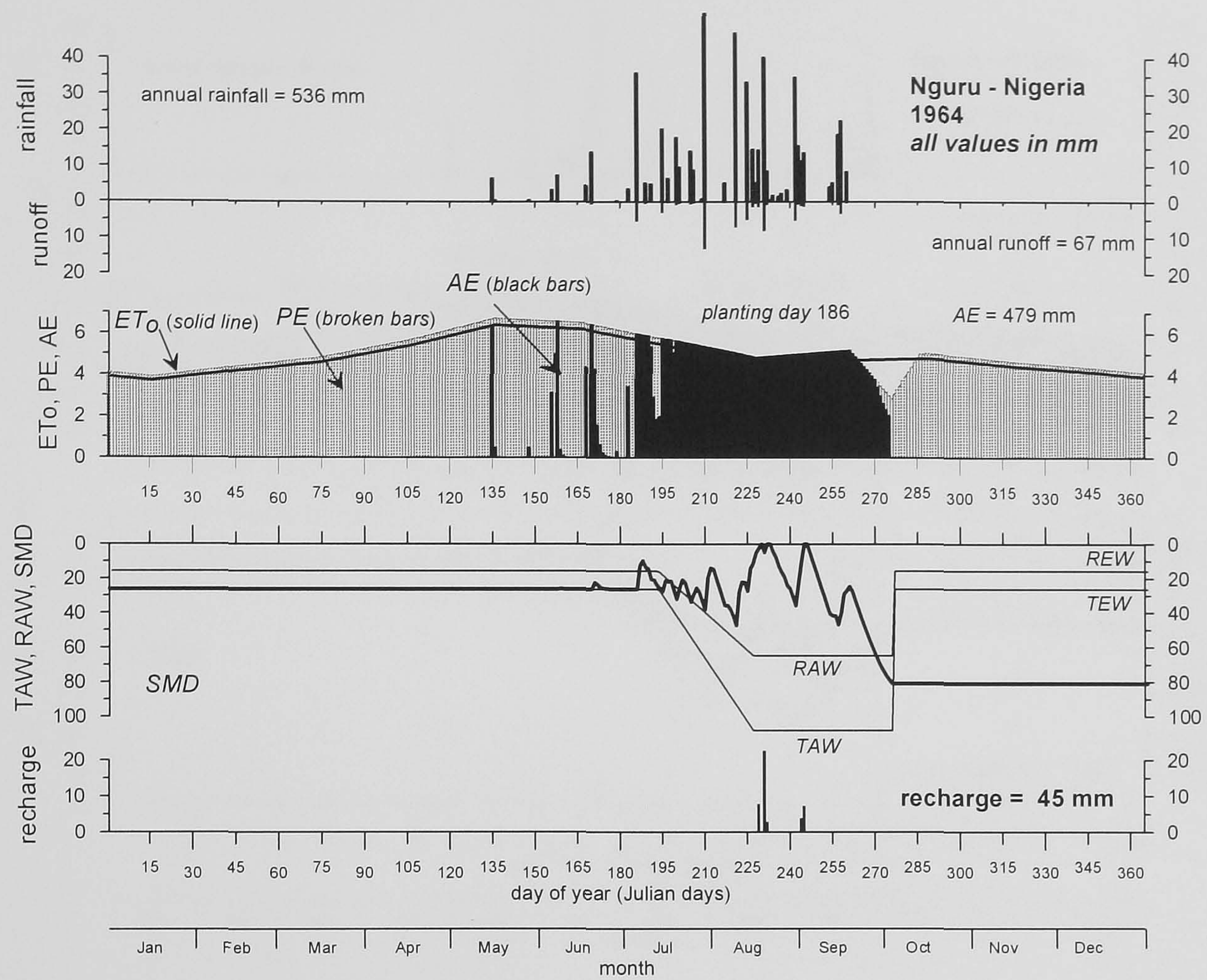
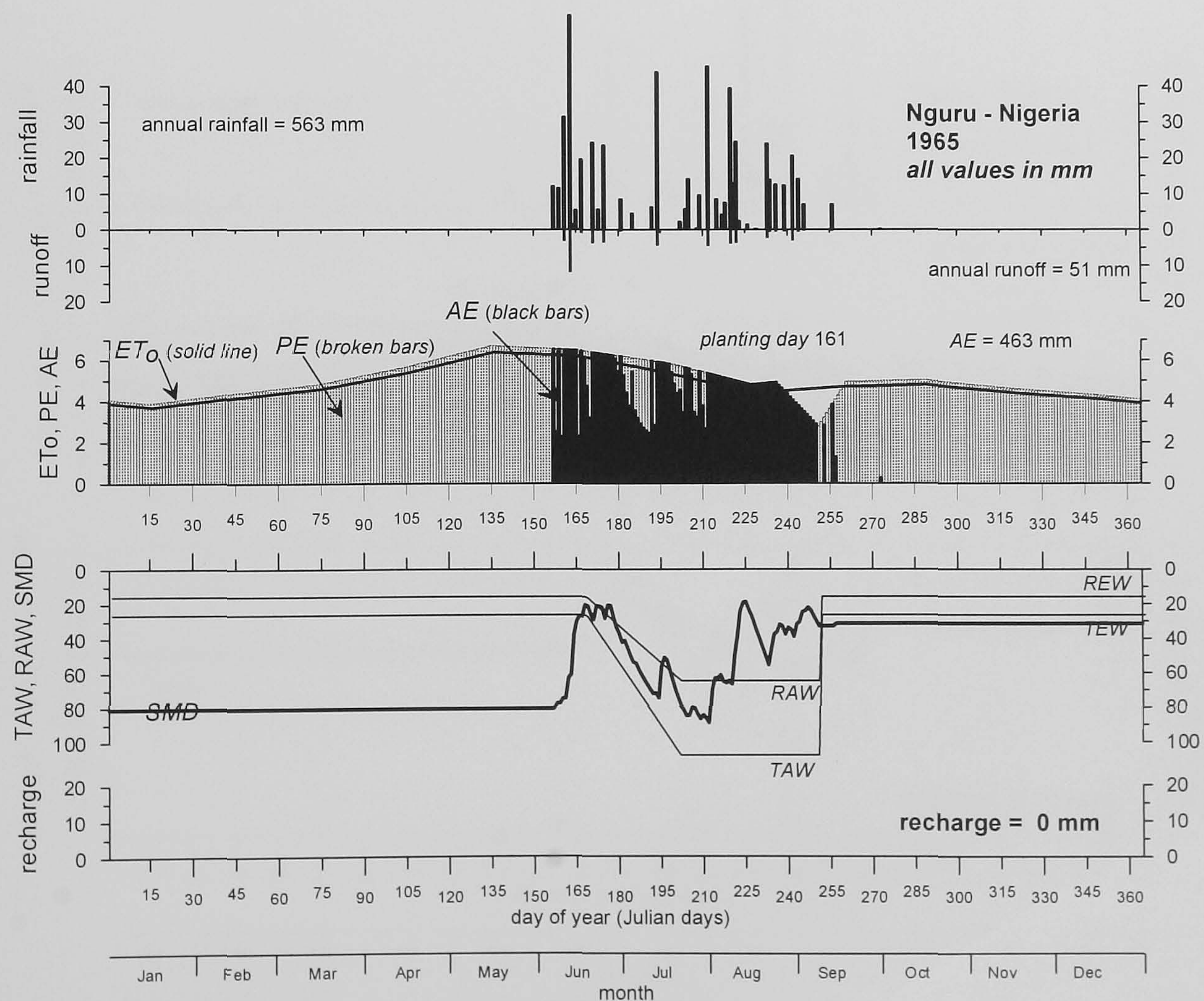
APPENDIX E

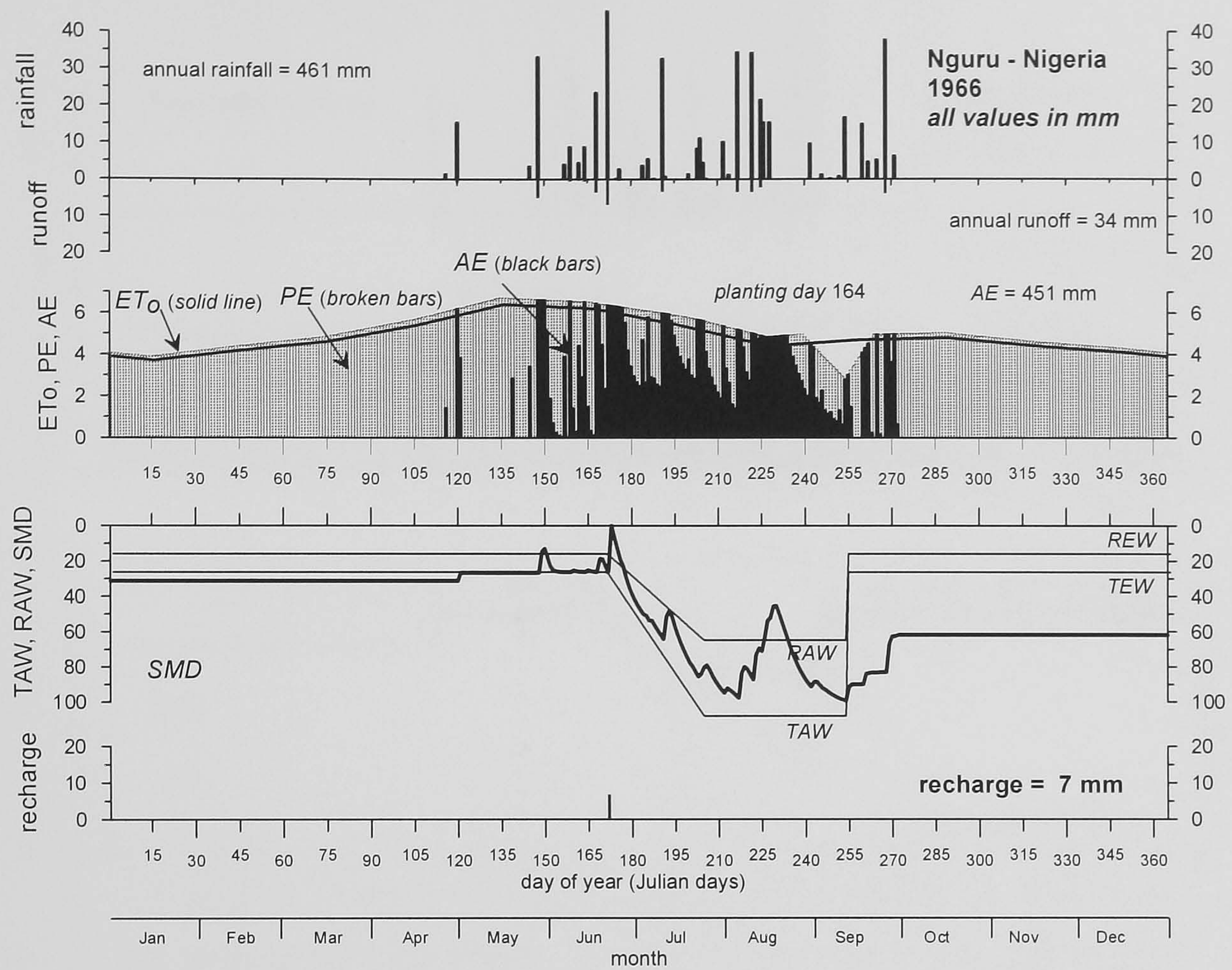
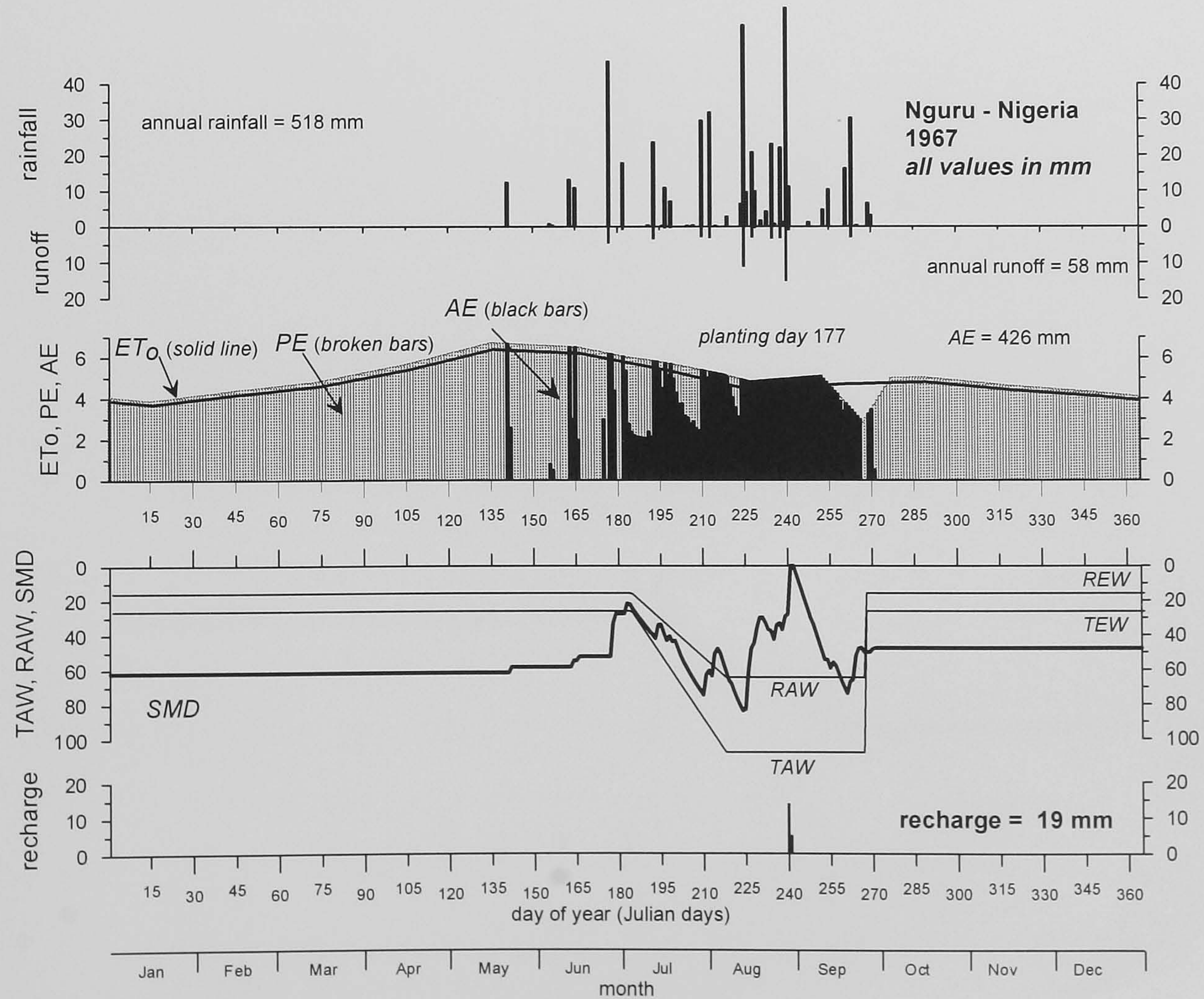
SELECTED WATER BALANCE COMPONENTS FROM THE SAMBA MODEL – NGURU (1962-97)

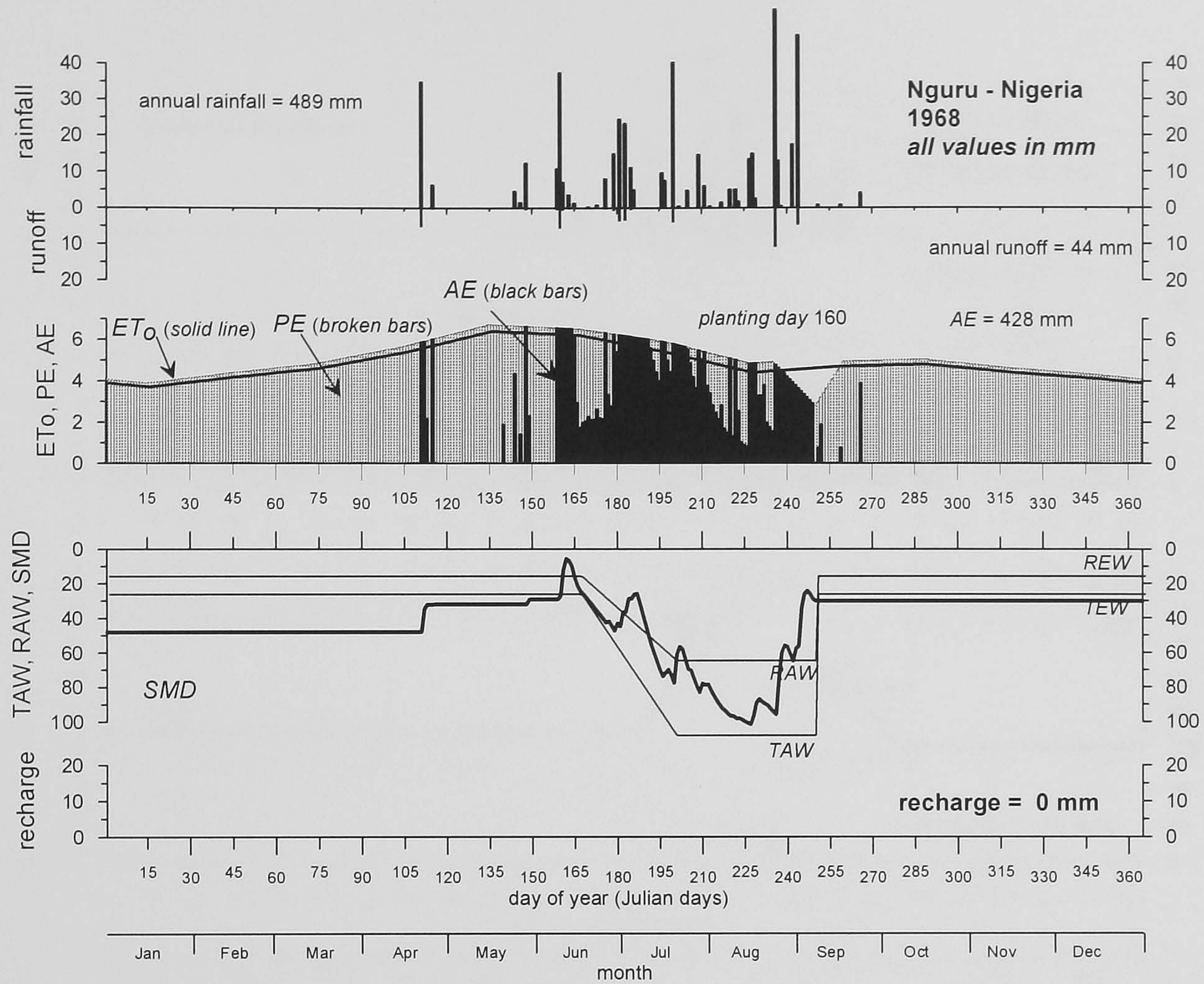
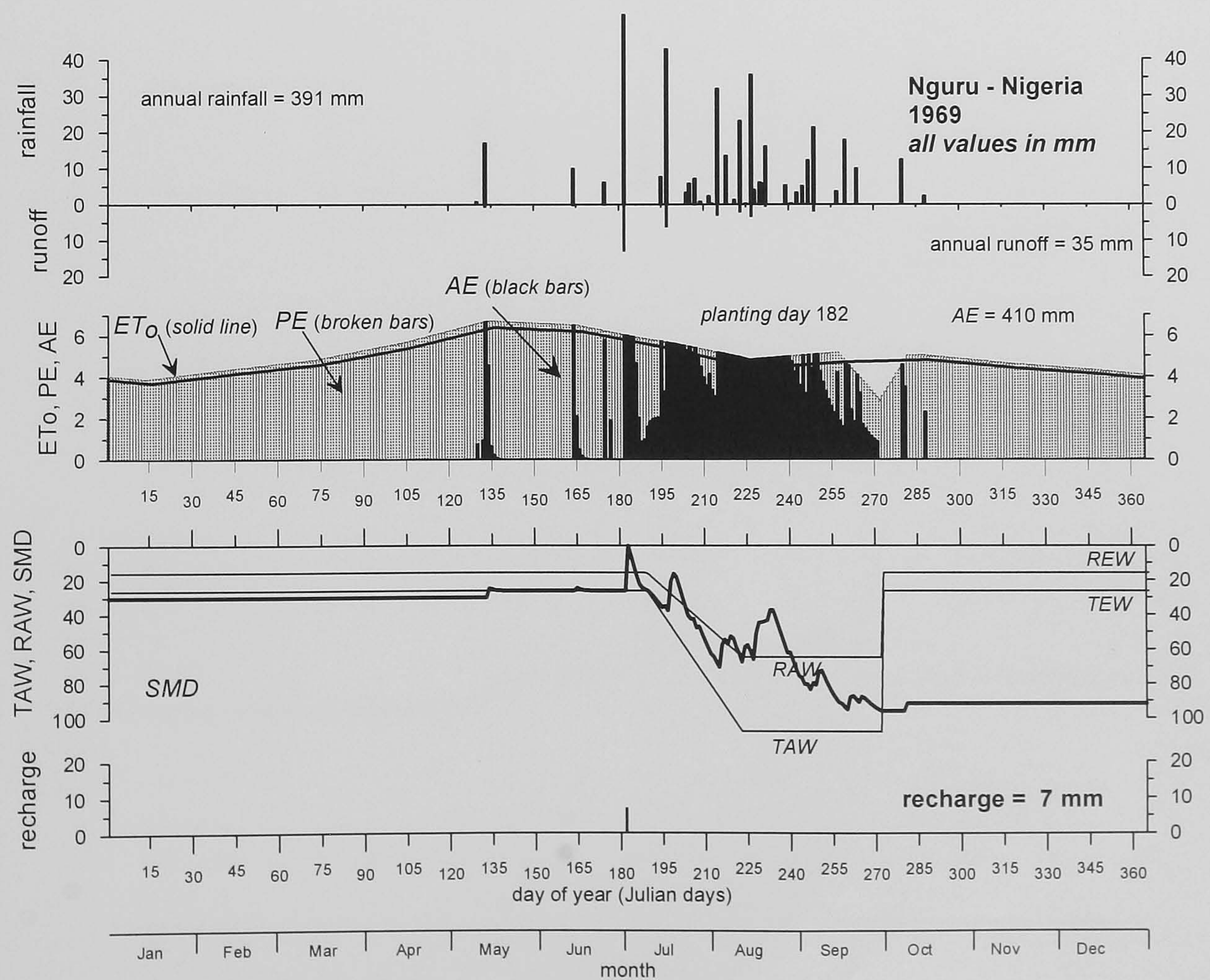
Table E. 1. Annual rainfall and principal outputs using the initial conditions and parameters described in Section 4.2. Rounded values. Initial conditions for 1961 assumed PWP (SMD = 108mm)

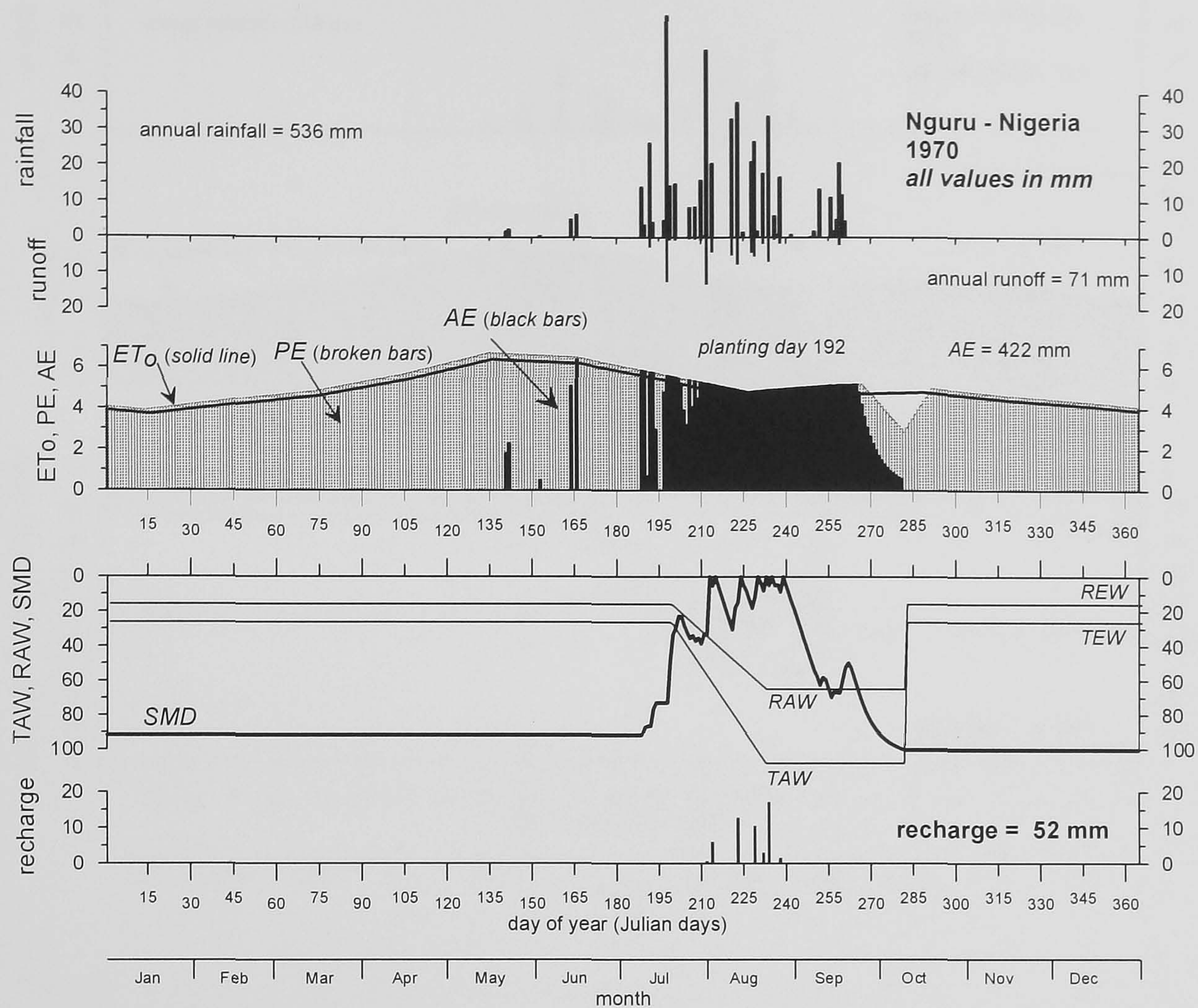
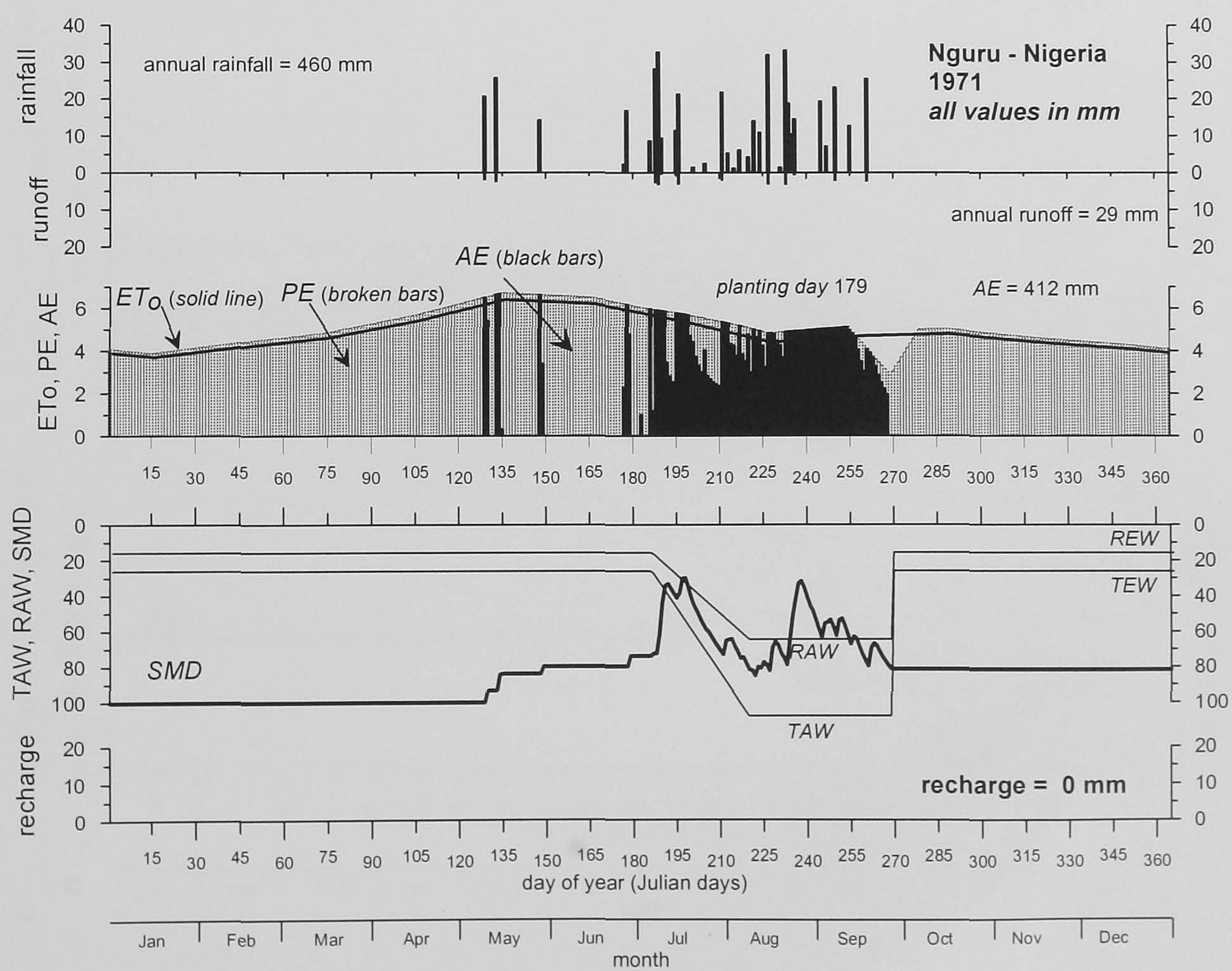
Year	Rainfall (mm)	Runoff (mm)	PE (mm)	AE (mm)	Rech (mm)	SMD at start of rain season (mm)	SMD at end of rain season (mm)	Annual Balance (change in soil water storage) (mm)
1962	441	32	1849	413	0	94	98	-4
1963	650	73	1851	505	0	98	26	72
1964	536	67	1853	479	45	26	81	-55
1965	563	51	1851	463	0	81	31	50
1966	461	34	1851	451	7	31	62	-31
1967	518	58	1850	426	19	62	48	14
1968	489	44	1855	428	0	48	30	17
1969	391	35	1849	410	7	30	92	-62
1970	536	71	1850	422	52	92	100	-8
1971	460	29	1850	412	0	100	82	19
1972	248	11	1855	247	0	82	92	-11
1973	258	12	1849	262	0	92	107	-15
1974	602	75	1849	453	47	107	82	26
1975	557	65	1849	443	25	82	58	24
1976	431	38	1854	383	0	58	49	8
1977	509	75	1849	397	94	49	105	-56
1978	497	66	1849	390	41	105	105	0
1979	588	61	1850	468	0	105	45	60
1980	340	28	1853	375	0	45	105	-60
1981	429	36	1850	345	3	105	61	44
1982	409	31	1850	395	0	61	78	-17
1983	235	5	1850	244	0	78	92	-14
1984	332	29	1853	318	0	92	106	-14
1985	419	22	1850	378	0	106	88	18
1986	241	6	1850	235	0	88	87	1
1987	250	15	1849	254	0	87	107	-19
1988	321	20	1854	299	0	107	105	2
1989	339	13	1849	324	0	105	103	1
1990	418	55	1849	361	4	103	105	-2
1991	327	15	1850	308	0	105	100	5
1992	411	23	1853	386	0	100	98	2
1993	333	26	1849	316	0	98	107	-9
1994	634	94	1850	400	60	107	26	81
1995	384	53	1849	393	15	26	105	-78
1996	421	41	1853	349	29	105	102	2
1997	545	64	1849	412	59	102	94	8

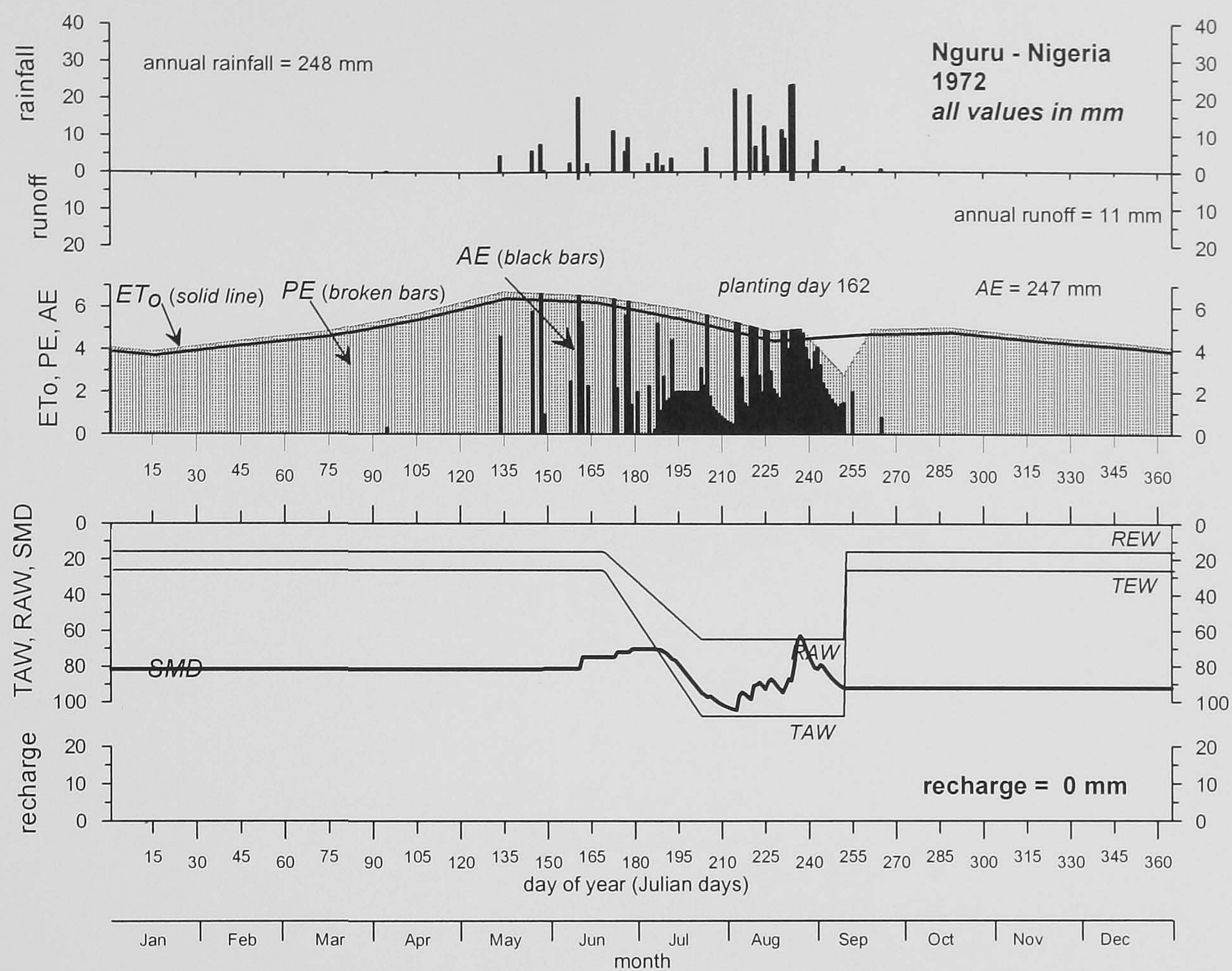
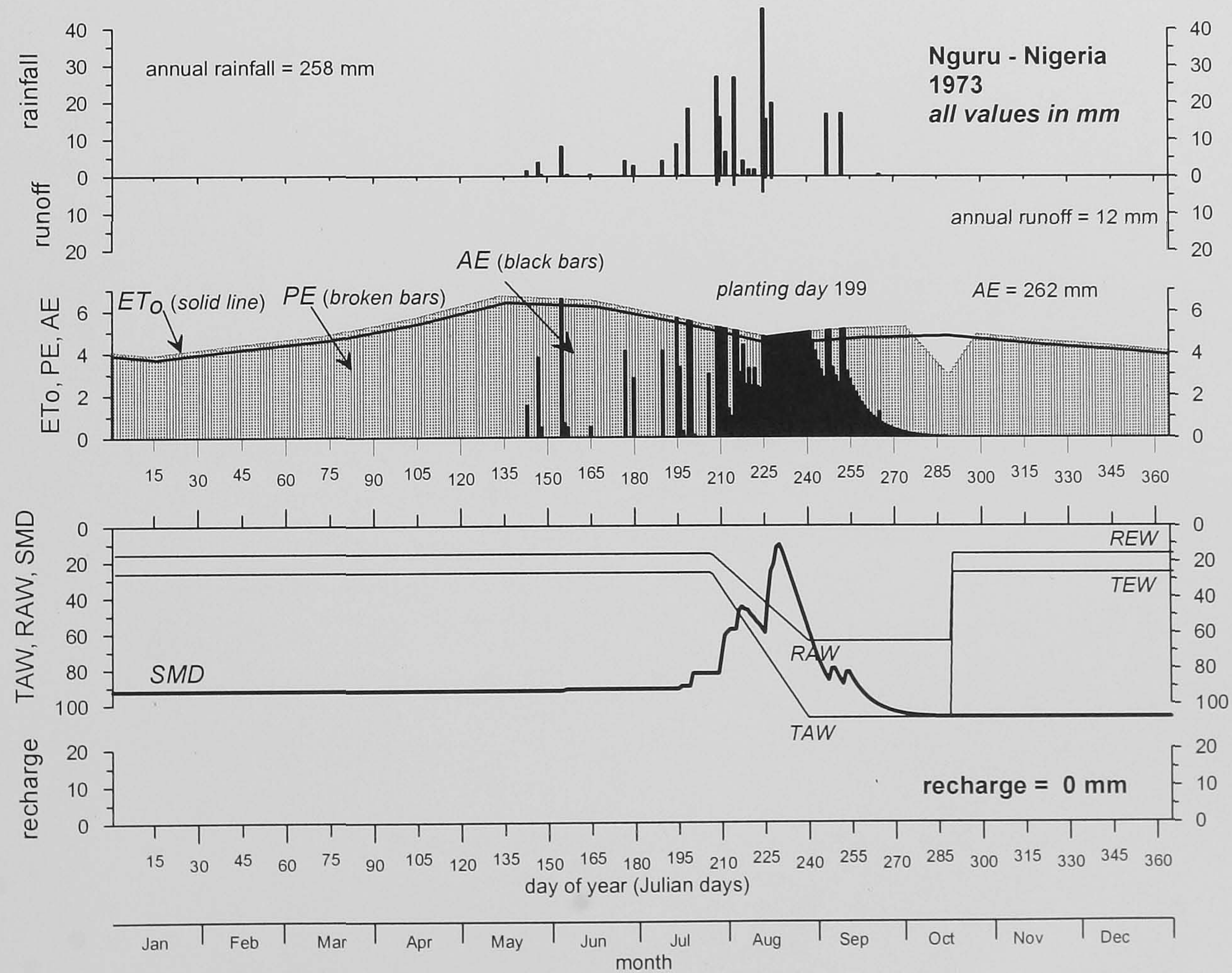


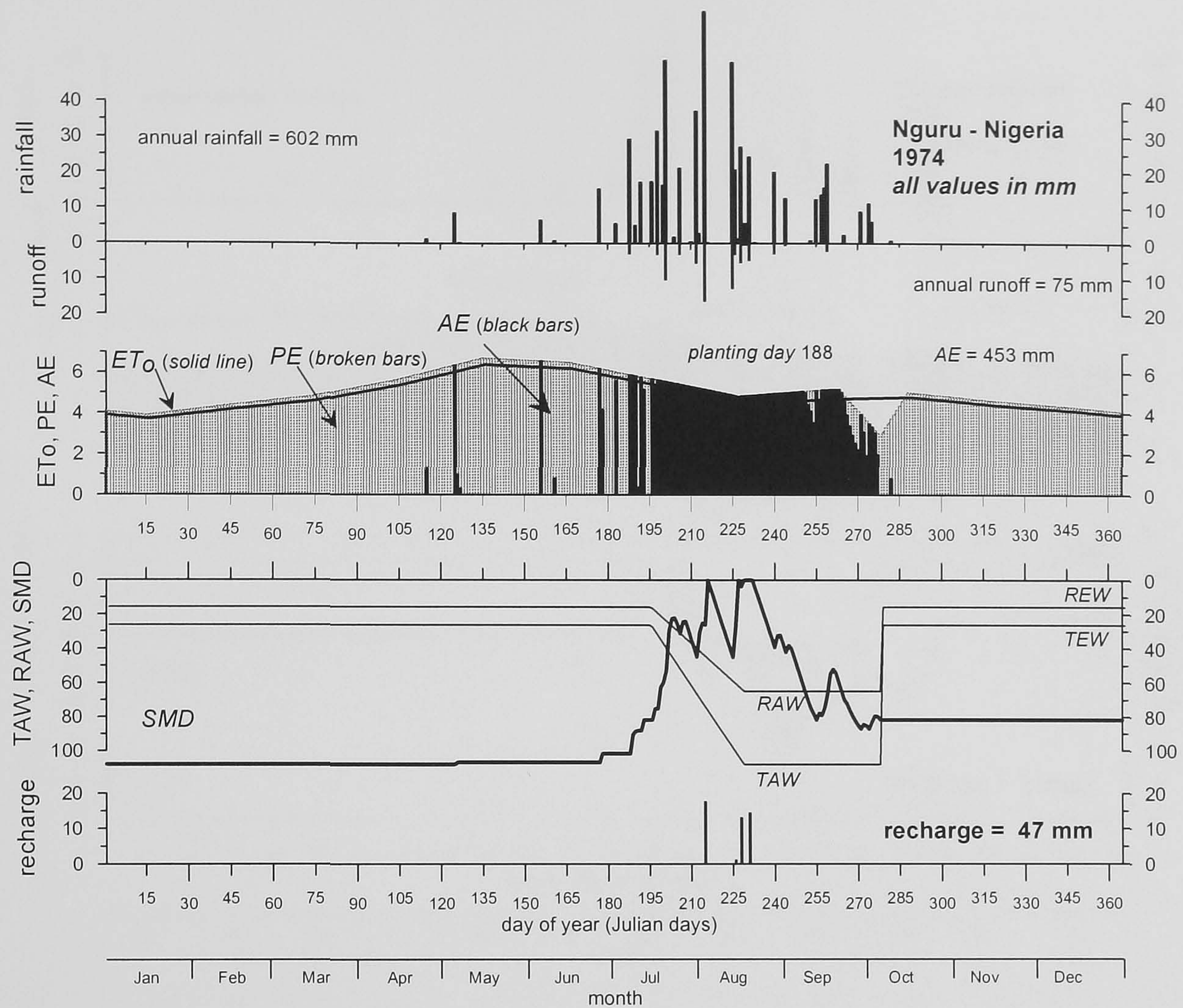
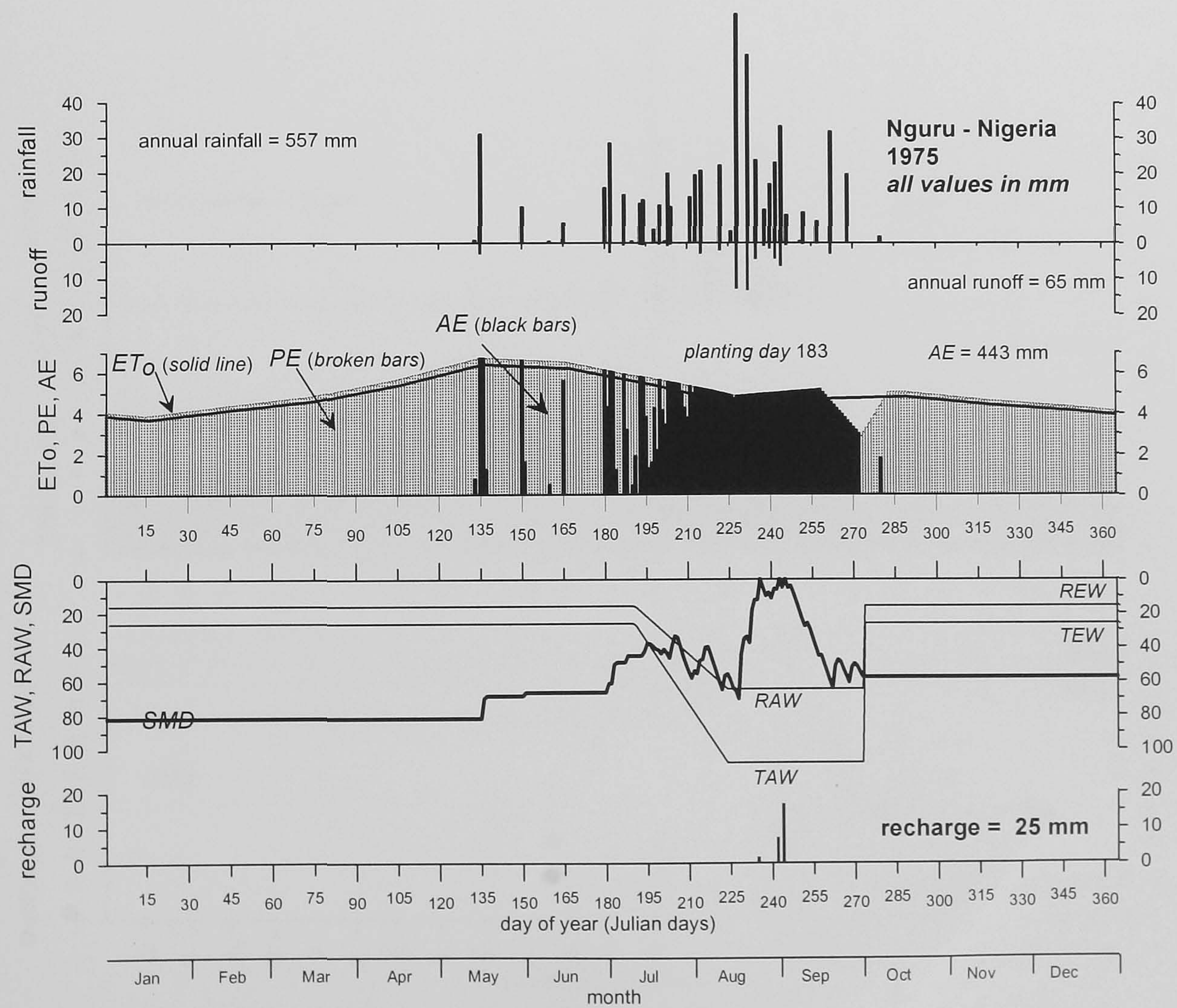


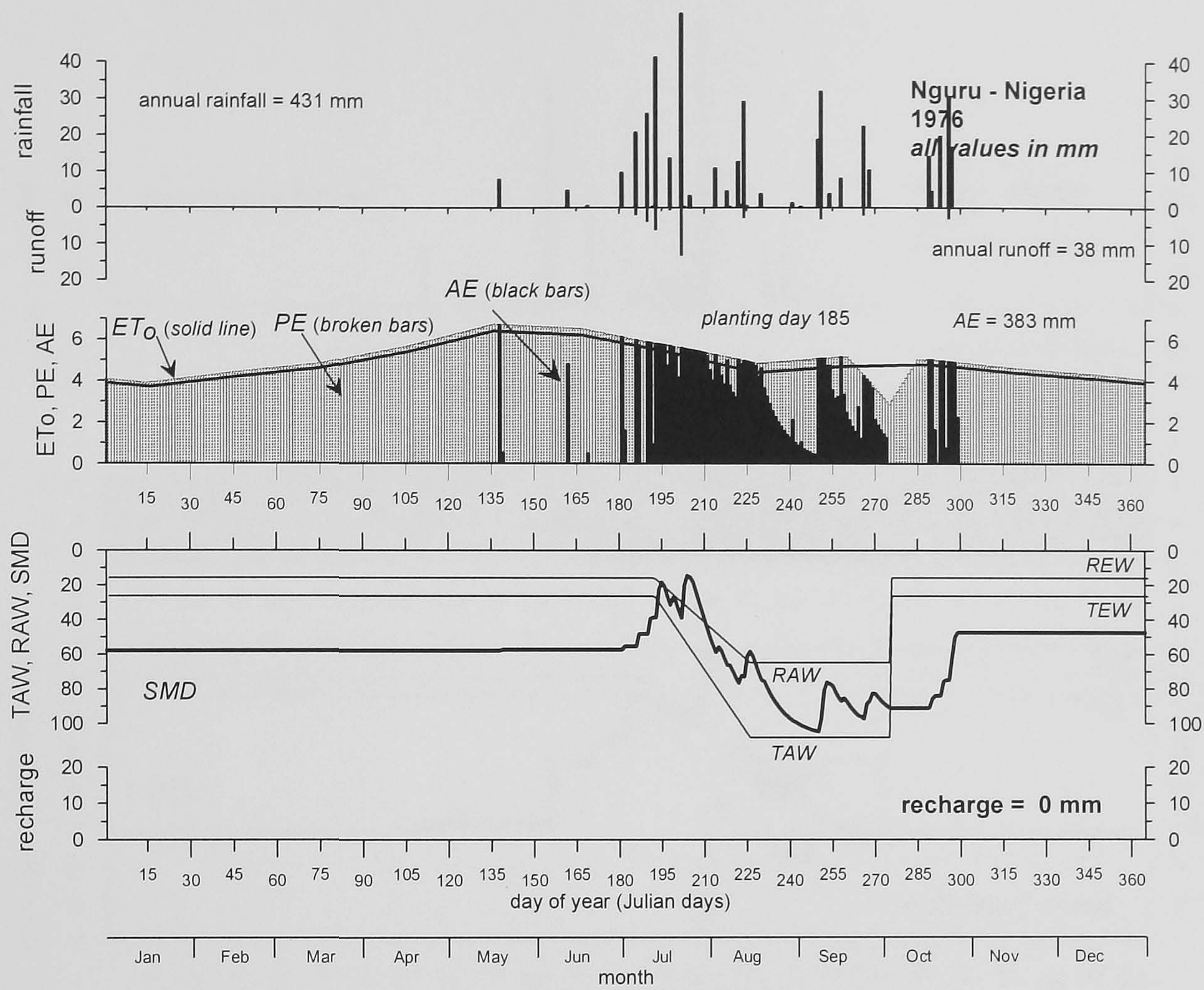
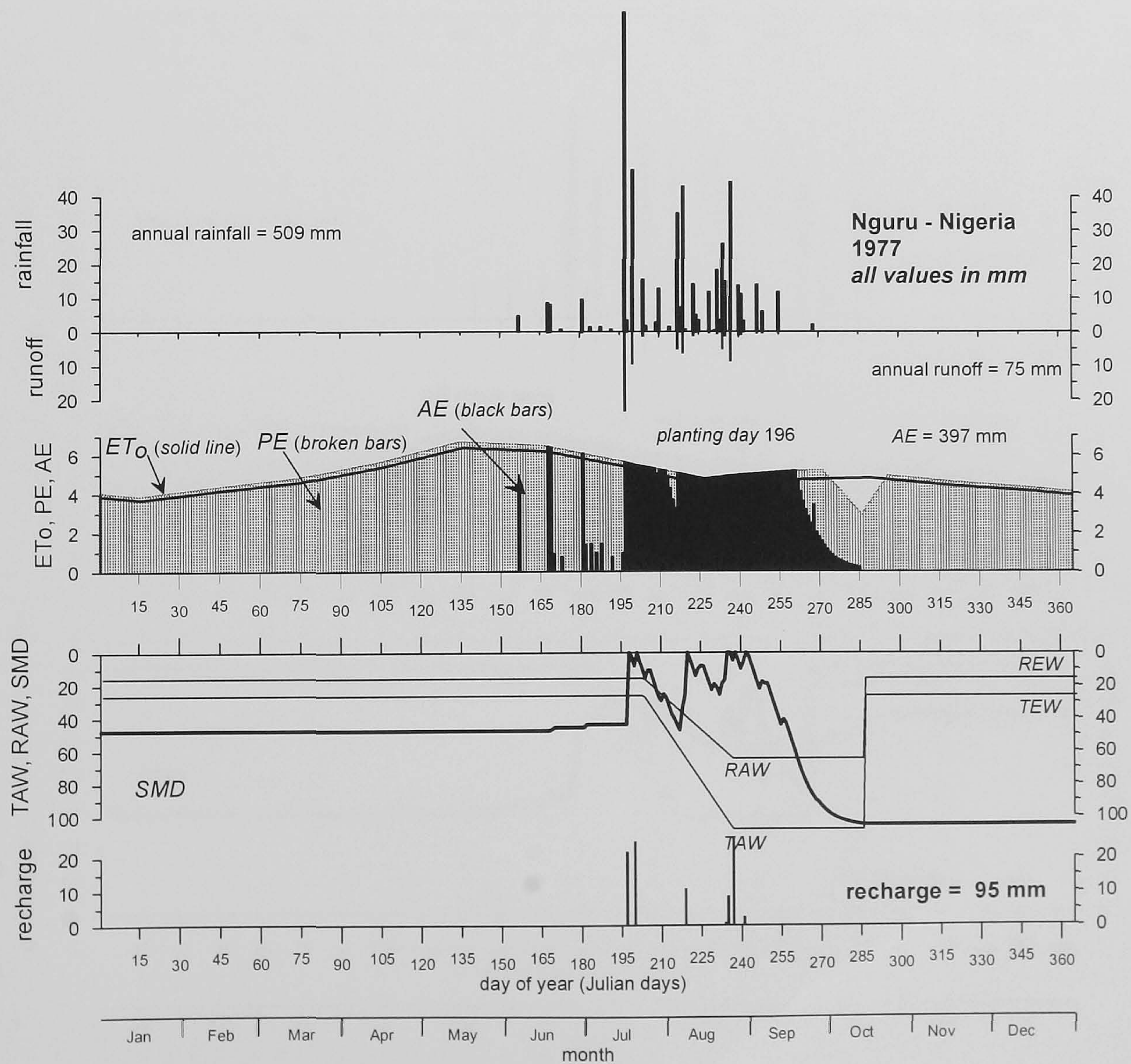


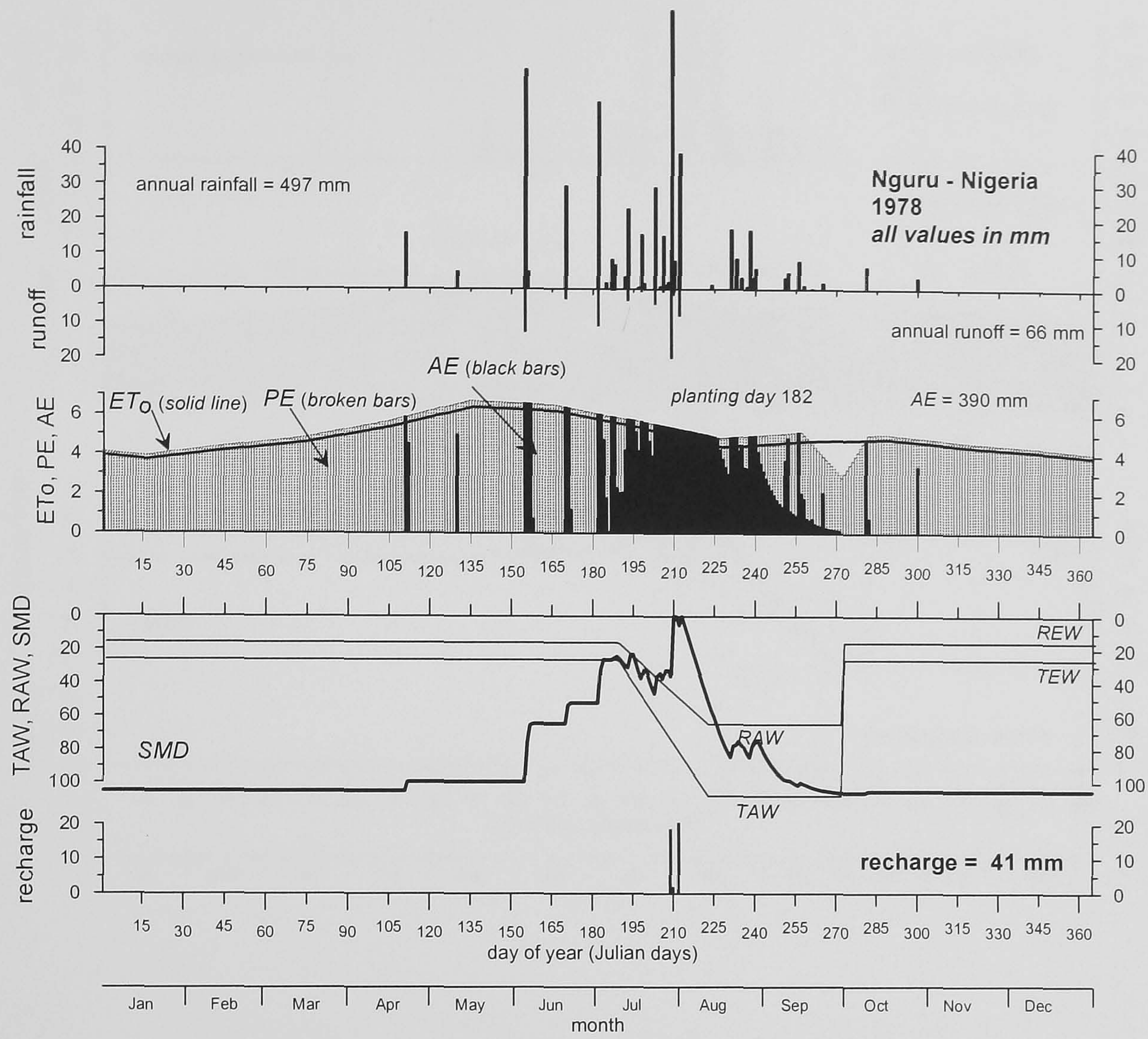
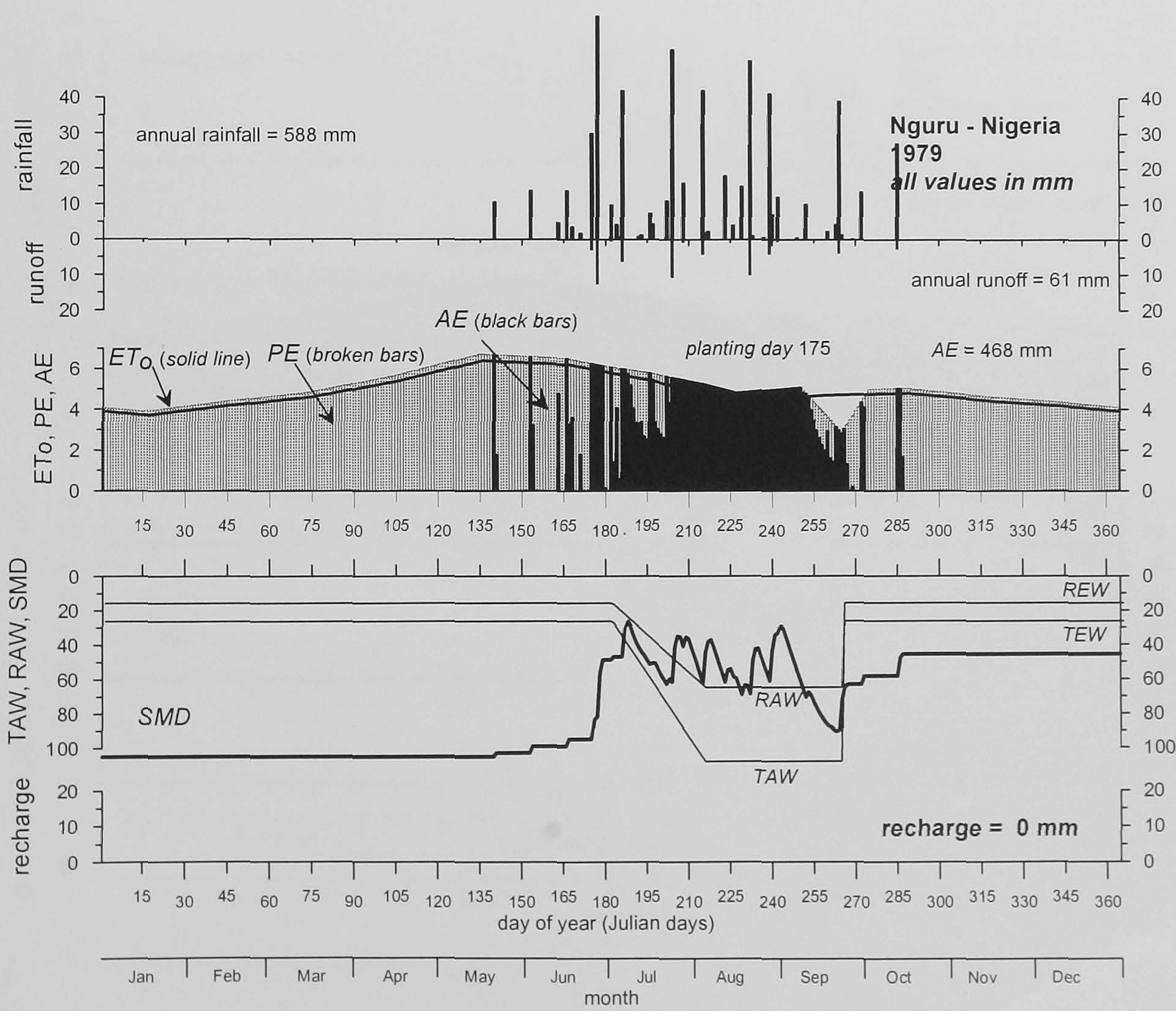


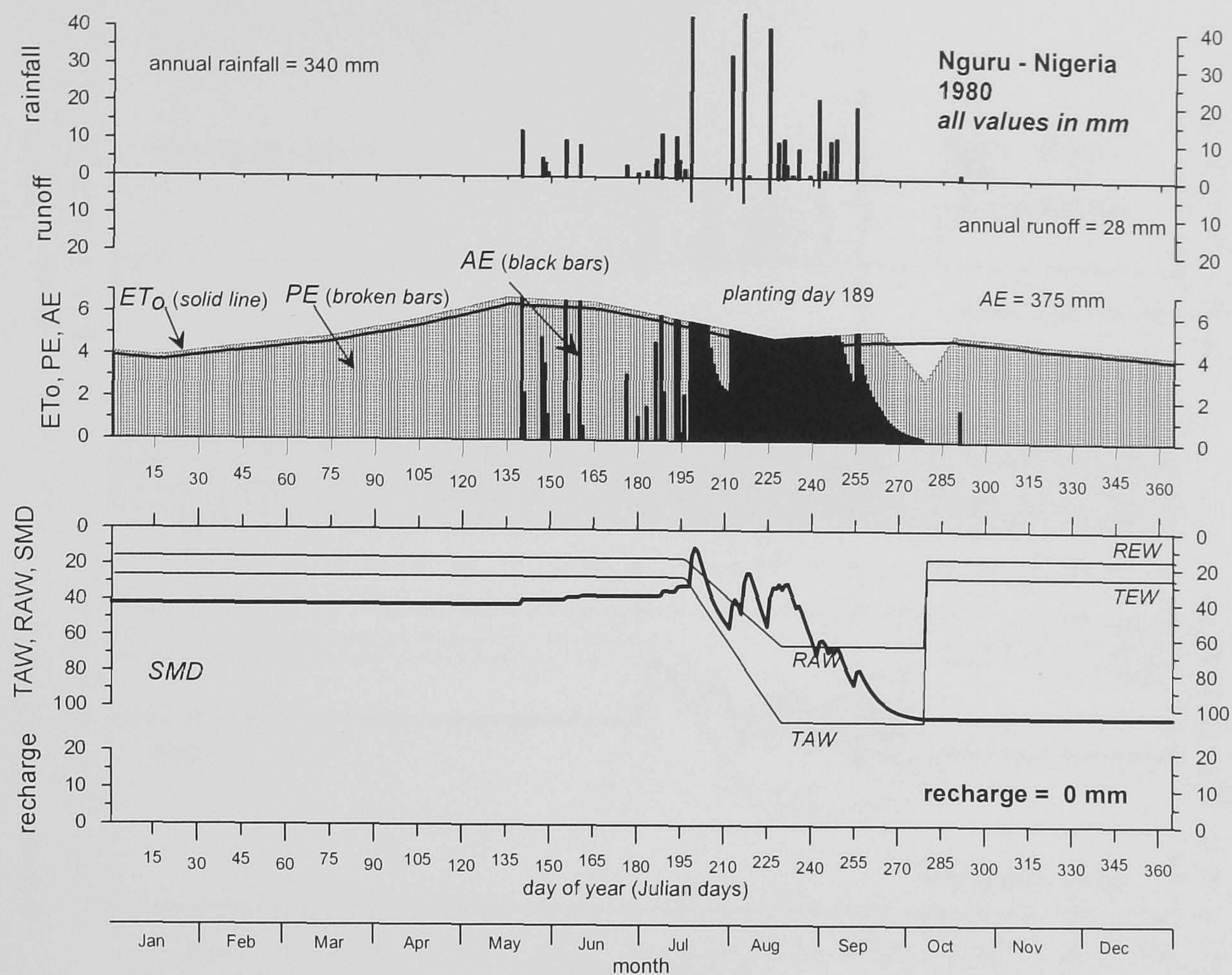
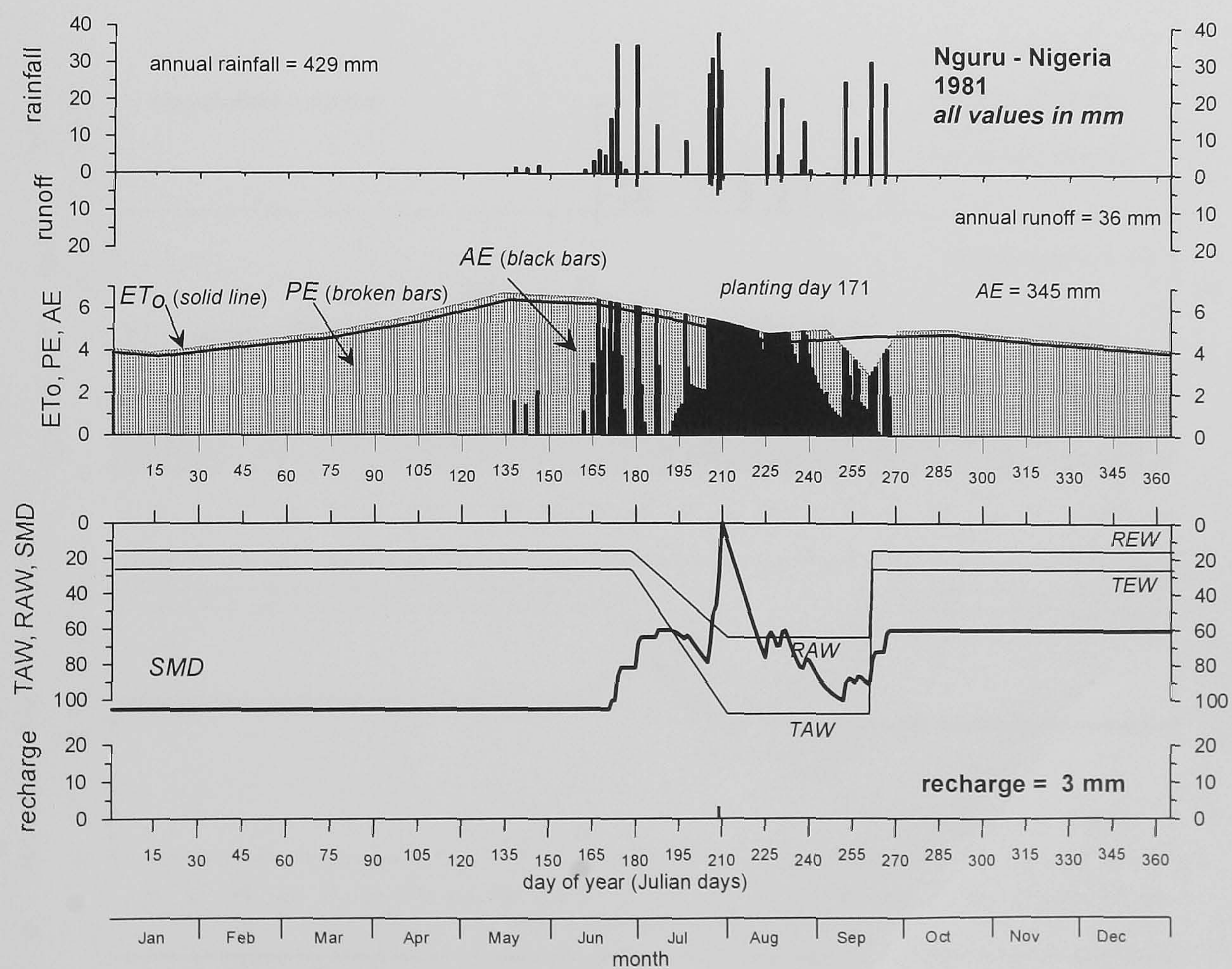


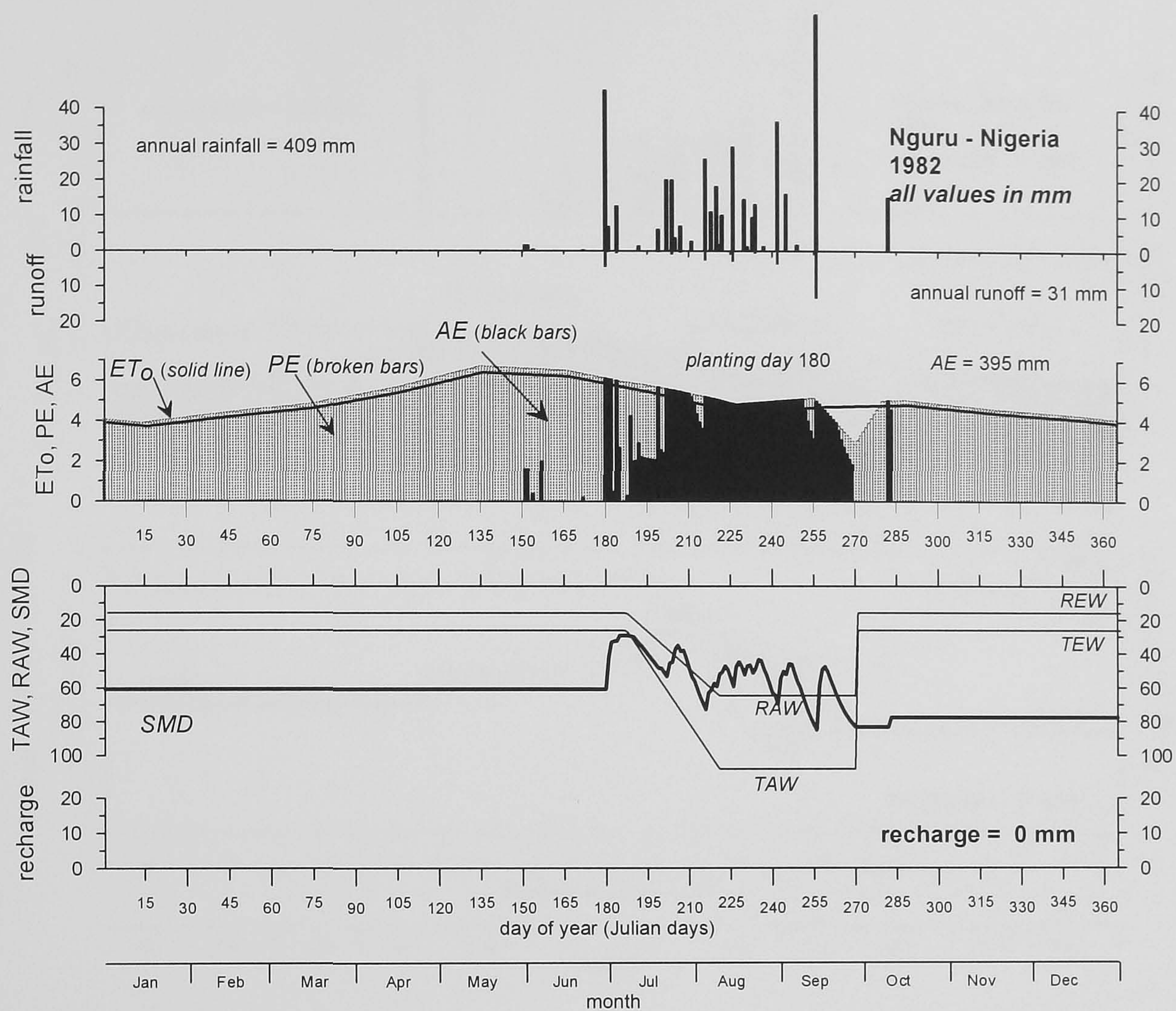
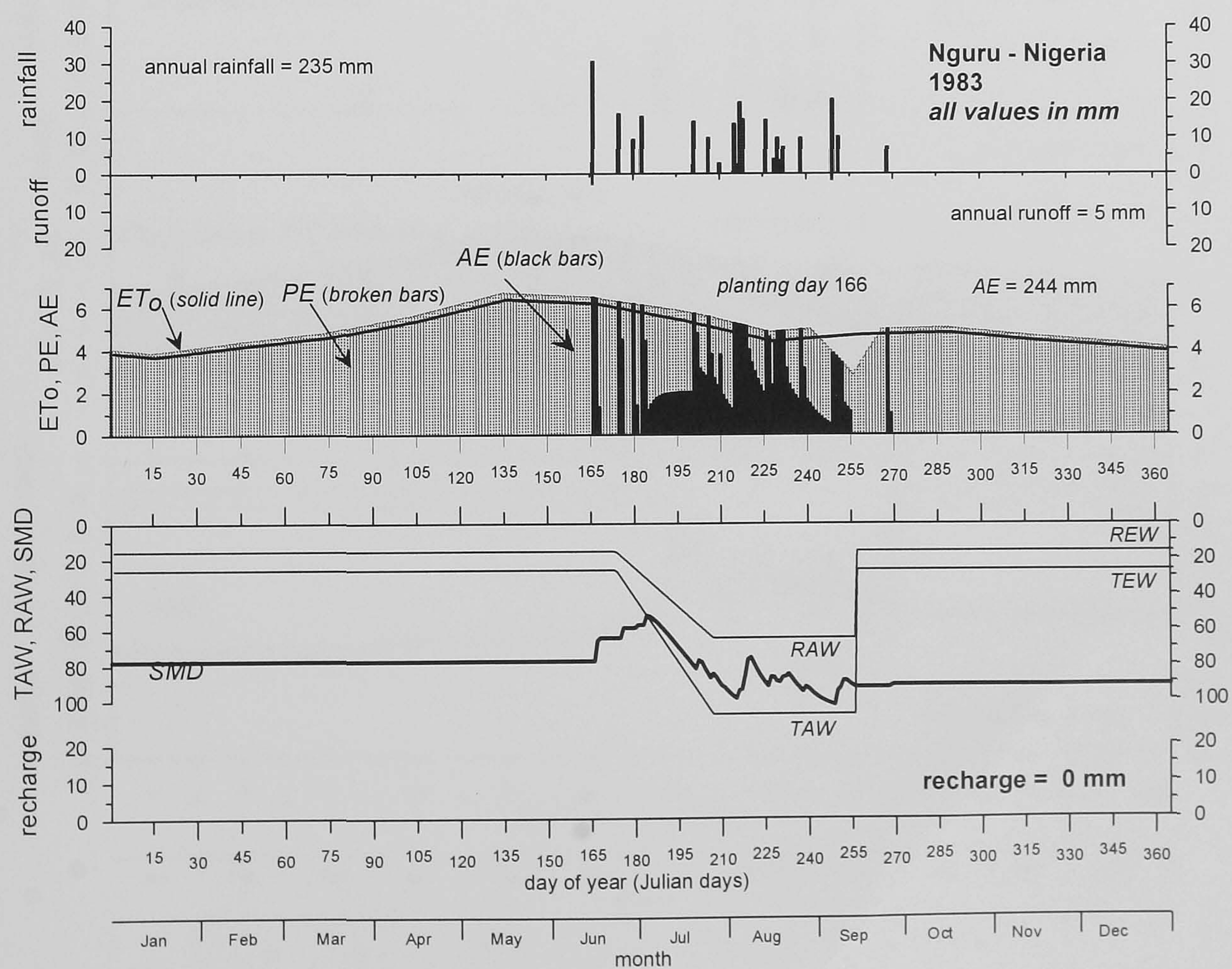


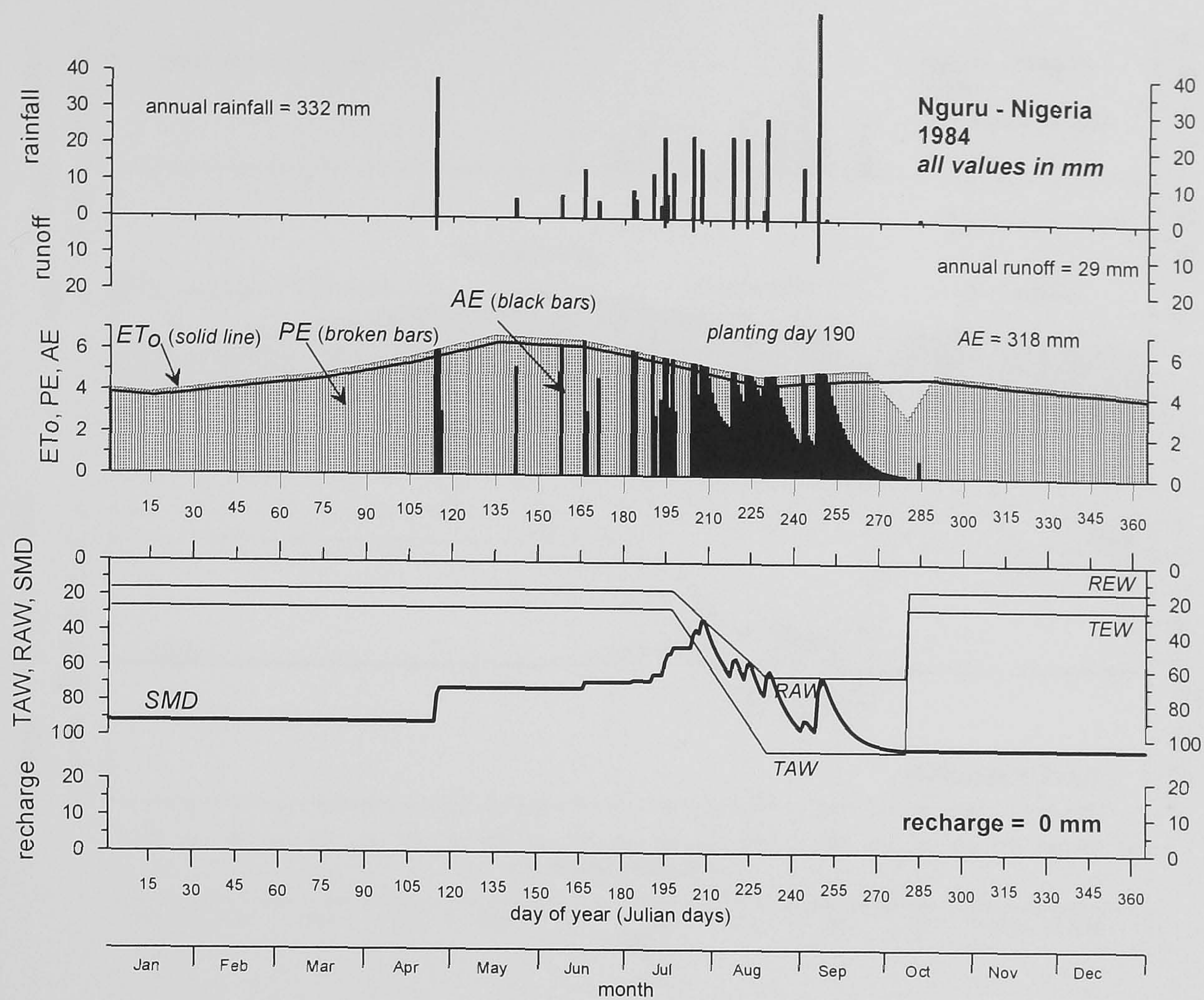
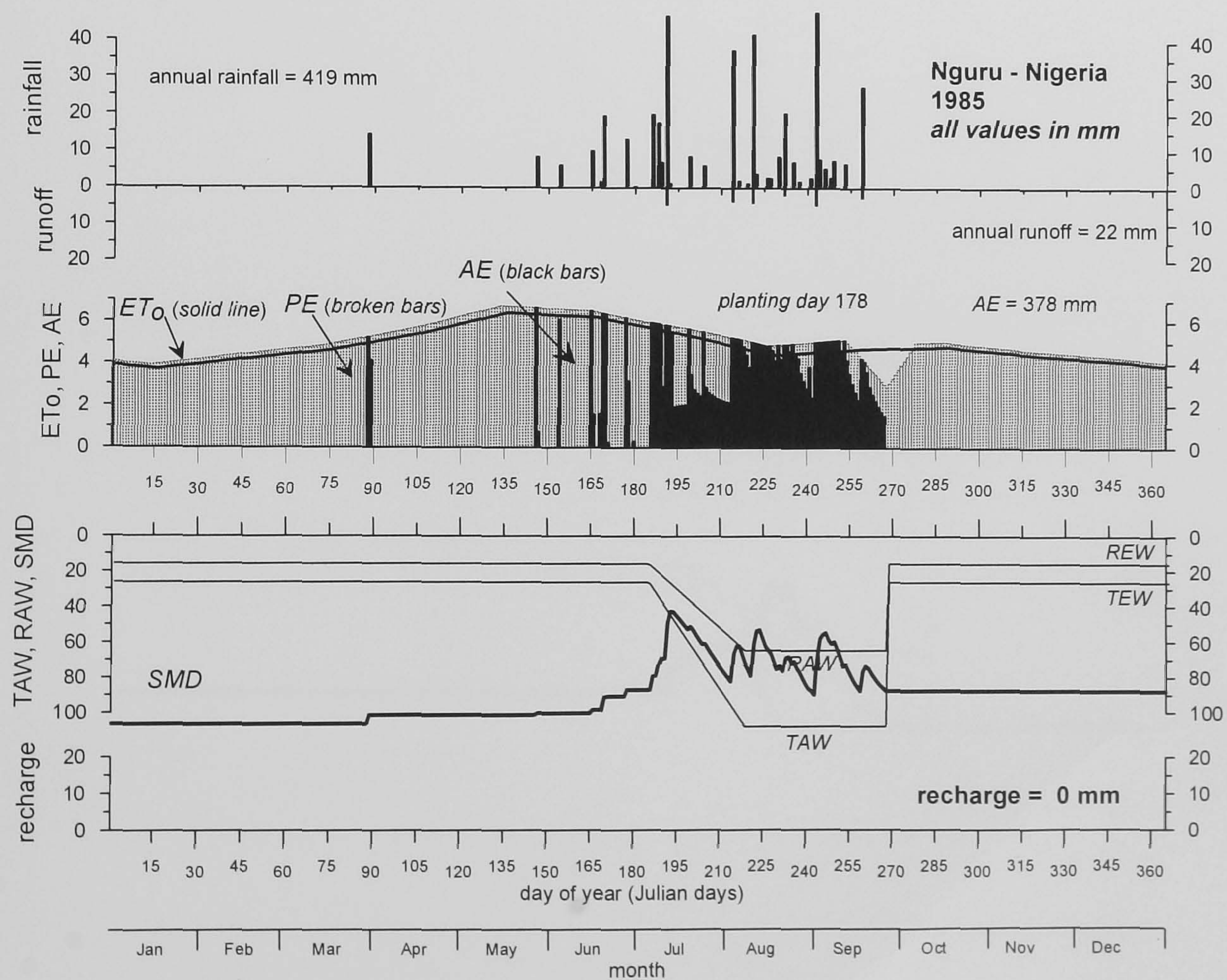


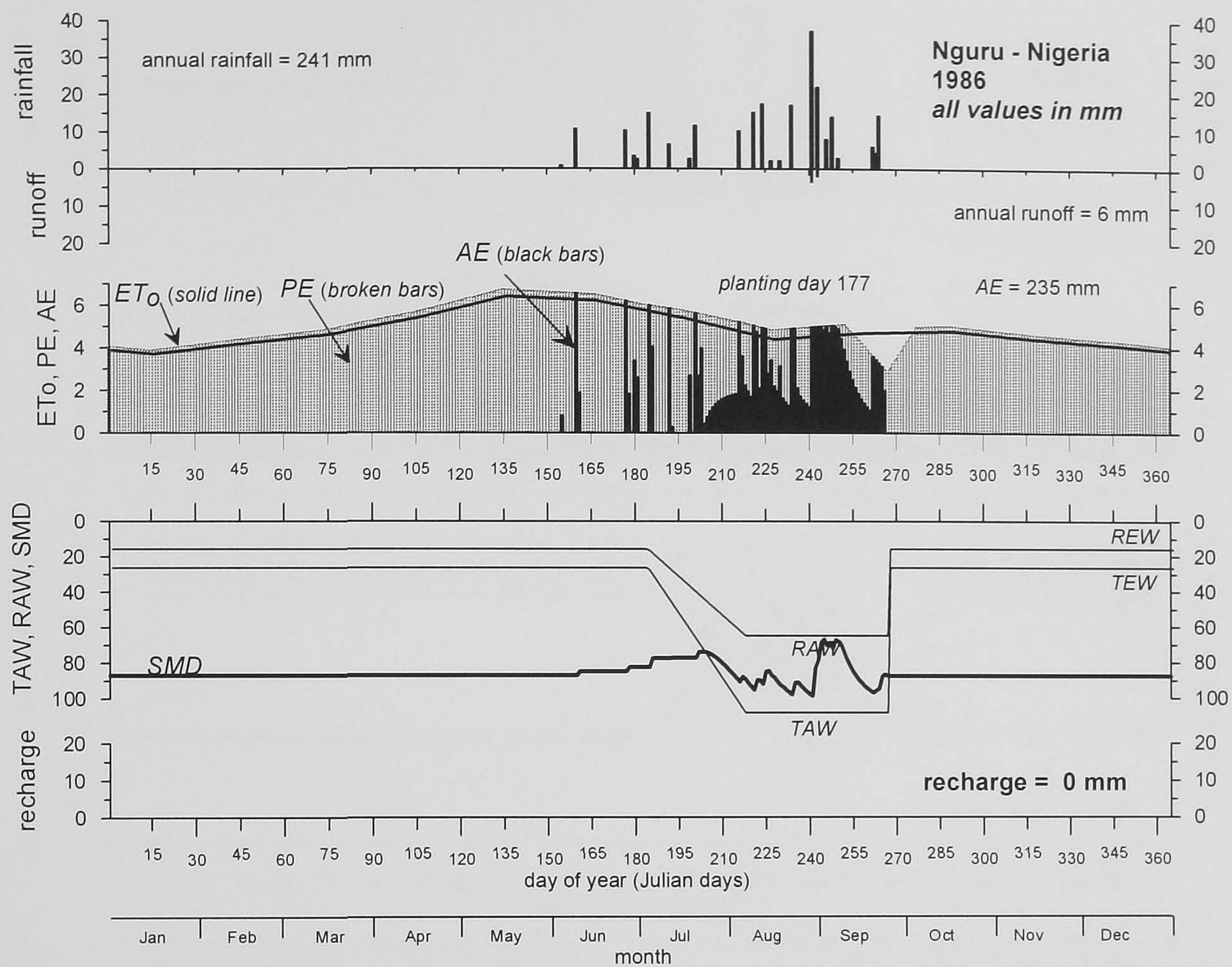
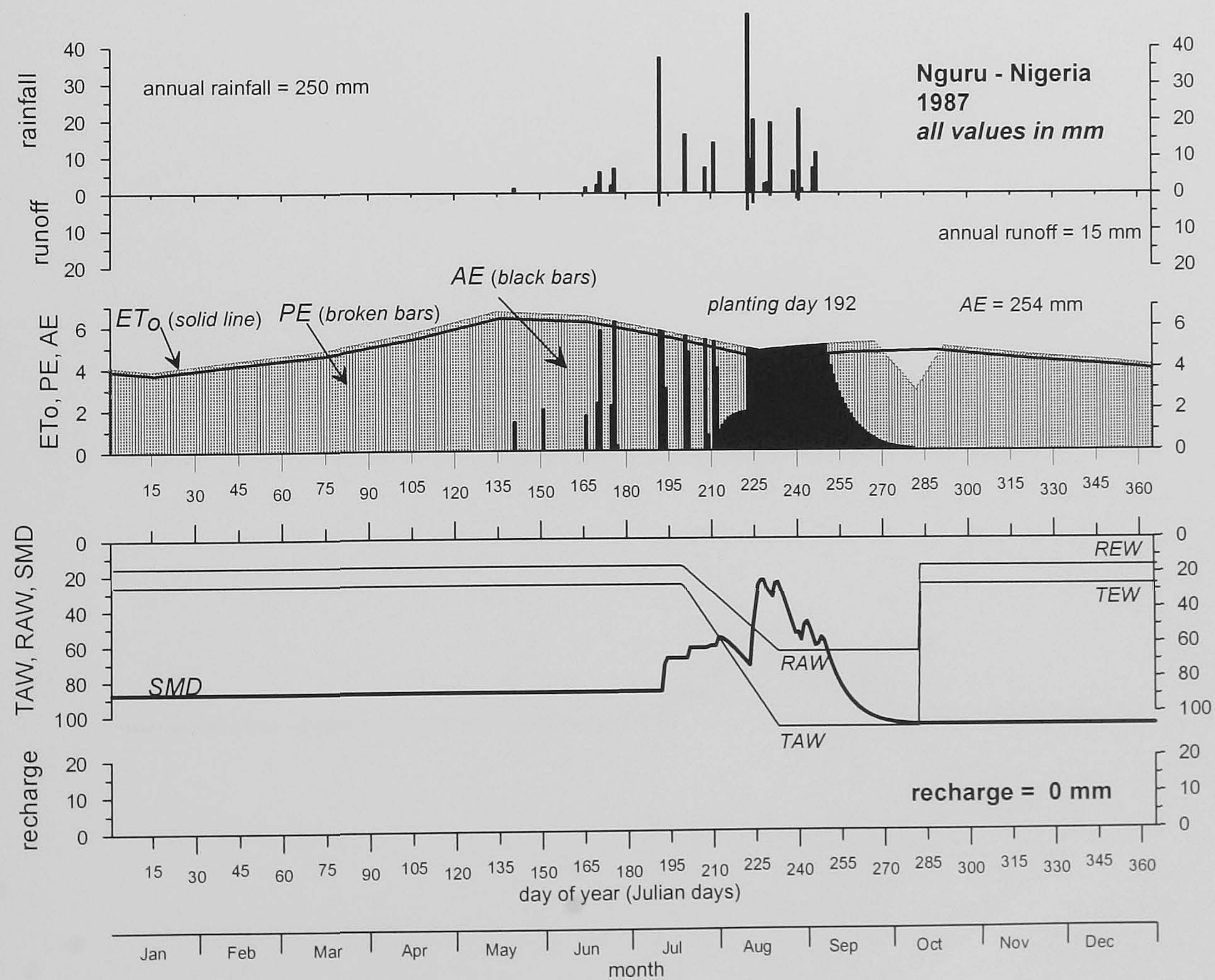


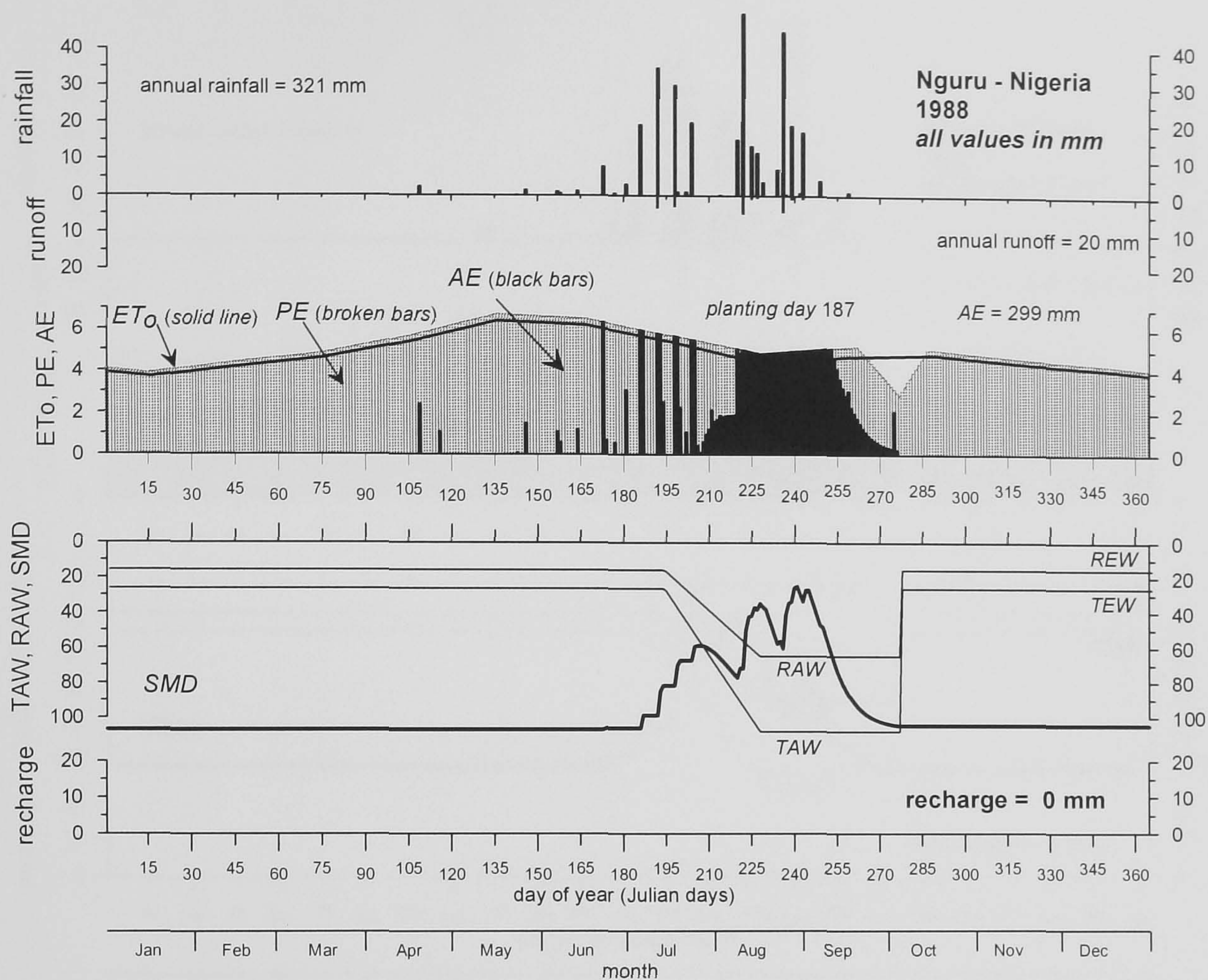
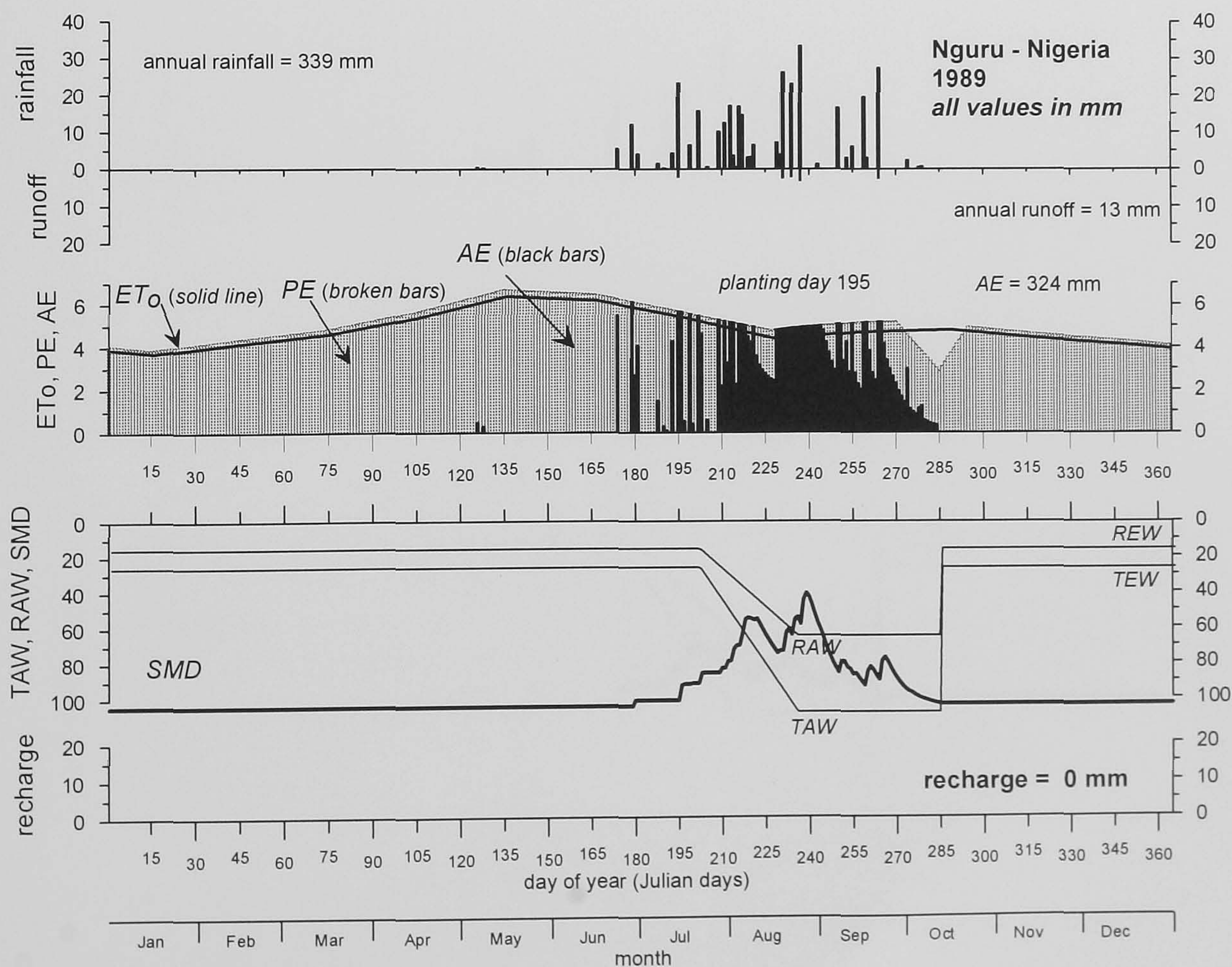


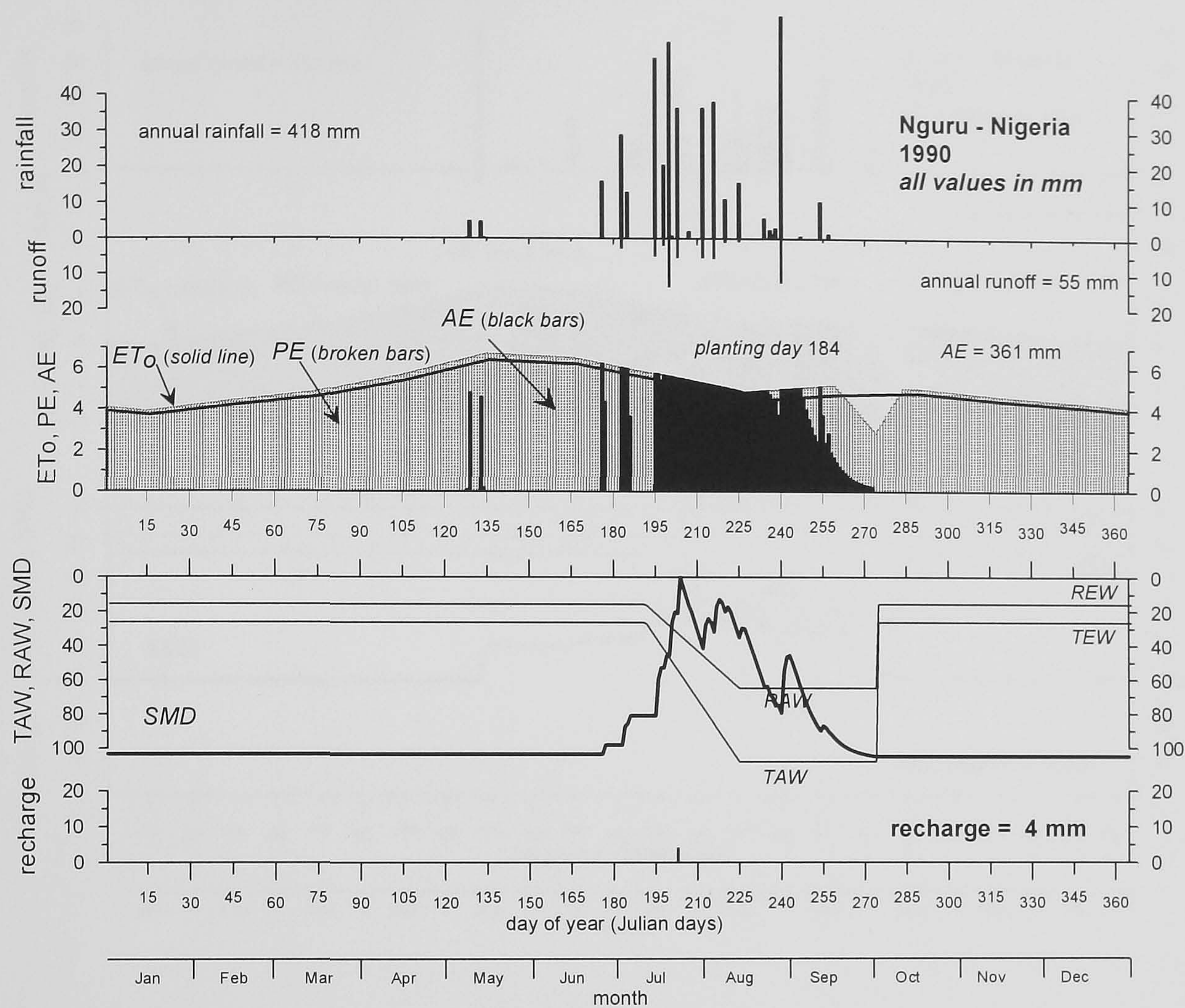
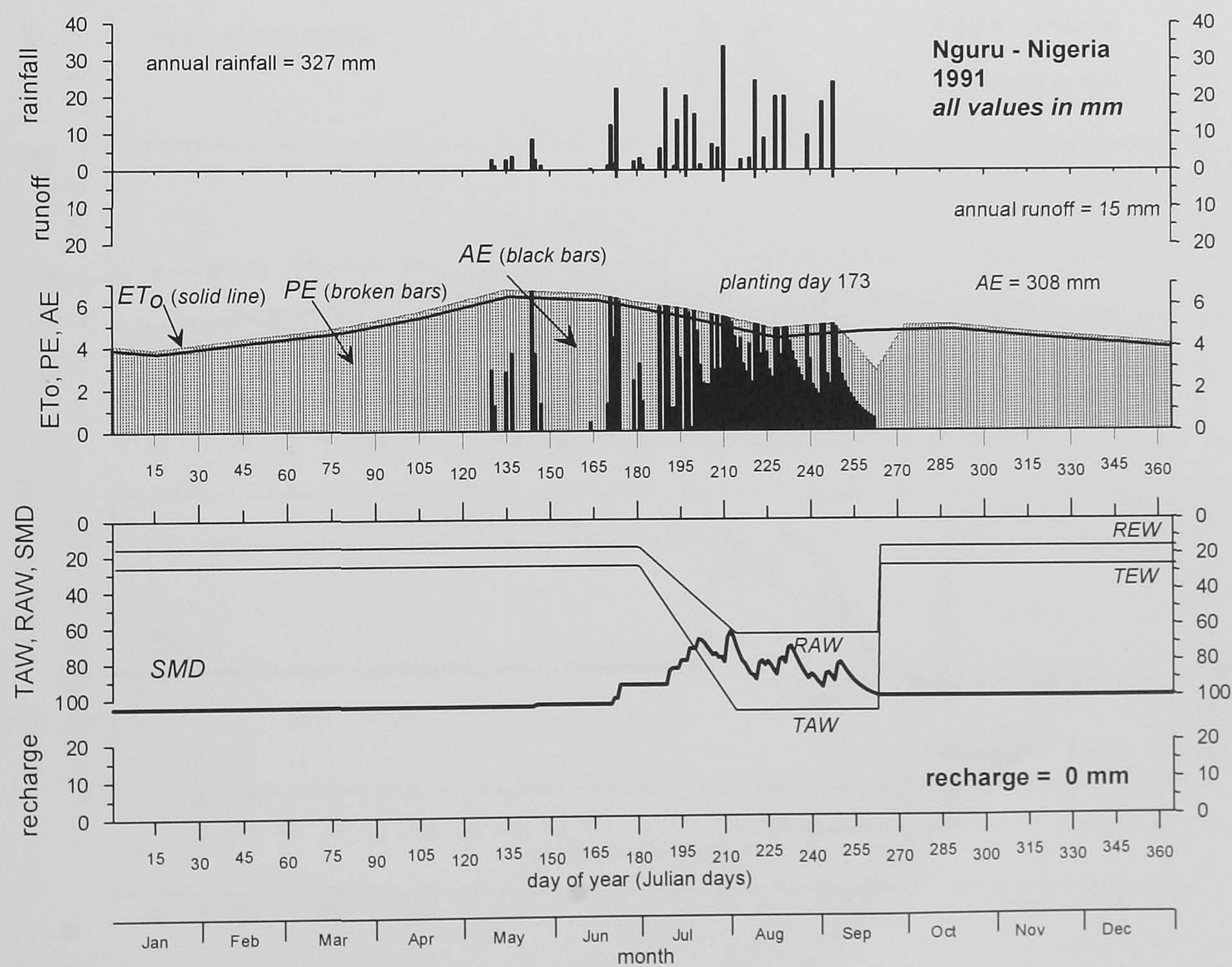


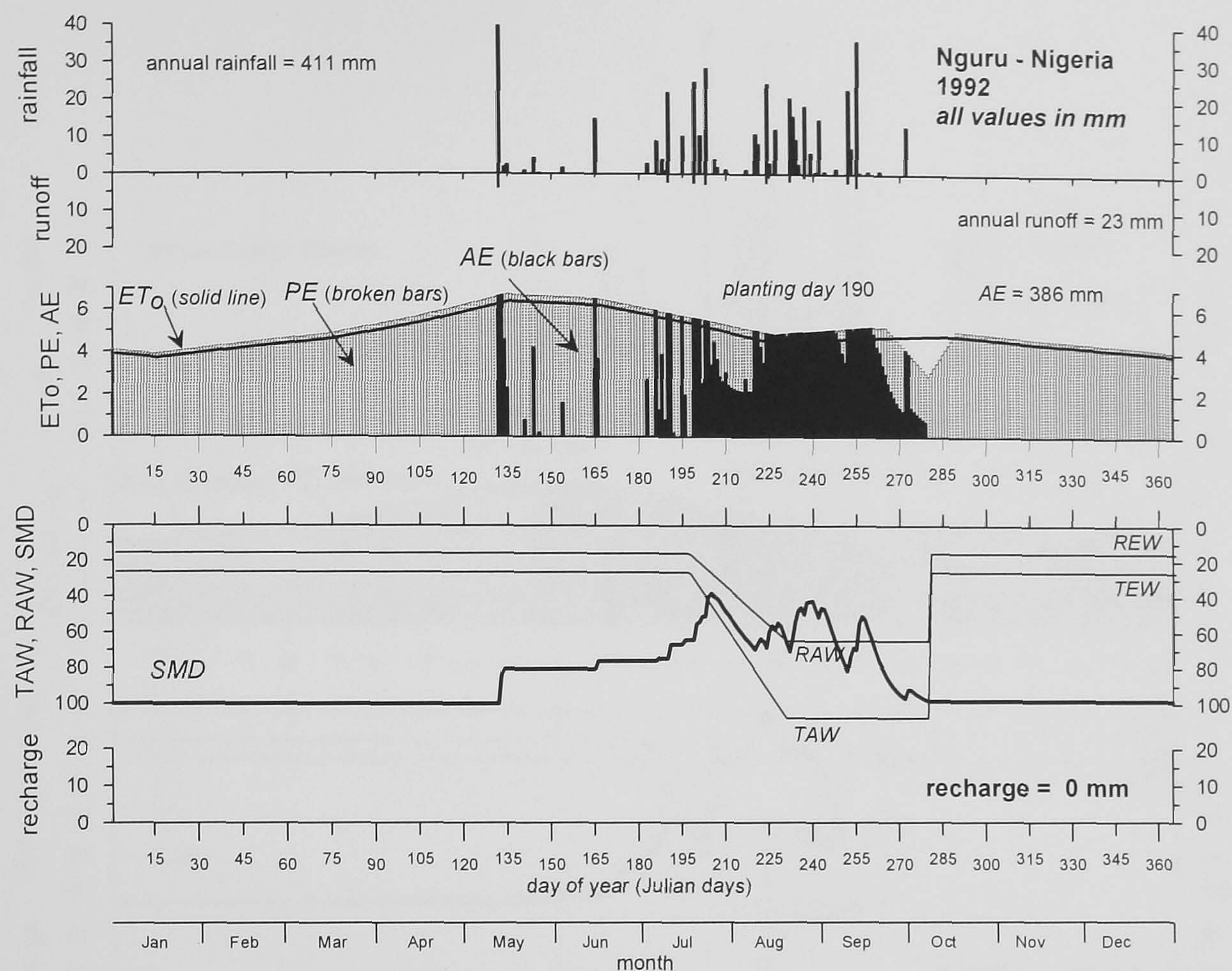
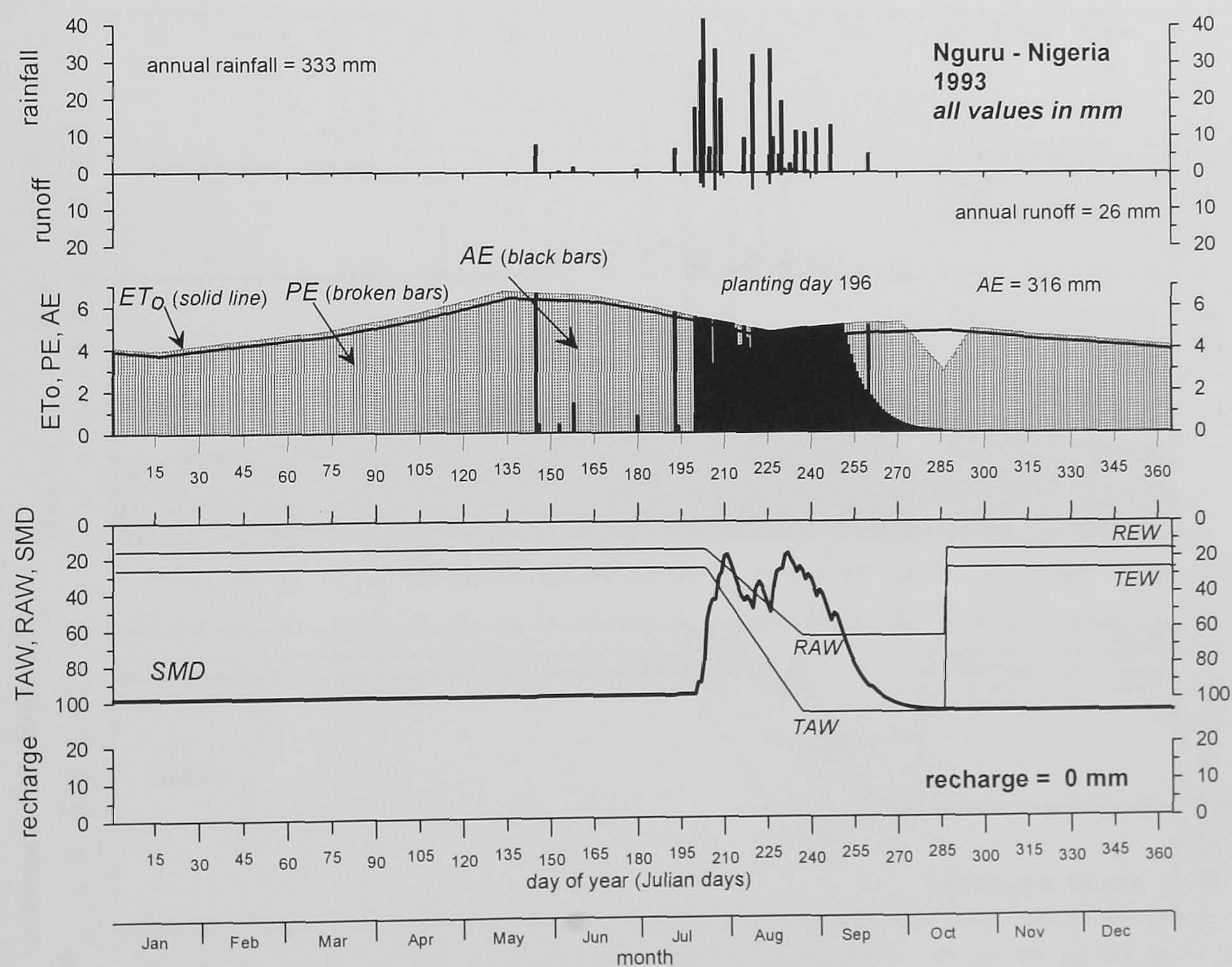


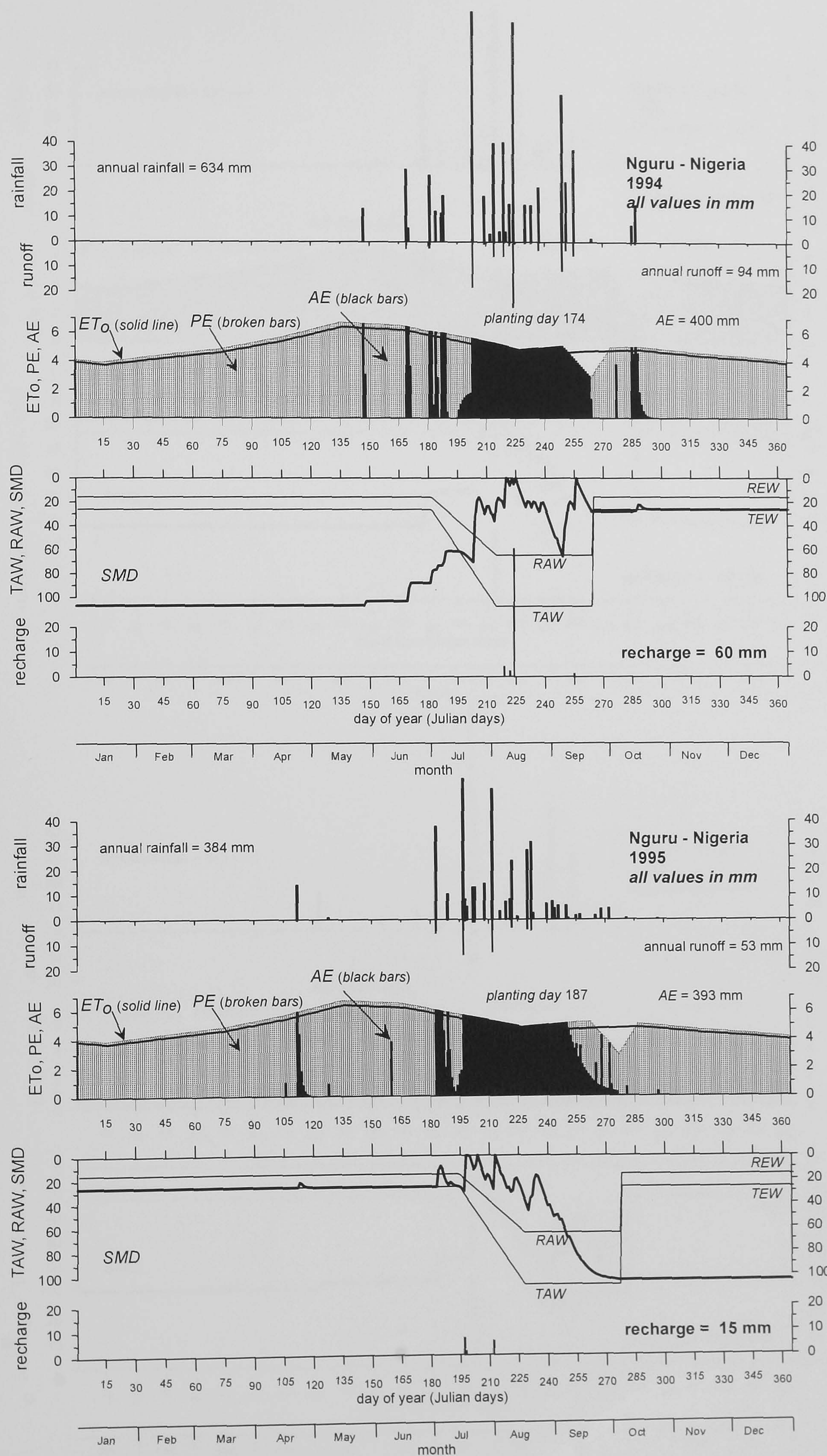


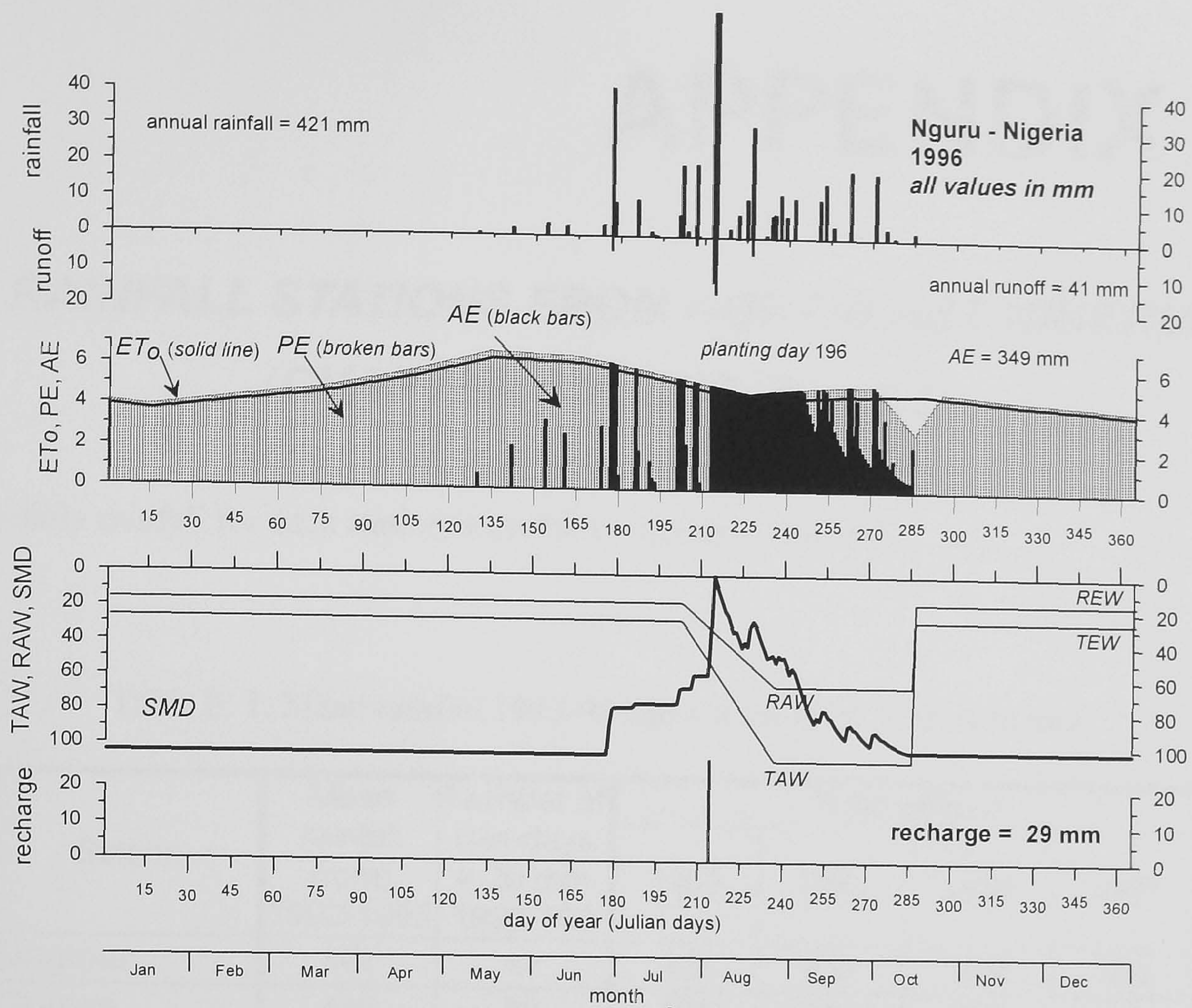
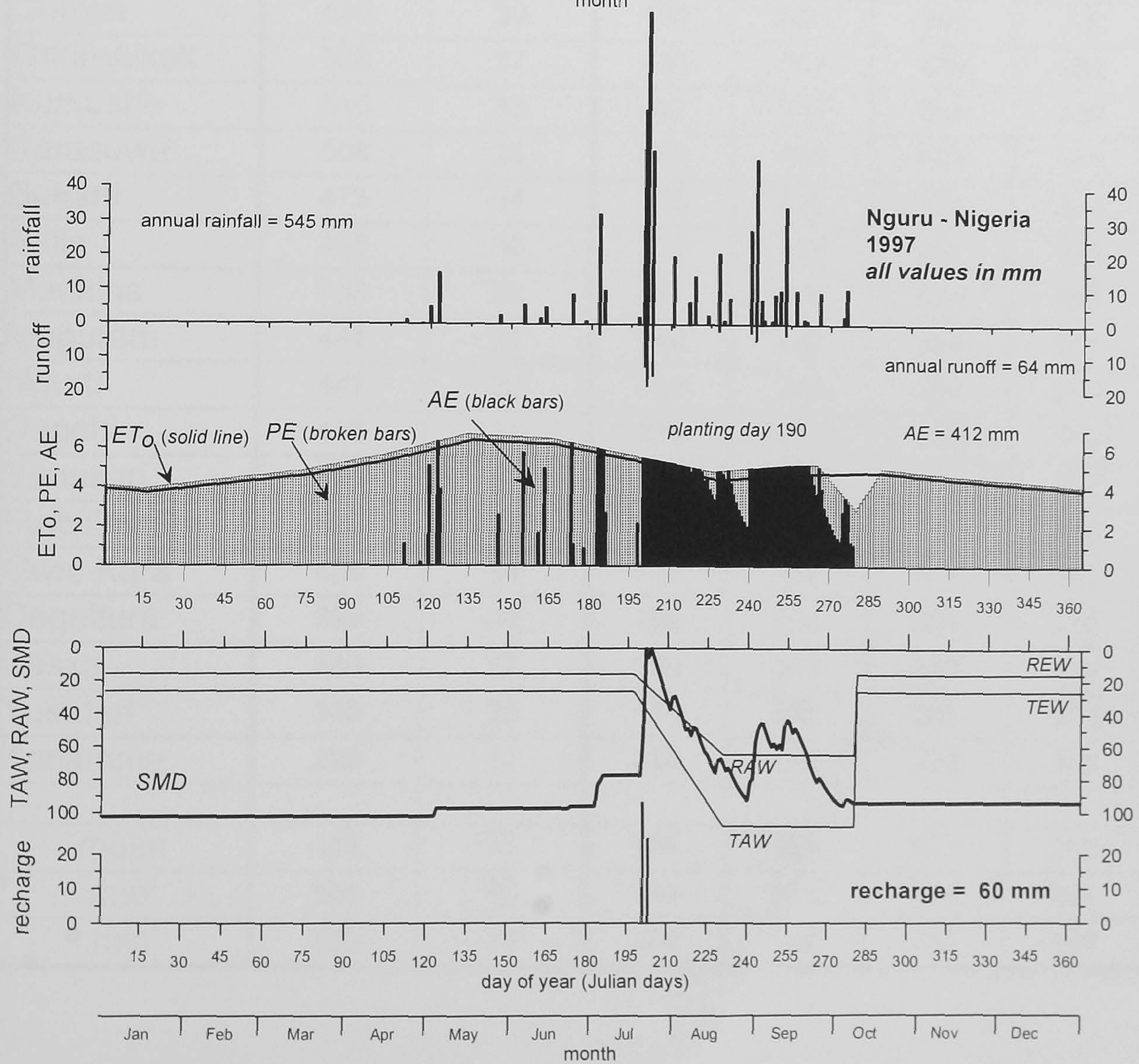












APPENDIX F

RAINFALL STATIONS FROM NORTHEAST NIGERIA (CHAPTER SIX, SECTION 6.5)

The daily rainfall for each station is on the floppy disk attached.

Table F. 1. Mean rainfall 1992-95 and annual total. Values in mm.

Station	Mean rainfall (mm) 1992-1995	Number of rain-days > 20 mm 1992-1995	Total annual			
			1992	1993	1994	1995
Dagona	561	42	416	436	1042	350
Gumsa	548	39	450	508	869	366
Garin Alkali	526	37	430	287	484	904
Kurkushe	516	35	482	410	941	230
Karasuwa	508	36	352	484	683	511
Bukarti	473	34	313	423	771	384
Balle	466	30	315	434	667	446
Machina	456	35	527	278	603	418
Muguram	444	31	544	409	624	201
Nguru	441	29	412	333	634	384
Dapchi	438	31	353	268	666	465
Yunusari	431	32	300	152	886	387
Futchimiram	429	36	362	416	730	208
Gwio Kura	420	33	472	313	657	237
Degeltura	399	30	493	282	603	216
Kaska	380	27	249	348	548	374
Yusufari	348	26	174	387	562	268
Kanamma	220	13	278	192	222	188
mean	445	32	384	353	677	363
max	561	42	544	508	1042	904
min	220	13	174	152	222	188

Figure F. 1. Annual rainfall at 18 stations (1992-1995).

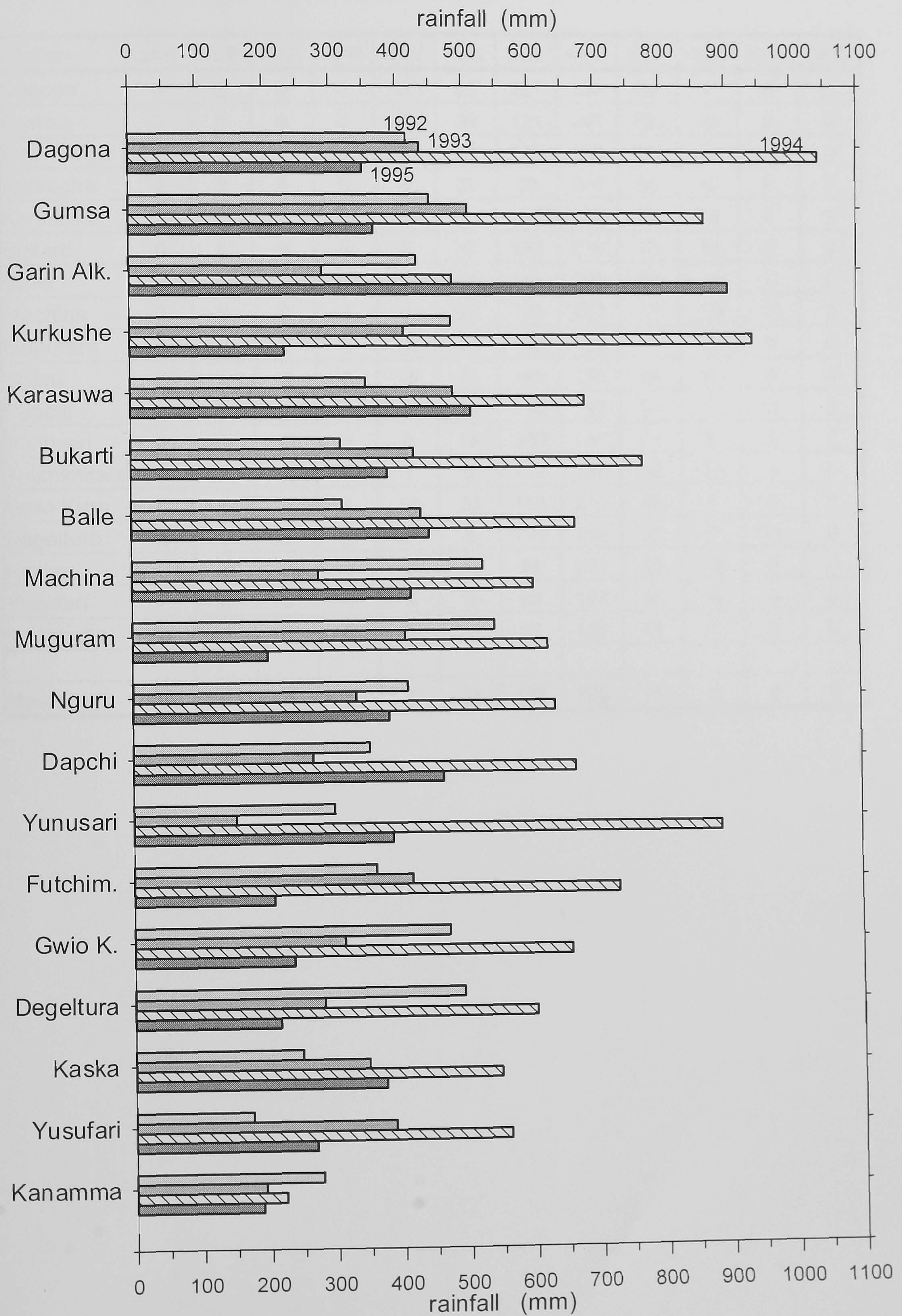


Table F. 2. Mean monthly rainfall 1992-95. Values in mm.

Station	JAN	FEB	MAR	APR	MAY	JUN	JUL	AUG	SEP	OCT	NOV	DEC
Dagona	0	0	0	0	6	90	221	184	52	7	0	0
Gumsa	0	0	0	0	17	36	121	247	109	19	0	0
Garin Alkali	0	0	0	0	6	53	175	200	86	6	0	0
Kurkushe	0	0	0	0	5	29	30	346	95	10	0	0
Karasuwa	0	0	0	1	4	25	161	240	64	13	0	0
Bukarti	0	0	0	0	5	37	133	216	70	12	0	0
Balle	0	0	0	0	9	41	151	193	69	3	0	0
Machina	0	0	0	1	1	23	138	217	67	10	0	0
Muguram	0	0	0	5	24	13	138	182	71	13	0	0
Nguru	0	0	0	4	18	21	161	165	64	8	0	0
Dapchi	0	0	0	0	5	21	116	197	91	8	0	0
Yunusari	0	0	0	2	0	15	190	145	71	9	0	0
Futchimiram	0	0	0	0	1	8	117	187	98	18	1	0
Gwio Kura	0	0	0	4	14	30	113	210	48	0	0	0
Degeltura	0	0	0	2	0	6	116	166	90	7	11	0
Kaska	0	0	0	3	2	19	98	201	45	13	0	0
Yusufari	0	0	0	0	3	3	93	197	44	9	0	0
Kanamma	0	0	0	0	6	11	51	111	41	1	0	0
Mean	0	0	0	1	7	27	129	200	71	9	1	0