Institut für Hydrologie

der Albert-Ludwigs-Universität Freiburg i.Br.

Matti Gerspacher

Rainfall runoff relationships

of the semiarid Kuiseb basin

Referent: Prof. Dr. Ch. Leibundgut
Koreferent: Dr. Ch. Külls

Diplomarbeit unter Leitung von Prof. Dr. Ch. Leibundgut
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Notation

Notation	unit	symbol
Baseflow coefficient	none	K
Baseflow Index	none	BFI
Baseflow recession constant	day-1	K
Capacity of surface store	mm	C
Coefficient of efficiency	none	Reff
Distance	km	d
Elevation	m	h
Global radiation	J/cm²	RG
Impervious threshold	mm	ImpT
Infiltration coefficient	day-1	COEFF
Infiltration loss	m³/s	L
Infiltration shape	none	SQ
Inflow rate	m³/s	I
Inverse distance factor	none	a
Observed runoff	m³/s	Qobs
Outflow rate	m³/s	Q
Partial area	%	A
Pervious fraction	%	Perv
Potential evapo-transpiration	mm/d	PET
Precipitation	mm	P
Rainfall interception store capacity	mm	INSC
Recharge coefficient	day-1	CRAK
Relative Humidity	%	rH
Saturated vapour pressure	hPa	E
Simulated runoff	m³/s	Qsim
Soil moisture storage capacity	mm	SMSC
Standard deviation	none	δ
Storage demand	m³	D
Storage depth	m	h
Storage volume	m³	S
Surface flow recession constant	day-1	KS
Temperature	°C	T
Time step	s	dt
Vapour pressure	hPa	E0

Extended Summary

Ephemeral rivers in arid regions show unpredictable, episodic occurrence of flow. Episodic, non-seasonal flood events are mainly generated by high intensity storm events which are highly variable in space and time. Spatial variability of rainfall inputs give rise to the problem that runoff in ephemeral rivers can not always be referred to an entire catchment.

Runoff generation in such arid environments differs substantially from humid regions. This is due to different meteorological inputs, different properties of mostly poorly developed soils, and to phenomena which are specific to arid regions. These phenomena include sparse, non-seasonal plant cover and transmission losses to the channel alluvium, mostly due to vertical percolation of flood water into channel beds. The dominant runoff generation process is usually surface runoff in terms of Hortonian overland flow, while at the same time some humid processes (e.g. baseflow) can be absent.

The combined effects of these features of arid zone hydrology cause a high degree of inconsistency in the relationships between rainfall and streamflow, which again reflects the general high variability and irregular occurrence of flows.

Moreover, runoff hydrographs in ephemeral rivers differ fundamentally from those in humid environments. Rapidly rising hydrographs, characterized by high peak flows, steep rising and falling limbs and no-flow conditions before and after an event usually characterize flood events in ephemeral streams.

All this leads to the need for a rainfall-runoff modelling approach that differs from the techniques developed for humid regions.

The subject of this study is rainfall-runoff modelling for the headwaters of the macro-scale

catchment of the ephemeral Kuiseb River in the semi-arid to hyper-arid environment of western Namibia.

For these purposes, an existing model for “whole of catchment modelling”, namely e2, has been applied.

The model offers various component models for rainfall-runoff modelling of which the SIMHYD and the AWBM model were selected for further application.

The SIMHYD model is a daily conceptual mass balance model that estimates daily flow from daily rainfall and evapo-transpiration data. The basic concept of SIMHYD subdivides the basin of interest into pervious and impervious areas and thereby identifying areas which directly contribute to surface runoff.

The AWBM model is a catchment water balance model relating runoff to rainfall with daily or even hourly data, calculating losses from rainfall for flood hydrograph modelling. AWBM subdivides the catchment into three areas with each area representing a surface store to mimic partial runoff areas.

Both rainfall-runoff models are calibrated against observed flow data over a period of more than eight years. Calibration examines both the entire period, and discrete runoff events with a general focus on monthly totals. Subsequently, validation for another eight years is conducted, again with a focus on monthly totals rather than on reproduction of discrete runoff events.

As data scarcity is identified as the main problem of rainfall-runoff modelling in arid regions, most of the models that have so far been applied in the southern African region are based on monthly time intervals.

The challenge of applying a daily time step model like e2, therefore, firstly requires adequate quantification of the main water inputs on a daily basis. For this purpose rainfall for the Kuiseb basin is regionalized by means of a rainfall decay function and the inverse distance squared method. At first a dot matrix is superimposed on the catchment and daily rainfall is calculated for every point within the matrix based on the long term rainfall records of four stations within the basin. e2 input requirement of area precipitation for each sub- catchment is achieved by averaging daily rainfalls of all points within one sub-catchment.

Regionalization of potential evapo-transpiration (PET) is conducted by creation of daily time series for Gobabeb in the west of the catchment, for Windhoek in the east of the catchment, and a representative fictive station in the centre of the basin.

Firstly, PET was calculated for Windhoek according to Turc. Then, ET data sets were generated for Gobabeb and Windhoek by fitting a sine function to the annual PET-regimes and then extrapolating these data sets in time. To allow for annual variation within the extrapolated data sets, a scattering around the values of the sine function was achieved by letting standard deviations vary in a normally distributed manner. The data set of the fictive station is obtained by interpolation between Gobabeb and Windhoek with regard to the particular elevation above sea level of each station. Area potential evapo-transpiration is gained by allocating representative areas to each of the three stations.

The results of regionalization, for both rainfall and PET, indicate that the spatial distribution of both parameters is displayed satisfactorily.

The results of model calibration show that both rainfall-runoff models underestimate both total flow volumes, and peak flows, for the entire period of calibration. In marked contrast to this, both models yield excessive flows on some isolated events within the calibration period.

Validation results prove to be poor for both models with the AWBM model performing worse than the SIMHYD model.

It is concluded that moderate model performance, to this extent, can only be due to inadequate inputs. Detailed examination showed that overestimation of modelled flow, and generation of modelled flow in no-flow periods follows from excessive regionalized area precipitation for the sub-catchments delineated in the e2 model. This, in turn, has its cause in high readings at the recording rainfall gauges. Small-scale precipitation cells hit the rainfall gauges, but without rainfall over the whole catchment and, subsequently, these gauged rainfalls are wrongly extrapolated to the entire catchment area. Conversely, underestimation of modelled flow, and the absence of modelled flow in periods where flow records exist, stem from non-existent rainfall records on occasions where the watershed as a whole obviously received sufficient rainfall to generate runoff.

Hence, it cannot be assumed that rainfall recorded at the gauging stations also correlates to the entire catchment. Moreover, the assumption that all rainfall that occurs within the catchment is also recorded to some extent at the rainfall gauges also seems to be wrong.

These findings raise doubts about whether rainfall regionalization, and therefore rainfall-runoff modelling based on such regionalization, is a sound basis for the assessment of a model's performance and the examination of rainfall runoff relationships within such arid, variable environment.

Keywords:
Arid, Rainfall-Runoff, Modelling, e2 Model, Regionalization, Kuiseb River

Zusammenfassung –Deutsch

Ephemere Flüsse in ariden Gebieten weisen häufig ein episodisches, nicht vorhersagbares Auftreten von Abflüssen auf. Diese episodischen Abflussereignisse, die keiner Saisonalität unterliegen, werden vor allem durch zeitlich und räumlich hochvariable Gewitterereignisse hervorgerufen, welche ihrerseits hohe Niederschlagsintensitäten aufweisen.

Aus der räumlichen Variabilität der Niederschläge entsteht das Problem, dass der Abfluss in ephemeren Flüssen nicht immer einem ganzen Einzugsgebiet zugeordnet werden kann.

Die Abflussbildung in derart ariden Gefilden unterscheidet sich wesentlich von jener in humiden Gebieten. Dies liegt an unterschiedlichen meteorologischen Eingangsgrößen, unterschiedlichen pedologischen Eigenschaften der meist karg entwickelten Böden und an Phänomen, die für aride Gebiete spezifisch sind. Diese Phänomene beinhalten eine spärliche, nicht saisonale Vegetationsdecke oder auch das Vorkommen so genannter Transmission Losses in das Gerinnealluvium welche meist infolge vertikaler Perkolation von Flusswasser ins Gerinnebett auftreten. Der dominante Abflussprozess in ariden Gebieten ist meist Oberflächenabfluss in Form von Horton´schem Oberflächenabfluss. Gleichzeitig jedoch kann auch das Fehlen manch typisch humiden Prozesses, wie zum Beispiel Basisabfluss, charakteristisch sein.

Die gebündelten Auswirkungen dieser hydrologischen Eigenschaften von Trockengebieten führen dazu, dass die Niederschlags-Abfluss-Beziehungen in diesen Regionen sehr unbeständig sind. Diese Unbeständigkeit spiegelt sich in der hohen Variabilität und dem unregelmäßigen Auftreten der Abflüsse wider.

Darüber hinaus unterscheiden sich Abflussganglinien ephemerer Flüsse grundlegend von denen in humiden Umgebungen. Rasch ansteigende Ganglinien mit hohen Spitzenabflüssen, gekennzeichnet durch steile ansteigende und abfallende Äste sowie durch keinen Abfluss vor und nach einem Ereignis, sind charakteristische Merkmale von Abflussereignissen in ephemeren Flüssen.

All dies erfordert somit einen Ansatz zur Niederschlags-Abfluss Modellierung, der sich von den Methoden, die speziell für humide Regionen entwickelt wurden, unterscheidet.

Gegenstand dieser Studie ist eine Niederschlags-Abfluss-Modellierung für den oberen Teil des makroskaligen Einzugsgebiets des Kuiseb, einem ephemeren Gerinne in der hyper- bis semi-ariden Umgebung des westlichen Namibias.

Zu diesem Zwecke wurde ein bestehendes Modell, e2, das der ganzheitlichen Simulation von Einzugsgebieten dient, angewendet. Das Modell beinhaltet verschiedene Komponentenmodelle zur Niederschlags-Abfluss-Modellierung. Zur weiteren Anwendung wurden aus dieser Palette das SIMHYD und das AWBM Modell ausgewählt.

SIMHYD ist ein tagesbasiertes konzeptionelles Modell, das auf dem Gesetz der Massenerhaltung beruht. Tägliche Abflüsse werden aus Zeitreihen täglicher Niederschläge und Evapotranspiration ermittelt. Das grundlegende Konzept von SIMHYD besteht darin, das betrachtete Einzugsgebiet in permeable und impermeable Flächen zu unterteilen, und somit eine Ausweisung jener Flächen zu ermöglichen, die unmittelbar zum Oberflächenabfluss beitragen.

AWBM ist ein Einzugsgebietsmodell, welches eine Wasserbilanz beinhaltet und Abflüsse auf der Basis täglicher oder gar stündlicher Daten aus Niederschlägen errechnet. Dabei werden Verlusttherme von Niederschlägen zur Modellierung von Abflussganglinien berechnet. Das AWBM Modell unterteilt das Einzugsgebiet in drei Flächen, wobei jeder dieser Flächen ein Oberflächenspeicher zugewiesen wird. Dies wiederum dient der Ausweisung von Flächen mit unterschiedlichem Abflussbildungsverhalten.

Die Kalibrierung beider Modelle erfolgt über eine Zeitreihe gemessener Abflüsse über einen Zeitraum von über acht Jahren. Der Fokus der Kalibrierung liegt sowohl auf der Betrachtung der gesamten Zeitreihe als auch auf der Betrachtung einzelner Ereignisse. Dabei liegt der Schwerpunkt stets mehr auf der Betrachtung von Monatssummen, als darauf, einzelne Ereignisse nachzubilden.

Da der Mangel an Daten bereits als das Hauptproblem arider Niederschlags-Abfluss-Modellierung erkannt wurde, haben die meisten Modelle die bisher im südlichen Afrika angewandt wurden, auf Monatsbasis gearbeitet.

Die Herausforderung einer tagesbasierten Modellierung wie mit e2 besteht also zunächst darin, die Eingangsgröße Niederschlag sinnvoll tagesbasiert zu quantifizieren. Zu diesem Zweck wird eine Niederschlagsregionalisierung für das Kuiseb Einzugsgebiet durchgeführt, die mit einer Niederschlags-Abnahme-Funktion und mit inversen Distanzen arbeitet. Zunächst wird dem Einzugsgebiet ein

Punktraster überlagert und im Folgenden für jeden Punkt dieses Rasters der Tagesniederschlag, basierend auf der langfristigen Niederschlagsmessung von vier Stationen im Einzugsgebiet, errechnet. Der Anforderung des e2 Modells nach Gebietsniederschlägen für jedes Teileinzugsgebiet wird Genüge getan, indem das Mittel der Tagesniederschläge aller Punkte im entsprechenden Teileinzugsgebiet gebildet wird.

Die Regionalisierung der potentiellen Evapotranspiration wird durchgeführt, indem, jeweils für Gobabeb im Westen, für Windhoek im Osten und für eine fiktive Station im zentralen Teil des Einzugsgebiets eine tagesbasierte Zeitreihe generiert wird.

Zuerst wurde die potentielle Evapotranspiration für Windhoek nach dem Ansatz von Turc errechnet. Anschließend wurden Datensätze der Evapotranspiration für Gobabeb und Windhoek erstellt, indem eine Sinusfunktion an die Jahresgänge der potentiellen Evapotranspiration angepasst wurde. Diese Sinusfunktion wird daraufhin über die Zeit extrapoliert. Um eine jährliche Variabilität der extrapolierten Daten zu simulieren, wird ein Rauschen um die Werte der Sinusfunktion erzeugt, indem die Standardabweichungen der Sinusfunktion normalverteilt variieren. Der Datensatz der fiktiven Station wird erzeugt, indem eine höhenabhängige Interpolation zwischen den Daten von Gobabeb und Windhoek durchgeführt wird.

Flächenhafte Verdunstungsdaten werden dadurch erzeugt, dass jeder der drei Stationen eine repräsentative Fläche des Einzugsgebiets zugeteilt wird.

Die Ergebnisse der Regionalisierung, sowohl für den Niederschlag als auch für die potentielle Evapotranspiration, werden hinsichtlich ihrer räumlichen Verteilung als zufrieden stellend erachtet.

Die Kalibrierungsergebnisse zeigen, dass beide Niederschlags-Abfluss-Modelle die Abflussvolumina wie auch die Spitzenabflüsse für den gesamten Kalibrierungszeitraum unterschätzen. Im Gegensatz dazu steht die Erkenntnis, dass beide Modelle in demselben Zeitraum bei vereinzelten Ereignissen deutlich überhöhte Abflüsse produzieren.

Die Ergebnisse der Validierung sind für beide Modelle dürftig, wobei das AWBM Modell im Vergleich schlechter abschneidet.

Eine derart mäßige Modellgüte kann nur auf unzulängliche Modellinputs zurückzuführen sein. Eine genauere Untersuchung der Eingangsdaten deutet darauf hin, dass die Überschätzung der Modelle, wie auch die Bildung von simuliertem Abfluss in Zeiten keines gemessenen Abflusses, auf überhöhte regionalisierte

Gebietsniederschläge der Teileinzugsgebiete zurückzuführen ist. Dies hat wiederum seine Ursache in hohen gemessenen Niederschlägen. Sehr kleinräumige Niederschlagszellen treffen also die Stationen, ohne aber das gesamte Gebiet zu überregnen und dieser gemessene Niederschlag wird im Folgenden fälschlicherweise auf das ganze Gebiet extrapoliert.

Im Umkehrschluss stammt eine Unterschätzung der simulierten Abflüsse sowie ein Ausbleiben simulierter Abflüsse zu Zeiten, in denen Abfluss gemessen wurde, von fehlenden Niederschlagsaufzeichnungen in Perioden, in denen das gesamte Einzugsgebiet offensichtlich ausreichende Niederschläge empfangen hat, um Abfluss zu erzeugen.

Somit kann nicht angenommen werden, dass gemessener Niederschlag auch die Niederschlagssituation im ganzen Einzugsgebiet widerspiegelt. Die Annahme, dass Niederschlag, der im Einzugsgebiet fällt, zumindest abgeschwächt auch an den Stationen aufgezeichnet wird, erscheint ebenso fehlerhaft.

Grundsätzlich erhebt sich aus diesen Erkenntnissen der Zweifel, ob eine Niederschlagsregionalisierung und eine darauf basierende Niederschlags-Abfluss-Modellierung in einer ariden, derart variablen Umgebung eine gute Basis zur Abschätzung der Modellgüte sowie zur Untersuchung von Niederschlags-Abfluss-Beziehungen bilden.

1 Introduction

As population pressure increases in the worlds arid countries, water demands for domestic, agricultural and industrial usage increase relentlessly.
Small rainfall volumes and high rates of potential evaporation, characteristics of arid environments, inevitably lead to a shortage of surface water. As surface water is hardly available, groundwater has become the most important water resource in many arid regions. Sustainable management of this resource is, therefore, essential. In this context, the term sustainable implies that abstraction rates do not exceed the rate of groundwater recharge. In any case, water resource management is dependent on a sound scientific understanding of the hydrological processes involved in these extreme climate zones.
Groundwater recharge in arid regions in turn mostly takes place as indirect recharge, in the form of percolation of flood water into alluvial aquifers. Hence, the first step to estimate groundwater recharge must be to adequately quantify episodic surface flows which are the primary feed to alluvial aquifers.
Hydrological models have, in many cases, proved to be useful tools to gain better scientific understanding of hydrological processes, and to support decision making by water resource managers.
This study tries to quantify runoff volumes for the headwaters of the Kuiseb River in Namibia by application of the Australian e2 catchment modelling software.
The Kuiseb basin, with mean annual rainfall of almost zero at the coast, can, as it features most of the hydrological processes typical for arid regions, be regarded as an exemplary arid watershed.
Most models are designed for humid environments, and the models so far applied in the southern African region are mostly based on monthly time intervals. Thus, the challenge is to apply an existing model to an arid environment, and to work with a daily model which does not overcome problems of high variability in terms of aggregation.
Prior to ultimate rainfall-runoff modelling, an attempt is made to regionalize rainfall and potential evapo-transpiration with the intention of adequately quantifying model inputs. Rainfall regionalization for the Kuiseb watershed is carried out here for the first time.

Regionalization of potential evapo-transpiration is achieved by application of a sine function to the annual regime with a scattering around the sine function according to a normal distribution.

The e2 catchment modelling software is part of the toolkit product of the CRC for Catchment Hydrology (CRCCH), Australia. The initial version of the model was developed by Robert M. Argent in 2004.

2 General aspects

2.1 Arid Hydrology

Most formal definitions of the term aridity are based on comparisons between precipitation and some measure of potential evaporation. In the classification published by UNESCO in 1979 the degree of aridity is based on the ratio of mean annual precipitation to mean annual potential evaporation estimated by the Penman approach (PILGRIM et al., 1988). The UNESCO classification defines three degrees of aridity: < 0.03 for the hyper-arid zone, 0.03-0.20 for the arid zone and 0.20-0.50 for the semi-arid zone. Thus, according to this classification, arid regions exist where annual potential evapo-transpiration is at least twice the yearly rainfall. Other classifications were defined, for instance, by Köppen in 1922, who assumes a tight correlation between temperature and ET, or by de Martonne in 1926.
NOIN et al. (1998) found that, according to the UNESCO classification, some 33 percent of Earth's terrestrial surface is arid. This area contains an estimated 840 million people, or about 15 percent of the world's population in 1994. Table 2.1 depicts the spatial distribution of drylands in different regions.

Table 2.1: Relative areas of drylands in different regions of the world (WILLIAMS, 2000).

Region	Hyper-arid	Arid	Semi-arid	Total
Africa	20.1	20.4	16.9	57.4
America	0.4	4.9	11	16.3
Middle East	18.3	49.7	16	84
Asia	1	10.5	13.9	25.4
Australia	0	49	20	69
Europe	0	0.1	2.3	2.4
World	5.7	14.1	13.2	33

These figures demonstrate the necessity of a sound scientific understanding of the hydrology of arid regions and the need for improved techniques for modelling runoff, recharge and other aspects of hydrology (PILGRIM et al., 1988). The lack of observed data in arid regions increases the need for synthesizing data and, at the same time, makes the task more difficult. Even in humid regions, were data availability is high, hydrological modelling involves many assumptions, simplifications and averaging over space and time (PILGRIM et al., 1988). Hence, for a rational interpretation of modelling results from arid regions, it is essential to be aware that greater errors and uncertainty are likely to occur in such zones.

By way of a general introduction to arid zone hydrology, the following passage describes the typical characteristics of water cycle components in arid regions. The hydrology of arid regions is characterized by rainfall inputs which show great variability in both time and space. Rainfall events often occur in infrequent local convective thunderstorms with high intensities over short periods. The total annual rainfall in arid regions can result from a few single rainfall events. WILLIAMS (2000) argues that, compared to sub-humid or semi-arid areas, where rain falls more or less seasonally, in arid and hyper-arid regions rainfall occurs unpredictably or episodically.

Potential evaporation in arid regions is high, while the actual evaporation is usually limited by the amount of water available. Moreover, high evaporation rates may cause an upward movement of water in the soil and therefore cause salination. The contribution of transpiration to ET is smaller than in humid regions, simply because the plant cover is sparse. According to PILGRIM et al. (1988), evaporation and transpiration may account for up to 95% of rainfall. This fact simplifies rainfall-runoff modelling in arid regions, because the probability is high that soil water stores are mostly empty at the beginning of a storm event (given sufficient time between two events). The effect is to minimize the errors in estimating current soil water storage.

Another feature of arid hydrology can be the absence of some humid processes, e.g. baseflow.

Runoff in arid regions is often episodic and can not always be referred to an entire catchment because rainfall events often cover just a fraction of the catchment. Many small rainfall events don’t ever generate runoff at all, because smaller amounts of water are directly returned to the atmosphere from the land surface by evaporation. Extreme rainfall events generate runoff with rapidly rising hydrographs, characterized by high peak flows, steep rising and falling limbs and no-flow conditions before and after an event. According to Jacobson (1997) peak discharges are often reached within minutes, and tributary or even mainstem flow may occur while large portions of the channel network remain dry. Concentration times are often short – the dominating runoff generation process is Hortonian overland flow, where the rate of rainfall exceeds the potential infiltration rate – and the channel network is often dense. Additional contribution to short concentration times is made by the sealing of surfaces e.g. by salt crusts which may effectively hinder the infiltration process. Hence, the soil type and superficial soil properties play a primary role in runoff production, even more so since saturation of the top soil layers hardly ever occurs (PILGRIM et al., 1988).

Sediment loads during runoff events are often high, because erosion rates are increased by the sparse vegetation. This often leads to hydrographic degeneration: the riverbed alters frequently.

Direct groundwater recharge by precipitation hardly ever occurs in truly arid regions. The water needs to be quickly transported to the subsurface; otherwise it will just contribute to evaporation and/or transpiration. Smaller rainfall events often cause an infiltration wetting front that only reaches shallow depths and can therefore usually not contribute to recharge. Hence, indirect recharge during extreme rainfall events, e.g. by transmission losses, bank storage, and cracks in rock areas, plays a major role. In particular infiltration of floodwater into alluvial sediments of the channel beds, known as transmission loss, was identified to be a key process for indirect recharge in arid regions (LANGE, 2005). In the flood hydrograph, transmission losses are typically characterized by a sudden drop-off in the hydrograph itself, and also in the flood volume further downstream (LEISTERT, 2005). Although the water table is typically below streambeds and disconnected from the surface drainage system, a temporary saturated hydraulic connection may occur during flood events, allowing groundwater recharge to emerge.

Another feature of arid environments is sparse plant cover, mainly consisting of xerophytes and ephemeral grasses and small leafy plants (PILGRIM et al., 1988). Density of vegetation may vary considerably in different regions and may also differ within one region after prolonged dry periods from those after a wet period. This leads to a variation in soil water demands over both time and space. Moreover, the absence of organic matter on the surface of grounds can have an effect on hydrological processes like interception, infiltration, evapo-transpiration, and runoff.

PILGRIM et al. (1988) insisted that rainfall-runoff modelling in arid regions must take account of the channel transmission losses which vary from point to point along an ephemeral channel with the degree of saturation of the channel alluvium, and the sealing of the channel surface prior to the runoff event.

PILGRIM et al. (1988) stresses that the hydrological processes discussed can not apply in a general sense to all arid catchments, and therefore any standard modelling approach for arid hydrology would be counter-productive and fail to develop valid models and results. This leads to the conclusion that any modelling techniques, hydrological models and parameter sets which are developed or derived for one region, will not necessarily be applicable to another. “The only sound basis for development of rational models in a particular region is observed rainfall and streamflow data for that region, complemented by careful observation and assessment of the region’s characteristics.”(PILGRIM et al., 1988)

2.2 Hydrological modelling

Hydrological models in general can be seen as simplified representations of natural systems. They serve to simplify complex natural systems, to simulate where there are no hydrological data, to make predictions for future hydrological scenarios, and to improve the scientific understanding of hydrological processes.

According to HUGHES (2004) hydrological models can be seen as mathematical representations of the processes involved in the transformation of climate inputs such as precipitation, solar radiation and wind, through surface and subsurface transfers of water and energy into hydrological outputs, typically flow in rivers, soil moisture content, and water levels in aquifers.

A common classification firstly divides hydrological models into mathematical, physical and analogue models. Physical models try to copy natural systems under laboratory conditions, so as to mimic the characteristics of a real system at a convenient scale. The application of physical models is more common in hydraulic investigations than in hydrology.

Analogue models represent the processes which occur in a system of interest by other, analogous processes. Neither analogue nor physical models will be discussed further here.

Mathematical models turn conceptual hydrological models, i.e. the idea of a hydrological system, into mathematical and logical expressions to simulate the behaviour of a natural system. Mathematical models can further be divided into stochastic and deterministic models. Deterministic models are based on the principle of cause and effect, i.e. at any time the same input generates the same output. On the contrary, the same inputs to stochastic models can generate varying outputs because at least one parameter is chosen at random.

Deterministic models are usually further classified according to their spatial resolution. Lumped models contain just one spatial unit which represents the whole system, as a rule the catchment of interest. Distributed models allow for detailed spatial distribution of the models parameters, usually dependent on the size of the grid employed. Semi-distributed models strike a balance between lumped and fully distributed models; model parameters can vary for different areas of the catchment, e.g. for sub-catchments or areas with the same hydrological behaviour.

Another classification divides mathematical models into physically based (White

Box), conceptual (Grey Box) and empirical (Black Box) models. Empirical models operate by using transfer functions which transfer a known input into a known output. They have no physical background and can consequently not describe particular processes within the system.

In marked contrast to this, physically-based models ideally describe all occurring physical processes in detail. However, this is hardly ever achievable; micro-scale processes (i.e. macro pore flow) often lack of the parameters needed for proper physically-based modelling.

Conceptual models combine some physical meaning with a certain degree of empiricism.

The main problem when working with conceptual and physically-based models is the issue of scaling (LEISTERT, 2005). The exaltation of parameters is usually done in the micro-scale or meso-scale, and the question that arises is how to transfer this knowledge to larger scales such as a catchment in a heterogeneous environment.

Working with spatial data requires careful verification of the scale that will be used for the transformation of natural data into a raster format.

Natural areas with curved perimeters are projected onto a raster with straight and rectangular perimeters which inevitably results in non-equivalence between the shape of the natural area and its projection. With increasing grid resolution this non-equivalence will decrease, making high grid resolutions desirable. But the larger the data set, the longer the computing time; hence, a compromise between data accuracy and adequate applicability has to be found (KLOCK, 2001).

2.3 State of the art

As water resources in arid countries are restricted, hydrological models are needed partly because it is impossible to observe streamflow and groundwater in sufficient detail to provide water resource management authorities with the information needed to quantify the availability of water as a natural resource. They are also required to estimate the effects of anthropogenic modifications on the natural environment and on the availability of water resources (HUGHES, 2004).

Despite this requirement for hydrological models to solve practical water resource problems, HUGHES (2004) argues that the proportion of contributions that have focused on the practical application of models for the solution of real-world problems is surprisingly low. Contrary to this trend, the limited research resources in South Africa have resulted in a fairly focused and practically orientated approach to rainfall-runoff modelling. Referring to HUGHES (2004) again, the emphasis has generally been on conceptual understanding and practical application, and less on mathematical techniques. This reference also points out that most of the models from South Africa tend to be of the more complex type, with a relatively large number of parameters, even for monthly time-step models. The reason may be a tradition of conceptual approaches to modelling in South Africa, in preference to mathematical alternatives.

A large number of parameter values, on the other hand, imply a great deal of parameter interaction. This leads to difficulties in achieving unique optimum solutions and, possibly, several combinations of parameter values that generate similar results (equifinality of model parameters).

Rainfall runoff modelling in arid regions is affected by the general characteristics of hydrological processes in arid regions. PILGRIM et al. (1988) identified the scarcity of observed data as the major problem for rainfall runoff modelling in arid regions. The paucity of long-term monitoring networks can often be explained by insufficient financial and human resources being available in arid countries. This is true, although technological advances have been made in data collection, computing power and software engineering which have led to an enormous increase in the information potentially available for hydrological modelling. LANGE & LEIBUNDGUT (2000) note that, because of sparse hydrometric gauging networks, practically all arid catchments are ungauged. Even where gauging stations exist, measurement problems may occur and stations may be harmed by violent surface flows during high magnitude floods, so further reducing the quality of the observed data. River cross sections in ephemeral river beds are frequently unstable. Despite this fact, only calibrated rainfall-runoff models have been applied to arid catchments so far; these require high quality runoff data for calibration. “To overcome problems with model calibration a distributed, non-calibrated rainfall-runoff model was developed for the 1400 km^2 catchment of Nahal Zin, Israel.” (LANGE et al., 2000). All parameters were determined in the field; hence no calibration had to be performed. The uncalibrated model proved to be a useful tool to simulate high magnitude events in arid catchments.

Developments in remote-sensing technology have experienced substantial progress (HUGHES, 2004). This leads to the issue of scale differences between available and required information; algorithms in models are often based on small scale properties (e.g. hydraulic conductivity) which are spatially highly variable. Hence the problem that arises is how to quantify these parameters at a modelling scale of several square kilometres. This problem has yet not been satisfactorily solved (HUGHES, 2004). Due to the high degree of spatial variability of rainfall inputs coupled with complex associations between soil characteristics and topography, HUGHES (1997) stresses that the development of generalizations about patterns of runoff generation in arid regions can be very difficult, even at smaller scales. At larger scales the variability of streamflow is increased by a spatially varying permeability of channel beds and high evaporation rates.

Due to the difficulty of satisfactorily quantifying the main water inputs, many of the water resource modelling approaches in the southern African region are based on monthly time intervals (HUGHES, 1997). The aggregation of rainfall data into monthly totals does indeed reduce the degree of spatial variability, but HUGHES (1997) also argues that, at the same time, “… a great deal of intensity information that can be critical to runoff generation processes in semi-arid areas is lost.”

Generally, most hydrological models were designed for humid environments and hardly reflect the hydrological situation in arid regions, which differ substantially from those in humid regions.

Despite the limitations of scarce data and limited resources there have been models developed in the southern African region that have turned out to `work` well. In the following some of these models are to be introduced.

An important process of the inter-annual water balance of some Namibian basins is that wet seasons lead to an improved vegetation cover which, in the following seasons, is the reason for increased infiltration, and more effective evapo-transpiration losses with a consequent reduction in the relative amount of runoff (HUGHES, 1997). These non-seasonal vegetation cover dynamics were incorporated into the NAMROM model developed by Mostert and may last for over three years following a wet season. This could also be an explanation for the high degree of inconsistency in relationships between rainfall and streamflow (HUGHES, 1997). The NAMROM model was designed to specifically address the properties of Namibian basins. It includes the dynamic, non-seasonal vegetation cover mentioned above, and also transmission losses to alluvial aquifers. “The model is based on a single equation for total effective precipitation, using four parameters; antecedent weighting factor (seasonally varying), initial loss, sub-catchment loss factor and loss exponent.” A regression equation, containing two regression parameters, is then developed for observed runoff and total effective precipitation. Hence, the model is more of a statistical regression type with weighting parameters having some perceived physical meaning. HUGHES (1997) says its general applicability is largely untested, because the NAMROM model has so far only been applied to a number of basins within Namibia.

HUGHES (2004) states: “The best example of a model that has been extensively applied in the southern African region is the Pitman model, also referred to by its commercial name, WRSM2000.” The Pitman model was first described by Pitman in 1973. According to HUGHES (1997), the model is an explicit soil moisture accounting model that represents interception, soil moisture and ground water storages, including model functions to represent the inflows and outflows from these. The basic conceptualization has been preserved throughout the years in all subsequent versions that have been re-coded by the original author and others, although, additional components, and functionality have been added. (HUGHES, METZLER, 1998). The model is of a conceptual type and is available both as a semi-distributed and as a fully-distributed model. The semi-distributed version was initially included in the HYMAS modelling system developed at Rhodes University.

A new version of the model, which takes the dynamic vegetation growth processes into account, is referred to as NamPit (HUGHES, METZLER, 1998).

A comparative study of the Pitman, the NamPit and the NAMROM models conducted by HUGHES & METZLER (1998) at semiarid catchments within Namibia under the southern African FRIEND (Flow Regimes from International Experimental and Network Data) programme indicates that, in general terms, the NAMROM model performed more successfully than the two versions of the Pitman model, and that the NamPit version performed better than the original version.

The variable Time Interval (VTI) model is a daily time-step (or less) model, developed at the IWR, Rhodes University to reach a compromise solution between complex, fully-distributed, physically-based models and simplified lumped approaches (HUGHES, 1997). The result is a physically based semi-distributed model, containing representations of the processes thought to prevail in the southern African region (semi-arid to humid). VTI has a modular structure, where each module describes a separate component of the hydrological cycle. HUGHES (1997) declares that this model is inevitably more difficult to apply than the Pitman model, partly because it has a far greater parameter space.

The problems with applying this model to semi-arid areas are similar to those experienced with the Pitman model; although the VTI model has a channel transmission loss function, it is very empirical and difficult to calibrate when the processes involved are not well understood and there is no real information available about observed losses.

Although transmission losses have been identified to be an important component for recharge in the region, there have been very few direct studies of the process itself (HUGHES, 1997). According to HUGHES (1997) losses to alluvial aquifers are well documented at various scales, but there is a lack of generalized quantitative approaches to estimating these losses. Point measurements of infiltration and alluvial moisture, measurements of the moisture in the alluvium during flood events and tracer experiments along short reaches can be applied at smaller scales to display the spatial and temporal variations of transmission losses (LANGE, 2005). Water balance estimations at larger scales may be applied if hydrometric data is available upstream and downstream in certain channel reach, and, if any lateral inflows can be quantified, too. LANGE (2005) has attempted to identify transmission losses during single events within a 150 km long arid channel reach of the Kuiseb River, Namibia, by applying a mathematical flow routing scheme with uncalibrated parameters. All parameters were derived independently from topographical maps, air photos and field measurements. The routing model does not include transmission losses. Thus, these losses can be identified as the difference between measured and modelled streamflow where the simulated hydrograph exceeds the values of the measured one. Results indicate that significant transmission losses are likely to occur at high discharge peaks and are clearly smaller during small to medium events. According to LANGE (2005) two different processes might explain this behaviour. First, the flooding of large overbank areas offers additional large storage volumes for runoff losses and in addition enhanced evaporation may occur from these areas. Secondly, the sealing of alluvial surfaces by a silt layer which is only forced open at higher discharge.

HUGHES (1997) points out that the losses from non-alluvial rivers could also be substantial, although there is, as yet, no evidence for this.

An application of the E2 model executed by ARGENT (2006) yielded satisfying flow prediction results. In this Whole-of-catchment modelling of the Port Phillip and Western Port Bay catchments in Australia, runoff was generated from over 150 sub-catchments with five different sets of regionalised rainfall runoff parameters based on land use and geographical position.

2.4 Conclusion

In many of the world's arid regions, rivers stay dry most of the year. This is due to episodic rainfalls. However, if these rainfall events yield enough water, runoff can be generated. Such rivers, with the unpredictable, non-seasonal occurrence of flow and no flow conditions between runoff events, are called ephemeral rivers.

General hydrological features of arid regions are high intensity rainfalls which are highly variable in space and time, high rates of potential evapo-transpiration ETP and runoff events with steep rising and falling limbs. In ephemeral rivers no-flow conditions characterize the interim periods between consecutive runoff events. Furthermore, indirect recharge, mainly caused by transmission losses in the riverbeds, is the dominating recharge process in truly arid regions.

These particular hydrological and meteorological properties, combined with soil properties which differ from those in humid regions cause differences in runoff generation processes, too. Baseflow is often non-existent and interflow hardly ever occurs; the dominating runoff process is usually surface runoff.

The application of hydrological models in arid regions is hampered by poor records of rainfall, evapo-transpiration, and runoff. The situation for gauging networks for evapo-transpiration is often even worse than for rainfall networks. Servicing of gauging networks is hindered by the difficult access to gauging stations and insufficient human and financial resources. Additionally, gauging stations can be harmed by high magnitude floods, by vandalism and by altering streambeds.

The scaling problem, generally inherent in hydrological modelling, is made more acute by the poor availability of data in arid regions. Thus, development of specialized local approaches for regionalisation is of particular importance in arid regions.

Due to the high variability of hydrological processes, most models which have so far been applied in the southern African region are based on monthly time intervals, and use aggregation to reduce the degree of variability. Processes particular to arid regions, e.g. the non-seasonal vegetation cover dynamics in Namibia, have also been incorporated into local rainfall-runoff models. Losses to alluvial aquifers have already been verified by means of water balance estimations, modelling approaches and by point measurements of infiltration and alluvial moisture.

3 The e2 modelling framework

3.1. Overview

“E2 provides a flexible approach to whole-of-catchment modelling, supporting creation of integrated models through selection and linking of component models of a complexity appropriate to the management or research questions being addressed, and the available data and knowledge.”(ARGENT et al., 2006).

3.2. Introduction

Most models are difficult to apply when the problem scenario changes and different, new tasks are asked of a given model. Models with fixed algorithms and structures are often not flexible enough to adequately satisfy changing model demands (ARGENT et al., 2006).

To better meet these needs, a flexible catchment modelling framework, named e2, was created in Australia within the Cooperative Research Centre for Catchment Hydrology (CRCCH) “Catchment Toolkit”. The idea behind e2 is to provide a flexible structure that allows users to select a level of model complexity appropriate to the problem at hand and the available data and knowledge (CRC for catchment Hydrology, Australia 2004-2005). This is in sympathy with the idea of NASH & SUTCLIFFE (1970) to work with a “simple” model in which complexity can be increased or decreased to obtain good values for the model efficiency R^2 and the parameter sensitivity, and consequently sound fits between observed and computed data.

It is designed to allow modellers and researchers to construct a model by selecting and linking components from a range of options. e2 is also extensible using a plug-in approach, where specialist functionality, such as new models, can be plugged-in to the core framework (ARGENT et al., 2006).

Both the plug-in approach and the option to work with component models from other toolkit products make e2 an attractive model to work with.

e2 has a particular conceptual structure made up of tens to hundreds of sub-catchments. e2 uses a hierarchical, nested structure, especially with respect to the spatial scale. The sub-catchments exist of Functional Units (FUs) – areas with similar hydrological properties - which, hydrologically, function in the same manner. These FUs satisfy the requirement for sub-area variability in the sub-catchments. A finer definition of sub-catchments can be made and hence the model is spatially scalable (ARGENT et al., 2006). Each FU can have component models, representing processes of runoff generation, constituent generation and filtering, attached to it.

The main model structure is “node-link”, where sub-catchments feed water and material fluxes into nodes, from which they are routed along links. Sub-catchment processes are then made up of a combination of runoff generation, constituent generation and filtering, whereas processes along flow links consist of routing and in-stream processing.

Depending upon the component or plug in model selected, E2 calculates flow predictions and constituent loads at defined points in a river network over time, operating down to daily or sub-daily time steps and reporting on monthly to decadal scales (ARGENT et al., 2006).

3.3. e2 – Model Structure

“The fundamental structure of e2 uses sub-catchments, nodes and links.” (ARGENT et al., 2006).

A spatial classification is obtained by dividing the catchment into sub-catchments. “The combined effects of the processes occurring in a sub-catchment are directed to a sub-catchment outlet, represented by a node.” (ARGENT et al., 2006).

Nodes and links subsequently provide for the movement of flow and material in the system and equally for routing and transformation of material in-stream.

e2 is built upon the Invisible Modelling Environment (TIME) sharing several of its characteristics.

Another core concept in e2 is the choice and combination of adequate component models. These component models represent fundamental hydrological processes like runoff generation or routing at a consistent level of detail.

“By selection of appropriate component models and specification of system network geometry, e2 can be used to create a range of whole-of-catchment models that differ in complexity but which use the same sets of input data.”(ARGENT et al., 2006).

3.4. Sub-Catchment Processes

Functional Units (FUs) in sub-catchments act in a hydrologically similar manner – hence, each FU is represented by a particular model with particular parameters.
However, there is no direct processing of flow or material from one FU to an adjacent one – the combination of processes occurring in a sub-catchment only act at the sub-catchment outlet node (ARGENT et al., 2006). “If such processing is required due to the nature of the problem situation, then a finer definition of sub-catchments can be made. In this way, e2 is scalable, being able to represent systems from backyards to continents.” (ARGENT et al., 2006).
Every single FU can have component models for:

- Runoff generation
- Constituent generation, and
- Filtering, where filtering includes transformation processes between source and outlet.

Hence, each FU can have its own rainfall-runoff model. The currently available rainfall-runoff component models are:

- AWBM
- Baseflow Separation
- Observed Flow
- SimHyd
- Sacramento
- SMAR

All these Rainfall-Runoff models are run on a daily basis, although e2 works with any time step required by a component model.
In this study, little importance will be attached to the processes of constituent generation and filtering, which can also be modelled using e2.

3.5. Nodes and links

“The outlet nodes of various sub-catchments are joined by links, along which the flow of water and constituents can be modified through routing, the effects of sources and sinks, and in-stream processing such as storage, decay and enrichment.” (ARGENT et al., 2006).

Nodes are considered to have no spatial extent. Reservoirs, as they do have a spatial extent, are represented as links rather than nodes.

The only types of behaviour represented by nodes are extraction and the representation of water demands.

In contrast, links can have both routing and processing models assigned to them. Various routing models are available, i.e. simple lags or Muskingum-Cunge routing among others. For in-stream processing, models are limited to exponential decay, and sediment and nutrient deposition.

Dams are treated as certain types of link models. “e2 has an elegant dam model available that has a depth-volume-area relationship, losses, and minimum and maximum release curves.

3.6. e2 – The Software

“e2 is a 32-bit Windows™ application based upon TIME, the invisible modelling environment, which is a model development system that relies heavily on the use of metadata and which has a component-based approach to software construction. TIME has been under development for some three years, and has a developer base of over 30 developers across Australia.” (ARGENT et al., 2006).

The architecture of e2 consists of three layers: user interface, modelling engine and handling of data input-output, which is a fairly standard approach in recent software.

The modelling engine, user interface, on-disc persistence mechanism and the calibration tools rely on the use of software interfaces, software reflection and various software design patterns for flexibility and to allow for extensibility of the modelling framework. The software engine exists of four main elements: nodes, links, sub-catchments and functional units and it relies on a standardized software representation of mass balance and unit consistency throughout the system. The problem of handling units and the related issue of mass balance is inherent to the modelling over a variety of spatial and temporal scales with altering constituents and different models. To confront this issue the modelling engine represents physical quantities in S.I. base units, allowing component models to have their parameters in other units as long as their output to the modelling engine conforms to the S.I. standard.

As already mentioned, e2 was designed for extensibility and flexibility and is also extensible through a plug-in approach.

If new component models are appropriately coded, they are recognized by e2 and can be loaded into the model through a plug-in menu (ARGENT et al., 2006).

“The basic operation of e2 uses projects, which are able to contain one or more scenarios.” (ARGENT et al., 2006). In this way e2 is based on a structure of projects and scenarios, where a project is the wrapper for a series of scenarios and therefore the keeper for a saved list of parameters which can be applied to models in the scenarios. Projects within the model are defined by the catchment network; hence the specification of the catchment network, by network calculation from a Digital Elevation Model (DEM) or by manual network configuration, is the first step, and one of the main tasks when setting up an e2 project.

Using the DEM method, a stream network is first calculated using a stream threshold setting for the area; then, sub-catchments are automatically created for areas above any junction. “For coarser or finer networks, and less or more sub-catchments, the `stream threshold` value is simply changed up or down.” (ARGENT et al., 2006). If the drainage network is not properly represented by the surface topography, or where no DEMs are at hand, the manual network definition uses a mouse click-and-drag approach.

“Scenarios are built through a wizard that steps users through the processes of specification, model assignment, data attachment and parameterization.” (ARGENT et al., 2006).

For the analysis of output from a model run, different tools are available (i.e. graphs, statistics, computation, unit conversion and maps).

3.7. Component Models

The flexible approach inherent in the e2 modelling framework supports the selection and linking of component models appropriate to the hydrological problem addressed and the available data and knowledge (ARGENT et al., 2004/05). The component models are grouped according to their function:

- Rainfall Runoff
- Constituent Generation
- Filters
- Links – routing and storage
- Links – processing
- Nodes

In the following, the common component models for rainfall runoff and links – routing and storage available for selection in the e2 modelling framework, will be introduced in more detail. The component models for the remaining functions are of no particular interest for this study and are thus not explained further.

3.7.1 Rainfall Runoff

The structure of e2 allows for the assignment of a rainfall runoff model for each FU. The user can also choose no model (Nil Runoff). Complexity and input requirements may alter from one model to another. The outputs of all rainfall runoff models are the same with two time series – one for the surface flow and one for the groundwater flow (ARGENT et al., 2004/05). Input requirements consist of daily time series of area rainfall and area PET, except for AWBM, which requires actual evapo-transpiration. Both data sets need to be continuous and overlapping. Daily flows are required for calibration.

3.7.1.1 Observed Runoff

This model is used to input an observed runoff sequence instead of runoff generated by a model.

3.7.1.2. Australian Water Balance Model (AWBM)

The AWBM is a mass balance model based on conceptual relationships. It is a catchment water balance model relating runoff to rainfall with daily or even hourly data, calculating losses from rainfall for flood hydrograph modelling. The original model was coded by Walter Boughton in FORTRAN and converted to C# by J.M.Perraud and forms part of the TIME library of models. The owner of the TIME version is the CRC for Catchment Hydrology (ARGENT et al., 2004/05).

The model has 8 parameters and it contains 5 stores. Three surface stores serve the simulation of partial runoff areas. The remaining stores consist of a base flow store and a surface runoff routing store. The structure of AWBM is shown in figure 3.7.1.

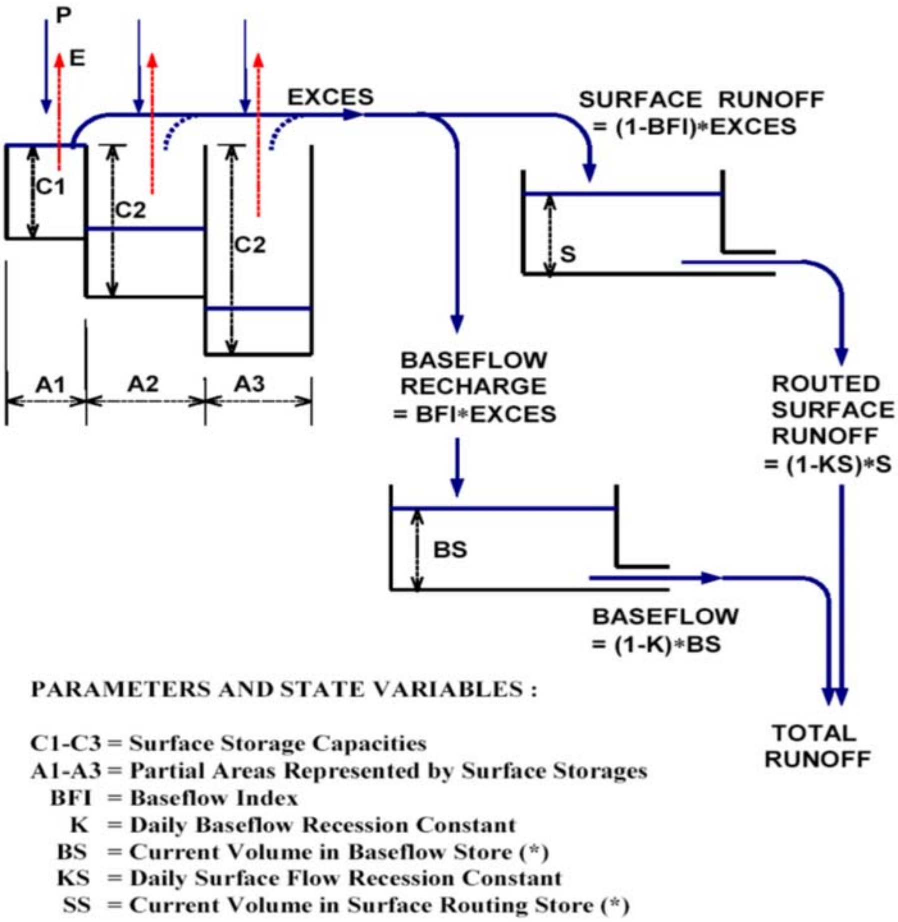

Figure 3.7.1: Structure of the AWBM model. (ARGENT et al., 2004/2005)

At each time step, the water balance for each surface store is calculated independently from the others at daily or hourly time steps. If the content of the moisture stores exceeds their storage capacity, runoff is generated and the stores are reset to their capacity (ARGENT et al., 2004/05). As runoff occurs from any store, part of it becomes recharge of the base flow store, given that there is base flow in the stream flow. The fraction of the runoff recharging the base flow store is BFI*runoff, where BFI is the base flow index. The remainder of the runoff, (1-BFI)*runoff, is surface runoff. The base flow store is drained at a rate of (1-K)*BS, where BS is the current moisture in the base flow store and K is the base flow recession constant. The surface store acts simultaneously to the base flow store and is depleted at the rate of (1-KS)*SS, where SS is the current moisture in the surface runoff store and KS is the surface runoff recession constant (ARGENT et al., 2004/05).

It is important to realize that AWBM requires actual evapo-transpiration as an input whereas most other models take potential evapo-transpiration (PET) as an input.

The model parameters are depicted in table 3.7.1.

Table 3.7.1: AWBM model parameters (ARGENT et al., 2004/2005)

Parameter	Description	Default	Minimum	Maximum
A1	Partial area of surface store 1	0.134	0	1
A2	Partial area of surface store 2	0.433	0	1
C1	Capacity of surface store 1 [mm]	7	0	50
C2	Capacity of surface store 2 [mm]	70	0	200
C3	Capacity of surface store 3 [mm]	150	0	500
BFI	Base flow index	0.35	0	1
K	Base flow recession [day^{-1}]	0.95	0	1
KS	Surface flow recession [day^{-1}]	0.35	0	1

Generally, the model is most sensitive to the recession constants and the base flow index.

3.7.1.3 Baseflow Separation

"The Baseflow separation filter identifies the base flow component of flow and when subtracted from the total flow gives the quick flow." (ARGENT et al., 2004/05). The Baseflow Separation model is a mathematical filtering algorithm. The current version was developed by J.M.Perraud and is part of the TIME library of models, owned by the CRC for Catchment Hydrology.

The model applies a digital filter and the only parameter is α, the filter parameter which lies between 0 and 1.

3.7.1.4 SIMHYD

SIMHYD is a lumped daily conceptual rainfall runoff model that estimates daily flow from daily rainfall and evapotranspiration data for the area (ARGENT et al., 2004/05). The model contains 3 stores for interception loss, soil moisture and ground water and has 7 parameters. SIMHYD, a mass balance model, is a simplified version of the HYDROLOG and MODHYDROLOG models, containing significantly fewer parameters.

The model's current version was developed by F.Chiew in Fortran and converted to C# by J.M. Perraud; it is part of the TIME library of models, owned by the CRC for Catchment Hydrology (ARGENT et al. , 2004/05). The model structure of SIMHYD is depicted in figure 3.7.2.

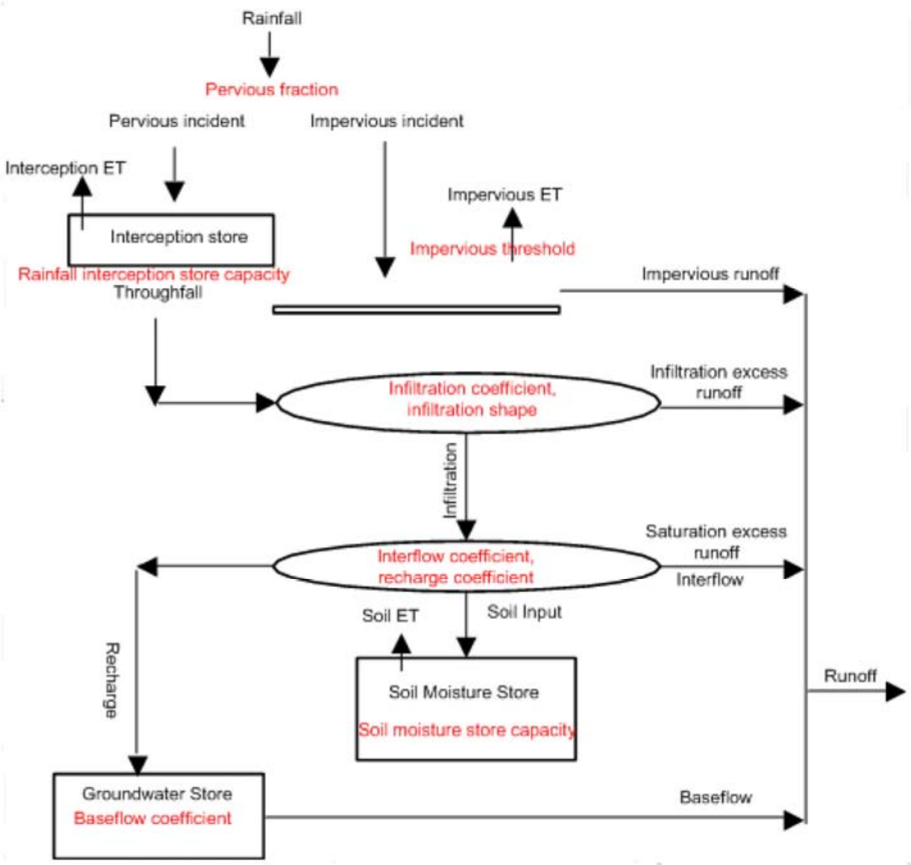

Figure 3.7.2: Structure of the SIMHYD model (ARGENT et al., 2004/2005)

A unique feature of SIMHYD is that it divides the catchment area into pervious and impervious fractions. This is done by the pervious fraction parameter which gives the amount of pervious surfaces within the catchment in percent. Impervious runoff is directly generated and is only diminished by ET from impervious surfaces.

Rainfall on pervious surfaces however fills the interception store which is emptied by evaporation. The rainfall interception storage (RISC) parameter determines the storage capacity of the interception store.

An infiltration function determines the infiltration capacity for the portion of rainfall that is not retained by the vegetation cover. The excess rainfall that exceeds the infiltration capacity becomes surface runoff. A soil moisture function then divides the infiltrated water into interflow, recharge and water that remains in the soil moisture store.

“Interflow is first estimated as a linear function of the soil wetness (soil moisture level divided by soil moisture capacity).” (ARGENT et al., 2004/05). Hence, the equation that simulates interflow mimics both the interflow and saturation excess runoff processes, where the soil wetness indicates saturated areas in the catchment from which saturation excess runoff may occur.

Recharge is then estimated as a linear function of the soil wetness, as well as evapotranspiration from the soil moisture store. Evapotranspiration cannot exceed PET. If the capacity of the soil moisture store is exceeded it overflows into the ground water store. Baseflow is controlled by a linear recession function.

The basic equations of the model are:

Impervious ET = min (pet, (1-pervious Fraction)*pervious Threshold, impervious Incident)

Interception ET = min (pervious Incident, pet, Rainfall Interception store capacity)

Infiltration capacity = pervious Fraction*infiltration Coefficient*exp (-Infiltration shape*Soil moisture fraction)

Infiltration = min (Throughfall, Infiltration capacity)

Interflow Runoff = Interflow Coefficient*soil Moisture Fraction*Infiltration

Infiltration after Interflow = Infiltration-Interflow Runoff

Recharge = recharge Coefficient*soil Moisture Fraction*Infiltration after Interflow

Soil Input = Infiltration after Interflow-Recharge

Thus the model reflects runoff generation processes from three different sources: infiltration excess runoff, interflow (and saturation excess runoff), base flow (ARGENT et al., 2004/05).

The models parameters are depicted in table 3.7.2

Table 3.7.2: SIMHYD model parameters (ARGENT et al. , 2004/2005)

Parameter	Description	Default	Minimum	Maximum
K	Baseflow coefficient, baseflow linear recession parameter.	0.3	0	1
ImpT	Impervious threshold, depression storage capacity.	1	0	5
COEFF	Infiltration coefficient, maximum infiltration loss [day-1]	200	0	400
SQ	Infiltration shape, part of the infiltration exponent	3	0	10
SUB	Interflow coefficient, constant of proportionality in the interflow equation. [day-1]	0.1	0	1
Perv	Pervious fraction.	0.9	0	1
INSC	Rainfall interception store Capacity [mm]	1.5	0	5
CRAK	Recharge coefficient, constant of proportionality in the gw recharge equation [day-1]	0.2	0	1
SMSC	Soil moisture store capacity [mm]	320	1	500

3.7.1.5 Simple Urban Runoff Model (SURM)

“SURM is a daily conceptual rainfall-runoff model that estimates daily stream flow from daily rainfall and areal evapotranspiration data.” (ARGENT et al. , 2004/05). The model’s origin is the same as for the SIMHYD model, and like SIMHYD, the SURM model contains 3 stores and 7 parameters.

3.7.1.6 Sacramento

“The Sacramento model is a catchment water balance model that relates runoff to rainfall with daily data.” (ARGENT et al. , 2004/05).

Two surface stores serve to mimic surface evaporation, surface runoff and interflow. The remaining 3 base flow stores represent two types of baseflow and evaporation from the soil store. The model has 16 parameters.
Sacramento is a lumped mass-balance model based on conceptual relationships.

3.7.1.7 Soil Moisture Accounting runoff Model (SMAR)

SMAR is a lumped conceptual rainfall-runoff water balance model that, according to ARGENT et al (2004/05), outputs "…daily estimates of surface run-off, groundwater discharge, evapotranspiration and leakage from the soil profile for the catchment as a whole", using soil moisture as a central issue. The model has 9 parameters and consists of a water balance and a routing component in series.

3.7.2 Links – Routing Models

In general, routing models serve to describe flood wave characteristics such as the time lag between a flood entering and leaving a channel reach, the changes in shape and amplitude of the hydrograph, and lateral inflows into the reach.
Most routing equations are based on the mass balance and an equation that relates outflow rates to changes in storage volume:

$$dS/dT = I_{(t)} - Q_{(t)}$$ **equation 3.7.1**

$$S = f\,(Q_{(t)})$$ **equation 3.7.2**

where I is the inflow rate in m^3/s, Q is the outflow rate in m^3/s and S the storage volume of the river reach in m^3.
Assuming that the flow rates over a time step are linearly interpolated between two points in time, and after some mathematical conversions we arrive at the equation:

$$(S_2 - S_1) / \Delta t = ((I_1+I_2) / 2) - ((Q_1+Q_2) / 2)$$ **equation 3.7.3**

, substituting $S_2 = f(Q_2)$ gives:

$$F(Q_2) = f(Q_2) + Q_2/2 - S_1 - \Delta t\ ((I_1+I_2-Q_1) / 2) = 0 \qquad \textbf{equation 3.7.4}$$

In this equation, the outflow rate Q_2 may be isolated, and its solving algorithm and starting values may be modified from one routing scheme to another (ARGENT et al., 2004/05).

So far, none of the routing models inherent in e2 allow for seepage within a channel reach.

3.7.2.1 Straight Through Routing

In this case no routing scheme is applied. This implies that neither delay, nor attenuation and time lag are taken into account. Inflow and outflow volumes match one another and a flow wave enters and leaves a channel reach within the same time step.

3.7.2.2 Lagged Flow Routing

The simplest routing model available, which operates by simply delaying flow within a link by a certain time step. Hence, it is parameterized by one single value of lag expressed as a certain number of time steps.

3.7.2.3 Laurenson Non-Linear

The Laurenson non-linear routing component is based on a storage-outflow relationship:

$$S_{(Q)} = K{*}Q^m \qquad \textbf{equation 3.7.5}$$

Where K is a dimensional empirical factor that acts as a storage delay parameter and m is a dimensionless empirical exponent serving as a measure of the non-linearity of the model.

3.7.2.4 Laurenson Non-Linear with Lag

This routing component is a combination of the Laurenson method described in the previous section (3.7.2.3) and a method of lagging the resulting outflows by a multiple of the routing time-step. According to ARGENT (2004/05), this routing scheme "…may be of particular interest as a fairly sound basis for routing constituents in a reasonably realistic manner."

3.7.2.5 Muskingum

The Muskingum routing procedure is based on the storage-outflow relationship:

$$S(Q) = K*(XI+(1-X)*Q) \quad \textbf{equation 3.7.6}$$

,where K is the travel time along the reach, and X is a dimensionless value expressing the relative effect of inflow and outflow on the storage of the reach. (ARGENT et al., 2004/05). The Muskingum parameters X and K can be seen as empirical and their values can be allocated when an outflow hydrograph is at hand. Without observed flow data however, the estimation is rather difficult; it can be done by a method introduced by Cunge, where the parameter X is related to the physical channel properties.

Conceptually the reach storage is expressed as the sum of a prism and a wedge storage.

3.7.2.6 EMSS routing

The EMSS routing method is based on the Muskingum Cunge routing procedure, but also takes into account lateral inflows to the channel reach.

3.7.2.7 Muskingum with Losses

This modified version of the Muskingum model calculates losses at every time step, based on a loss rate value. The losses are "…set to occur at the end of the time step, to ensure that mass balance is preserved up to that point." (ARGENT et al., 2004/05).

3.7.2.8. Storage model

Storages in e2 are represented as links, and are thus just another routing scheme at a more abstract level and have therefore to occur in the list of routing models.

The e2 storage model works by maintaining mass balance, and operates on a monthly daily or sub-daily time step. (ARGENT et al., 2004/05).

The central water balance equation of the e2 storage scheme assumes that the change in storage height is small compared to the storage fluxes. “This simplification avoids an iterative solution across each time step.” (ARGENT et al., 2004/05). This assumption is true for large storage volumes but needs some correction in cases with small storage volumes to ensure that the storage volume does not become negative.

Within the water balance, rainfall and inflow are added first, before losses in terms of infiltration and evaporation are removed; this ensures that losses cannot exceed the storage volume. The water balance equation of the e2 storage model is given by:

$$S_t = \max(0, S_{t-1} + ((I_t + I_{t-1})/2 + (R-E)*A(h_{t-1}) - L(h_{t-1}) - \max(OU(h_{t-1})*\min(D, OC(h_{t-1}))))*dt)$$

equation 3.7.7

Where:

S_t : storage at the end of the time step [m³]

S_{t-1}: storage at the start of the time step [m³]

It: inflow at the start of the time step [m³/s]

It-1: inflow at the end of the time step [m³/s]

R: rainfall [m/s]

E: evaporation [m/s]

A(ht-1): surface area for the storage depth at the start of the time step [m²]

L(ht-1): infiltration losses for the storage depth at the start of the time step [m³/s]

D: storage demand [m³]

OU(ht-1): uncontrolled (minimum) outflow for the storage depth at the start of the time step [m³/s]

OC(ht-1): controlled (maximum) outflow for the storage depth at the start of the time step [m³/s]

h: storage depth [m]

dt: time step [s]

The input requirements for the storage model include: time series for evaporation and rainfall, the storage volume, area and level of the storage with minimum and maximum outlet tables, and, as an option, a table in which the storage level is depicted against infiltration.
The outputs of the storage model consist of time series of: storage volume, net evaporation volume, demand volumes and released flow. (ARGENT et al., 2004/05).

3.8. Calibration Tool

e2 contains a flexible calibration tool to support calibration of flow from both sub-catchments and sub-networks, where sub-networks are a small group of sub-catchments, nodes and links.
Similar to the scenario wizard, the user is guided through sub-catchment or network selection, parameter grouping and scaling, and model running by a `wizard` in the calibration tool.
An important aspect of the models calibration tool is the option to group and scale parameters manually for various sub-catchments.
"A range of efficiency criteria, inherited from TIME, are available and new methods can be added as required." (ARGENT et al., 2006).

3.9. Plug Ins

Plug-ins expand the model's features to include a range of input, manipulation or output tools beyond the basic modelling of flow and/or constituents.
Models of the appropriate type are automatically recognized and made available for assignment by e2 for the common processes – hence users with specific needs can have their own custom built component models added to the basic version.
"Connection and integrated operation of e2 with other models is also handled through plug-ins." (ARGENT et al., 2006). e2 model integration currently works with the 2CSalt model, SedNet and IQQM. Other plug-ins can be drawn from general and specific sources to provide functions like raster and time-series calculation or to incorporate the River Analysis Package (RAP) routines for Hydraulic and Time Series Analysis.

3.10 Conclusion

e2 provides a flexible framework for catchment modelling. The facility to choose component models for runoff and constituent generation, filtering and routing contributes to this fundamental idea. As a result, the hydrological processes mentioned can be described at different levels of complexity, according to which component models are chosen.

Additionally, e2 is spatially scalable in the sense that users can increase or decrease the number of sub-catchments, and the number of functional units.

In regard to rainfall-runoff modelling, again complexity can be added or subtracted from the model as recommended by NASH&SUTCLIFFE (1979). This is achieved by assigning appropriate component models for rainfall-runoff modelling to each FU. Thus, areas with different runoff generation and runoff concentration processes can be described separately by adequate models.

Unfortunately not all component models for rainfall-runoff modelling were readily available when this study was carried out. The AWBM and the SIMHYD model were chosen for application.

AWBM offers a hydrotope-like approach with three surface stores which are represented by three areal fractions of the catchment. In this way the model can accommodate, for example, classification of dominant soils, classes of slope, or types of land use.

The SIMHYD model divides the catchment into areas with pervious and impervious properties. Runoff from impervious areas is directly generated as surface runoff whereas surface runoff from pervious areas can occur as infiltration excess and saturation excess runoff.

Both models are characterized as spatially lumped; due to the sub-catchment structure of e2, however, they act more in a semi-distributed manner.

The e2 storage model allows for losses, and is therefore of potential interest for further application at the Kuiseb River.

Summarizing, e2 offers a variety of potential applications at different levels of complexity within a flexible framework for catchment modelling.

4 Study area

4.1 General

The Kuiseb is an ephemeral stream in the western part of Namibia. The river has its source in the Komashochland at an elevation of about 2000 m above sea level near the country's capital Windhoek. It has a catchment area of 14700 km^2, and a length of approximately 560 km.

The geographical location of the Kuiseb catchment within Namibia is shown in figure 4.1.1. A map of Namibia is depicted, settlements are shown as red dots. The Kuiseb basin is pictured with an internal classification according to classes of elevation above sea level. The satellite image in figure 4.1.1 shows how the Kuiseb River divides the northern gravel plains from the dune fields in the south.

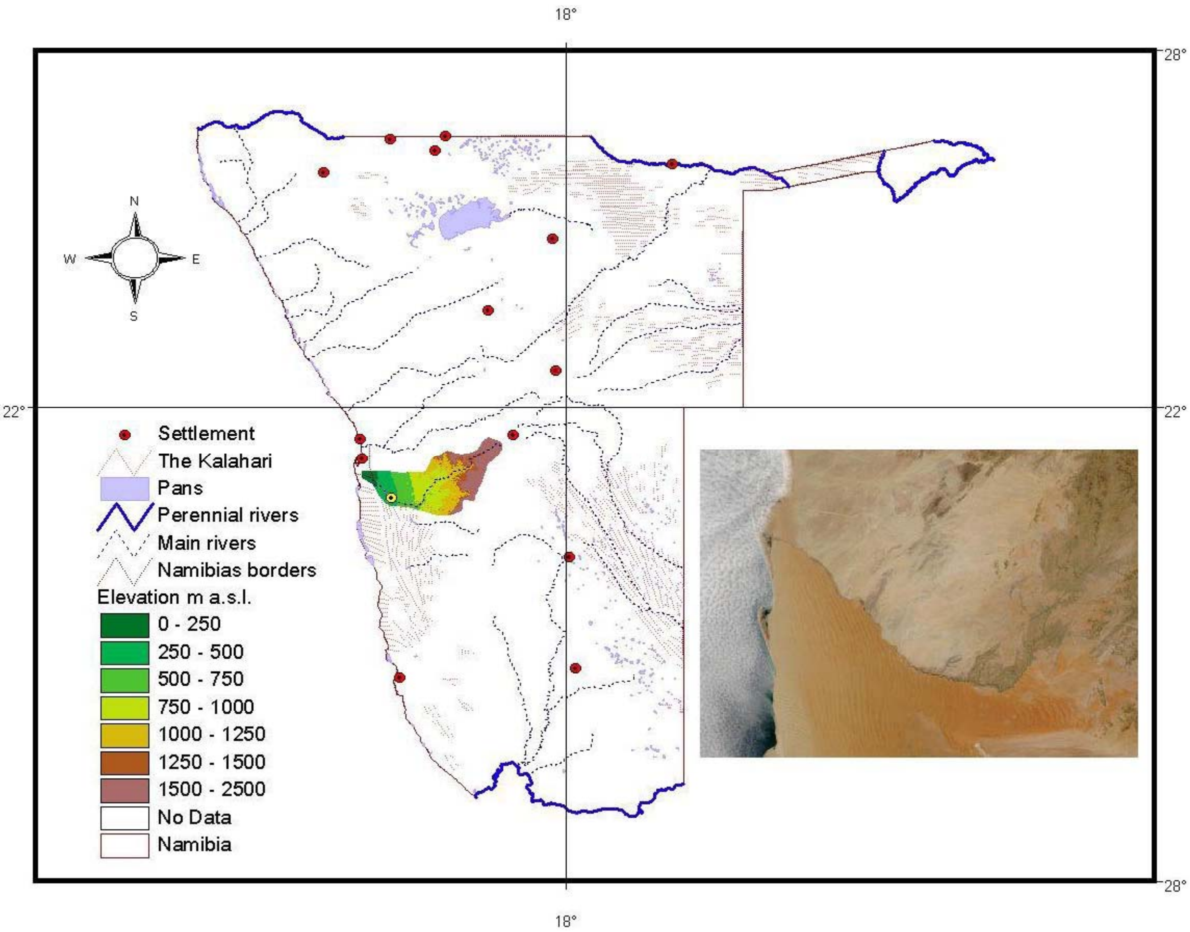

Figure 4.1.1: Geographical location of the Kuiseb catchment within Namibia

The Kuiseb channel runs over the escarpment that separates the inland plateau from the coastal plains, then crosses the Namib Desert from east to west and finally reaches the Atlantic Ocean. The Namib Desert runs the length of the country and extends inland approximately 150 km to the Great Western escarpment (JACOBSON, 1997). Within the Namib, the Kuiseb channel separates the dune fields in the south from the gravel plains in the northern part. According to Hattle (1985), about 1/3 of the catchment consists of the desert plain which only yields runoff in exceptionally wet years.

4. 2. Climate

The mean annual rainfall of Namibia is 284 mm, ranging from only 50mm in the western coastal area to 700 mm/a in the north-western part of the country. Within the Kuiseb basin, mean annual rainfall ranges from 20 mm/a on the coast to 360 mm/a in the headwater area. Rainfalls in the headwater areas have their origin in bodies of humid air which flow from the Indian Ocean across southern Africa (SCHMIDT & PLOETHNER, 1999).

The characteristic meteorological situation arises from a high pressure area on the Namibian coast combined with the effects of the cold Benguela current drifting northward, and broadly dominates the region's climate. (SCHMIDT & PLOETHNER, 1999). The spatial distribution of mean annual rainfalls within the Kuiseb catchment is shown in figure 4.2.1.

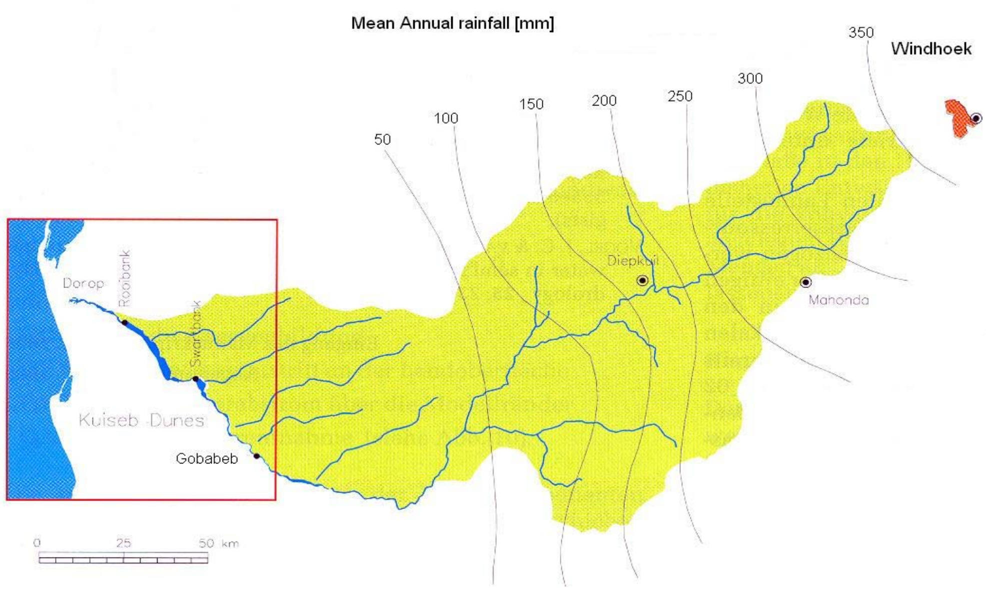

Figure 4.2.1: Mean annual rainfall distribution in the Kuiseb basin (SCHMIDT & PLOETHNER, 1999).

The highly infrequent precipitation in the desert area, in the form of rain or offshore mist, is only just sufficient to feed the demands of local flora and fauna.

Figure 4.2.1 shows the increase in mean annual rainfall in the Kuiseb basin from west to east. It rises as the elevation above sea level increases.

Most of the rainfall events are recorded during the hot summer months; they occur as small scale convective storms. Precipitation peaks between January and April.

JACOBSON (1997) points out that the annual evaporative losses are high throughout the region, reaching a mean pan evaporation rate of 3168 mm/a in the Central Namib and culminating in annual rates of about 4000 mm. Annual evaporative losses thus exceed mean annual rainfalls by a factor of about 200.

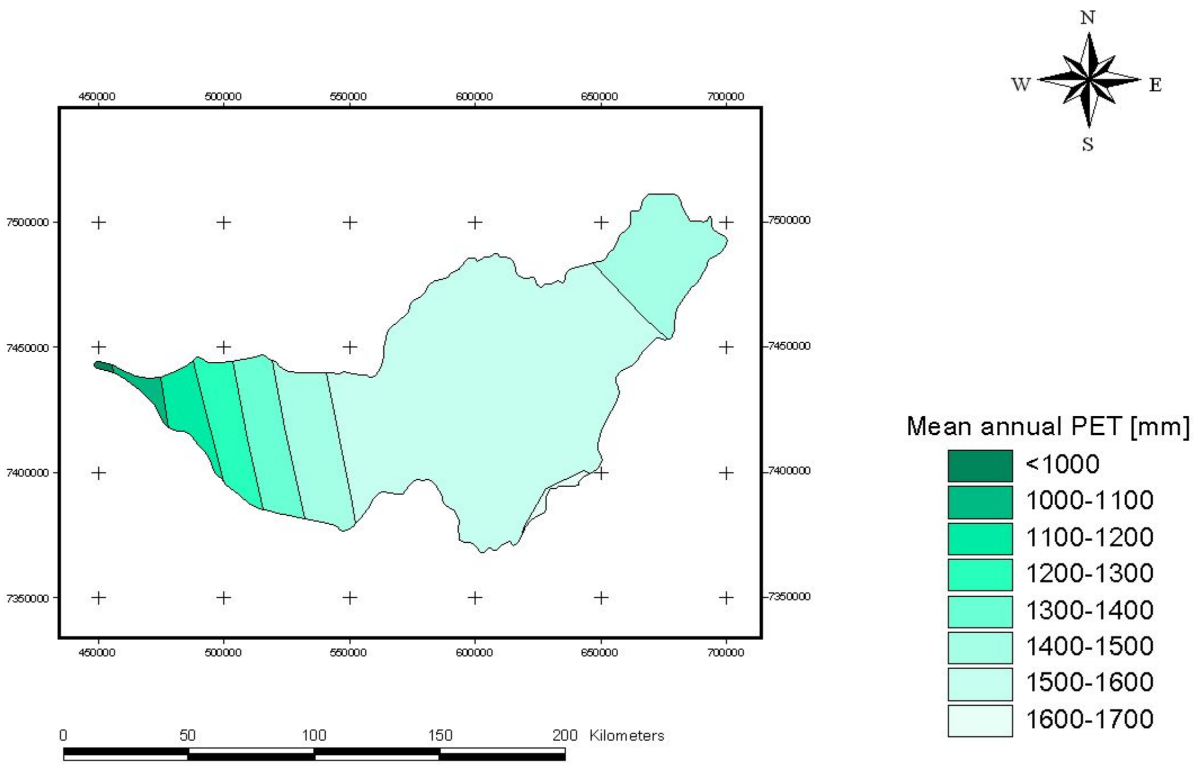

Figure 4.2.2: Distribution of mean annual PET in the Kuiseb basin

Figure 4.2.2 depicts the spatial distribution of mean annual PET, with PET at its highest in the central part of the Kuiseb basin.

4. 3. Geology

In general, the Namib Desert is underlain by Precambrian bedrocks including granites, gneisses and schists. Outcrops of these bedrocks can be found all over the central Namib region. (SCHMITZ, 2004).

According to JACOBSON (1997) the Kuiseb begins on the interior plateau of Namibia at about 2000 m above sea level, where the geology mainly consists of schists and sandstones. To the west of the headwaters the catchment geology is dominated by schists, sandstones and quartzites. After crossing the escarpment the geology is characterized by schists, granite, schists and dolomite, granite and sand and calcrete as can be seen in figure 4.3.1.

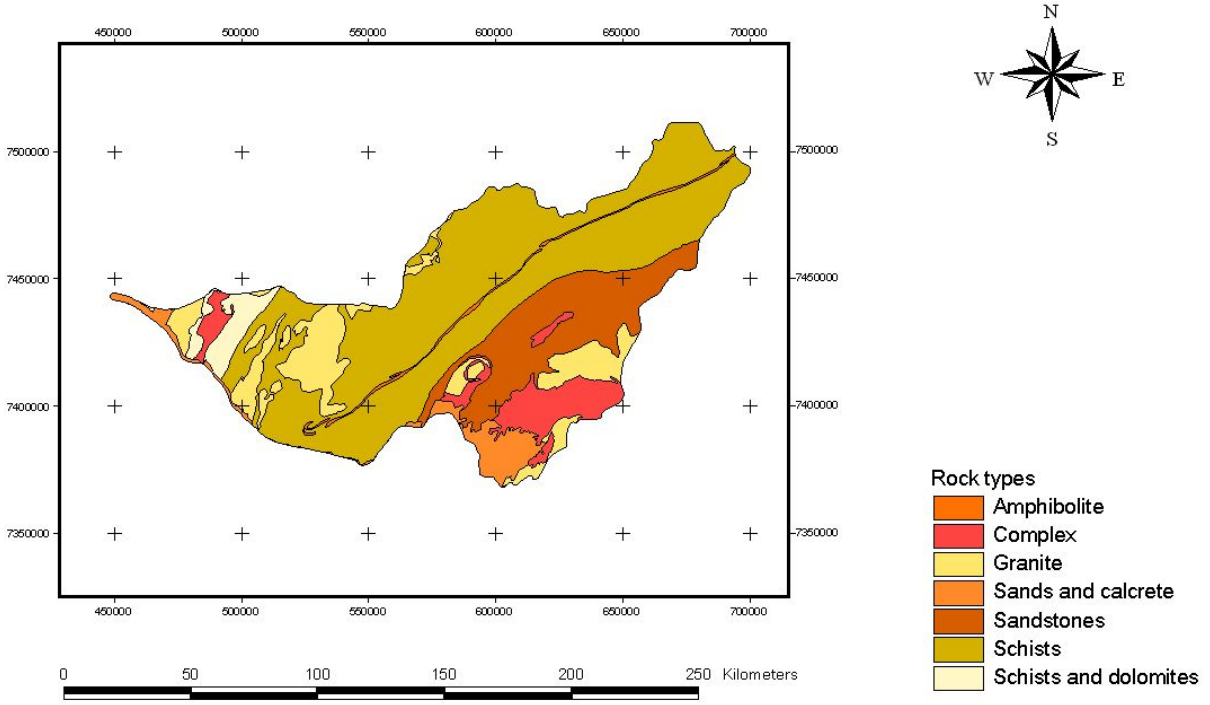

Figure 4.3.1: Geology in the Kuiseb basin, classified for rock types

The alluvial aquifer in the lower Kuiseb catchment consists of sediments of the Namib group. The aquifer is intermittent with its lateral extent restricted by basement outcrops on both sides of the channel. Its vertical extent is restricted by underlying bedrock material.

Recharge to the alluvial aquifer occurs by means of vertical percolation of flood water (transmission losses) and by through-flow within the alluvial aquifer itself (SCHMITZ, 2004). South of the riverbed, paleochannels are incised into the Tsondab-sandstone formation and into the basement.

4. 4. Soils

In the upper Kuiseb catchment the dominating soils are lithic Leptosols. The FAO (2007) characterizes Leptosols as very shallow soils over hard rock or highly calcareous material but also deeper soils which are extremely gravelly and/or stony. Leptosols are particularly common for in mountainous regions and show implicit low water holding capacities. (FAO, 2007). Leptosols are weakly developed soils, often with an incomplete solum. The term lithic implies that the soil profile is only 10 cm deep. Eutric Leptosols ,which arise only sparsely on the western boundary of the upper catchment (see figure 4.4.1), show soil depths between 20 and 50 cm and a base saturation > 50%.

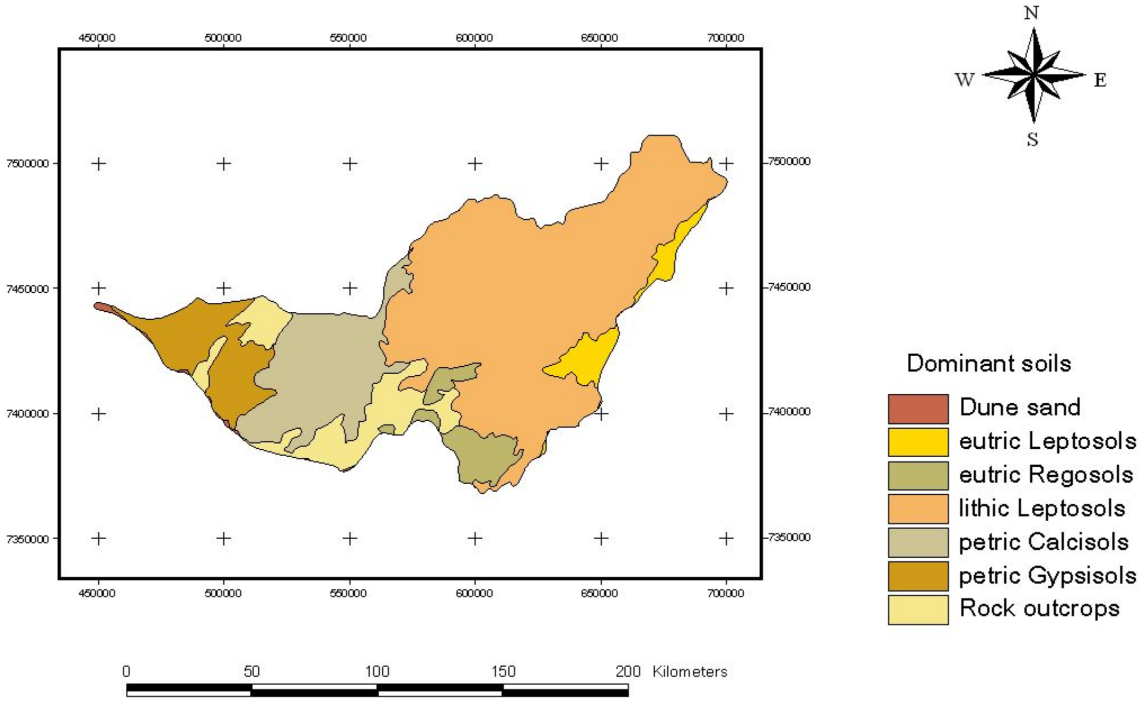

Figure 4.4.1: Dominant soils in the Kuiseb basin, classified for soil groups

As figure 4.4.1 depicts, the soils in the lower Kuiseb basin consist of eutric Regosols, petric Calcisols, petric Gypsisols, rock outcrops and dune sands.

Regosols are a taxonomic rest group. The FAO (2007)defines them as soils of some depth in unconsolidated material (excluding coarse textured materials with fluvic properties) which have no diagnostic horizon. Regosols are particularly common in arid areas, in the dry tropics and in mountain regions but can also occur in climates without permafrost at all elevations (FAO, 2007). The great variation among Regosols as a group makes it almost impossible to give a generalised account of Regosol characteristics. Low coherence of the matrix material makes most Regosols in sloping areas prone to erosion. The typical texture of Regosols is sandy to silty with high porosities resulting in small storage capacities and high permeabilities. Regosols which are exemplary for poorly vegetated areas typically display depths between 20 ad 50 cm and poor nutrient pools. Calcisols**,** which are generated by secondary limestone incorporations and show low depths and good drainage capacities, can also be found in the lower basin. The FAO (2007)defines Calcisols as soils which show substantial secondary accumulation of lime, or more precisely, Calcisols are soils which show a calcic or petrocalcic layer within the uppermost 100 cm. Usually Calcisols consist of alluvial, colluvial and aeolian deposits of base-rich weathering material. “Most Calcisols have a thin (<10 cm) brown or pale brown surface horizon over a slightly darker subsurface horizon and/or a yellowish brown subsoil that is speckled with white calcite mottles. The organic matter content of the surface soil is low, in line with the sparse vegetation and rapid decomposition of vegetal debris.” (FAO,2007).

The Gypsisols of the lower basin are characterized by a gypsic or petrogypsic horizon within 100 cm from the surface which accumulates by the precipitation of calcium and sulphate that percolates through the soil profile. As the Calcisols, Gypsisols mainly consist of unconsolidated alluvial, colluvial or aeolian deposits of base-rich weathering material (FAO,2007). Generally, Gypsisols feature a wide range of hydraulic properties. “Saturated hydraulic conductivity values vary from 5 to >500 cm/d. Infiltration of surface water is almost zero in severely encrusted soils. By contrast, very high percolation losses occur in soils in which dissolution of gypsum has widened fissures, holes and cracks to interconnected subterranean cavities.” (FAO,2007).

The rock outcrops are not soils in the common sense but hard rock or huge blocks of hard rock that occur on the terrain’s surface.

The distribution of the soil characteristics corroborate well with the fact that only the upper portions of the catchment, from the escarpment inland, contribute significant runoff to the lower reaches of the river in most years. (JACOBSON, 1997).

4. 5 Hydrology

On the reach between the headwaters and the escarpment, the Kuiseb has eroded a shallow, winding valley into the bedrock.

Coastward of the escarpment, the river has incised a deep, narrow canyon into the basement rocks. JACOBSON (1997) observed that within this canyon the river often runs on bare rocks and has no alluviation which is due to the steep gradient, on average 0.0034 m/m, and the narrow channel. At about 65 km from the coast the river valley broadens to occupy a wide, shallow valley which becomes indistinctive at about 20 km distance from the coastline.

JACOBSON (1997) argues that the ephemeral rivers crossing the Namib desert are unusual in that many of them are well-gauged, having had gauging stations in place since the 1960s, and for which a database has been accumulated managed by the Namibian Department of Water Affairs. He also found that the ephemeral rivers crossing the Namib Desert are among the most hydrologically variable fluvial systems yet described. The mean annual coefficient of variation for runoff (CV_{MAR}) among 28 stations representing 7 rivers crossing the Namib, averaged 1.55, compared to a global average of approximately 0.45.

The mean annual runoff volume of the Kuiseb River at Schlesien adds up to approximately 15 400 000 m³. The mean annual runoff volume at Gobabeb amounts approximately 10 600 000 m³. Thus, almost 5 000 000 m³ of water percolate into the channel alluvium every year. This underlines the importance of the Kuiseb River regarding to indirect recharge by means of transmission losses.

Only the upper portions of the catchment, from the escarpment inland, contributes significant runoff to the lower reaches of the river in most years. The coastal desert plain contributes very little runoff to the rivers apart from exceptionally wet years. (JACOBSON, 1997).

Figure 4.5.1 shows an image of the Kuiseb basin, the course of the river itself and its tributaries. Three gauging stations at the Kuiseb namely Swartbank, Gobabeb and

Schlesien and one gauging station at the Kuiseb`s main tributary, the Gaub River, are also depicted.

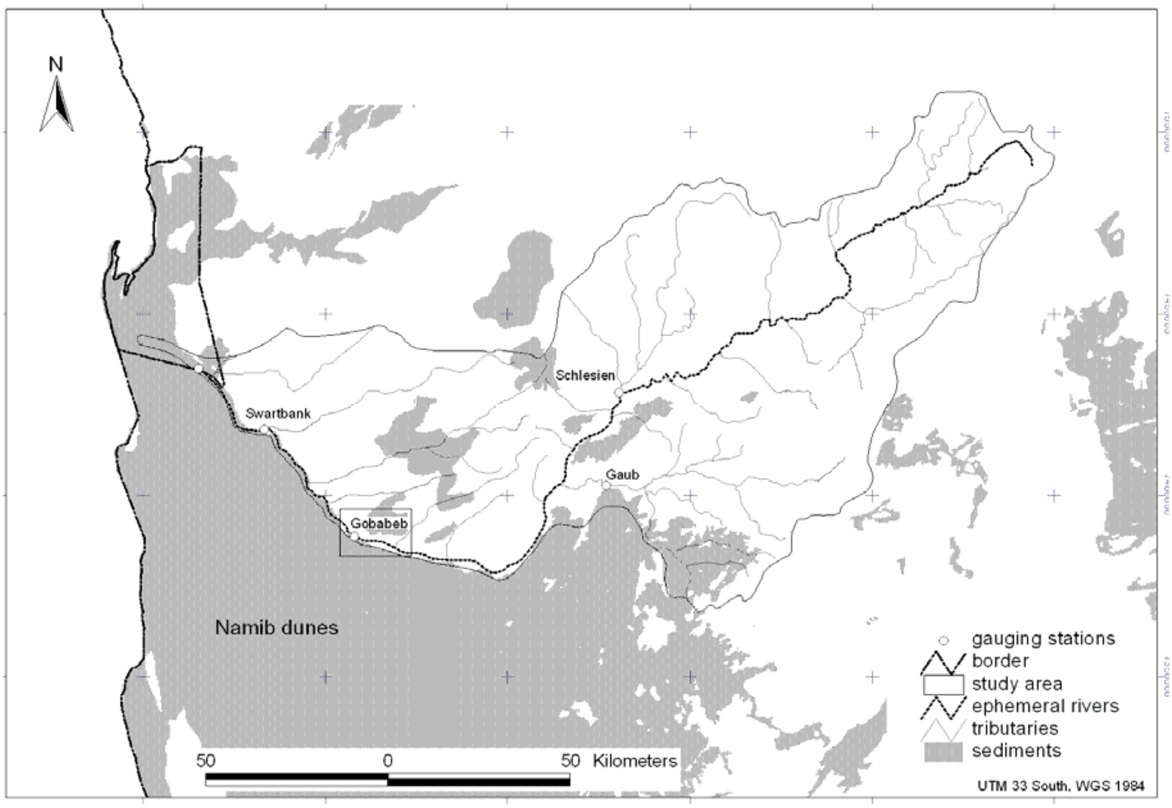

Figure 4.5.1: The Kuiseb basin with the Kuiseb River and its tributaries

4.6 Conclusion

The Kuiseb basin is located in a hyper-arid area near the Namibian coast. Annual rainfalls in the catchment range from almost zero at the coast up to 360 mm in eastern part. Rainfall events occur unpredictably and are highly variable over space and time.

With its 14 700 km² the Kuiseb basin can be seen as a typical, arid, macro-scale catchment. The arid character of the catchment is underlined by shallow, poorly developed soils. Most of the soils show low water holding capacities and infiltration rates can go down to zero if soils are encrusted.

The geology mainly consists of schists, granites and lacustrine sediments.

A special feature of the Kuiseb catchment is that is well gauged compared to other arid catchments.

5 Preprocessing

5.1. Digital Elevation Model (DEM)

Topographic information of the entire catchment, available as an existing DEM at the Institute of Hydrology Albert-Ludwigs University Freiburg (IHF) has been re-projected in UTM. The chosen cell size for the DEM is 1000 m *1000 m, although finer spatial resolutions were available. However, finer resolution would have made model runs costly in terms of computing time.

In e2, the stream network can be defined with the help of a DEM or a catchment map.

"The "DEM based" method requires a DEM, and divides a catchment into sub-catchments based on a user-specified measure of upstream area, usually for first order streams." (ARGENT et al., 2004/2005).

In this case, under the DEM based method, a number of nodes are used to define the sub-catchments (ARGENT et al., 2004/2005).

The nodes have to be loaded in a pre-defined format (e.g. .MIF, .SHP, .tsd).

For this purpose, two shapefiles were created. The first shapefile includes nodes at the roughly estimated locations of the gauging stations of Rooibank, Swartbank, Gobabeb, Schlesien, and additional nodes at the junction of the Kuiseb River and the Gaub River plus one node in the uppermost Kuiseb catchment. The second shapefile only has one node at the main outlet of the Kuiseb catchment.

After setting the stream threshold to a large value of 500 km^2, 33 sub-catchments were automatically defined by the model.

The designation of sub-catchments in the model, and the related nodes are shown in Annex 2.

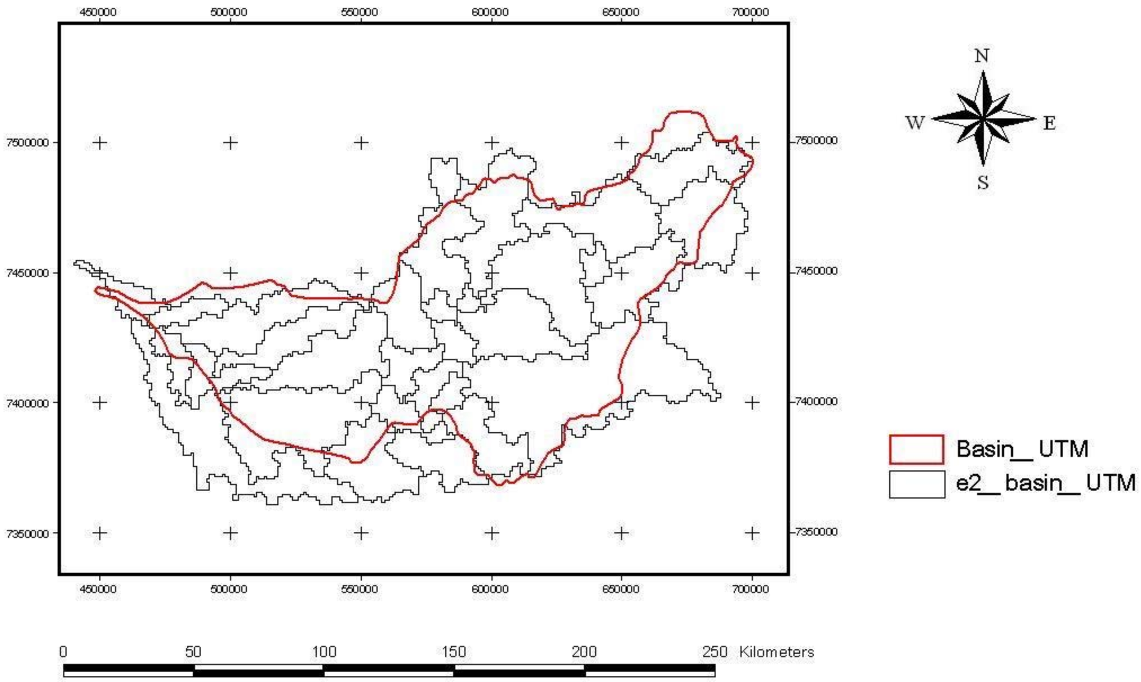

Figure 5.1.1: Comparison of real catchment and catchment computed by E2.

This however resulted in a computed catchment area of 19 329 km^2, compared to the real catchment size of 14 300 km². In figure 5.1.1, the Kuiseb catchment is shown red, compared to the computed catchment displayed in black. Near the estuary mouth, the model calculated a sub-catchment with its tributary to the south of the Kuiseb channel. This area is effectively covered by sand dunes and thus can not yield any tributary flow.

5.2. Land Use and soils

Within the e2 scenario wizard all possible types of functional units, which occur in the catchment of consideration, have to be defined. The following functional units were defined with the aid of a land-use map:

1. Namib grassland
2. Sparse grass & shrubland
3. Sparse shrubland
4. Riparian vegetation

5. Riverine woodland
6. Mountain shrubland

In the next step, areas must be assigned to functional units in each sub-catchment. The manual approach was rejected so the areas were assigned using the available land-use raster. To do this, the land-use raster must contain codes which match the types of functional unit in the model. This had already been taken into account when the land-use map was generated. A table allows the codes in the raster to be matched to the types of functional unit.

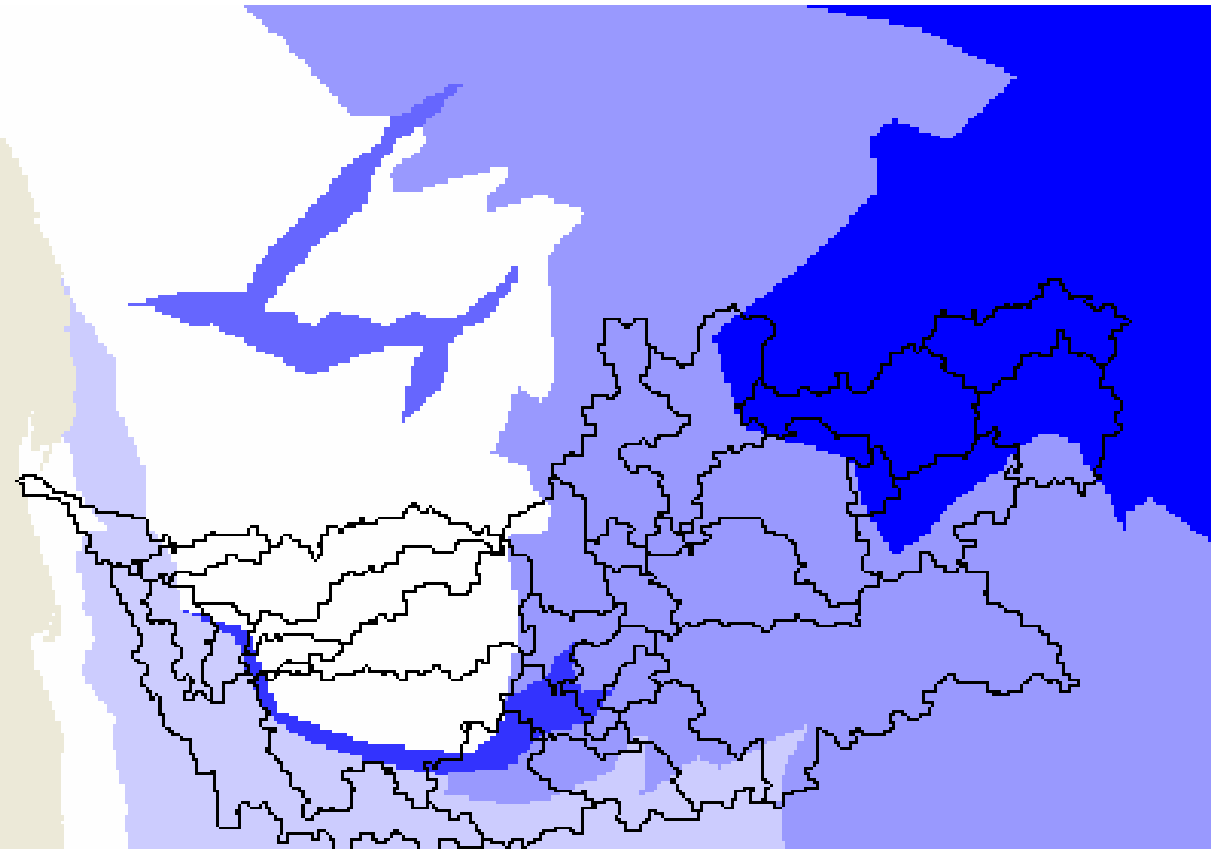
Figure 5.2.1: Computed Kuiseb catchment and assigned types of functional units

The functional units in figure5.2.1 match the codes in the list shown above.

5.3 Rainfall Regionalization

All rainfall-runoff models in the e2 modelling framework require continuous and overlapping daily time series of precipitation for every sub-catchment.

Existing rainfall data for the Kuiseb catchment consists of four stations (Tantus, Middelplaas, Rostock and Schlesien), none of which have recorded precipitation continuously during past decades.

This makes it necessary to regionalize point measurements of rainfall into sub-catchment rainfall values. The location of the four stations is shown in figure 5.3.2, which depicts the rainfall gauges as grey points.

5.3.1 General

The density of rainfall gauging networks in arid regions is often low. The WMO recommends a gauging network density of about 10000 km² per station for rainfall gauging in arid regions. The number of monitoring stations is high, compared to other arid basins. Nevertheless, having only four rainfall stations for continuous long-term monitoring of rainfall gives sparse coverage when we consider the high spatial variability of rainfall events. Small convective storm cells may cross the catchment without being recognized by the monitoring network at all.

Thus, regionalizing rainfall without the aid of rainfall radar is likely to yield poor results regarding the spatial resolution and occurrence of rainfall events.

5.3.2 Timeframe for Regionalization

The time series that provides the basis for rainfall regionalization was reduced to a phase during which at least two of the four stations were recording simultaneously. This period was between the 1st October 1952 and the 30th April 1983. The total of recording stations between October 1951 and October 1987 is shown in figure 5.3.1.

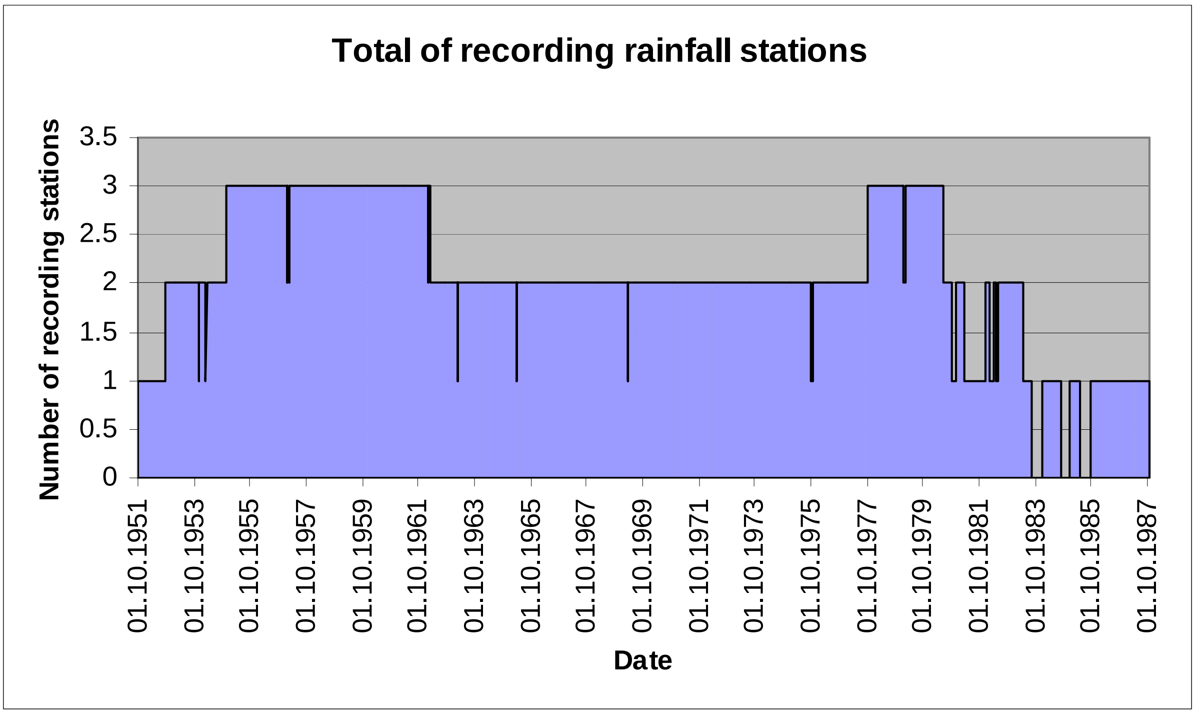

Figure 5.3.1: Total of recording rainfall stations from October 1951 to October 1987

5.3.3 Methodology

In the next step, a dot matrix including a total of 315 points was superimposed on the catchment area. Precipitation from the four recording stations was then regionalized for every point in the dot matrix. The dot matrix is shown in Annex 1.

At times when only one station was recording, a rainfall decay function was applied to describe the decline in rainfall values with increasing distance from the recording station. The rainfall decay function describes a decrease in rainfall volumes with increasing distance of the gauging station of interest, according to a normal distribution.

In cases when two or more stations were recording the inverse distance weighted method is applied.

In general, Inverse Distance Weighting (IDW) is a method for multivariate interpolation, or in other words, a process of assigning values to unknown points by using values from known points. IDW methods are based on the fundamental assumption that the interpolating surface is influenced most by nearby points and less by more distant points. Hence, the interpolating surface is a weighted average of the scatter points and the weight assigned to each scatter point diminishes as the distance from the interpolation point to the scatter point increases.

In a first step, the recorded rainfalls were transferred to sea level elevation in accordance with the formula:

$$P_{Sea\ level} = P_{Station} + (2000\ m – Station\ elevation) * MF$$ **equation 5.3.3.1**

The monthly factors for correcting rainfall for altitude are obtained by plotting mean monthly precipitation values for the different stations against their elevation for each month. Then, a regression is performed, with the slope of the regression line representing the monthly factor MF.

The next step is to convert rainfall into area precipitation by interpolating the rainfall at gauges sites (sea level) on to the dot matrix points, and then to estimate area precipitation by averaging the interpolated rainfall at grid points within the catchment. Interpolation is achieved by application of the inverse distance-squared method according to MAIDMENT (1993).

The estimate for the j^{th} grid point then is:

$$P_j = a * \Sigma\ d_{ij}^{-2} * P_i$$ **equation 5.3.3.2**

Where d_{ij} is the distance from gauge i to the grid point j and a is the inverse of the sum of the inverse distance-squared values for all gauges:

$$a = (\Sigma\ d_{ij}^{-2})^{-1}$$ **equation 5.3.3.3**

This can also be expressed as:

$$P_{IDW} = (P_1/d_1^2 + P_2/d_2^2 + P_n/d_n^2) / (d_1^{-2} + d_2^{-2} + d_n^{-2})$$ **equation 5.3.3.4**

Where P_{IDW} is the resulting precipitation in mm at a grid point, $P_{1,2,...n}$ are the precipitation values in mm at the recording gauges, and $d_{1,2,....n}$ are the relevant distances between each gauge and the calculated grid point.

The distances between all the four recording stations and each scatter point on the dot matrix were calculated using Arcview GIS 3.3.

The resulting precipitation, regionalized to sea level, now has to be adjusted to the real elevations at each point within the dot matrix. The elevation above sea level of each grid point was calculated with Arcview GIS 3.3.

Adjusting the regionalized precipitation to the real altitude of the grid points is carried out using the formula:

$$P_{realaltitude} = P_{IDW} – (2000\ m – Station\ elevation) * MF \quad \textbf{equation 5.3.3.5}$$

Where $P_{realaltitude}$ is the adjusted precipitation in mm at a grid point.

The final precipitation for each sub-catchment area is calculated by generating the arithmetic mean of all grid point values located in the sub-catchment of interest:

$$P_{Sub\text{-}catchment} = 1/n * \Sigma P_i \quad \textbf{equation 5.3.3.6}$$

Where $P_{Sub\text{-}catchment}$ represents the area precipitation in mm, n is the number of grid points located within this sub-catchment, and P_i is the precipitation at each grid point.

Four sub-catchments near the catchment outlet in the west are not covered by the dot matrix on which regionalization is based. According to ARGENT (2006), missing records are typically filled by neighbouring stations. The selection of the best neighbouring site can be on the basis of proximity or correlation within the wettest months. Thus, area precipitation for these four sub-catchments was assumed to be best represented by the precipitation of the nearest raster points.

5.3.4 Results

Mean annual rainfalls for sub-catchment time series range from 27.26 mm/a in sub-catchment 4 which is located in the west to 79.9 mm/a in sub-catchment 11 in the east. Mean annual rainfalls at the rain gauges range from 61 mm/a at Rostock to 233.8 mm/a at Tantus. Hence, regionalized rainfall reflects the increase in annual rainfall volumes in the catchment from west to east. Figure 5.3.2 shows the distribution of relative rainfall amounts for a rainfall event on 3[rd] May 1980.

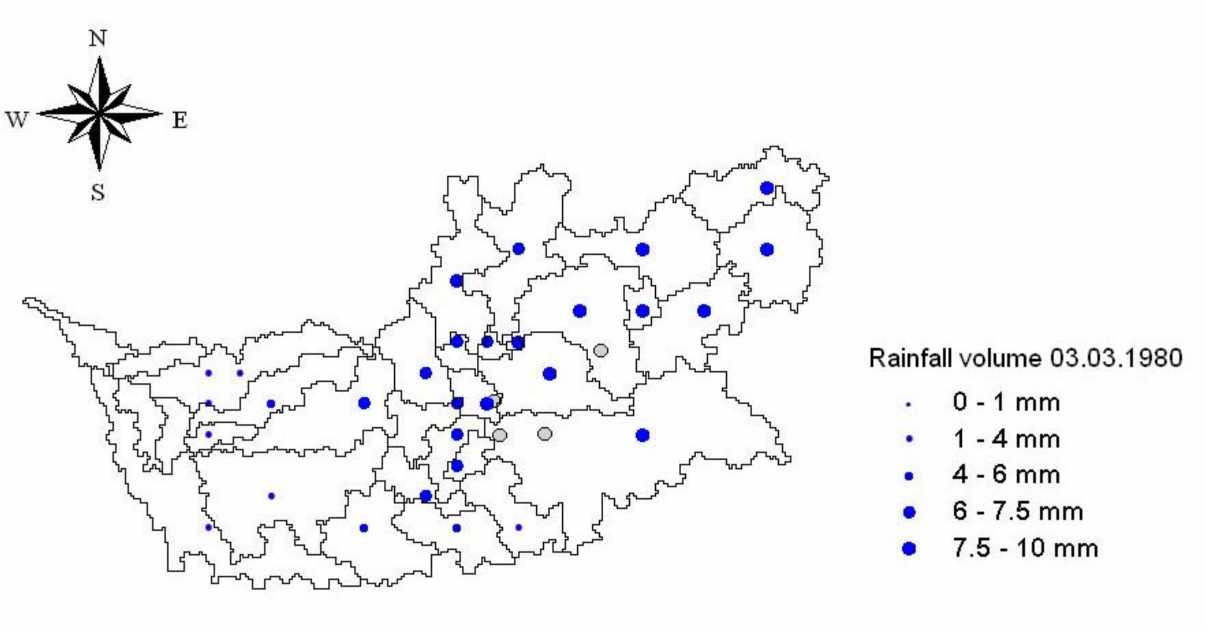

Figure 5.3.2: Spatial distribution of regionalized rainfall on 03.03.1980; recording rainfall gauges in grey.

The size of the blue points indicates the amount of rainfall; grey points represent the four recording rainfall gauges.

5.4 ET-Regionalization

Most of the component models within e2 for rainfall-runoff modelling require continuous daily time series for PET. ARGENT (1997) argues that, in the context of rainfall-runoff modelling, the area potential evapotranspiration, rather than point potential evaporation should be used. He also points out that the inter-annual variability of PET is relatively low compared to rainfall and that the day to day variation in PET has little influence on the water balance on a daily time scale.

Available data consisted of two time series; one time series included monthly values of temperature and vapour pressure for Windhoek over a period of nineteen years (1961 to 1980), another time series provided approximately seventeen months of daily ETP values for Gobabeb.

5.4.1 Methodology

The conceptual idea was to generate three “ET-stations” to represent the strong climatic and topographic gradient that runs through the catchment from west to east. The intention was to divide the catchment into three zones with each zone having its area PET represented by one of the stations.

In the course of this, the Gobabeb PET-series represents the conditions in the Namib at lower altitudes, but had to be extrapolated in time to provide a sufficiently long time series for modelling.

Windhoek ET stands for the mountainous fraction of the headwaters. Monthly data from Windhoek however had first to be disaggregated to daily values, and then modified to fit the time intervals of rainfall inputs (1952 to 1983).

A fictive station with a synthesized time series of daily ET represents the transition between Gobabeb and Windhoek.

Generation of each of the three ET data sets is done by fitting a sine function to the annual PET-regimes and then extrapolating these data sets in time. To allow for annual variation within the extrapolated data sets, a scattering around the values of the sine function is achieved by letting standard deviations vary according to a normal distribution.

5.4.2 Gobabeb dataset

Daily potential evapotranspiration for Gobabeb from the 1st November 2002 to the 15th April 2004 had already been calculated, applying multiple approaches (Blaney-Criddle, Thorntwaite, Turc, Hargreave, Penman). A comparative graph plots all approaches over time, indicating that annual PET has a sinusoidal regime in Gobabeb. The comparative plot is shown in figure 5.4.1:

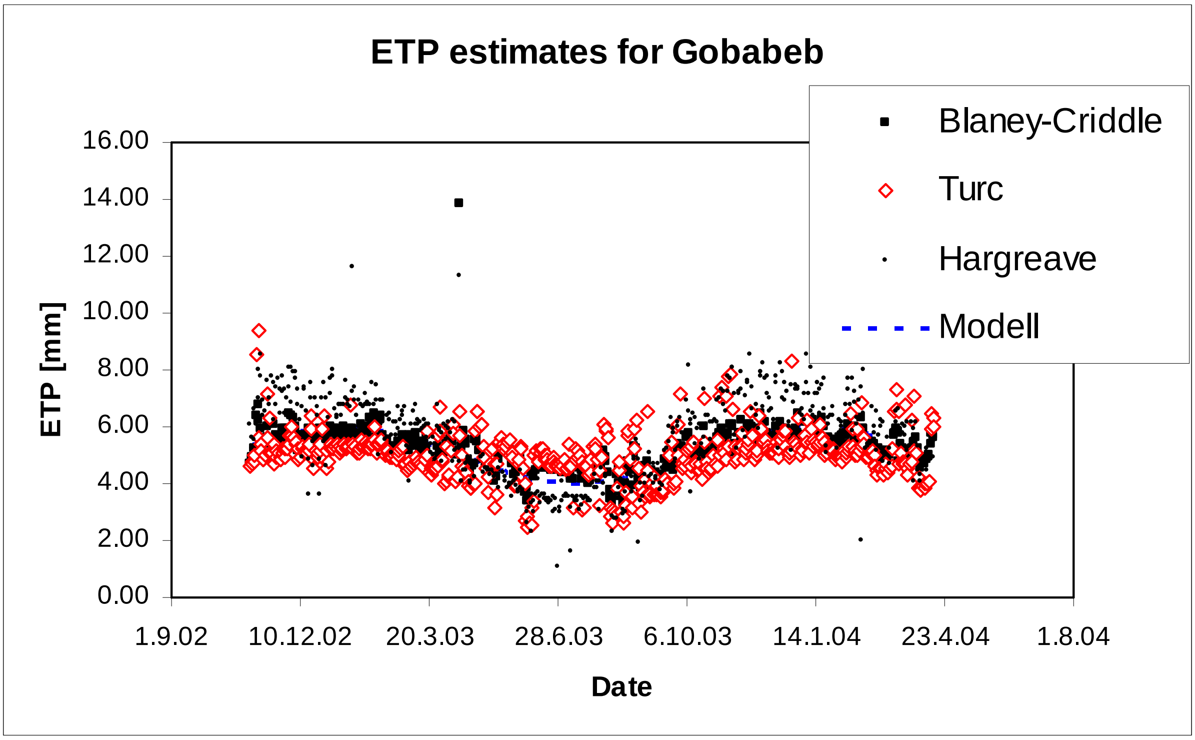

Figure 5.4.1: Comparative plot of different methods for the calculation of PET at Gobabeb

A sine function is chosen to fit the point cloud representing the annual PET regime at Gobabeb. The notion of taking the sine function as annual regimes for each of the years modelled was rejected, and it was decided to allow for some variation of the annual sine function to mimic inter-annual variability. The variation of the sine function is achieved by allowing the standard deviation of the annual ET regime to vary according to a normal distribution. The density function of the standard normal distribution is shown in figure 5.4.2.

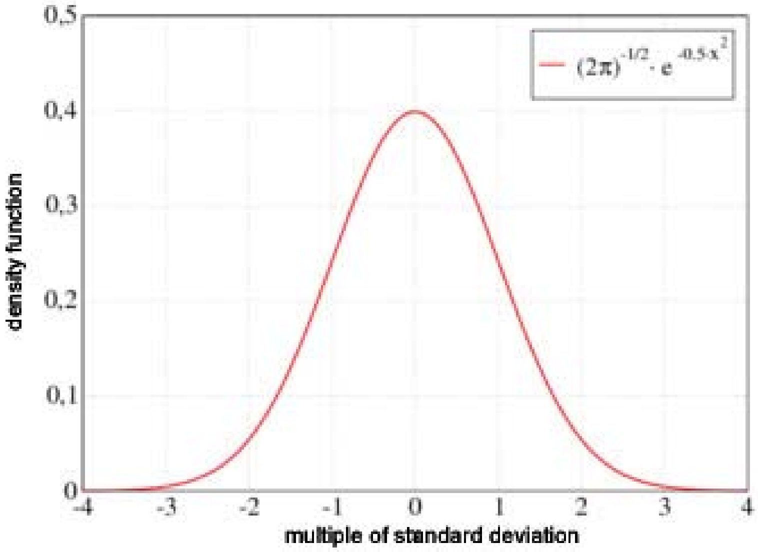

Figure 5.4.2.2: Density function of the standard normal distribution

The function that describes the density function shown in figure 5.4.2.2 is:

$$Y = 1 / \sqrt{2 * \Pi} \; * \exp(-0{,}5 * x^2)$$ **equation 5.4.2.1**

, and dissolved for x gives:

$$x = +/- \sqrt{-2 * \ln(y * \sqrt{2\prod})}$$ **equation 5.4.2.2**

Y-values are chosen at random for a range between y>0 and y<0.4. This results in two outputs, one positive and one negative, one of which is picked at random for further calculation.

In the next step, the function shown above is “cut” to avoid unrealistic small and high values of PET, forcing x-values into range between -1.5 and 2.

Such generated x-values are then multiplied by the monthly standard deviations calculated from the Penman time series for Gobabeb, to yield “new” monthly standard deviations which vary according to the normal distribution. Adding these “new” monthly standard deviations to the PET-values yielded by the sine function model, results in a stochastic scattering of PET-values around the sine function for Gobabeb.

5.4.3 Windhoek dataset

Monthly data of temperature and vapour pressure in Windhoek had to be transformed into monthly ET-values and then disaggregated into a daily time series. In a first step, saturated vapour pressure is calculated using the Magnus formula:

$$E_{(t)} = E_0 * \exp (c_1 * t / c_2 + t)$$ **equation 5.4.3.1**

Monthly potential evapotranspiration was calculated according to TURC (1961), using monthly means for temperature and vapour pressure:

$$PET_{TURC} = 0.0031 * C * (R_G + 209) * (T/(T+15))$$ **equation 5.4.3.2**

Where $C = 1+((50-rH)/70)$, for $rH < 50\%$ or $C=1$, for $rH > 50\%$

rH = Mean Monthly relative Humidity [%]

R_G = Mean Monthly global radiation [J/cm²]

T = Mean Monthly Temperature [°C]

The annual regime of mean monthly PET-rates, calculated from the available time series (1961 to 1980), indicates that annual PET regimes can be described by a sine function, too.

Figure 5.4.3 depicts the regime of monthly means for calculated PET. It is evident that the annual regime is again best presented by a sine function.

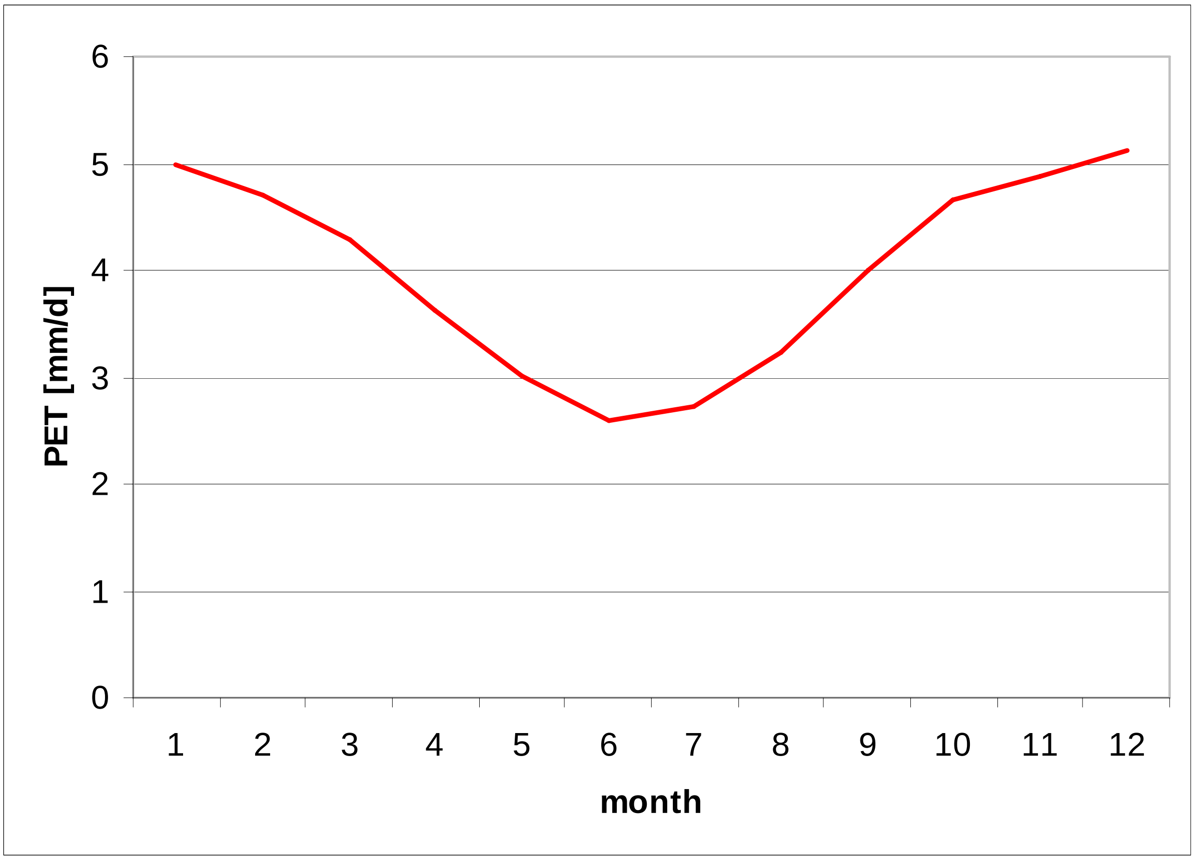

Figure 5.4.3: Regime of mean annual PET for Windhoek, calculated according to Turc

A final daily PET time series for Windhoek was then generated as described above for the time series at Gobabeb.

Owing to the lack of daily time series for PET at Windhoek, a different method had to be used to calculate the monthly standard deviation for the available dataset. Thus, the assumption is made that the ratios of monthly means for PET between Gobabeb and Windhoek also reflect the ratio of the monthly standard deviations for the two stations:

$$PET_{Windhoek}/PET_{Gobabeb} = \delta_{Windhoek}/\ \delta_{Gobabeb} \qquad \textbf{equation 5.4.3.3}$$

The resulting monthly standard deviations for Windhoek are then multiplied by the results of the density function of the standard normal distribution as already described in the previous section, creating “new” standard deviations which, again, show daily variations following the normal distribution.

In a last step, these "new" standard deviations were added to the daily values for PET which result from the sine function model for Windhoek, yielding final daily PET values required by the e2s rainfall runoff models.

5.4.4 Fictive station dataset

The dataset for the fictive station, which is situated approximately in the middle of the catchment, was generated by simply calculating the altitude gradient for PET.
Daily PET values for Windhoek (1686 m.a.s.l.) were subtracted from daily PET values for Gobabeb (461 m.a.s.l.) and then divided by the difference in altitude (1225 m.a.s.l.) between the two stations:

$$Gradient_{PET} = (PET_{Gobabeb} - PET_{Windhoek}) \,/\, \Delta h \qquad \textbf{equation 5.4.4.1}$$

Where $PET_{Gobabeb}$ is the calculated daily PET for Gobabeb [mm/d], $PET_{Windhoek}$ is the calculated daily PET for Windhoek [mm/d] and Δh is the difference in altitude [m].
The fictive station itself is located at an elevation of 885 m.a.s.l.. The final daily PET values are provided by the equation:

$$PET_{Fictive} = PET_{Gobabeb} - (Gradient_{PET} * (h_{Fictive\ station} - h_{Gobabeb}) \qquad \textbf{equation 5.4.4.2}$$

or alternatively:

$$PET_{Fictive} = PET_{Windhoek} + (Gradient_{PET} * (h_{Fictive\ station} - h_{Windhoek}) \qquad \textbf{equation 5.4.4.2}$$

Where h is elevation above sea level [m].

5.4.5 Conclusions

An important requirement of E2 models are catchment characteristics. Terrain characteristics are usually incorporated in terms of a digital elevation model (DEM). A DEM with a spatial resolution of 1000 m*1000 m which was available at the IHF was reprojected in UTM. The E2 model delineated 33 sub-catchments from the DEM, enclosing a total catchment area of 19 329 km². Compared to the actual catchment area of 14 300 km² this corresponds to a 35,17% overestimation of catchment area. However, e2 nicely reproduced the actual stream network of the basin.

The delineation of functional units in the e2 modelling framework is usually based upon another catchment characteristic: the types of land-use. For this purpose a land use raster was created, including six different land use types. Even if land use information is not directly used to differentiate between types of functional units, the type of land use will influence surface runoff characteristics, evapotranspiration rates and interception losses, and is thus worth consideration.

According to ARGENT et al. (2004/2005) the most important step in calibrating any hydrological model is data preparation. Therefore, the best possible data sets will speed up the calibration process. Input data to hydrological models has in most cases to be interpolated in some way prior to use.

Time series of four rainfall gauges were available for the regionalization of precipitation. The deterministic interpolation method of choice is the inverse distance weighted or inverse distance squared method. Results indicate that the spatial distribution of rainfall quantities within the basin is satisfactorily represented by regionalized rainfalls.

Due to low data availability, a method had to be found to create continuous daily time series of PET for area evapotranspiration in the Kuiseb basin over the modelling timeframe.

Following the climatic gradient from west to east, the basin was subdivided into three lumped zones with similar PET inputs.

It had proved that annual ET-regimes at Gobabeb can be described by a sine function. A plot of the annual regime of mean monthly PET-rates, calculated over a period of 19 years, indicates that the annual regime at Windhoek can also be described by a sine function. The sine function adjusted for Windhoek PET shows similar amplitude and a lower base compared to Gobabeb.

Results of the PET-regionalization represent the climatic gradient in the basin with decreasing PET-rates from west to east.

6 Calibration

6.1 Introduction

Model calibration is the process of choosing the “best” set of parameter values. It is a process of optimising a model’s parameter values to improve the “fit” between observed and simulated catchment processes.

The chosen approach to calibration is to consider a proportion of the available data for calibration and use the remainder for verification. This also offers a way to assess model performance outside the calibration period.

The “fit” is assessed visually by comparing measured and modelled hydrographs. An additional comparison of the quality of different model results can be carried out using the model efficiency according to NASH & SUTCLIFFE (1970) which is calculated in accordance with the formula:

$$R_{eff} = 1 - \frac{\sum (Qobs - Qsim)^2}{\sum (Qobs - \overline{Qobs})^2} \qquad (-\infty < R_{eff} \leq 1) \qquad \textbf{equation 6.1.1}$$

Where R_{eff} is the model efficiency, Q_{obs} is the measured runoff at a modelling time step, Q_{sim} is the simulated runoff at a modelling time step and $\overline{Q_{obs}}$ is the mean of all measured runoffs throughout the time of modelling.

A perfect fit between measured and modelled hydrograph consequently leads to a model efficiency of R_{eff}=1.

Within the calibration no routing model is applied (straight through routing in e2).

6.2 Flow data for calibration

Rainfall runoff models are calibrated against flow data. “In calibrating against flow data the assumption is made that these data have no errors.” (ARGENT et al., 2004/2005). Flow data for calibration of the headwaters consists of a time series of approximately 9 years (2.11.1962 – 16.02.1971) observed at the gauging station at Schlesien.

6.3 SIMHYD

The SIMHYD parameters are introduced in section 3.7.1.4.

6.3.1 Sensitivity analysis

A worst case scenario was performed manually prior to the calibration. Worst case scenarios are sensitivity tests which are performed to find out about how sensitive the model is to certain parameters. "This is useful to understand how the model functions and also what parameters need more attention than others. If the model is significantly affected by a particular parameter, then the focus of calibration should be on that parameter." (PODGER, 2004).

The worst case scenario was performed by setting one parameter to its upper and lower limits in turn, while the remaining parameters were kept at their default values. The spread in results for each parameter is then regarded as a measure of how severely the model reacts to a particular parameter, i.e. as a measure of that parameter's sensitivity. In advance, the baseflow coefficient (K) and interflow coefficient were set to zero to take into account the hydrological characteristics of the Kuiseb basin, where neither baseflow nor interflow are likely to occur.

Results of the worst case scenario indicate that the model is most sensitive to the soil moisture storage capacity (SMSC), infiltration shape (SQ), infiltration coefficient (COEFF) and pervious fraction (PERV). Rainfall interception storage capacity (RISC), Impervious threshold (ImpT) and recharge coefficient (CRAK) had minor impacts on model outputs and can therefore be regarded as less sensitive.

However, in many rainfall-runoff models the behaviour of many parameters is closely linked to the values of other parameters, i.e. the models are non-linear. Hence, the sensitivity of particular parameters may be dependent on the values of the other parameters.

The Rainfall Runoff Library RRL, another toolkit product from the CRC for Catchment Hydrology offers an additional option for sensitivity analysis. Both, SIMHYD and AWBM are included in RRL. Various objective functions are available in RRL to display sensitivity analysis for single parameters. The chosen objective function is the sum of the squares of errors; results for SMSC, SQ, and COEFF are shown in Annexes 3 to 5.

6.3.2 Calibration with RRL

Exploration of parameter sets and single site calibration for the headwaters is achieved using RRL.

Calibration in RRL can be done manually or automatically. Automatic calibration in RRL offers seven optimisation algorithms of which the genetic algorithm was chosen. The primary objective function applied to the automatic calibration procedure is the “Nash-Sutcliffe criterion” or Coefficient of efficiency R_{eff} (see **equation 6.1.1**). The secondary objective function of choice is “Runoff difference in %”.

Automatic calibration yielded final results of R_{eff}=0.128 on a daily basis and R_{eff}=0.602 on a monthly basis. Table 6.3.1 shows different interim findings during the calibration process. The parameter sets are shown in columns. Set 1 to set 4 were calibrated automatically, and set 5 shows final parameter values after manual calibration.

Table 6.3.1: Interim and final results of SIMHYD calibration with RRL

Parameter	set 1	set 2	set 3	set 4	set 5
Baseflow coefficient	0.396	0.498	0.498	0.396	0
Impervious threshold	3.1	4.4	4.4	3.1	5
Infiltration Coefficient	251	240	240	251	250
Infiltration shape	5.2	0.6	0.6	5.2	10
Interflow coefficient	0.027	0.388	0	0.027	0
Pervious fraction	0.98	1	0.99	0.99	0.93
RISC	2.7	2.9	2.7	2.7	2
Recharge coefficient	0.29	0.38	0.38	0.38	0
SMSC	435	433	50	50	100
R_{eff} (daily)	0.027	0	0.093	0.128	-0.696
R_{eff} (monthly)	0.327	0.25	0.571	0.602	0.614

Final parameter values in RRL (set 5) were manually calibrated on the basis of two different assumptions:

1. Daily R_{eff} cannot exceed 0.2. Hence the focus is on an optimum monthly $R_{eff.}$
2. A conceptual hydrological understanding of the Kuiseb catchment has to influence the choice of parameter values.

In accordance with the latter assumption, baseflow and interflow coefficients are again set to zero. The recharge coefficient is also set to zero because the occurrence of direct recharge is regarded as implausible.

Focusing on optimum monthly results leads to an increase in SMSC and to a decrease in pervious fraction. Temporarily, monthly Reff values of about 0.79 are obtained by setting pervious fraction down to 0.9. However, visual calibration indicates that this leads to a clear overestimation of peak flows. Hence, this parameter set is rejected. Parameter values found in set 5 (see table 6.3.1) indicate to be the best compromise between best monthly Reff and visual fit of the observed and calculated flows. Observed and calculated flows resulting from parameter set 5 are shown in figure 6.3.1.

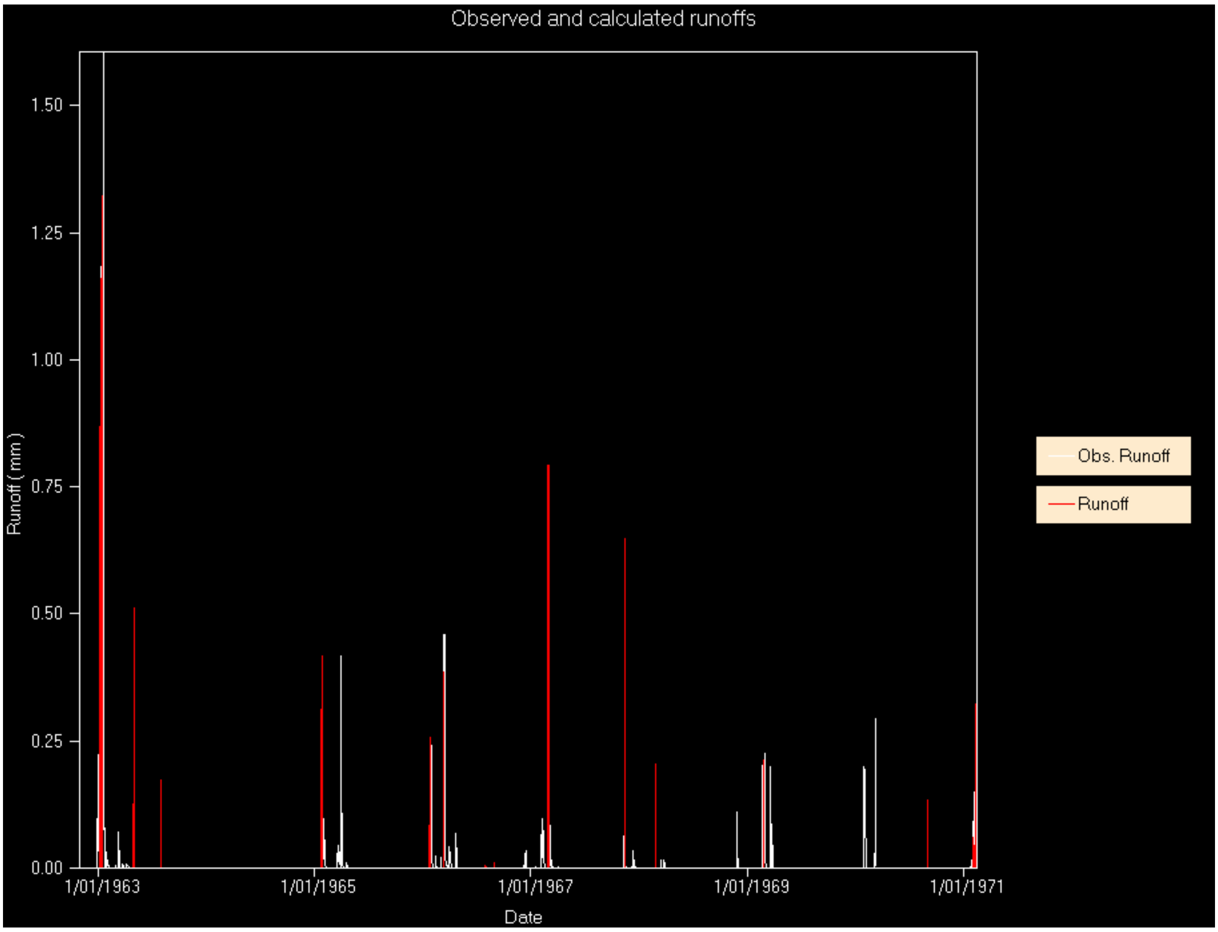

Figure 6.3.1: Observed and calculated runoffs for the calibration period in RRL

Comparison of computed and measured hydrographs in figure 6.3.1 shows that the model tends to overestimate small runoff events. Temporal occurrence of runoff events is also incorrect, but can be explained by the lack of a routing routine.

6.3.3 Calibration with E2

For calibration, the flexible calibration tool implied in e2 can be applied. Within the calibration tool either sub-catchment flow or link flow can be calibrated. By calibrating sub-catchment flow, sub-network selection can be done manually, i.e. only a portion of the catchment is calibrated. The network is cut below the node which represents the gauging station at Schlesien. Therefore only the upstream area from this node is the object of further calibration. Figure 6.3.2 shows the reduced network of the calibration tool with a crop made below the node representing Schlesien. The remaining sub-network now consists of 12 sub-catchments.

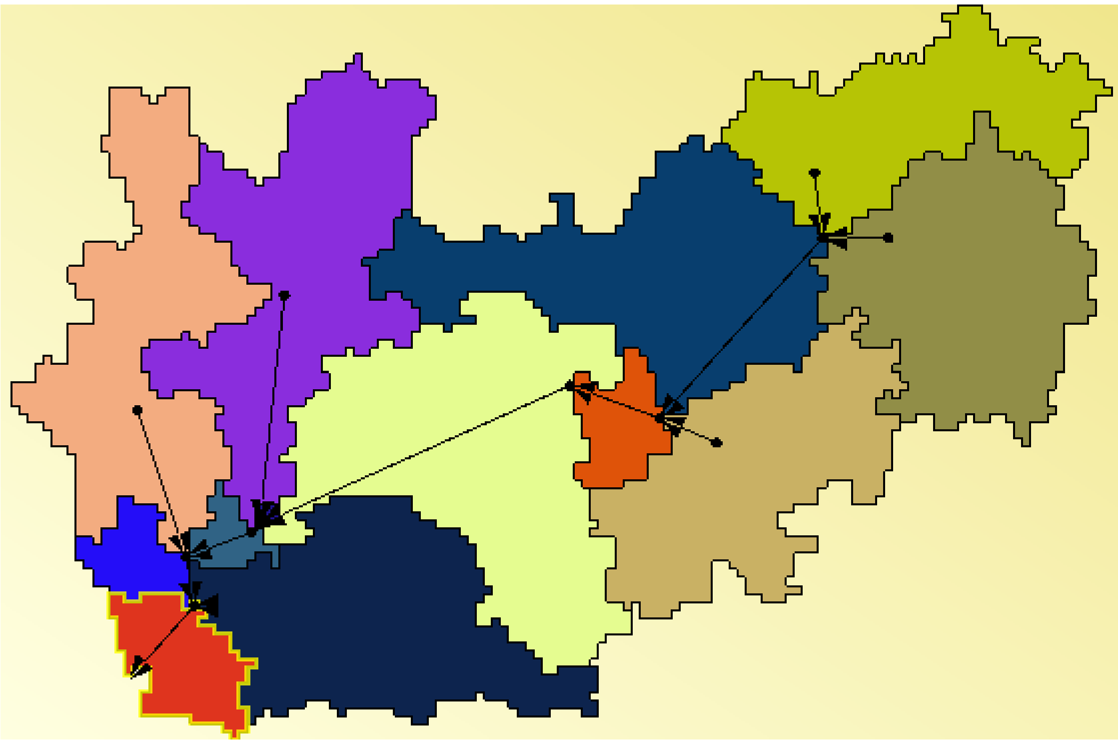

Figure 6.3.2: Sub-network of E2s calibration tool. Crop done below the node representing Schlesien

Subsequent grouping of parameters allows for different groups within every single FU in every sub-catchment. A sub-division according to FUs is not performed; consequently model parameters are solely grouped according to sub-catchments. Results from the calibration tool are then transferred back to the actual model.

Initially parameter values obtained by calibration in RRL (set 5 in table 6.3.1) are transferred to the e2 model.

Figure 6.3.3 depicts the hydrograph measured at Schlesien and the hydrograph produced by e2. Daily R_{eff} for this simulation is -0,169 whereas monthly R_{eff} is 0,289.

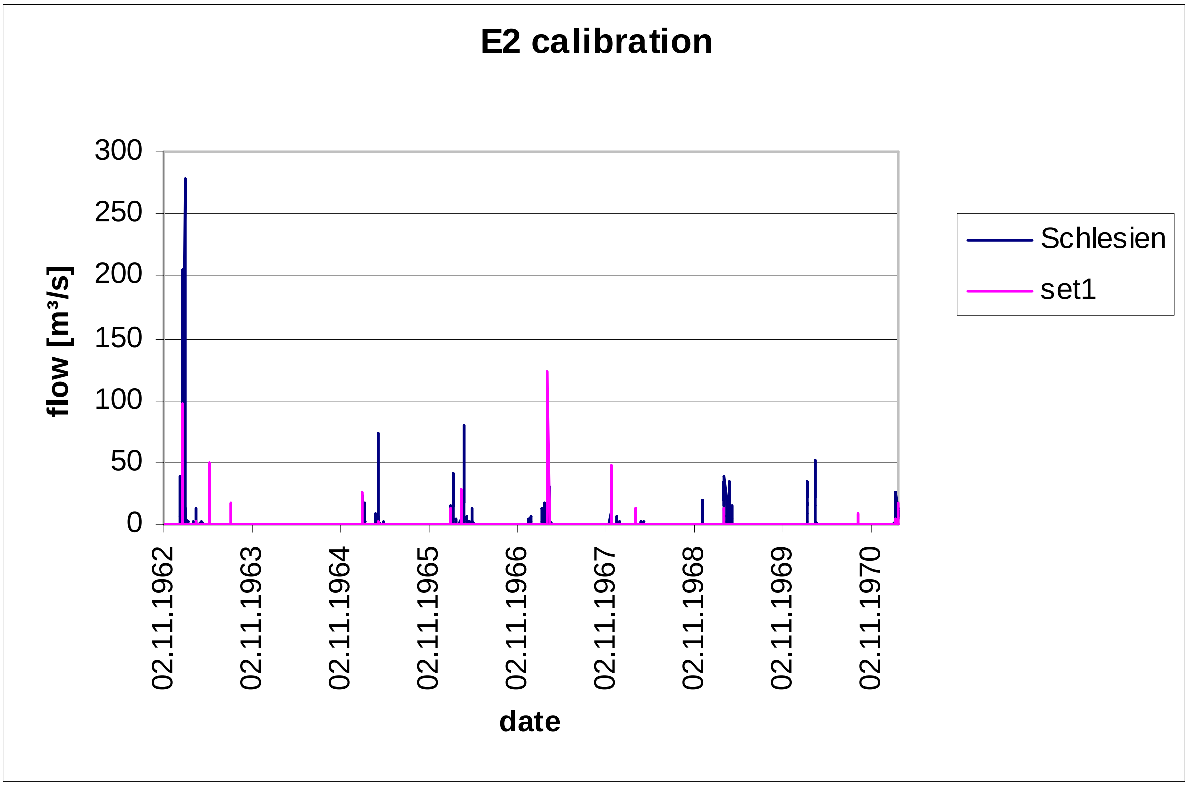

Figure 6.3.3: E2 calibration results with initial SIMHYD parameter set 1, obtained in RRL

Visual comparison of the hydrographs shown in figure 6.3.3 indicates that the model underestimates peak flows. Especially the first event with an approximate peak flow of 275 m^3/s is unsatisfactorily simulated. On the other hand the model does not reproduce all events. Furthermore the model produces runoff in times where no measured runoff occurs.

To compensate the underestimation resulting from parameter set 1, the impervious threshold was decreased to 3 mm. Additional visual calibrations led to the values seen in parameter set 2. On the one hand the error in daily R_{eff} is reduced by application of parameter set 2, on the other hand, this parameter set leads to a considerable degradation in monthly R_{eff}.

Table 6.3.2 shows interim results of the calibration process in E2. Different parameter sets are presented in columns. Set 9 represents the final parameter set with an optimum monthly Reff of 0.483.

Table 6.3.2: Interim and final results of SIMHYD calibration with E2

Parameter	set 1	set 2	set 3	set 4	set 5	set 6	set 7	set 8	set 9
Baseflow coefficient	0	0	0	0	0	0	0	0	0
Impervious threshold	5	3	3	3	3	3	3	3	3
Infiltration coefficient	250	300	300	250	250	250	150	250	200
Infiltration shape	10	5	5	10	5	5	5	10	5
Interflow coefficient	0	0	0	0	0	0	0	0	0
Pervious fraction	0.93	0.97	0.97	0.9	0.9	0.9	0.9	0.9	0.9
RISC	2	2	2	2	2	1	2	2	2
Recharge coefficient	0	0	0	0	0	0	0	0	0
SMSC	100	100	150	80	80	80	80	100	30
Reff (daily)	-0.169	-0.025	-0.025	-0.398	-0.398	-0.398	-0.398	-0.513	-0.514
Reff (monthly)	0.289	0.127	0.467	0.46	0.46	0.46	0.46	0.477	0.483

Thus, parameter set 3 shows an increase in SMSC up to 150 mm with a view to reduction of excessive runoff volumes. As a result daily Reff does not alter whereas monthly R_{eff} now is 0.467. This presents an improvement as the calibration focus is on an optimum monthly Reff.

In parameter set 4, pervious fraction is reduced to 0.9. To compensate for the increase in runoff volume, which results from a higher fraction of impervious areas and a decrease in SMSC, the infiltration shape is set back to its initial value of 10.

Modification of parameter values for Infiltration shape, Infiltration coefficient and RISC executed in parameter sets 5, 6 and 7 failed to show any impact on monthly or daily R_{eff}. Compared to the daily and monthly R_{eff} values achieved with parameter set 3, none of sets 4-7 showed an improvement. Results of the following calibration runs are shown in parameter set 8. Impervious threshold, pervious fraction and RISC are regarded as fixed. Increases in infiltration shape, in infiltration coefficient, and in SMSC yielded a monthly R_{eff} of 0.477. However, daily R_{eff} at -0.513 is now at its worst.

Finally in parameter set 9, SMSC is reduced to a minimal 30 mm. Additional runoff volume is compensated by a reduction in infiltration coefficient and infiltration shape. This improves monthly R_{eff} to 0.483, and daily R_{eff}, now at -0,514, is only slightly worse compared to parameter set 8.

Figure 6.3.4 shows the modelled hydrograph which results from the application of final parameter set 9.

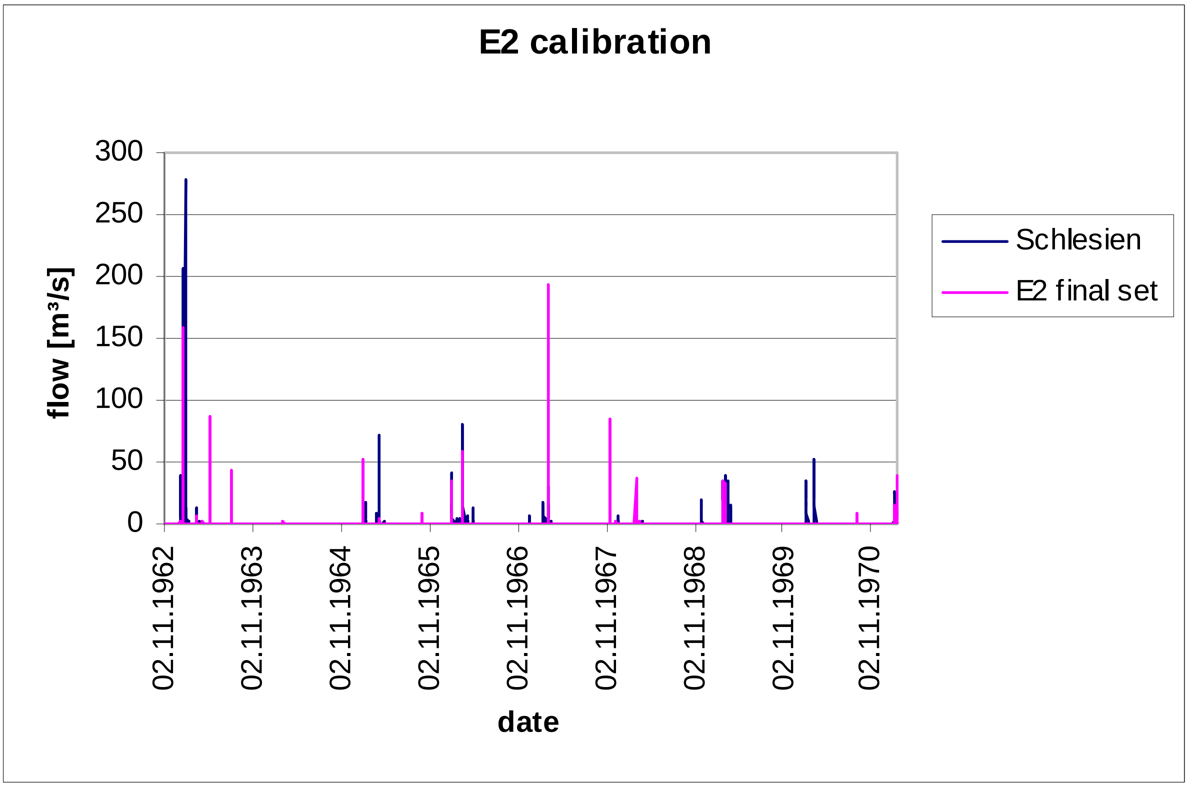

Figure 6.3.4: E2 calibration results with final SIMHYD parameter set 9

6.3.4 Single events

In the following, three discrete single events are examined more precisely. The focus is on the arrival time of flood events, their magnitudes and peak flows, and on the total flow volume.

The first event is a medium size multiple peak event with a maximum peak flow of almost 40 m³/s.

The second event actually consists of several single events. Two major floods are the subject of investigation. The second major flood represents a high magnitude event with a maximum peak flow of about 80 m³/s.

The third event is again of medium size, but with a preceding smaller event.

All simulation results are achieved by applying the final parameter set 9.

6.3.4.1 Runoff event 26.02. – 09.03.1969

Prior to the event there was a phase of about two and a half months in which no flow was recorded.

Figure 6.3.5 displays a discrete runoff event of medium size. Daily R_{eff} for this event is -2.167. It is evident that the measured runoff event starts three days earlier than the simulated event. The actual event covers a time period of 12 days whereas the simulated event only lasts for 3 days.

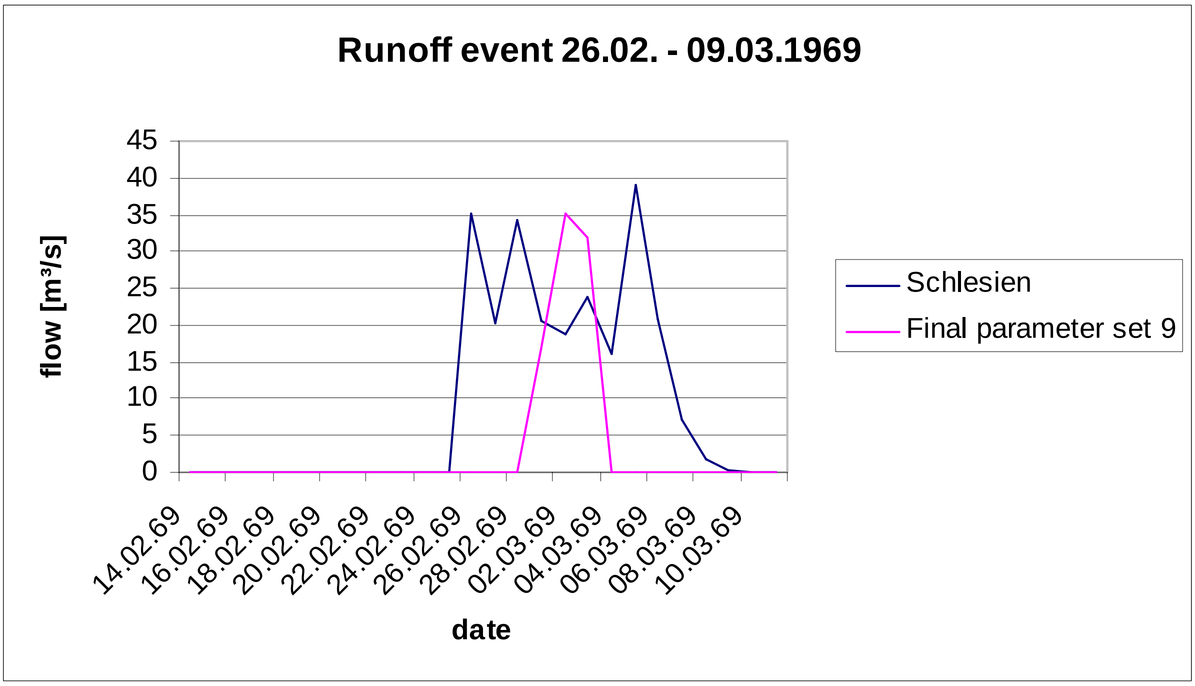

Figure 6.3.5: Measured and modelled runoff event 14.02.1969 – 11.03.1969

The measured runoff event features multiple peaks while the simulated hydrograph consists of a single peak event. Maximum peak flow of the measured hydrograph is 39.09 m^3/s compared to 35.13 m^3/s for the simulated event. Again, as for the whole calibration time series, flow volumes are significantly underestimated. This is confirmed by a measured total flow volume of 20.546.337 m^3 compared to a simulated total flow volume of 7.256.427 m^3.

6.3.4.2 Runoff events 29.01. – 05.05.1966

Prior to the event that started on 29.01.1966, no flow had been recorded for almost nine months. Hence, soil moisture storages were empty and the channel bed dry.

Figure 6.3.6 depicts two major runoff events between 29.01.1966 and 05.05.1966. The event between 17.03.1966 and 29.03.1966, with a peak discharge of 80.04 m³/s, can be regarded as a high magnitude runoff event within the Kuiseb basin. The peak discharge yielded by the model is 58.5 m³/s, approximately 70% of the actual peak discharge. It is conspicuous that the model is unable to reproduce runoff events smaller than 10 m³/s.

Coefficient of efficiency Reff for the entire 97 days is -0,533.

Reproduction of total runoff volumes is again poor. Measured runoff volume total is 33 341 343 m³ whereas simulated runoff volume is 9 528 827 m³.

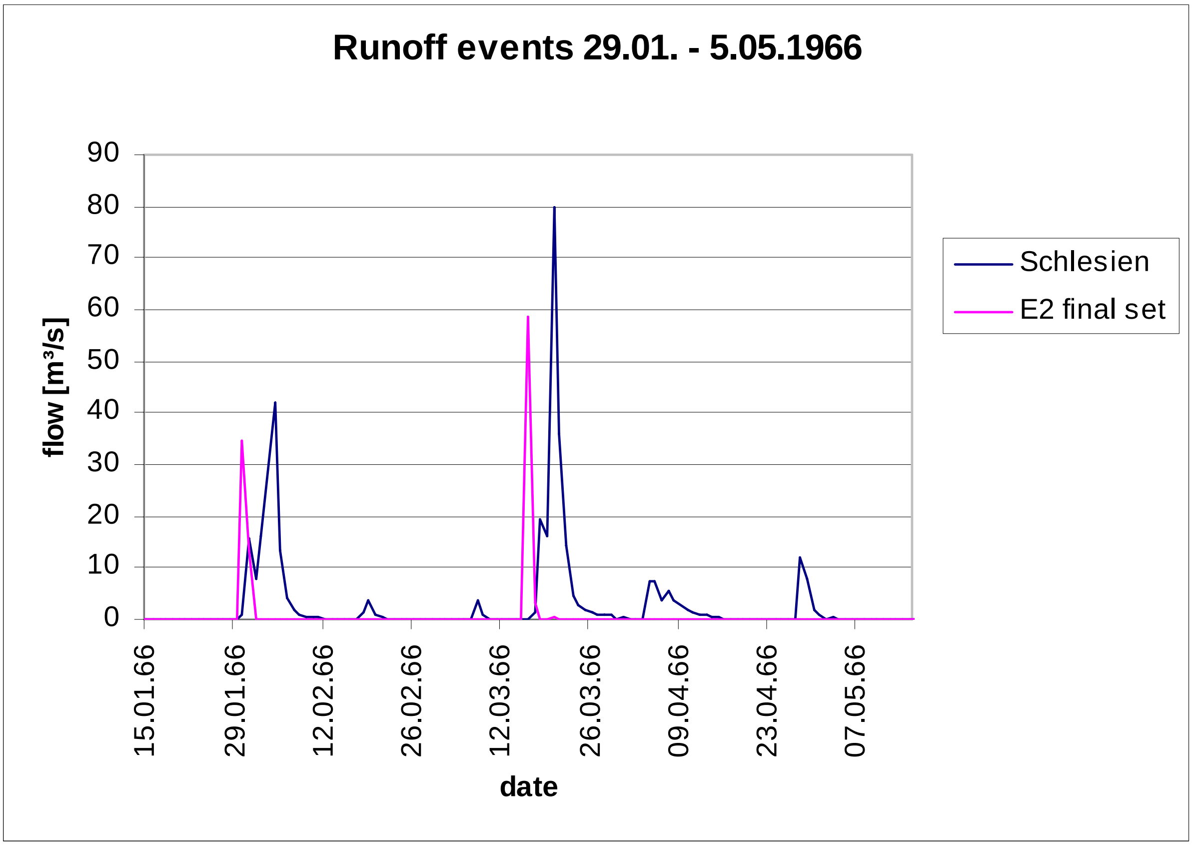

Figure 6.3.6: Measured and modelled runoff events 15.01. – 15.05.1966

Additionally simulated peak flows anticipate measured ones by 5 days for the first major event and by 4 days for the latter major event. This is due to the lack of a routing scheme and the associated delay of flow waves. The absence of a routing scheme also implies that the flow wave is not subject to attenuation. As a result flow recession is not adequately mimicked.

6.3.4.3 Runoff events 15.12.1966 – 23.03.1967

Runoff events in December 1966 and March 1967 did not exceed 31 m³/s. The three events in this period are therefore regarded as smaller runoff events. Prior to the event that starts on 06.03.1967, two smaller runoff events occured with maximum peak flows of 6.05 m³/s and 17.00 m³/s. The second of the earlier events yielded a total runoff volume of 8 130 852 m³ and ended 9 days before the third event. Therefore, it can be assumed that there was already moisture in soil water stores and in the channel alluvium at the start of the third event.

Figure 6.3.7 shows the three events observed at Schlesien, and the event reproduced by the e2 model.

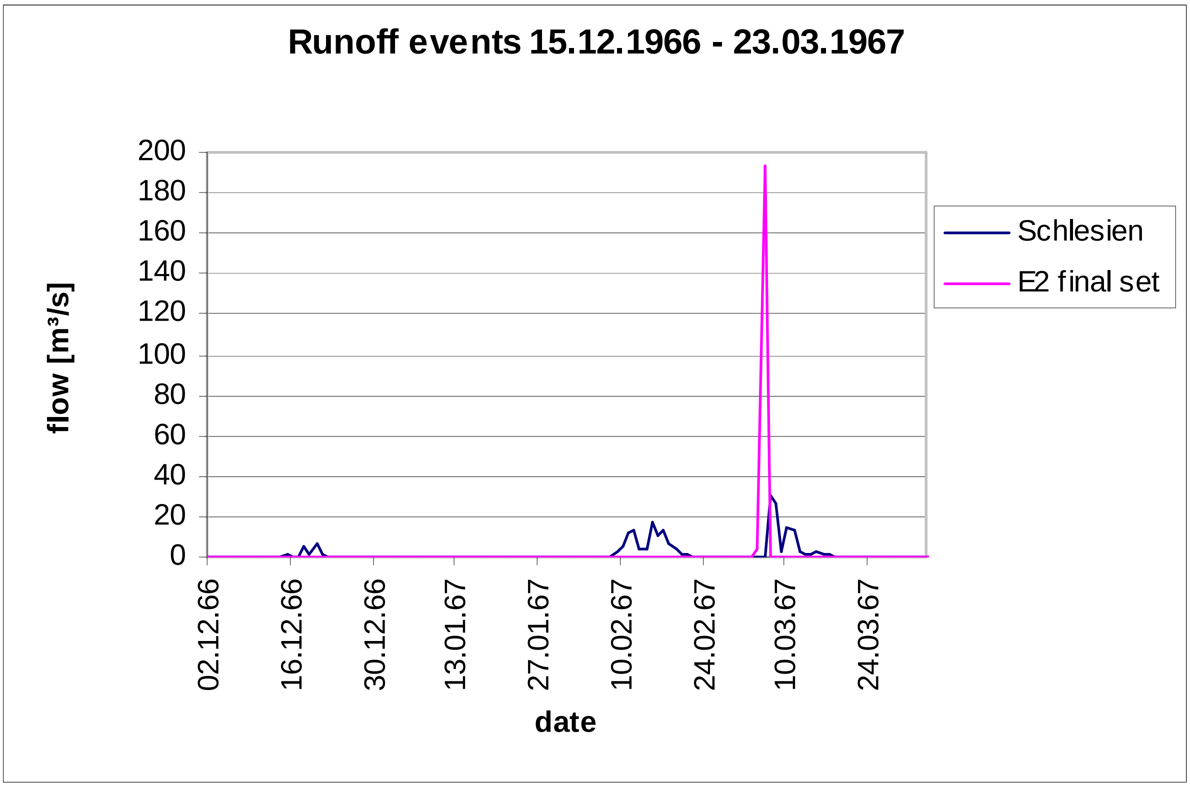

Figure 6.3.7: Measured and modelled runoff 02.12.1966 – 02.04.1967

The coefficient of efficiency R_{eff} for the 99 days investigated is -13.99. This indicates that the deviation of modelled to measured values is significantly higher than the deviation between measured values and their mean. On the one hand, this is because the model fails to yield any runoff for the first two events. On the other hand, overestimation of peak flow and total runoff volume, and exaggeration of the modelled runoff, compared to the measured flow for the third event, contribute to poor R_{eff}. Overestimation of modelled flows is, however, contrary to earlier findings during calibration. Modelled peak flow totals 193.25 m^3/s; compared to the measured maximum peak flow of 31.01 m^3/s for the third event. Thus, the overestimation is more than 600%.

Investigation of Rainfall inputs for 06.03.1967 where modelled flow amounts 193.25 m^3/s, shows that regionalized rainfalls on this particular day are exceptionally high. The mean of the regionalized area precipitation for all sub-catchments on 06.03.1967 is 59,63 mm. Re-examination of observed rainfalls proves that only two stations were recording for this period. One of the recording stations observed a daily rainfall of 140.5 mm for 06.03.1967. This high reading explains the high rainfall volumes of regionalized area precipitation for the sub-catchments, which, in turn, inevitably lead to the high runoff volumes produced by the SIMHYD model.

The flow volume of the three measured events accounts 18 286 920 m^3 whereas modelled runoff volumes make up 17 037 736 m^3. Thus the model yields almost as much runoff within one high magnitude event covering two days than was recorded over three events in almost four months.

6.3.5 Results

Automatic calibration in RRL provides final results of R_{eff}=0.128 on a daily basis and R_{eff}=0.602 on a monthly basis. Manual calibration in RRL further improved daily Reff to –0.696 and monthly Reff to 0.614.

Devolvement of a final parameter set in RRL for the E2 modelling framework resulted in an initial daily R_{eff} of -0.169 and an initial monthly R_{eff} of 0.289. Numerous calibration runs improved daily R_{eff} to -0.514 and monthly R_{eff} to 0.483 for set 9, the final parameter set. The best compromise between optimum values for both monthly and daily R_{eff} is achieved by parameter set 3 with daily R_{eff} = -0.025 and monthly R_{eff} = 0.467.

Discrepancy between monthly and daily R_{eff} indicates that the model simulates monthly totals better than daily runoff volumes.

Figure 6.3.3.2 demonstrates that the final parameter values obtained from the calibration in e2 still underestimate higher peak flows while at the same time overestimating small runoff events. Some runoff events are not reproduced by the model at all whereas modelled events occur in times of no runoff.

Comparison of runoff volumes for the entire calibration period attests that the model on average underestimates flow volumes. Measured flow volumes for the calibration period add up to 251 307 156 m³, whereas modelled flow volumes amount 103 456 857 m³. This is an underestimation of total flow volumes for the calibration period by a factor of more than two.

The examination of three discrete runoff events shows that runoff peaks are reproduced moderately. For two of the three events considered, modelled peak flows are well ahead those observed. Peak flow volumes are underestimated for two events, but overestimated for one event. The same is true for total runoff volumes for discrete events. R_{eff} for the three events ranges from -2.167 to -13.99.

Poor results for modelling on a daily basis can in parts be due to the absence of a flow routing scheme within the rainfall-runoff modelling procedure. A future flow routing scheme could be able to mimic the missing time lag; correction of total flow volumes by the application of a routing scheme is, however, unlikely.

6.4 AWBM

AWBM parameters are introduced in section 3.7.1.2.

6.4.1 Sensitivity analysis

Similarly to the manual worst case scenario for the SIMHYD model, a worst case scenario is conducted for the AWBM model.

Results of the worst case scenario indicate that the model reacts the most sensitively to the base flow index (BFI), and the recession constants for surface and baseflow (KS and K). Minor sensitivity is detected towards the capacities of the surface stores (C_1, C_2, and C3) and towards the partial areas of each surface store (A_1, A_2). An additional sensitivity test is conducted in the RRL with sum of squares of errors as the objective function. Results, for BFI, KS, and K are shown in Annexes 6 to 8.

6.4.2. Calibration with RRL

Exploration of parameter sets and single site calibration for the headwaters is achieved using RRL.

In addition to automatic and manual calibration, RRL offers a custom calibration method for the AWBM model. This “AWBM Auto Calibration” works with two optimizer parameters namely convergence criterion and maximum average capacity. Unfortunately the custom calibration method did not yield feasible results.

Thus, the initial step of calibration consists of the application of the model’s default parameters in parameter set 1 with baseflow index BFI and baseflow recession constant K set to zero. This resulted in daily R_{eff} = 0.099 and monthly R_{eff} = 0.609 which is surprisingly good. Parameter sets with representative interim results from AWBM calibration in RRL are displayed in table 6.4.1.

Subsequently, automatic calibration is performed by applying the genetic algorithm with the “Nash-Sutcliffe criterion”, or coefficient of efficiency R_{eff} as the primary objective function. The secondary objective function of choice is “Runoff difference in %”. The interim results of automatic calibration are shown in parameter sets 2 and 3 in table 6.4.1. Despite numerous automatic calibration runs there is hardly any improvement in daily and monthly R_{eff}. In addition, automatic calibration yields nonsensical parameter values, such as high base flow indexes and baseflow recession constants or zero storage capacity for one of the surface runoff stores (see sets 1 and 2 in Table 6.4.1).

As a result, the automatic calibration approach is rejected, and calibration proceeded manually.

Table 6.4.1: Interim and final results of AWBM calibration with RRL

Parameter	set 1	set 2	set 3	set 4	set 5	set 6
A1	0.134	0.1	0.1	0.2	0.2	0.134
A2	0.433	0.05	0.05	0.1	0.1	0.1
BFI	0	0.569	0.192	0	0	0
C1	7	25.3	48.6	5	2	2
C2	70	19.6	0	16	34	34
C3	150	319.6	260.8	218	102	100
K	0	0.718	0.957	0	0	0
KS	0.7	0.847	0.882	0.9	0.884	0.7
R_{eff} (daily)	0.099	0.001	0.063	0.173	0.162	-0.044
R_{eff} (monthly)	0.604	-0.094	0.077	0.421	0.483	0.675

The first presentable results of manual calibration are shown in parameter set 4. BFI and K are set to zero for all subsequent calibration runs to better represent the conditions prevailing in the Kuiseb basin. Daily R_{eff} is now 0.162 and monthly R_{eff} is 0.483. However, comparison of measured and computed flow indicates that a surface flow recession constant KS of 0.9 leads to unrealistically long flow recessions after an event. Additionally, the flow never fell to zero to represent of no-flow conditions between consecutive events.

A decrease in KS combined with altered storage capacities leads to an improvement in monthly R_{eff} as presented in parameter set 5. Monthly R_{eff} increases to 0.483 whereas daily R_{eff} degrades to 0.162.

The final parameter set 6 is achieved by again reducing KS and additionally setting A1 back to its default value. Monthly R_{eff} improved to 0.675, admittedly at the expense of daily R_{eff} which fell to -0.044.

Observed and calculated flows resulting from the final parameters in set 6 are shown in figure 6.4.1. The high magnitude flood at the start of the calibration period is underestimated by modelled flow. Furthermore, the model does not reproduce all measured events while, at the same time, it generates flow on some occasions when there was none. Especially smaller runoff events are not taken into consideration by the model.

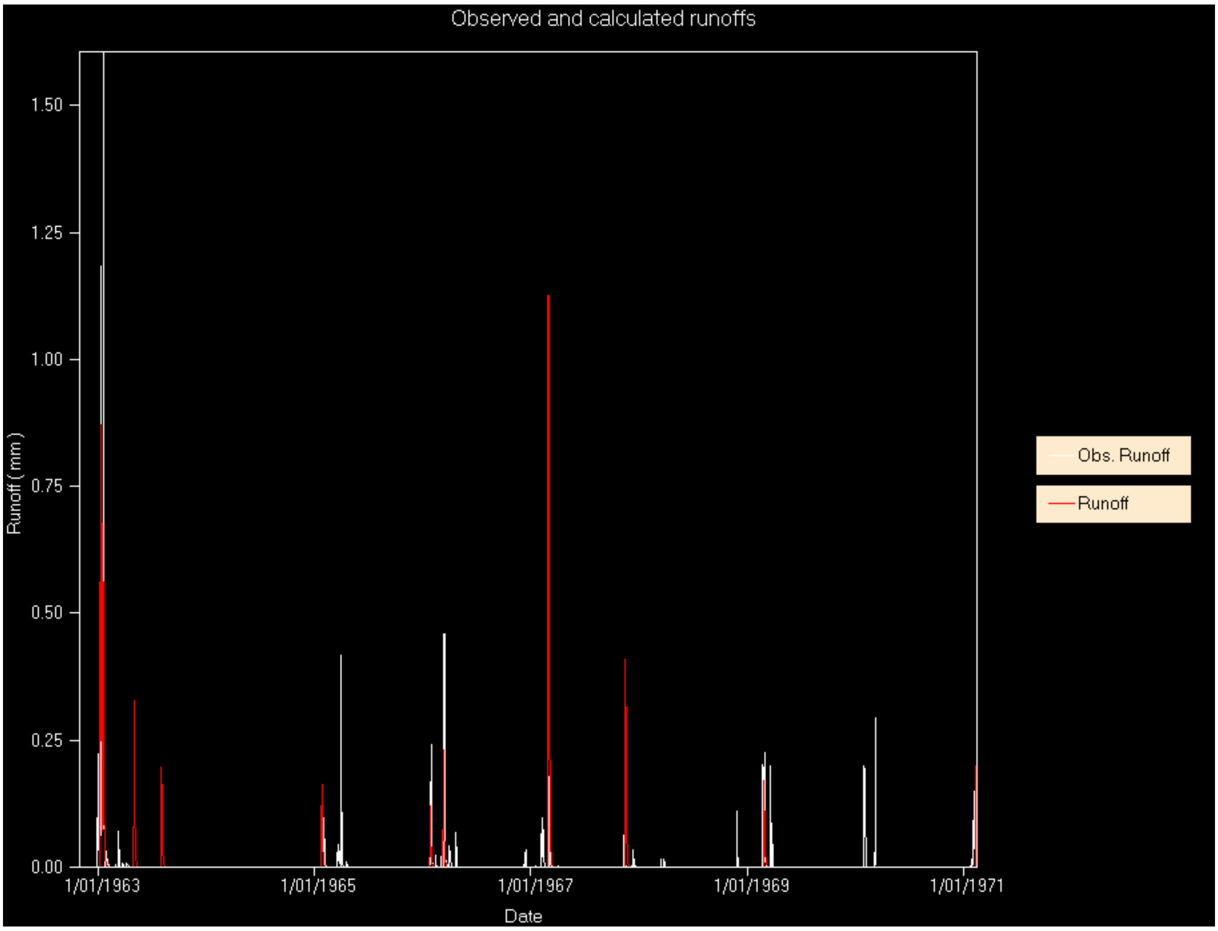

Figure 6.4.1 : Observed and calculated runoff of AWBM over the entire calibration period in RRL

Some of the modelled events are underestimated, others are clearly overestimated. This, together with poor daily Reff values, indicates that the model is better suited to simulating the hydrological behaviour of the system in a balanced long term examination than to reproducing single events.

6.4.3 Calibration with E2

The flexible calibration tool available in E2 is introduced in section 6.3.3. The calibration tool is also applied for calibrating AWBM.

Again, as for SIMHYD calibration, the network is cut below the node which represents the gauging station at Schlesien. As a result only the upstream area from this node is the object of further calibration. The reduced network is shown in figure 6.3.2.

At first the default parameters are the object of calibration. Again, BFI and K were first set to zero, as in parameter set 1. This resulted in daily R_{eff} = 0.125 and monthly R_{eff} = 0.287. A visual comparison of modelled and measured hydrographs admittedly showed no modelled events exceeding a peak flow volume of 50 m³/s. Moderate R_{eff} values are therefore due rather to shared no flow periods in the simulated and modelled hydrographs than to adequate modelling of flow events. Representative interim results of parameter sets and the final, set 6 parameters from e2 calibration are shown in table 6.4.2.

Table 6.4.2: Interim and final results of AWBM calibration with E2

Parameter	set 1	set2	set3	set4	set5	set6
A1	0.134	0.134	0.15	0.2	0.2	0.2
A2	0.433	0.1	0.05	0.1	0.1	0.1
BFI	0	0	0	0	0	0
C1	7	2	5	7	2	0
C2	70	34	30	50	35	40
C3	150	100	100	100	100	100
K	0	0	0	0	0	0
KS	0.7	0.7	0.6	0.7	0.4	0.5
R_{eff} (daily)	0.125	0.135	0.099	0.154	-0.322	-0.247
R_{eff} (monthly)	0.287	0.43	0.384	0.413	0.58	0.622

Application of the parameter set finally obtained from RRL calibration (see set 2 in table 6.4.2) improved daily R_{eff} to 0,135 and monthly R_{eff} to 0,43 respectively.

Figure 6.4.2 depicts the modelled hydrograph for parameter set 2.

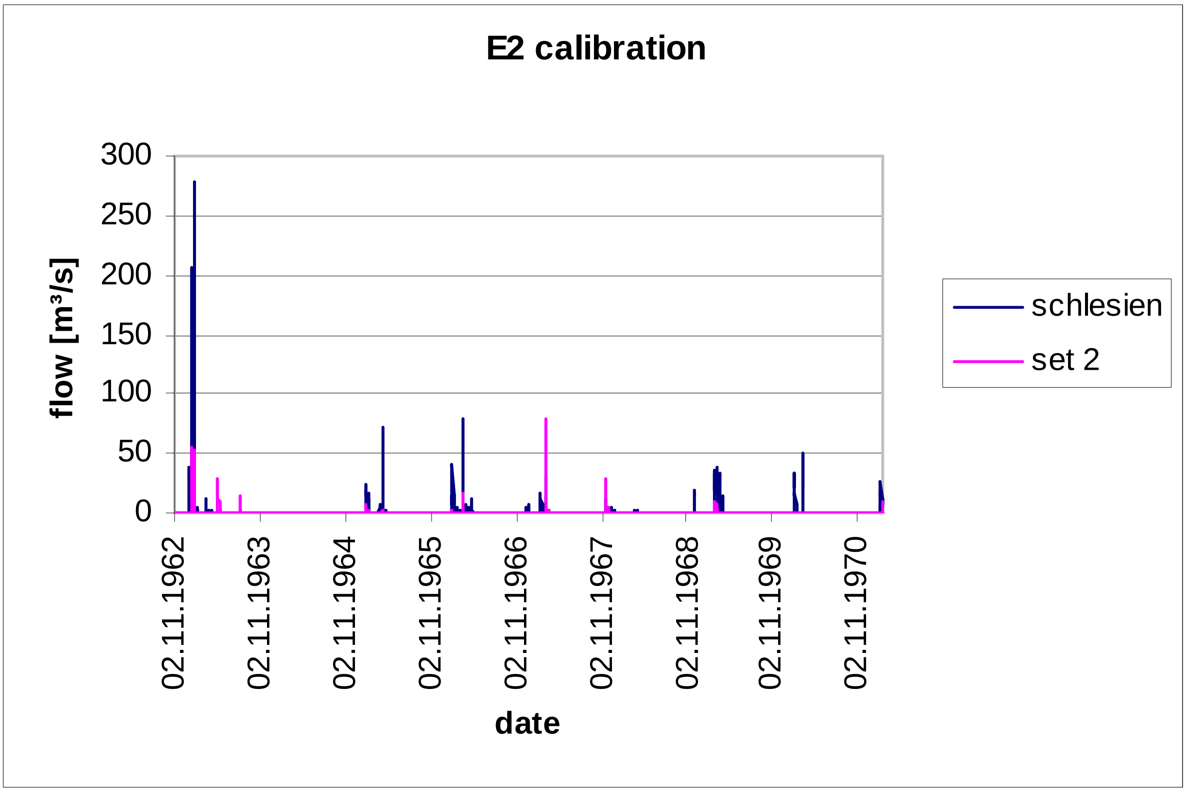

Figure 6.4.2: e2 calibration results with initial AWBM parameter set 2, obtained in RRL

Except for one event, modelled flow still significantly underestimates the flow measured at Schlesien. This is validated by a total modelled flow volume of 81 762 549 m³ over the calibration period compared to a measured total flow volume of 251 307 156 m³. Thus, the total modelled volume is only one third of that measured.

Variations of parameter values in parameter sets 3 and 4 were an attempt to increase total flow volume without overestimating particular events, or even generating runoff where there is none.

Parameter set 3 shows decreasing values for A2, KS and C2. However, the consequence is deterioration in both daily and monthly R_{eff}. Parameter set 4 did not show substantial progress in either R_{eff} values or the modelled hydrograph.

Only a decrease in the surface flow recession constant KS down to 0.4 resulted in an observable increase in runoff volumes. Reduction of the storage capacities of the first and the second surface flow stores also contributed to this. These alterations are incorporated in interim parameter set 5. Monthly R_{eff} for set 5 is 0.58 while, at the same time, daily R_{eff} worsens to -0.322.

To compensate for overestimation of some modelled events using parameter set 5, the surface flow recession constant was increased slightly from 0.4 to 0.5. In parameter set 6, the storage capacities of the first and second surface flow storages are altered: the capacity of the first store C1 is reduced to zero. Thus, the area of the first surface store A1 now represents an area which directly contributes to runoff. This is in line with the conceptual hydrological picture of the Kuisebs headwaters, where outcrops of basement rocks act as impervious areas which can directly contribute to runoff.

To counterbalance the excess runoff, C2 is increased to 40 mm.

Modelled runoff resulting from final parameter set 6 is displayed in figure 6.4.3. The first high magnitude event is still significantly underestimated by the model, but better fits for this event would have resulted in overestimation of all later events. The modelling of the runoff event that started on 06.03.1967, as with the SIMHYD model, shows strong overestimation.

Despite this, total flow volumes for the entire calibration period yielded by the model are below those measured. The total volume of modelled flows is 154 844 252 m³ compared to a total measured volume of 251 307 156 m³. This implies that over a period of approximately eight years and 3 months the model underestimates flow volume by almost 100 000 000 m³.

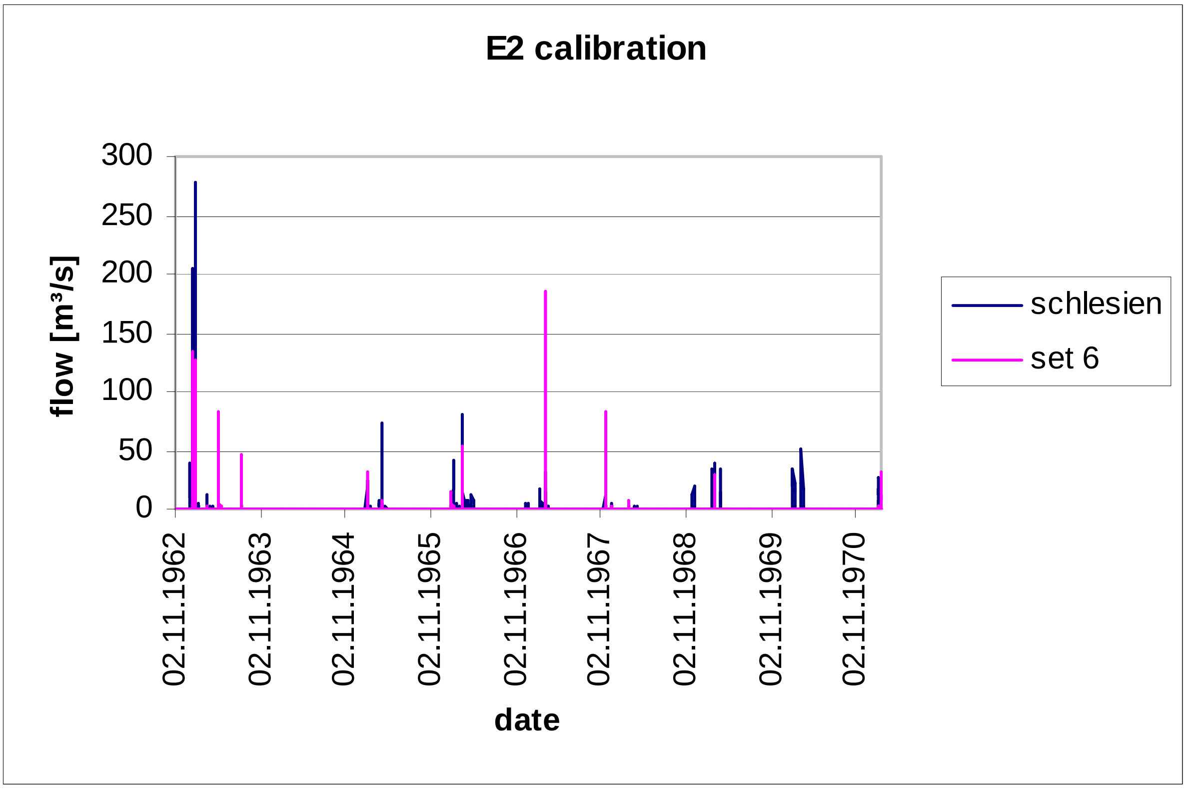

Figure 6.4.3: e2 calibration results with final AWBM parameter set 6

Although parameter set 6 yields the best overall fit between measured and simulated hydrographs, the problem of simulated events in no-flow periods, and the occurrence of measured runoff events in the absence of their simulated counterparts is not satisfactorily solved.

However, a final monthly R_{eff} of 0.62 is acceptable and marks an improvement compared to the results from previous parameter sets.

6.4.4 Single events

For reasons of comparability between the SIMHYD and the AWBM model the events examined below are the same as those already studied in section 6.3.4.

The focus is again on the arrival time of flood events, their magnitudes and peak flows, and the total flow volumes.

All simulated hydrographs are obtained by the application of final parameter set 6.

6.4.4.1 Runoff event 26.02. – 09.03.1969

Prior to this medium size event there was a phase of about two and a half months were no flow has been recorded. Therefore, soils and channel alluvium are assumed to have been dry.

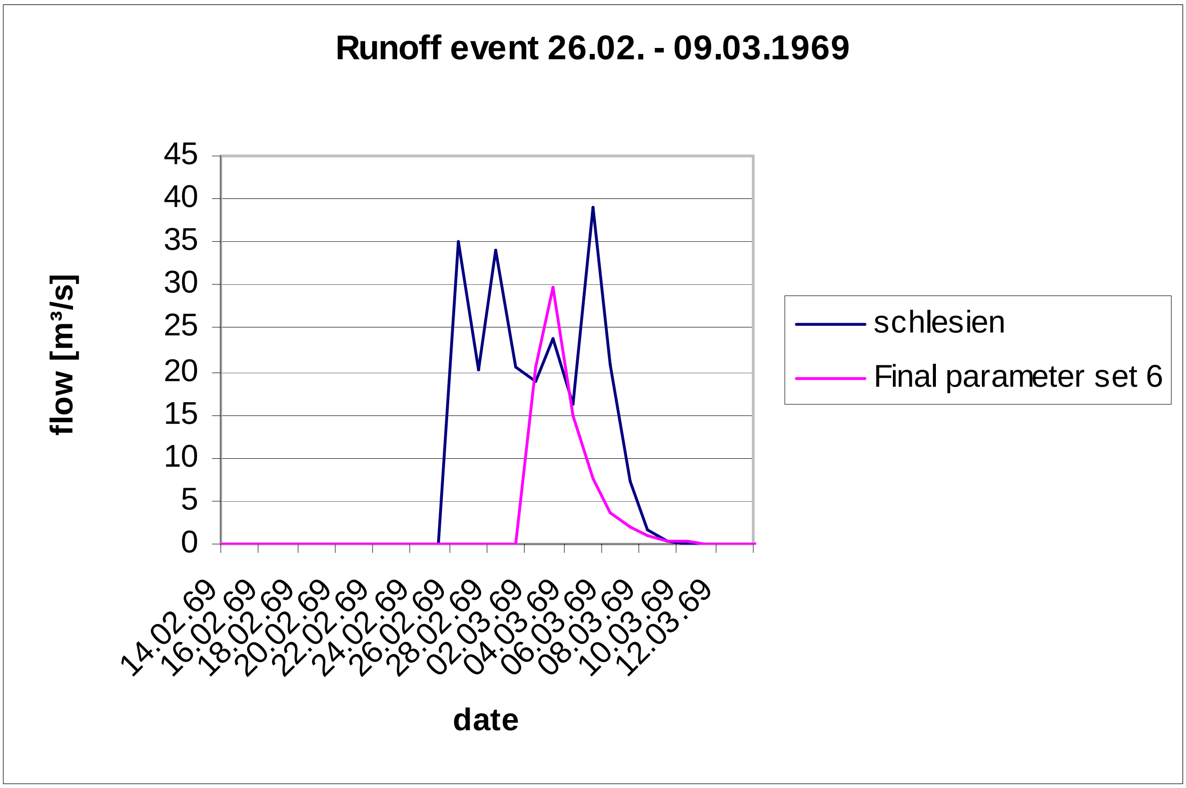

Figure 6.4.4: Measured and modelled runoff event 14.02.1969 – 13.03.1969

Daily Reff for the event shown in figure 6.4.4 is -1.68. Calculated runoff starts four days later than measured runoff. Total modelled flow volume of 6 890 852 m^3 is significantly lower than total measured flow volume of 20 546 337 m^3. Thus, the model flow is only approximately 30% of measured volume. Underestimation of peak flow rates is of minor magnitude with a maximum measured peak flow of 39.09 m^3/s compared to a modelled peak flow of 29.83 m^3/s.

The modelled hydrograph exhibits a rather long flow recession and as a consequence shows a duration of ten days, compared to a duration of twelve days for the measured event.

6.4.4.2 Runoff events 29.01. – 05.05.1966

Prior to the event that started on 29.01.1966 no flow had been recorded for almost nine months. Hence, soil moisture storages were empty and the channel bed was dry.

Daily R_{eff} for the entire period of examination is -0.33. Arrival times for both major events are acceptable. Modelled runoff for the first major flood is delayed by one day, while, for the second event, it is one day in advance. Four minor runoff peaks with less than 10 m³/s peak flow are not reproduced by the model. The modelled and measured flows are depicted in figure 6.4.5

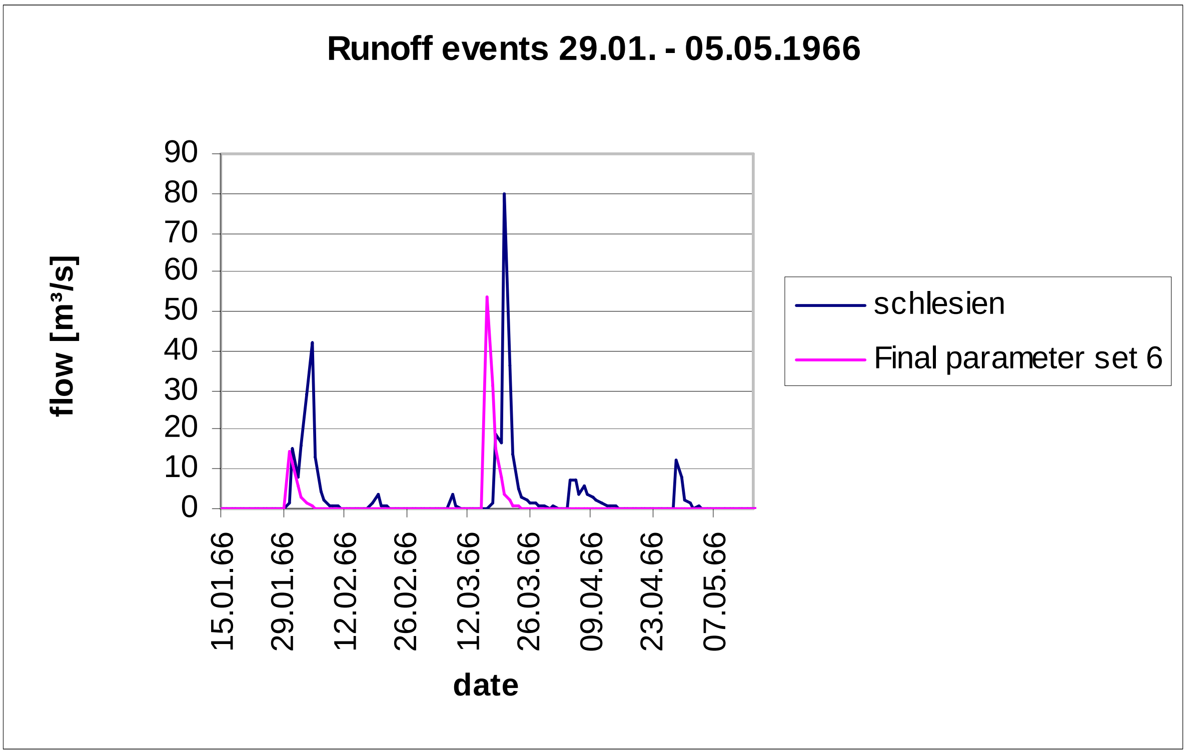

Figure 6.4.5: Measured and modelled runoff events 15.01.1966 – 15.05.1966

Peak flows are again underestimated by the model. This shows in modelled peak flows of 14.45 m³/s and 53.70 m³/s compared to measured peak flows of 80.04 m³/s and 41.98 m³/s. The modelled flow peaks occur four to five days earlier than those measured for both major events.

Total flow volumes are again reproduced rather poorly. Measured events yield a total volume of 33 341 343 m³ compared to 13 177 595 m³ from model calculation. Both flow volumes are calculated over the entire period of 97 days.

6.4.4.3 Runoff events 15.12.1966 – 23.03.1967

All three events in this period are regarded as smaller runoff events. Prior to the last event starting on 06.03.1967, two smaller runoff events occur with maximum peak flows of 6.05 m³/s and 17.00 m³/s. Thus, it is assumed that there was already moisture in soil water stores and in the channel alluvium at the start of the third event. Daily Reff for the entire 99 days is -13.63. This results from the models inability to reproduce the first and the second events and from moderate reproduction of the final event. Measured peak flow of the final event is 31.01 m³/s, and is seriously overestimated by a modelled peak flow of 184.15 m³/s on 06.03.1967.

As previously for the SIMHYD model in section 6.3.4.3, the conclusion can be drawn that overestimation of flow on 06.03.1967 is due to inadequate inputs in terms of regionalized rainfall.

Additionally, the modelled peak arrives one day earlier than that measured. The modelled and measured hydrographs are shown in figure 6.4.6.

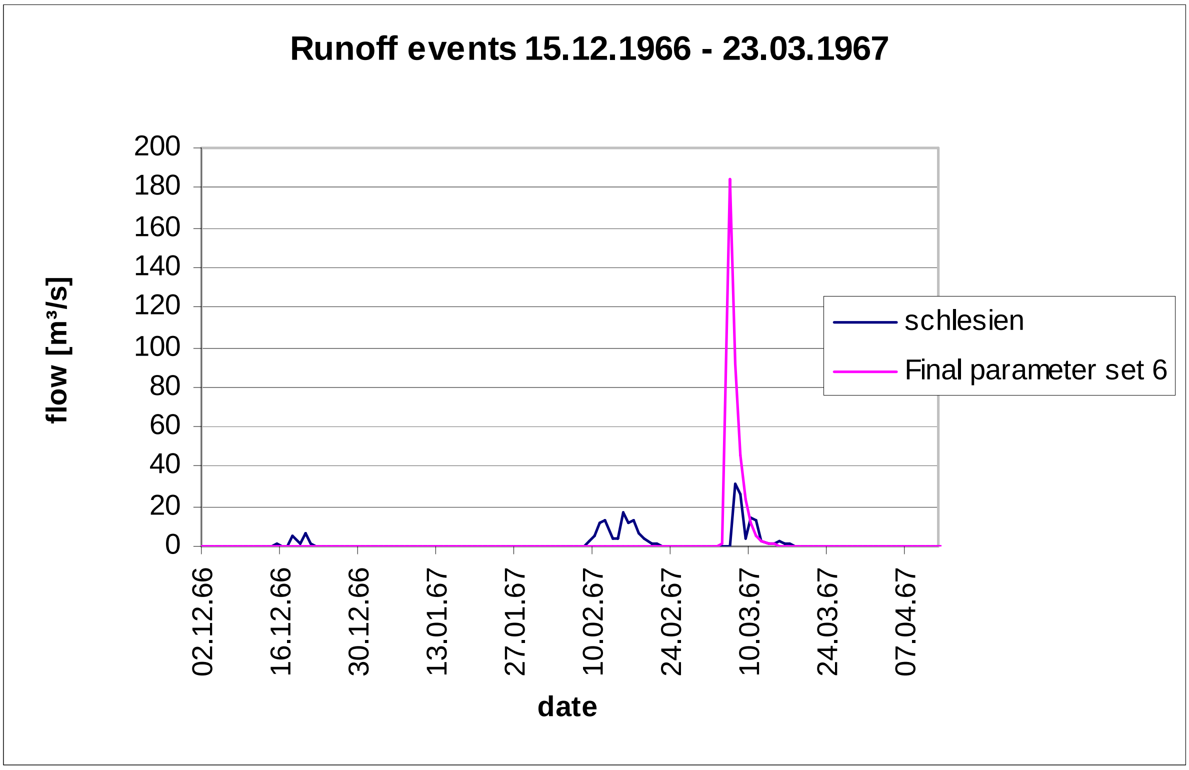

Figure 6.4.6: Measured and modelled runoff 02.12.1966 – 02.04.1967

The overall flow volume for all the three measured events is 18 286 920 m³ and is exceeded by a modelled flow volume of 31 899 680 m³.

6.4.5 Results

Sensitivity analysis in terms of a worst case scenario indicates that the model is most sensitive towards the base flow index and the flow recession constants for surface flow and base flow. Minor model sensitivity was verified for the capacities of the surface stores and for the partial areas of these stores.

Automatic calibration in RRL does not contribute to improvement of initial coefficients of efficiency; neither by means of genetic algorithm nor using a custom calibration routine for AWBM. However, manual calibration in RRL produces a final daily R_{eff} of -0.044 and a final monthly R_{eff} of 0.675. Focusing on monthly totals, this is satisfactory.

Subsequent calibration in e2 lead to massive underestimation of flow volumes and flow peaks for the first calibration runs. This is still true for the parameter set finally obtained in RRL, which, when transferred to e2, only reproduces approximatey 30% of measured flow volume.

Only after reducing the surface flow recession constant KS to 0.4 and lowering the capacities of surface flow stores, was noticeable increase in modelled runoff volumes achieved. This results in a monthly R_{eff} of 0.58.

In final parameter set 6, the capacity of the first surface flow store is set to 0 mm, making partial area A1 practically impervious. KS is increased again to 0.5, while BFI and K are maintained at zero throughout all calibration runs in e2. The final monthly R_{eff} is passable at 0.622 compared to the poor daily R_{eff} of -0.247. The main errors in the computed hydrograph are due to enormous overestimation of flow for two events, and to the absence of modelled flow for at least one event. Moreover, the model is incapable of satisfactorily reproducing the high magnitude flood at the outset of the calibration time series.

Examination of the whole calibration period modelled with final parameter set 6 indicates that the model generally underestimates flow volumes. This is emphasized by a total modelled volume of 154 844 252 m^3 compared to a total measured volume of 251 307 156 m^3 over the entire period. In two of three cases, examination of discrete events also shows an underestimation of modelled flows; only one event showed an overestimation in both modelled peak flows and modelled flow volumes.

6.5 Conclusion

Available flow data observed at the gauging station at Schlesien consisted of a time series of approximately 22 years. From that period, a time series of 8 years and 4 months is chosen for calibration (02.11.1962 – 16.02.1971).

The quality of model results is assessed by visual comparison of the modelled and measured hydrographs, and by means of model efficiency according to NASH & SUTCLIFFE (1970).

Sensitivity analyses conducted in RRL (another toolkit product from the CRC for Catchment Hydrology), and in terms of worst case scenarios, proved to be valuable tools for evaluating the impact of different parameters on model results.

RRL additionally proves to be a useful tool for exploration of parameter sets before applying them to e2. Various automatic calibration approaches are provided within RRL which eased parameter estimation for the SIMHYD model but not for AWBM. Manual calibration in RRL allows for direct control of calibration results by means of various objective functions, greatly speeding up the calibration process.

Calibration in RRL indicates that further calibration should focus rather on monthly totals than on the simulation of daily runoff and that a conceptual hydrological understanding of the modelled basin must be incorporated in the calibration process.

Final parameter sets obtained from RRL show a daily R_{eff} of -0.696 and a monthly R_{eff} of 0.614 for the SIMHYD model, compared to a daily R_{eff} of -0.044 and a monthly R_{eff} of 0.675 for the AWBM model.

However, a handicap of the RRL model is its inability to represent spatial variations in the input time series. Only one time series of rainfall and ET can be loaded. Thus, alteration of the input series results in different model outputs.

Within e2, a calibration tool eases calibration by permitting definition of a sub-network and the manual grouping of parameters. Final parameter sets obtained from calibration in e2 yield a daily R_{eff} of -0.514 and a monthly R_{eff} of 0.483 for the SIMHYD model, and a daily R_{eff} of -0.247 and a monthly R_{eff} of 0.622 for the AWBM model.

Final parameter sets for both SIMHYD and AWBM result in the underestimation of total flow volumes both over the entire calibration period, and for most discrete runoff events. In contrast to this, both models overestimate some discrete runoff events. Overestimation of discrete events is worse for the SIMHYD than for the AWBM model.

As already proved for the modelled event on 06.03.1967 in sections 6.3.4.3 and 6.4.4.3, overestimation of discrete modelled runoff events is generally caused by excessive rainfall inputs.

This assumption is underlined by the events on 07.05.1963 ($Q_{sim.}$ = 88.02 m³/s), and on 08.08.1963 ($Q_{sim.}$ = 43.91 m³/s). For both events, no flow was observed at the gauging station at Schlesien. Examination of rainfall records, however, shows that on 07.05.1963 the rainfall gauge at Middelplaas recorded 20.3 mm, while the rainfall gauge at Tantus recorded 12 mm. On 08.08.1963, Middelplaas recorded 13 mm compared to 16.5 mm at Tantus. As a result, the mean of regionalized area precipitation for all sub-catchments is 14.76 mm on 07.05.1963, and 9.54 mm on 08.08.1963. With these sub-catchment inputs, the model consequently generates runoff on both days although no runoff has been recorded. Thus, the conclusion has to be drawn that, on these occasions, small precipitation cells hit the two recording stations while at the same time the remainder of the catchment did not receive sufficient rainfall to generate runoff. Therefore, extrapolation of observed rainfall amounts to the entire catchment results in inaccurate regionalized rainfalls which, in turn, generate runoff in periods in which no flow was recorded.

Moreover, both models have problems to satisfactorily simulate the high magnitude event at the start of the calibration period, although the SIMHYD model performes slightly better. Additionally, neither SIMHYD nor AWBM generate runoff for the measured events on 02.02.-10.02.1970 or 13.03.-15.03.1970, respectively. The reverse of the assumption above is that local convective thunderstorms yielded sufficient rainfall to generate runoff within the catchment, but did not, or only strongly alleviated, hit the rainfall gauges. This explains both the underestimation in modelled peak flows for the high magnitude event around 24th January 1964, and the absence of modelled flows for the events in February and March 1970. This is underlined by total recorded rainfall volumes of 0.9 mm at Middelplaas and 13 mm between 2nd and 10th February 1970, and total recorded rainfall volumes of 0 mm at Middelplaas and 2.4 mm at Tantus for the period between 13th and 15th of March 1970.

7 Validation

7.1 Introduction

Validation procedure compares modelled simulations and observations using data that were not part of the calibration procedure.

The objective function used to provide a quantitative assessment of validation results is again model efficiency according to NASH & SUTCLIFFE (1970) as shown in equation 6.1.1.

According to ARGENT (2004/05) the extent to which the validation step really tests the model structure and parameters is dependent on the type of data available for testing.

Flow data for validation consists of a time series of approximately 8 years (26.08.1972 – 29.08.1980) observed at the gauging station at Schlesien.

The gap between the calibration and the validation time series is due to poor flow conditions in this period.

7.2 SIMHYD

The hydrograph produced by the SIMHYD model for the validation period is depicted in figure 7.2.1.

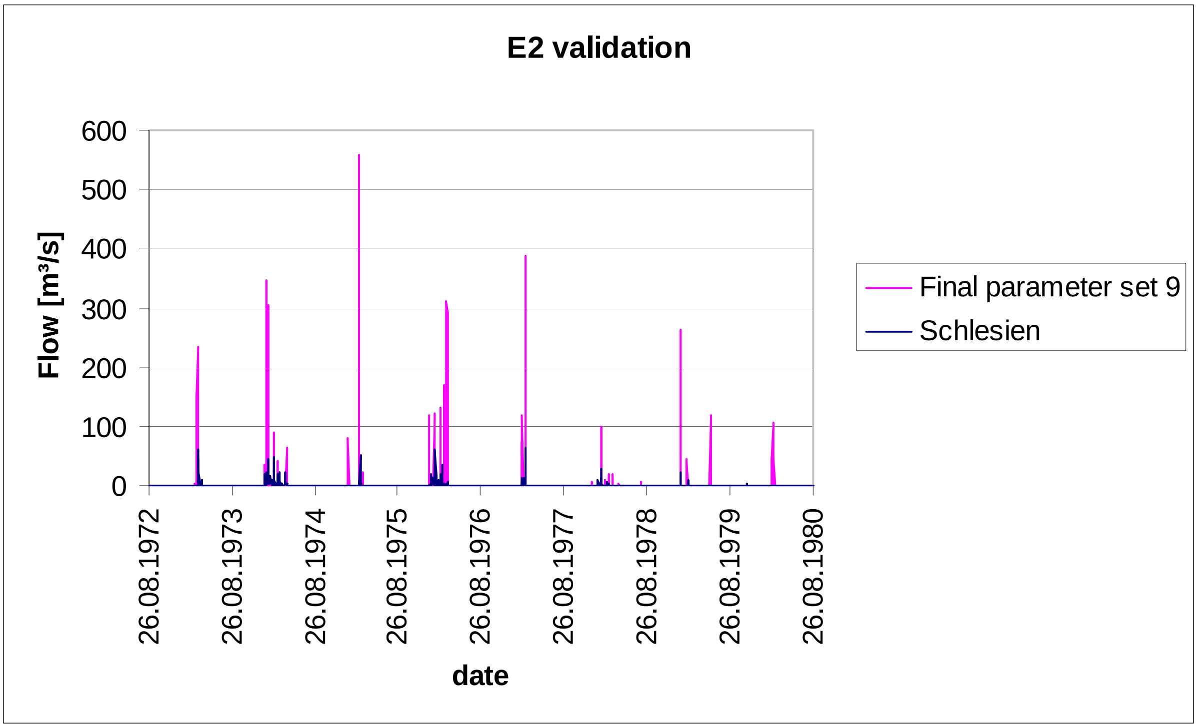

Figure 7.2.1: Validation results for SIMHYD, obtained with final parameter set 9

It is self-evident that the model overestimates peak flows for all events. As a result total modelled flow volumes are a multiple of measured ones.

More precisely the total simulated flow volume amounts 426 984 845 m^3 over the entire validation period compared to a total measured flow volume of 152 263 584 m^3. This conforms to an almost triple overestimation in total flow volume.

Daily R_{eff} for validation is -0.013, whereas monthly R_{eff} is -5.52.

7.3 AWBM

As for the SIMHYD model, validation results for AWBM are poor. Flow volumes and peak flows are constantly overestimated by the model. As a result daily R_{eff} is -30.56, whereas monthly R_{eff} is -15.54.

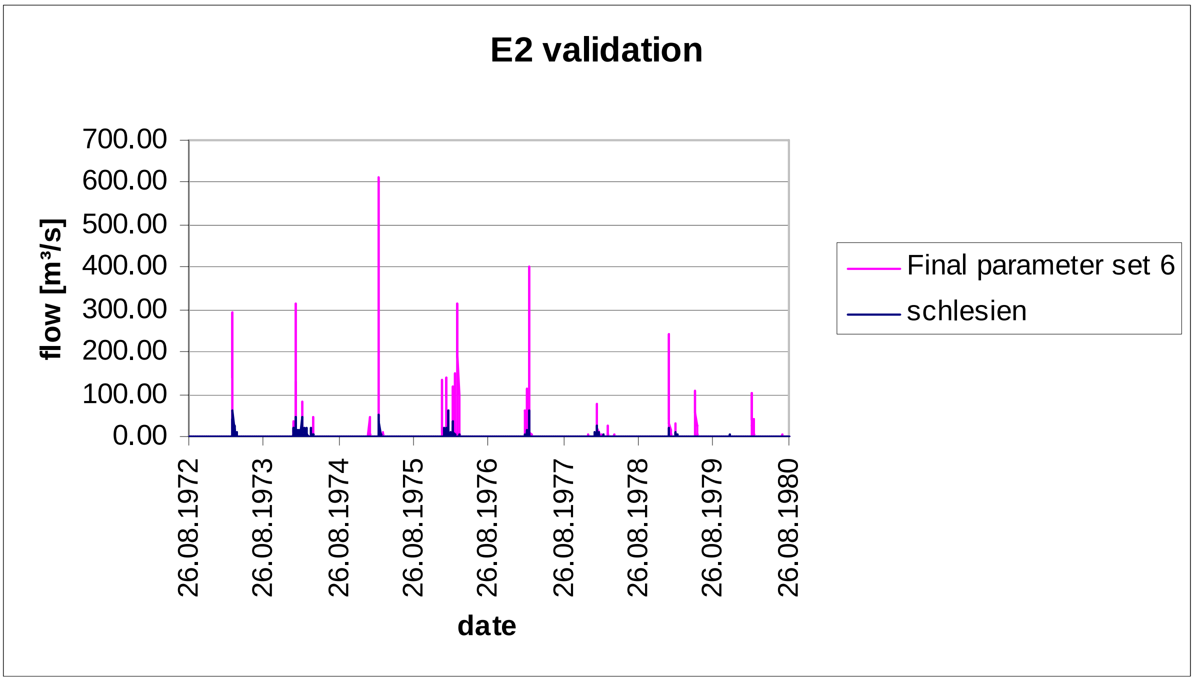

Figure 7.3.1: Validation results for AWBM, obtained with final parameter set 6

Comparison of total flow volumes of the entire validation period underlines the meagre results of model validation. Modelled flow volume amounts to 686 528 965 m^3 compared to 152 263 584 m^3 passing the gauging station at Schlesien within the same period. Furthermore, the model generates flow on five occasions where no measured flow is existent.

Again, this must be regarded as an unacceptable reproduction of flow for the validation period.

7.4 Conclusion

Comparison of validation results from the SIMHYD and AWBM models shows that AWBM performed worse than SIMHYD. This is true in regard to both peak flows and total flow volumes.

Generally, results from both models must be regarded as poor. Both models show multiple overestimations of total flow volumes. This leads to poor monthly R_{eff} values for both models.

As already demonstrated for the calibration period, bad model performance is again due to erroneous rainfall inputs. Apart from the errors inherent in the regionalization procedure, the basic problem is the regionalization approach itself. Apparently it cannot be assumed that rainfall recorded at the gauging stations also correlates to the entire catchment. Conversely, the assumption that all rainfall that occurs within the catchment is also recorded to some extent at the rainfall gauges seems to be wrong as well.

As a result, this leads to an overestimation in modelled flows for some events, whereas other events are significantly underestimated.

Within the validation period, maximum overestimation of peak flow is observed on the 8^{th} March 1975 with a modelled peak flow of 558,53 m³/s from SIMHYD and 613,93 m³/s resulting from AWBM. Peak flow for the relevant event measured at Schlesien does not exceed 50,33 m³/s. Observed rainfall for 8^{th} March 1975 amounts 85 mm at Tantus, and 7mm at Middelplaas. This leads to a mean regionalized area precipitation of 43.79 mm for all sub-catchments. These observations confirm the assumptions made above.

These findings raise the question of how reasonable it is to use a rainfall regionalization approach based on only four recording stations is in an arid macro-scale watershed like the Kuiseb basin. Results show that both temporal occurrence and regionalized rainfall volumes produce runoff that is not realistic. Consequently the value of an approach to rainfall-runoff modelling that is based on such regionalization is open to doubt.

As the inputs to the SIMHYD and the AWBM models proved to be faulty, further assessment of model performance is difficult.

8 Final conclusion and outlook

8.1 Final conclusion

Subject of this study was to apply the existing daily e2 model to the arid macro-scale watershed of the Kuiseb River for rainfall-runoff modelling.

The e2 catchment modelling software is a flexible framework for catchment modelling from the CRC for Catchment Hydrology (CRCCH), Australia.

The basic structure of the model is node-link, in which fluxes of water and constituents from sub-catchments appear at outlet nodes and are then routed along links. Although the model allows for component models of various processes, the focus of this study was solely on the model's ability to perform rainfall-runoff modelling.

From a set of available rainfall runoff models SIMHYD and AWBM were chosen for further application. AWBM offers a hydrotope-like approach with three surface stores which represent the three fractions of the catchment area. The SIMHYD model divides the catchment into areas with pervious and impervious properties. Runoff from impervious areas is directly generated as surface runoff whereas surface runoff from pervious areas can occur as infiltration excess and saturation excess runoffs.

In general e2 can be regarded as a user-friendly modelling framework that offers a multiplicity of options to custom build a whole of catchment model of the complexity required. Due to inadequacy of the model inputs used in this study, however, no appraisal can be made of how suitable e2 rainfall-runoff models are for the Kuiseb watershed.

To generate the required inputs for the rainfall runoff models, rainfall and evaporation had to be aggregated in both time and space.

The results of the regionalization of potential evapo-transpiration are regarded as satisfactory. The conceptual idea behind it was to fit a sine function to the annual regime of PET and allow for random scattering around the values of the sine function according to a normal distribution. This was done for Windhoek in the easterly part of the catchment at high altitude, and for Gobabeb which represents the conditions in the westerly part of the catchment at lower altitudes. An additional fictive data set was created for the central catchment to better represent the conditions in the transition between Gobabeb and Windhoek.

Rainfall regionalization is based on four rainfall gauges within the catchment. The regionalization approach for rainfall is a mixture of the inverse distance squared method and the application of a rainfall decay function. The rainfall decay function is applied solely in periods where only one rainfall gauge was active. Initially, rainfall was regionalized for each point of a dot matrix that was superimposed on the catchment. Then, sub-catchment area precipitation was obtained by averaging the regionalized rainfall of all points within a particular sub-catchment. The spatial distribution of regionalized rainfalls was found acceptable in comparison with the climatic gradient that runs through the catchment from west to east. However, it is inherent to these regionalization methods that rainfall is highest close to the rainfall gauges.

Within the subsequent process of model calibration, the rainfall runoff library (RRL), another toolkit product from the CRCCH, proved to be of great benefit to estimate parameter values before applying them to e2.

Calibration of parameters for SIMHYD and AWBM was eased by the calibration tool included in the e2 modelling software. Results of the calibration were acceptable with a monthly coefficient of efficiency R_{eff} of 0.483 for the SIMHYD model and a monthly R_{eff} of 0.622 for the AWBM model. However, poor reproductions of flow behaviour for discrete runoff events by both models suggest that rainfall inputs must be erroneous. Detailed examination subsequently proved that regionalized rainfall was too low on some occasions and excessive on others. This is not due to incorrect regionalization methodology, but rather to trying to use the regionalization approach within the arid macro-scale Kuiseb watershed with a rainfall gauging network of only four stations. As the occurrence of local convective thunderstorm cells is highly variable over space and time, small-scale rainfalls can hit the gauging network while a large proportion of the catchment does not receive any rainfall at all. Conversely, local storm events can generate runoff without ever hitting the rainfall gauging network.

These findings explain in advance the poor validation results for both models, while, at the same time, they withdraw the basis for a sound assessment of model performance.

Therefore, the conclusion is that the application of e2 in the Kuiseb basin is not limited by the model's structure, but rather by the input requirements of the model, and by the method used to obtain this input data.

8.2 Outlook

The high variability of hydrological processes and data scarcity in arid environments is widely acknowledged. Nevertheless, an attempt was made to conduct rainfall-runoff modelling with daily time-stepping based upon the regionalization of precipitation in the Kuiseb basin, Namibia.

However, results of the rainfall regionalization are poor, regarding the subsequent attempts to mimic flow behaviour with this data.

Firstly, the application of a routing scheme (e.g. e2 storage model) can improve the temporal simulation of discrete events within the existent rainfall-runoff models.

Further contribution to improved regionalization of rainfall would definitely result from a better meteorological network, or the use of remote sensing techniques like rainfall radar. Both approaches are unrealistic in so far as neither acquisition nor maintenance of such instruments is possible in many arid regions.

Future improvements in satellite-based precipitation analysis might increase the applicability of models that work with higher resolutions for space, and time.

At present, rainfall-runoff models which require area precipitation as inputs may not be the optimum solution for application in large arid basins. If the basin is of smaller size, and it is well-gauged, regionalization could possibly provide the data required by such models.

An alternative could be to employ models which do not require detailed rainfall information as inputs. An example is the NAMROM model, which is more of a statistical regression type. It relates runoff to rainfall by means of a single regression equation, controlled by two regression parameters.

The aim of this study was, among others, to quantify flood volumes with the intention of providing a better database for estimating groundwater recharge from the percolation of flood water into alluvial aquifers. A possible alternative method to estimate these losses to alluvial aquifers has already been presented by LANGE (2005) who applied a non-calibrated routing scheme to a reach of the Kuiseb River, deliberately excluding transmission losses from the model. Losses can then be estimated as the difference between measured and modelled streamflow.

In any case, until there is an improved database for rainfall-runoff modelling in arid regions, the focus for the assessment of groundwater recharge of alluvial aquifers should rather be on methods which take hydrometric data as inputs.

9 Literature

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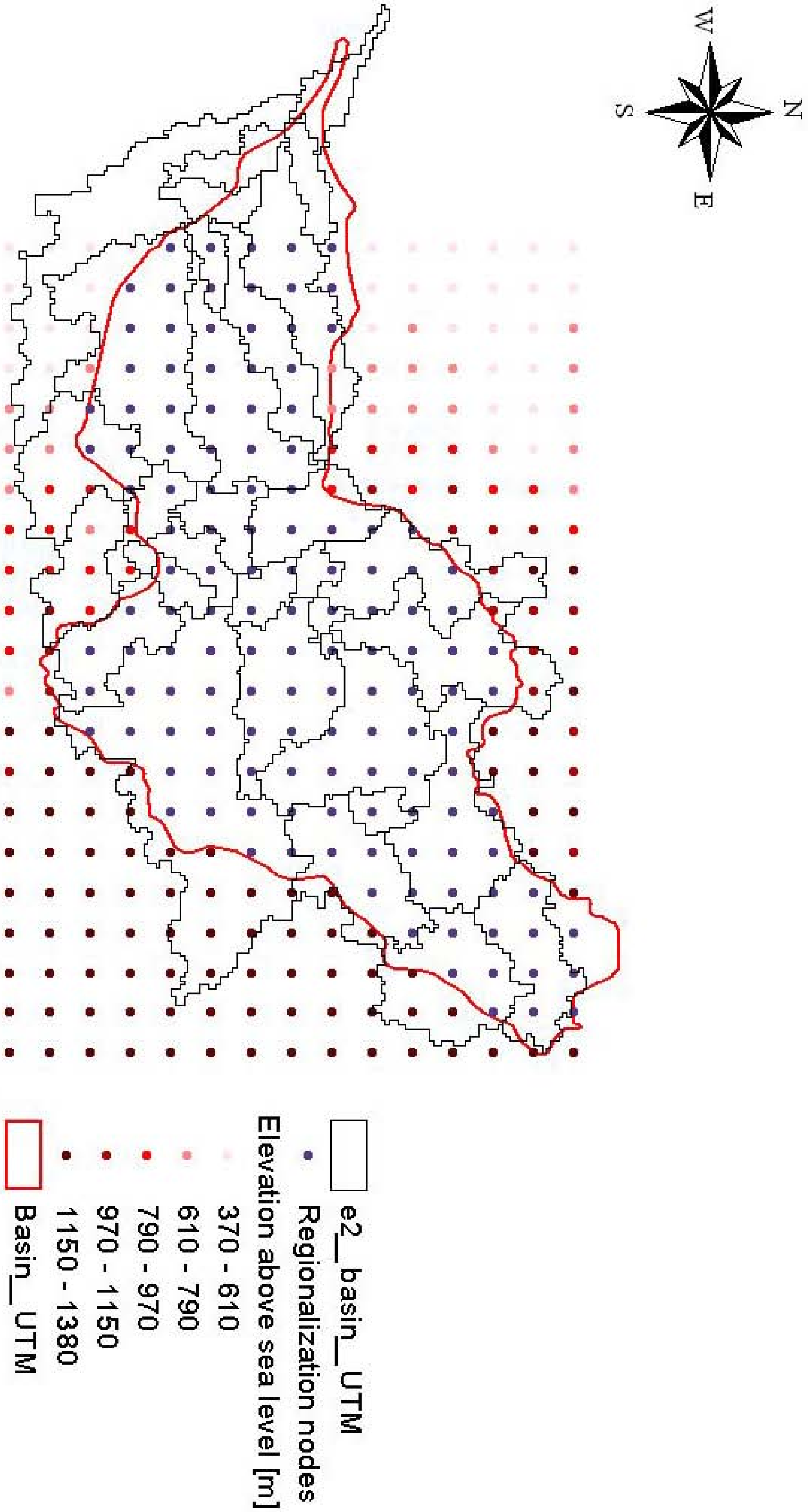
ANNEX 1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Regionalization nodes
Elevation above sea level [m]
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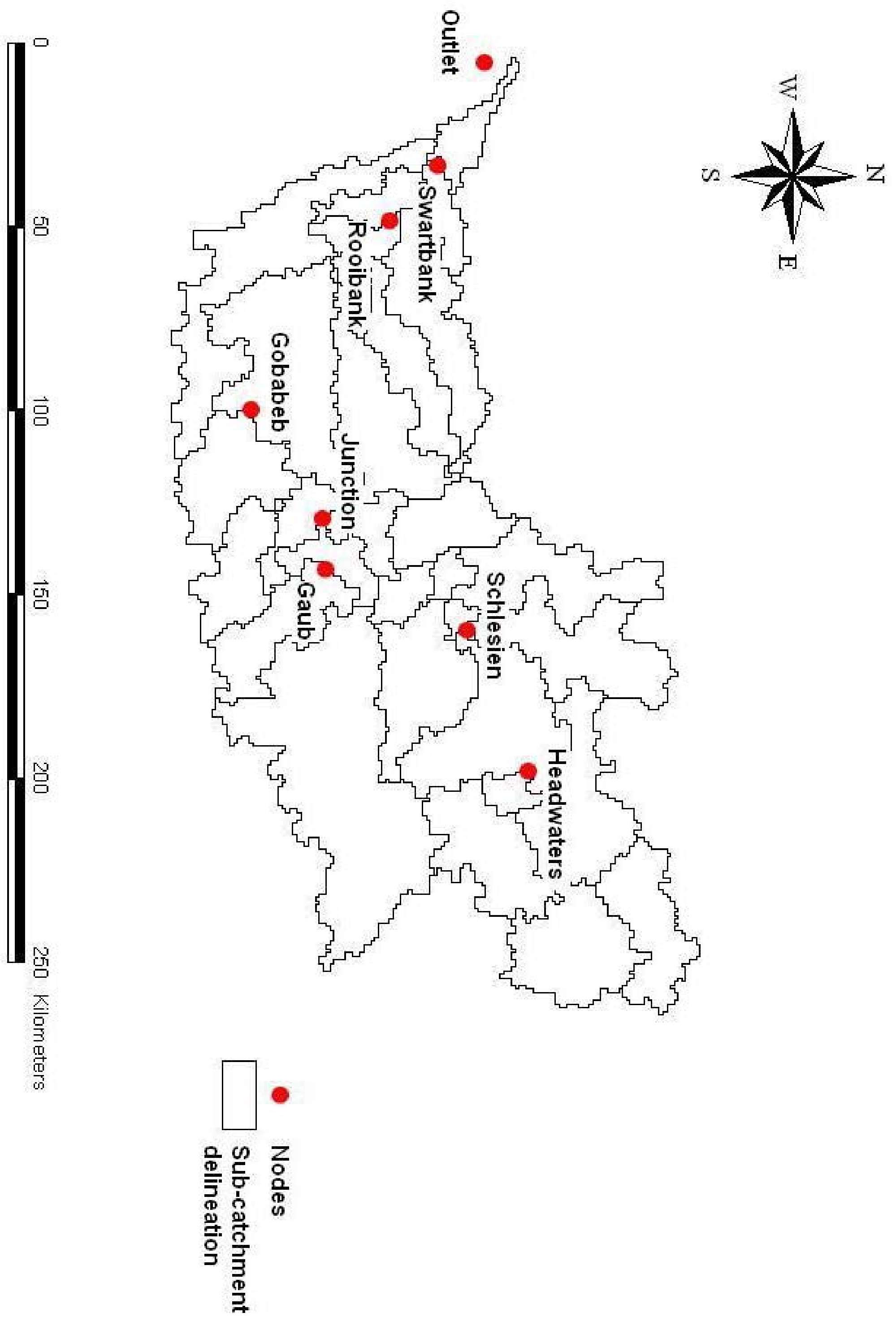
610 - 790
790 - 970
970 - 1150
1150 - 1380
Basin__UTM
N
S
W
E

ANNEX 2



W
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N
E
Outlet
Swartbank
Rooibank
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Gaub
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Headwaters
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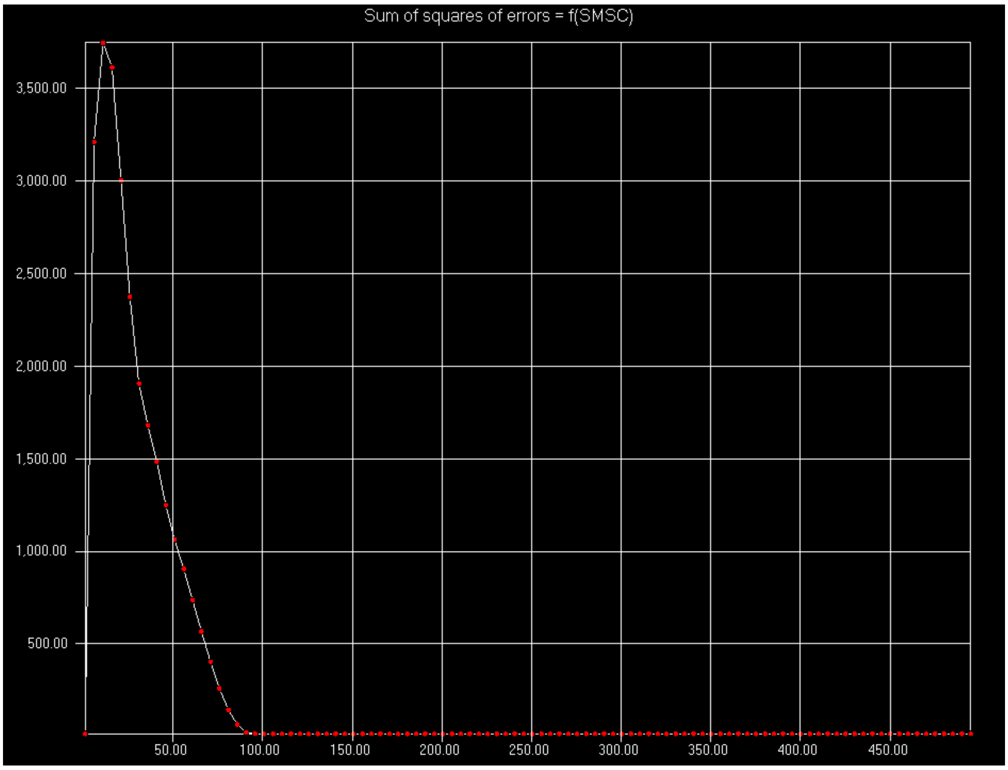
100
150
200
250
Kilometers
Nodes
Sub-catchment delineation

ANNEX 3



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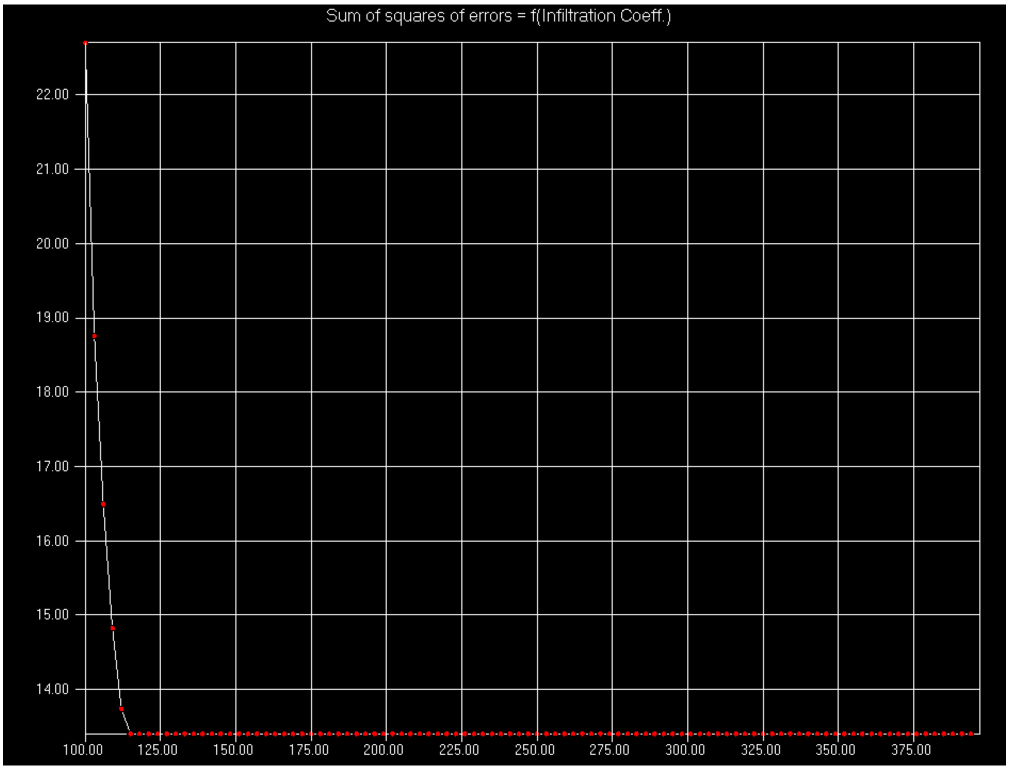

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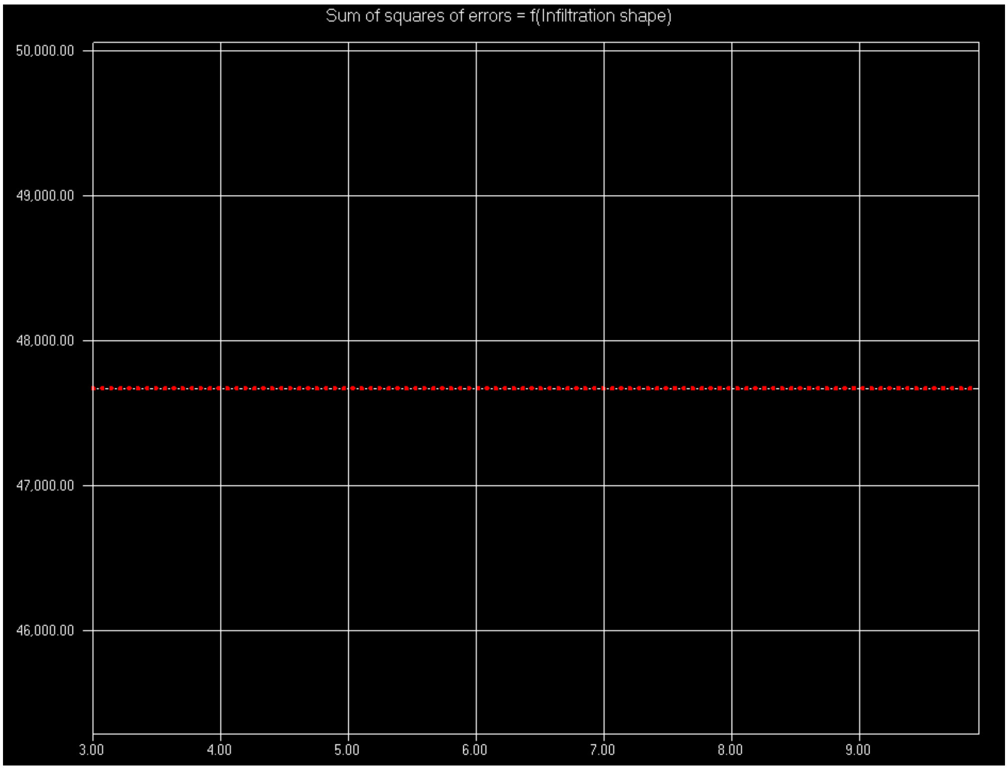

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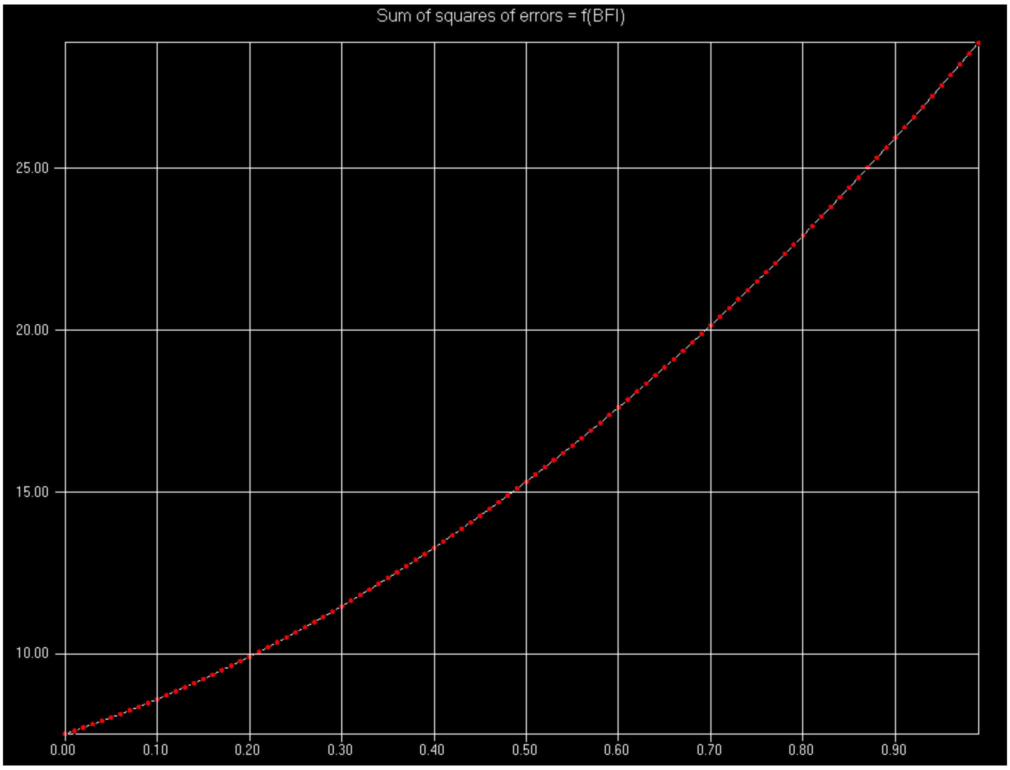

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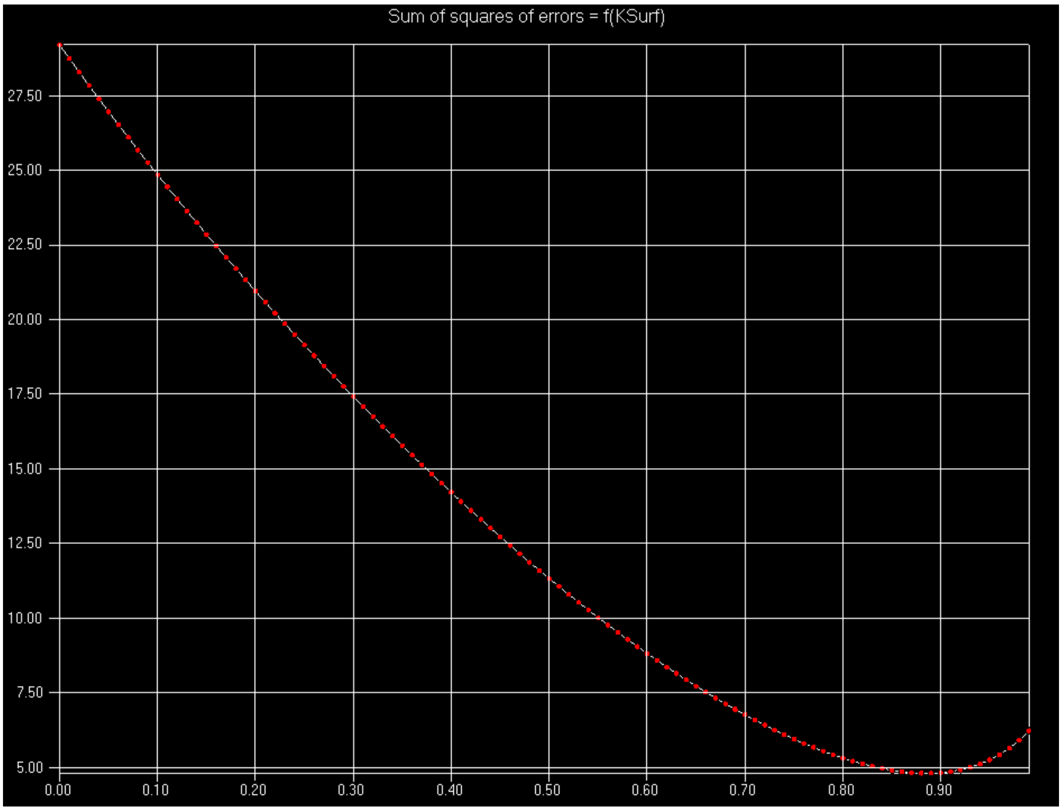

ANNEX 8