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Hydrological Processes of Arid and Semi-Arid Areas

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WADI HYDROLOGY
Hydrological Processes in Arid and Semi Arid Areas

Howard S. Wheater

Department of Civil and Environmental Engineering
Imperial College of Science, Technology and Medicine,
London SW7 2BU, UK

INTRODUCTION

The arid and semi-arid regions of the world are under severe and increasing pressure due to expanding populations, increasing per capita water use, and limited water resources. Increasing volumes of industrial and domestic waste provide a major threat to those scarce resources, and increasing development also creates new pressures for flood protection of lives and infrastructure. Effective management is essential, and this requires appropriate understanding of the hydrological processes in these areas.

Despite the critical importance of water in arid and semi-arid areas, hydrological data have historically been severely limited. It has been widely stated that the major limitation of the development of arid zone hydrology is the lack of high quality observations (McMahon, 1979; Nemec and Rodier, 1979; Pilgrim et al., 1988). There are many good reasons for this. Populations are usually sparse and economic resources limited; in addition the climate is harsh and hydrological events infrequent, but damaging. However, in the general absence of reliable long-term data and experimental research, there has been a tendency to rely on humid zone experience and modelling tools, and data from other regions. At best, such results will be highly inaccurate. At worst, there is a real danger of adopting inappropriate management solutions which ignore the specific features of dryland response.

Despite the general data limitations, there has been some substantial and significant progress in development of national data networks and experimental research. This has given new insights and we can now see with greater clarity the unique features of arid zone hydrological systems and the nature of the dominant hydrological processes. This provides an important opportunity to develop methodologies for flood and water resource management which are appropriate to the specific hydrological characteristics of arid areas and the associated management needs, and hence to define priorities for research and hydrological data. The aim of this introductory chapter is to review this progress and the resulting insights, and to consider some of the implications.

RAINFALL

Rainfall is the primary hydrological input, but rainfall in arid and semi-arid areas is commonly characterised by extremely high spatial and temporal variability. The temporal variability of point rainfall is well-known. Although most records are of relatively short length, a few are available from the 19th century. For example, Table I presents illustrative data from Muscat (Sultanate of Oman) (Wheater and Bell, 1983), which shows that a wet month is one with one or two raindays. Annual variability is marked and observed daily maxima can exceed annual rainfall totals.

For spatial characteristics, information is much more limited. Until recently, the major source of detailed data has been from the South West U.S.A., most notably the two, relatively small, densely instrumented basins of Walnut Gulch, Arizona (150km^2) and Alamogordo Creek, New Mexico (174km^2), established in the 1950s (Osborn et al., 1979). The dominant rainfall for these basins is convective; at Walnut Gulch 70% of annual rainfall occurs from purely convective cells, or from convective cells developing along weak, fast-moving cold fronts, and falls in the period July to September (Osborn and Reynolds, 1963). Raingauge densities were increased at Walnut Gulch to give improved definition of detailed storm structure and are currently better than 1 per 2km^2. This has shown highly localised rainfall occurrence, with spatial correlations of storm rainfall of the order of 0.8 at 2km separation, but close to zero at 15-20km spacing. Osborn et al. (1972) estimated that to observe a correlation of $r^2 = 0.9$, raingauge spacings of 300-500m would be required.

Recent work has considered some of the implications of the Walnut Gulch data for hydrological modelling. Michaud and Sorooshian (1994) evaluated problems of spatial averaging for rainfall-runoff modelling in the context of flood prediction. Spatial averaging on a 4kmx4km pixel basis (consistent with typical weather radar resolution) gave an underestimation of intensity and led to a reduction in simulated runoff of on average 50% of observed peak flows. A sparse network of raingauges (1 per 20km^2), representing a typical density of flash flood warning system, gave errors in simulated peak runoff of 58%. Evidently there are major implications for hydrological practice, and we will return to this issue, below.

The extent to which this extreme spatial variability is characteristic of other arid areas has been uncertain. Anecdotal evidence from the Middle East underlay comments that spatial and temporal variability was extreme (FAO, 1981), but recent data from South West Saudi Arabia obtained as part of a five-year intensive study of five basins (Saudi Arabian Dames and Moore, 1988), undertaken on behalf of the Ministry of Agriculture and Water, Riyadh, have provided a quantitative basis for assessment. The five study basins range in area from 456 to 4930 km^2 and are located along the Asir escarpment (Fig 1), three draining to the Red Sea, two to the interior, towards the Rub al Khali. The mountains have elevations of up to 3000m a.s.l., hence the basins encompass a wide range of altitude, which is matched by a marked gradient in annual rainfall, from 30-100mm on the Red Sea coastal plain to up to 450mm at elevations in excess of 2000m a.s.l.

The spatial rainfall distributions are described by Wheater et al. (1991a). The extreme spottiness of the rainfall is illustrated for the 2869km^2 Wadi Yiba by the frequency distributions of the number of gauges at which rainfall was observed given the occurrence of a catchment rainday (Table 2). Typical inter-gauge spacings were 8-10km, and on 51% of raindays only one or two raingauges out of 20 experienced rainfall. For the more widespread events, sub-daily rainfall showed an even more spotty picture than the daily distribution. An analysis of relative probabilities of rainfall occurrence, defined as the probability of rainfall occurrence for a given hour at Station B given rainfall at Station A, gave a mean value of 0.12 for Wadi Yiba, with only 5% of values greater that 0.3. The frequency distribution of rainstorm durations shows a typical occurrence of one or two-hour duration point rainfalls, and these tend to occur in mid-late afternoon. Thus rainfall will occur at a few gauges and die out, to be succeeded by rainfall in other locations. This is illustrated for Wadi Lith in Figure 2, which shows the daily rainfall totals for the storm of 16th May 1984 (Fig2a), and the individual hourly depths (Figs 2b-e). In general, the storm patterns appear to be consistent with the results from the South West USA and areal reduction factors were also generally consistent with results from that region (Wheater et al., 1989).

The effects of elevation were investigated, but no clear relationship could be identified for intensity or duration. However, a strong relationship was noted between the frequency of raindays and elevation. It was thus inferred that once rainfall occurred, its point properties were similar over the catchment, but occurrence was more likely at the higher elevations. It is interesting to note that a similar result has emerged from a recent analysis of rainfall in Yemen (UNDP, 1992), in which it was concluded that daily rainfalls observed at any location are effectively samples from a population that is independent of position or altitude.

It is dangerous to generalise from samples of limited record length, but it is clear that most events observed by those networks are characterized by extremely spotty rainfall, so much so that in the Saudi Arabian basins there were examples of wadi flows generated from zero observed rainfall. However, there were also some indications of a small population of more wide-spread rainfalls, which would obviously be of considerable importance in terms of surface flows and recharge. This reinforces the need for long-term monitoring of experimental networks to characterise spatial variability.

For some other arid or semi-arid areas, rainfall patterns may be very different. For example, data from arid New South Wales, Australia have indicated spatially extensive, low intensity rainfalls (Cordery et al., 1983), and recent research in the Sahelian zone of Africa has also indicated a predominance of widespread rainfall. This was motivated by concern to develop improved understanding of land-surface processes for climate studies and modelling, which led to a detailed (but relatively short-term) international experimental programme, the HAPEX-Sahel project based on Niamey, Niger (Goutorbe et al., 1997). Although designed to study land surface/atmosphere interactions, rather than as an integrated hydrological study, it has given important information. For example, Lebel et al. (1997) and Lebel and Le Barbe (1997) note that a 100 raingauge network was installed and report information on the classification of storm types, spatial and temporal variability of seasonal and event rainfall, and storm movement. 80% of total seasonal rainfall was found to fall as widespread events which covered at least 70% of the network. The number of gauges allowed the authors to analyse the uncertainty of estimated areal rainfall as a function of gauge spacing and rainfall depth.

Recent work in southern Africa (Andersen et al., 1998, Mocke, 1998) has been concerned with rainfall inputs to hydrological models to investigate the resource potential of the sand rivers of N.E.Botswana. Here, annual rainfall is of the order of 600mm, and available rainfall data is spatially sparse, and apparently highly variable, but of poor data quality. Investigation of the representation of spatial rainfall for distributed water resource modelling showed that use of convential methods of spatial weighting of raingauge data, such as Theissen polygons, could give large errors. Large sub-areas had rainfall defined by a single, possibly inaccurate gauge. A more robust representation resulted from assuming catchment-average rainfall to fall uniformly, but the resulting accuracy of simulation was still poor.

RAINFALL-RUNOFF PROCESSES

The lack of vegetation cover in arid and semi-arid areas removes protection of the soil from raindrop impact, and soil crusting has been shown to lead to a large reduction in infiltration capacity for bare soil conditions (Morin and Benyamini, 1977). Hence infiltration of catchment soils can be limited. In combination with the high intensity, short duration convective rainfall discussed above, extensive overland flow can be generated. This overland flow, concentrated by the topography, converges on the wadi channel network, with the result that a flood flow is generated. However, the runoff generation process due to convective rainfall is likely to be highly localised in space, reflecting the spottiness of the spatial rainfall fields, and to occur on only part of a catchment, as illustrated above.

Linkage between inter-annual variability of rainfall, vegetation growth and runoff production may occur. Current modelling in Botswana suggests that runoff production is lower in a year which follows a wet year, due to enhanced vegetation cover, which supports observations reported by Hughes (1995).

Commonly, flood flows move down the channel network as a flood wave, moving over a bed that is either initially dry or has a small initial flow. Hydrographs are typically characterised by extremely rapid rise times, of as little as 15-30 minutes (Fig 3). However, losses from the flood hydrograph through bed infiltration are an important factor in reducing the flood volume as the flood moves downstream. These transmission losses dissipate the flood, and obscure the interpretation of observed hydrographs. It is not uncommon for no flood to be observed at a gauging station, when further upstream a flood has been generated and lost to bed infiltration.

As noted above, the spotty spatial rainfall patterns observed in Arizona and Saudi Arabia are extremely difficult, if not impossible, to quantify using conventional densities of raingauge network. This, taken in conjunction with the flood transmission losses, means that conventional analysis of rainfall-runoff relationships is problematic, to say the least. Wheater and Brown (1989) present an analysis of Wadi Ghat, a 597 km^2 sub-catchment of wadi Yiba, one of the Saudi Arabian basins discussed above. Areal rainfall was estimated from 5 raingauges and a classical unit hydrograph analysis was undertaken. A striking illustration of the ambiguity in observed relationships is the relationship between observed rainfall depth and runoff volume (Figure 4). Runoff coefficients ranged from 5.9 to 79.8%, and the greatest runoff volume was apparently generated by the smallest observed rainfall! Goodrich et al. (1997) show that the combined effects of limited storm areal coverage and transmission loss give important differences from more humid regions. Whereas generally basins in more humid climates show increasing linearity with increasing scale, the response of Walnut Gulch becomes more **non-linear** with increasing scale. It is argued that this will give significant errors in application of rainfall depth-area-frequency relationships beyond the typical area of storm coverage, and that channel routing and transmission loss must be explicitly represented in watershed modelling.

The transmission losses from the surface water system are a major source of potential groundwater recharge. The characteristics of the resulting groundwater resource will depend on the underlying geology, but bed infiltration may generate shallow water tables, within a few metres of the surface, which can sustain supplies to nomadic people for a few months (as in the Hesse of the North of South Yemen), or recharge substantial alluvial aquifers with potential for continuous supply of major towns (as in Northern Oman and S.W. Saudi Arabia).

The balance between localised recharge from bed infiltration and diffuse recharge from rainfall infiltration of catchment soils will vary greatly depending on local circumstances. However, soil moisture data from Saudi Arabia (Macmillan, 1987) and Arizona (Liu et al., 1995), for example, show that most of the rainfall falling on soils in arid areas is subsequently lost by evaporation. Methods such as the chloride profile method (e.g. Bromley et al., 1997) and isotopic analyses (Allison and Hughes, 1978) have been used to quantify the residual percolation to groundwater in arid and semi-arid areas.

In some circumstances runoff occurs within an internal drainage basin, and fine deposits can support widespread surface ponding. A well known large-scale example is the Azraq oasis in N.E. Jordan, but small-scale features (Qaa's) are widespread in that area. Small scale examples were found in the HAPEX-Sahel study (Desconnets et al., 1997). Infiltration from these areas is in general not well understood, but may be extremely important for aquifer recharge. Desconnets et al. report aquifer recharge of between 5 and 20% of basin precipitation for valley bottom pools, depending on the distribution of annual rainfall.

The characteristics of the channel bed infiltration process are discussed in the following section. However, it is clear that the surface hydrology generating this recharge is complex and extremely difficult to quantify using conventional methods of analysis.

WADI BED TRANSMISSION LOSSES

Wadi bed infiltration has an important effect on flood propagation, but also provides recharge to alluvial aquifers. The balance between distributed infiltration from rainfall and wadi bed infiltration is obviously dependant on local conditions, but soil moisture observations from S.W. Saudi Arabia imply that, at least for frequent events, distributed infiltration of catchment soils is limited, and that increased near surface soil moisture levels are subsequently depleted by evaporation. Hence wadi bed infiltration may be the dominant process of groundwater recharge. As noted above, depending on the local hydrogeology, alluvial groundwater may be a readily accessible water resource. Quantification of transmission loss is thus important, but raises a number of difficulties.

One method of determining the hydraulic properties of the wadi alluvium is to undertake infiltration tests. Infiltrometer experiments give an indication of the saturated hydraulic conductivity of the surface. However, if an infiltration experiment is combined with measurement of the vertical distribution of moisture content, for example using a neutron probe, inverse solution of a numerical model of unsaturated flow can be used to identify the unsaturated hydraulic conductivity relationships and moisture characteristic curves. This is illustrated for the Saudi Arabian Five Basins Study by Parissopoulos and Wheater (1992a).

In practice, spatial heterogeneity will introduce major difficulties to the up-scaling of point profile measurements. The presence of silt lenses within the alluvium was shown to have important effects on surface infiltration as well as sub-surface redistribution (Parissopoulos and Wheater, 1990), and sub-surface heterogeneity is difficult and expensive to characterise. In a series of two-dimensional numerical experiments it was shown that "infiltration opportunity time", i.e. the duration and spatial extent of surface wetting, was more important than high flow stage in influencing infiltration, that significant reductions in infiltration occur once hydraulic connection is made with a water table, and that hysteresis effects were generally small (Parissopoulos and Wheater, 1992b). Also sands and gravels appeared effective in restricting evaporation losses from groundwater (Parissopoulos and Wheater, 1991).

Additional process complexity arises, however. General experience from the Five Basins Study was that wadi alluvium was highly transmissive, yet observed flood propagation indicated significantly lower losses than could be inferred from in situ hydraulic properties, even allowing for sub-surface heterogeneity. Possible causes are air entrapment, which could restrict infiltration rates, and the unknown effects of bed mobilisation and possible pore blockage by the heavy sediment loads transmitted under flood flow conditions.

A commonly observed effect is that in the recession phase of the flow, deposition of a thin (1-2mm) skin of fine sediment on the wadi bed occurs, which is sufficient to sustain flow over an unsaturated and transmissive wadi bed. Once the flow has ceased, this skin dries and breaks up so that the underlying alluvium is exposed for subsequent flow events. Crerar et al., (1988) observed from laboratory experiments that a thin continous silt layer was formed at low velocities. At higher velocities no such layer occurred, as the bed surface was mobilised, but infiltration to the bed was still apparently inhibited. It was suggested that this could be due to clogging of the top layer of sand due to silt in the infiltrating water, or formation of a silt layer below the mobile upper part of the bed.

Further evidence for the heterogeneity of observed response comes from the observations of Hughes and Sami (1992) from a 39.6 km^2 semi-arid catchment in S.Africa. Soil moisture was monitored by neutron probe following two flow events. At some locations immediate response (monitored 1day later) occurred throughout the profile, at others, an immediate response near surface was followed by a delayed response at depth. Away from the inundated area, delayed response, assumed due to lateral subsurface transmission, occurred after 21 days.

The overall implication of the above observations is that it is not possible at present to extrapolate from in-situ point profile hydraulic properties to infer transmission losses from wadi channels. However, analysis of observed flood flows at different locations can allow quantification of losses, and studies by Walters (1990) and Jordan (1977), for example, provide evidence that the rate of loss is linearly related to the volume of surface discharge.

For S.W. Saudi Arabia, the following relationships were defined:-

$$\text{LOSSL} = 4.56 + 0.02216\ \text{UPSQ} - 2034\ \text{SLOPE} + 7.34\ \text{ANTEC}$$
(s.e. 4.15)

$$\text{LOSSL} = 3.75 \times 10^{-5}\ \text{UPSQ}^{0.821}\ \text{SLOPE}^{-0.865}\ \text{ACWW}^{0.497}$$
(s.e. 0.146 log units (±34%))

$$\text{LOSSL} = 5.7 \times 10^{-5}\ \text{UPSQ}^{0.968}\ \text{SLOPE}^{-1.049}$$
(s.e. 0.184 loge units (±44%))

Where:-

LOSSL =	Transmission loss rate (1000m^3/km)	(O.R.1.08-87.9)
UPSQ =	Upstream hydrograph volume (1000$m^{3)}$	(O.R. 69-3744)
SLOPE =	Slope of reach (m/m)	(O.R.0.001-0.011)
ANTEC =	Antecedent moisture index	(O.R. 0.10-1.00)
ACWW =	Active channel width (m)	(O.R. 25-231)
and O.R. =	Observed range	

However, generalisation from limited experience can be misleading. Wheater et al. (1997) analysed transmission losses between 2 pairs of flow gauges on the Walnut Gulch catchment for a ten year sequence and found that the simple linear model of transmission loss as proportional to upstream flow was inadequate. Considering the relationship:

$$V_x = V_0\,(1-\alpha)^x$$

where V_x is flow volume (m^3) at distance x downstream of flow volume V_0 and α represents the proportion of flow lost per unit distance, then α was found to decrease with discharge volume:

$$\alpha = 118.8\ (V_0)^{-0.71}$$

The events examined had a maximum value of average transmission loss of 4076 $m^3\ km^{-1}$ in comparison with the estimate of Lane et al. (1971) of 4800-6700 $m^3\ km^{-1}$ as an upper limit of available alluvium storage.

The role of available storage was also discussed by Telvari et al. (1998), with reference to the Fowler's Gap catchment in Australia. Runoff plots were used to estimate runoff production as overland flow for a 4km^2 basin. It was inferred that 7000 m^3 of overland flow becomes transmission loss and that once this alluvial storage is satisfied, approximately two-thirds of overland flow is transmitted downstream.

A similar concept was developed by Andersen et al. (1998) at larger scale for the sand rivers of Botswana, which have alluvial beds of 20-200m width and 2-20m depth. Detailed observations of water table response showed that a single major event after a seven weeks dry period was sufficient to fully satisfy available alluvial storage (the river bed reached full saturation within 10 hours). No significant drawdown occurred between subsequent events and significant resource potential remained throughout the dry season. It was suggested that two sources of transmission loss could be occurring, direct losses to the bed, limited by available storage, and losses through the banks during flood events.

It can be concluded that transmission loss is complex, that where deep unsaturated alluvial deposits exist the simple linear model as developed by Jordan (1977) and implicit in the results of Walters (1990) may be applicable, but that where alluvial storage is limited, this must be taken into account.

GROUNDWATER RECHARGE FROM EPHEMERAL FLOWS

The relationship between wadi flow transmission losses and groundwater recharge will depend on the underlying geology. The effect of lenses of reduced permeability on the infiltration process has been discussed and illustrated above, but once infiltration has taken place, the alluvium underlying the wadi bed is effective in minimising evaporation loss through capillary rise (the coarse structure of alluvial deposits minimises capillary effects). Thus Hellwig (1973), for example, found that dropping the water table below 60cm in sand with a mean diameter of 0.53mm effectively prevented evaporation losses, and Sorey and Matlock (1969) reported that measured evaporation rates from streambed sand were lower than those reported for irrigated soils.

Parrisopoulos and Wheater (1991) combined two-dimensional simulation of unsaturated wadi-bed response with Deardorff's (1977) empirical model of bare soil evaporation to show that evaporation losses were not in general significant for the water balance or water table response in short-term simulation (i.e. for periods up to 10 days). However, the influence of vapour diffusion was not explicitly represented, and long term losses are not well understood. Andersen et al. (1998) show that losses are high when the alluvial aquifer is fully saturated, but are small once the water table drops below the surface.

Sorman and Abdulrazzak (1993) provide an analysis of groundwater rise due to transmission loss for an experimental reach in Wadi Tabalah, S.W. Saudi Arabia and estimate that on average 75% of bed infiltration reaches the water table. There is in general little information available to relate flood transmission loss to groundwater recharge, however. The differences between the two are expected to be small, but will depend on residual moisture stored in the unsaturated zone and its subsequent drying characteristics. But if water tables approach the surface, relatively large evaporation losses may occur.

Again, it is tempting to draw over-general conclusions from limited data. In the study of the sand-rivers of Botswana, referred to above, it was expected that recharge of the alluvial river beds would involve complex unsaturated zone response. In fact, observations showed that the first flood of the wet season was sufficient to fully recharge the alluvial river bed aquifer. This storage was topped up in subsequent floods, and depleted by evaporation when the water table was near-surface, but in many sections sufficient water remained throughout the dry season to provide adequate sustainable water supplies for rural villages. And as noted above, Wheater et al. (1997) showed for Walnut Gulch and Telvari et al. (1998) for Fraser's Gap that limited river bed storage affected transmission loss. It is evident that surface water/groundwater interactions depend strongly on the local characteristics of the underlying alluvium and the extent of their connection to, or isolation from, other aquifer systems.

HYDROLOGICAL MODELLING AND THE REPRESENTATION OF RAINFALL

The preceding discussion illustrates some of the particular characteristics of arid areas which place special requirements on hydrological modelling, for example for flood management or water resources evaluation. One evident area of difficulty is rainfall, especially where convective storms are an important influence. The work of Michaud and Sorooshian (1994) demonstrated the sensitivity of flood peak simulation to the spatial resolution of rainfall input. This obviously has disturbing implications for flood modelling, particularly where data availability is limited to conventional raingauge densities. Indeed, it appears highly unlikely that suitable raingauge densities will ever be practicable for routine monitoring. However, the availability of 2km resolution radar data in the USA can provide adequate information and radar could be installed elsewhere for particular applications. Morin et al. (1995) report results from a radar located at Ben-Gurion airport in Israel, for example.

One way forward is to develop an understanding of the properties of spatial rainfall based on high density experimental networks and/or radar data, and represent those properties within a spatial rainfall model for more general application. It is likely that this would have to be done within a stochastic modelling framework in which equally-likely realisations of spatial rainfall are produced, possibly conditioned by sparse observations.

Some simple empirical first steps in this direction were taken by Wheater et al. (1991a,b) for S.W.Saudi Arabia and Wheater et al. (1995) for Oman. In the Saudi Arabian studies, as noted earlier, raingauge data was available at approximately 10km spacing and spatial correlation was low. Hence a multi-variate model was developed, assuming independence of raingauge rainfall. Based on observed distributions, seasonally-dependent catchment rainday occurrence was simulated, dependent on whether the preceding day was wet or dry. The number of gauges experiencing rainfall was then sampled, and the locations selected based on observed occurrences (this allowed for increased frequency of raindays with increased elevation). Finally, start-times, durations and hourly intensities were generated. Model performance was compared with observations. Rainfall from random selections of raingauges was well reproduced, but when clusters of adjacent gauges were evaluated, a degree of spatial organisation of occurrence was observed, but not simulated. It was evident that a weak degree of correlation was present, which should not be neglected. Hence in extension of this approach to Oman (Wheater et al., 1995), observed spatial distributions were sampled, with satisfactory results.

However, this multi-variate approach suffers from limitations of raingauge density, and in general a model in continuous space (and continuous time) is desirable. A family of stochastic rainfall models of point rainfall was proposed by Rodriguez-Iturbe, Cox and Isham (1987, 1988) and applied to UK rainfall by Onof and Wheater (1993,1994). The basic concept is that a Poisson process is used to generate the arrival of storms. Associated with a storm is the arrival of raincells, of uniform intensity for a given duration (sampled from specified distributions). The overlapping of these rectangular pulse cells generates the storm intensity profile in time. These models were shown to have generally good performance for the UK in reproducing rainfall properties at different time-scales (from hourly upwards), and extreme values.

Cox and Isham (1988) extended this concept to a model in space and time, whereby the raincells are circular and arrive in space within a storm region. As before, the overlapping of cells produces a complex rainfall intensity profile, now in space as well as time. This model has been developed further by Northrop (1998) to include elliptical cells and storms and is being applied to UK rainfall (Northrop et al., 1999).

Recent work (Samuel, 1999) has been exploring the capability of these models to reproduce the convective rainfall of Walnut Gulch. In modelling point rainfall, the Bartlett-Lewis Rectangular Pulse Model was generally slightly superior to other model variants tested. Table 3 shows representative performance of the model in comparing the hourly statistics from 500 realisations of July rainfall in comparison with 35 years from one of the Walnut Gulch gauges (gauge 44).

Table 3 Performance of the Bartlett-Lewis Rectangular Pulse Model in representing July rainfall at gauge 44, Walnut Gulch

	Mean	Var	ACF1	ACF2	ACF3	Pwet	Mint	Mno	Mdur
Model	0.103	1.082	0.193	0.048	0.026	0.032	51.17	14.34	1.68
Data	0.100	0.968	0.174	0.040	0.036	0.042	53.71	13.23	2.38

where Mean is the mean hourly rainfall (mm), Var its variance, ACF1,2,3 the autocorrelations for lags 1,2,3, Pwet the proportion of wet intervals, Mint the mean storm inter-arrival time (h), Mno the mean number of storms per month, Mdur the mean storm duration (h).
This performance is generally encouraging (although the mean storm duration is underestimated), and extreme value performance is excellent.

Work with the spatial-temporal model is still at a preliminary stage, but Fig 5 (Samuel, 1999) shows a comparison of observed spatial coverage of rainfall for 25 years of July data from 81 gauges (for different values of the standard deviation of cell radius) and Fig 6 (Samuel, 1999) the corresponding fit for temporal lag-0 spatial correlation. Again, the results are encouraging, and there is promise with this approach to address the significant problems of spatial representation for hydrological modelling.

INTEGRATED MODELLING FOR WATER RESOURCE EVALUATION

Appropriate strategies for water resource development must recognise the essential physical characteristics of the hydrological processes. Surface water storage, although subject to high evaporation losses, is widely used, although temporal variability of flows must be adequately represented to define long term yields. It can be noted that in some regions, for example, the northern areas of southern Yemen, small scale storage has been developed as an appropriate method to maximise the available resource from spatially-localised rainfall. Numbers of small storages have been developed, some of which fill from localised rainfall. These then provide a short-term resource for a nomadic family and its livestock.

Groundwater is a resource particularly well suited to arid regions. Subsurface storage minimises evaporation loss and can provide long-term yields from infrequent recharge events. The recharge of alluvial groundwater systems by ephemeral flows can provide an appropriate resource, and this has been widely recognised by traditional development, such as the "afalaj" of Oman and elsewhere. There may, however, be opportunities for augmenting recharge and more effectively managing these groundwater systems. In any case, it is essential to quantify the sustainable yield of such systems, for appropriate resource development.

It has been seen that observations of surface flow do not define the available resource, and similarly observed groundwater response does not necessarily indicate upstream recharge. Figure 7 presents a series of groundwater responses from 1985/86 for Wadi Tabalah, S.W. Saudi Arabia, which shows a downstream sequence of wells 3-B-96, -97, -

98, -99 and -100 and associated surface water discharges. It can be seen that there is little evidence of the upstream recharge at the downstream monitoring point.

In addition, records of surface flows and groundwater levels, coupled with ill-defined histories of abstraction, are generally insufficient to define long term variability of the available resource.

To capture the variability of rainfall and the effects of transmission loss on surface flows, a distributed approach is necessary. If groundwater is to be included, integrated modelling of surface water and groundwater is needed. Distributed surface water models include KINEROS (Wheater and Bell, 1983, Michaud and Sorooshian, 1994) and the model of Sharma (1997, 1998). A distributed approach to the integrated modelling of surface and groundwater response following Wheater et al.(1995) is illustrated in Fig 8. This requires the characterisation of the spatial and temporal variability of rainfall, distributed infiltration, runoff generation and flow transmission losses, the ensuing groundwater recharge and groundwater response. This presents some technical difficulties, although the integration of surface and groundwater modelling allows maximum use to be made of available information, so that, for example, groundwater response can feed back information to constrain surface hydrological parameterisation. It does, however, provide the only feasible method of exploring the internal response of a catchment to management options.

In a recent application, this integrated modelling approach was developed for Wadi Ghulaji, Sultanate of Oman, to evaluate options for groundwater recharge management (Wheater et al., 1995). The catchment, of area 758 km^2, drains the southern slopes of Jebal Hajar in the Sharqiyah region of Northern Oman. Proposals to be evaluated included recharge dams to attenuate surface flows and provide managed groundwater recharge in key locations. The modelling framework involved the coupling of a distributed rainfall model, a distributed water balance model (incorporating rainfall-runoff processes, soil infiltration and wadi flow transmission losses), and a distributed groundwater model (Fig 9).

The representation of rainfall spatial variability presents technical difficulties, since data are limited. Detailed analysis was undertaken of 19 rain gauges in the Sharqiyah region, and of six raingauges in the catchment itself. A stochastic multi-variate temporal- spatial model was devised for daily rainfall, a modified version of a scheme orgininally developed by Wheater et al., 1991a,b. The occurrence of catchment rainfall was determined according to a seasonally-variable first order markov process, conditioned on rainfall occurrence from the previous day. The number and locations of active raingauges and the gauge depths were derived by random sampling from observed distributions.

The distributed water balance model represents the catchment as a network of two-dimensional plane and linear channel elements. Runoff and infiltration from the planes was simulated using the SCS approach. Wadi flows incorporate a linear transmission loss algorithm based on work by Jordan (1977) and Walters (1990). Distributed calibration parameters are shown in Figure 10.

Finally, a groundwater model was developed based on a detailed hydrogeological investigation which led to a multi-layer representation of uncemented gravels, weakly/strongly cemented gravels and strongly cemented/fissured gravel/bedrock, using MODFLOW.

The model was calibrated to the limited flow data available (a single event), and was able to reproduce the distribution of runoff and groundwater recharge within the catchment through a rational association on loss parameters with topography, geology and wadi characteristics. Extended synthetic data sequences were then run to investigate catchment water balances under scenarios of different runoff exceedance probabilities (20%, 50%, 80%), as in Table 4, and to investigate management options.

CONCLUSIONS

It has been shown that for many applications, the hydrological characteristics of arid areas present severe problems for conventional methods of analysis. Recent data are providing new insights. These insights must be used as the basis for development of more appropriate methods for flood design and water resource evaluation, and in turn, to define data needs and research priorities. Much high quality research is needed, particularly to investigate processes such as spatial rainfall, and infiltration and groundwater recharge from ephemeral flows.

For developments to maximise the resource potential, define long-term sustainable yields and protect traditional sources, it is argued that distributed modelling is a valuable, if not essential tool. However, this confronts severe problems of characterisation of rainfall, rainfall-runoff processes, and groundwater recharge, and of understanding the detailed hydrogeological response of what are often complex groundwater systems. Similarly, new approaches to flood design and management are required which represent the extreme value characteristics of arid areas and recognise the severe problems of conventional rainfall-runoff analysis.

Above all, basic requirements are for high quality data of rainfall, surface water flows and groundwater response to support regional analyses and the development of appropriate methodologies. Too often, studies focus on either surface or subsurface response without taking an integrated view. Too often, networks are reduced after a few years without recognition that the essential variability of wadi response can only be characterised by relatively long records. Quality control of data is vital, but can easily be lost sight of with ready access to computerised data-bases.

Superimposed on these basic data needs are the requirement for specific process studies, including sediment transport, surface water/groundwater interactions in the active wadi channel, evaporation processes and consumptive use of wadi vegetation, and the wider issues of groundwater recharge. These are challenging studies, with particularly challenging logistical problems, and require the full range of advanced hydrological experimental methods to be applied, particularly integrating quantity and quality data to deduce system responses, and making full use of remote sensing and geophysical methods to characterise system properties.

It must not be forgotten that in general, data networks are under threat world-wide, and a major priority for hydrologists must be to promote recognition of the value of data for water management, the importance of long records in a region characterised by high inter-annual variability, and of the particular technical and logistical difficulties in capturing hydrological response in arid areas. The current International Hydrological Programme rightly prioritizes hydrological data as the essential foundation for effective management. The results of both detailed research and regional analyses are required for the essential understanding of wadi hydrology which must underlie effective management.

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Table 1. Summary of Muscat rainfall data (1893-1959) (after Wheater and Bell, 1983)

Monthly rainfall mm.	Jan.	Feb.	Mar.	Apr.	May.	June	July	Aug.	Sept.	Oct.	Nov.	Dec.
Mean.	31.2	19.1	13.1	8.00	0.38	1.31	0.96	0.45	0.00	2.32	7.15	22.0
Standard deviation	38.9	25.1	18.9	20.3	1.42	8.28	4.93	2.09	0.00	7.62	15.1	35.1
Max	143.0	98.6	70.4	98.3	8.89	64.0	37.1	14.7	0.00	44.5	77.2	171.2
Mean number of raindays	2.03	1.39	1.15	0.73	0.05	0.08	0.10	0.07	0.00	0.13	0.51	1.60
Max daily fall, mm	78.7	57.0	57.2	51.3	8.9	61.5	30.0	10.4	0.00	36.8	53.3	57.2
Number of years record	63	64	62	63	61	61	60	61	61	60	61	60

Table 2. Wadi Yiba raingauge frequencies and associated conditional probabilities for catchment rainday occurrence

Number of Gauges	Occurrence	Probability
1	88	0.372
2	33	0.141
3	25	0.106
4	18	0.076
5	10	0.042
6	11	0.046
7	13	0.055
8	6	0.026
9	7	0.030
10	5	0.021
11	7	0.030
12	5	0.021
13	3	0.013
14	1	0.004
15	1	0.004
16	1	0.005
17	1	0.004
18	1	0.004
19	0	0.0
20	0	0.0
TOTAL	235	1.000

Table 4. Annual catchment water balance, simulated scenarios

Annual Catchment Water Balance (Mm^3)					
Scenario	Rainfall	Evaporation	Groundwater Recharge	Runoff	%Runoff
Wet	87.7	70.9	12.8	4.0	4.6
Average	87.0	72.3	11.2	3.5	4.0
Dry	52.6	45.3	5.5	1.7	3.2

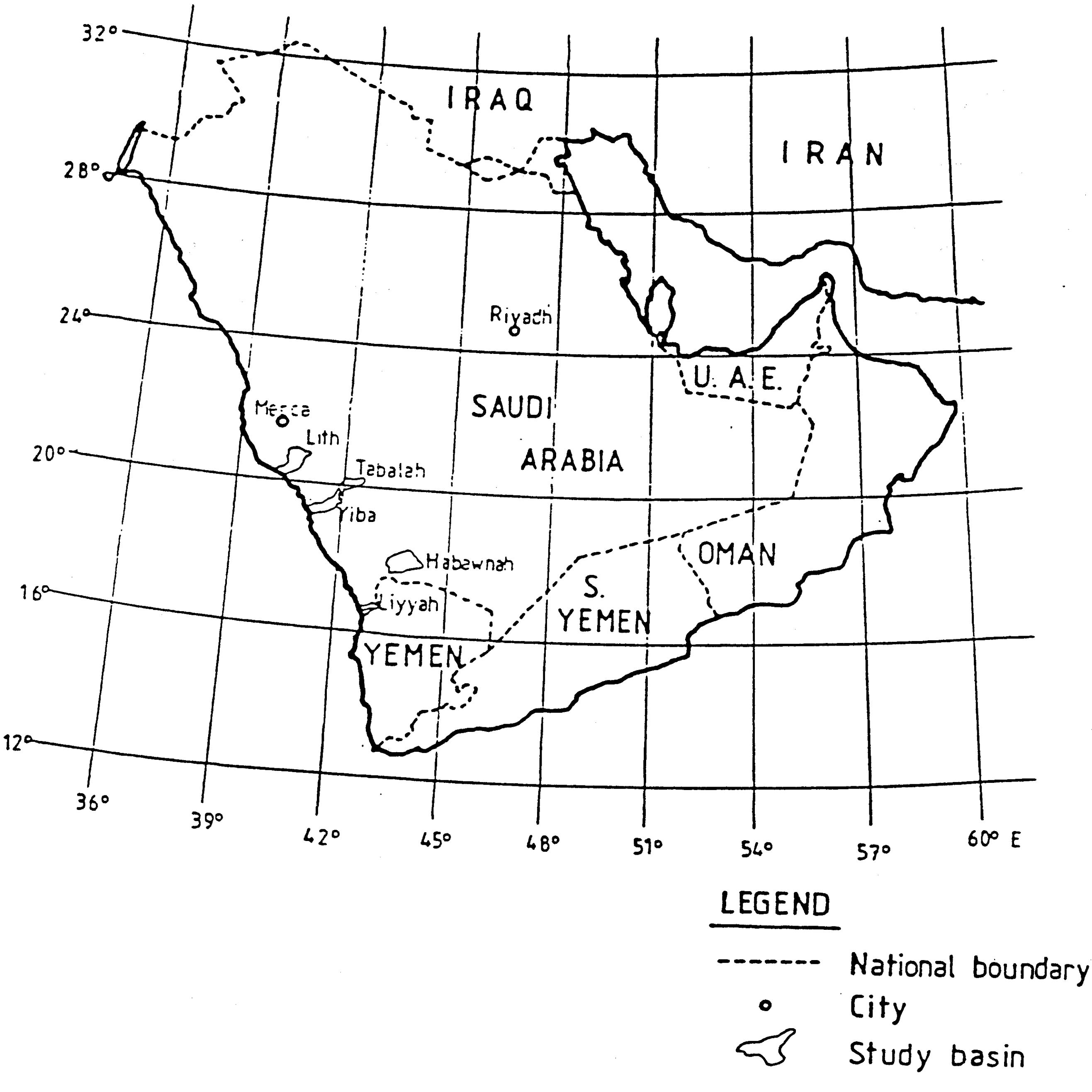


Figure 1

Location of Saudi Arabian study basins

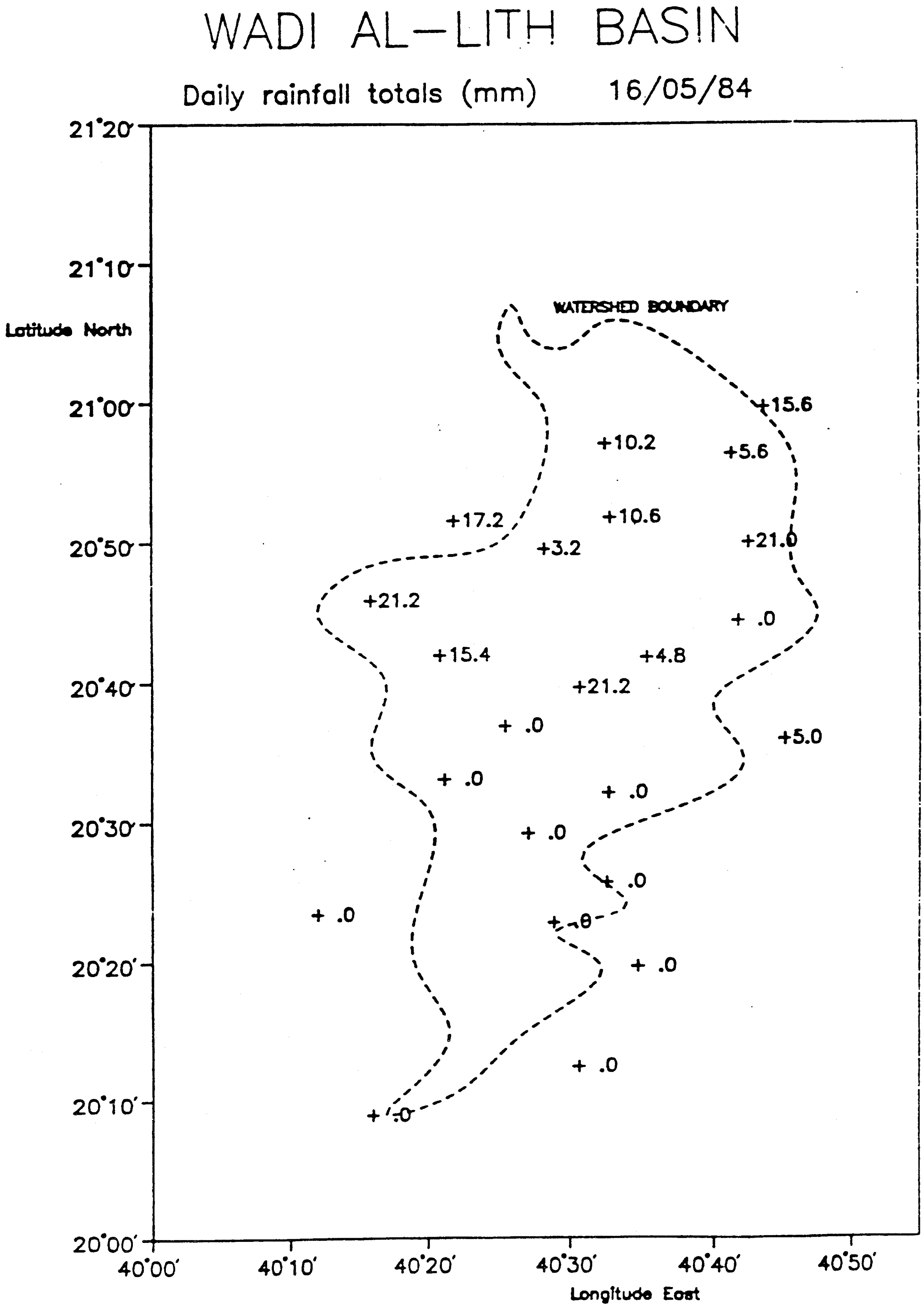


Figure 2a

Wadi Lith rainfall, 16th May 1984

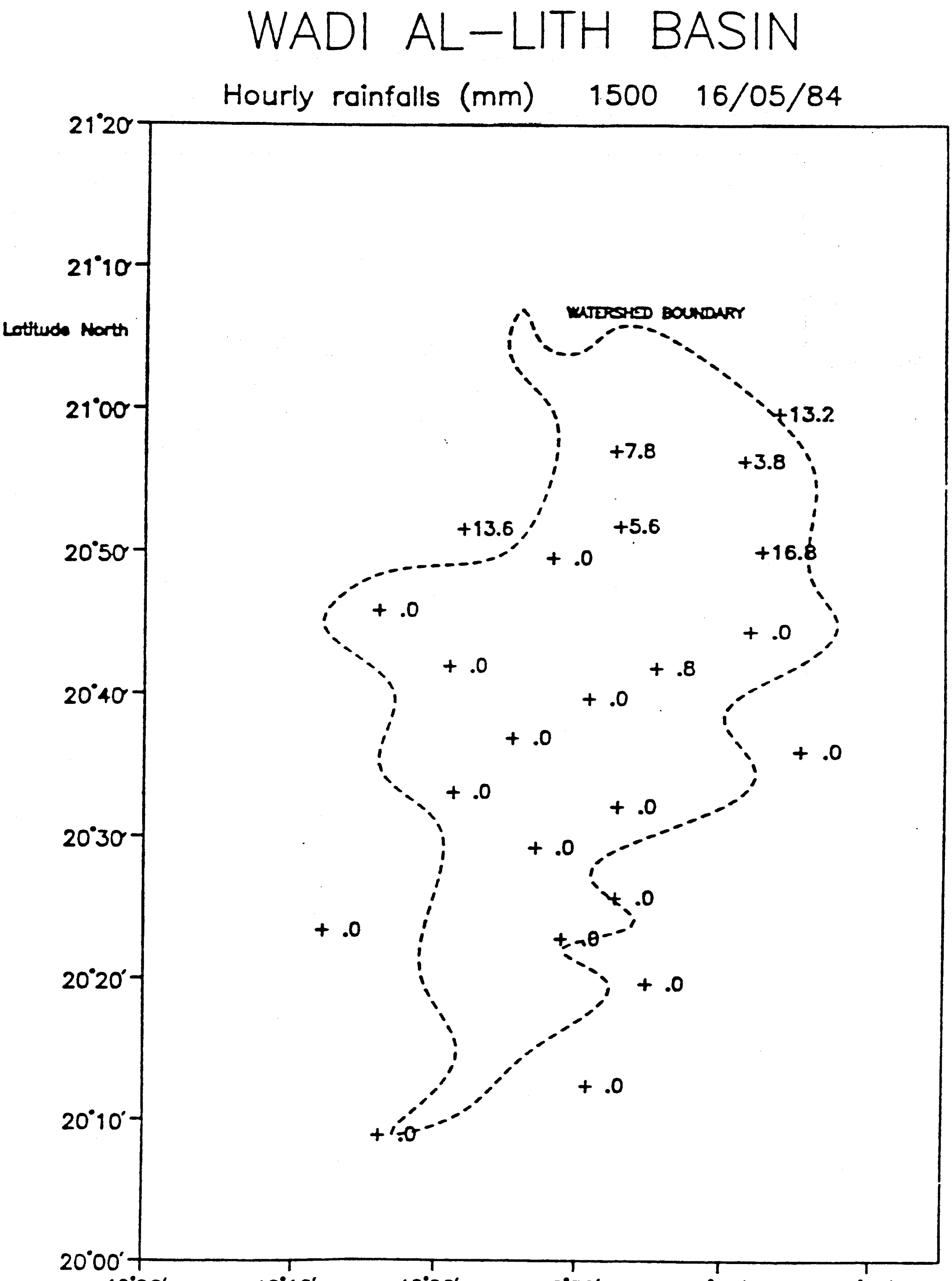

WADI AL-LITH BASIN
Hourly rainfalls (mm) 1500 16/05/84
21°20′
21°10′
Latitude North
21°00′
20°50′
20°40′
20°30′
20°20′
20°10′
20°00′
40°00′
40°10′
40°20′
40°30′
40°40′
40°50′
Longitude East
WATERSHED BOUNDARY
+13.2
+7.8
+3.8
+13.6
+5.6
+16.8
+ .0
+ .0
+ .0
+ .0
+ .8
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0

Figure 2b

WADI AL-LITH BASIN

Hourly rainfalls (mm) 1600 16/05/84

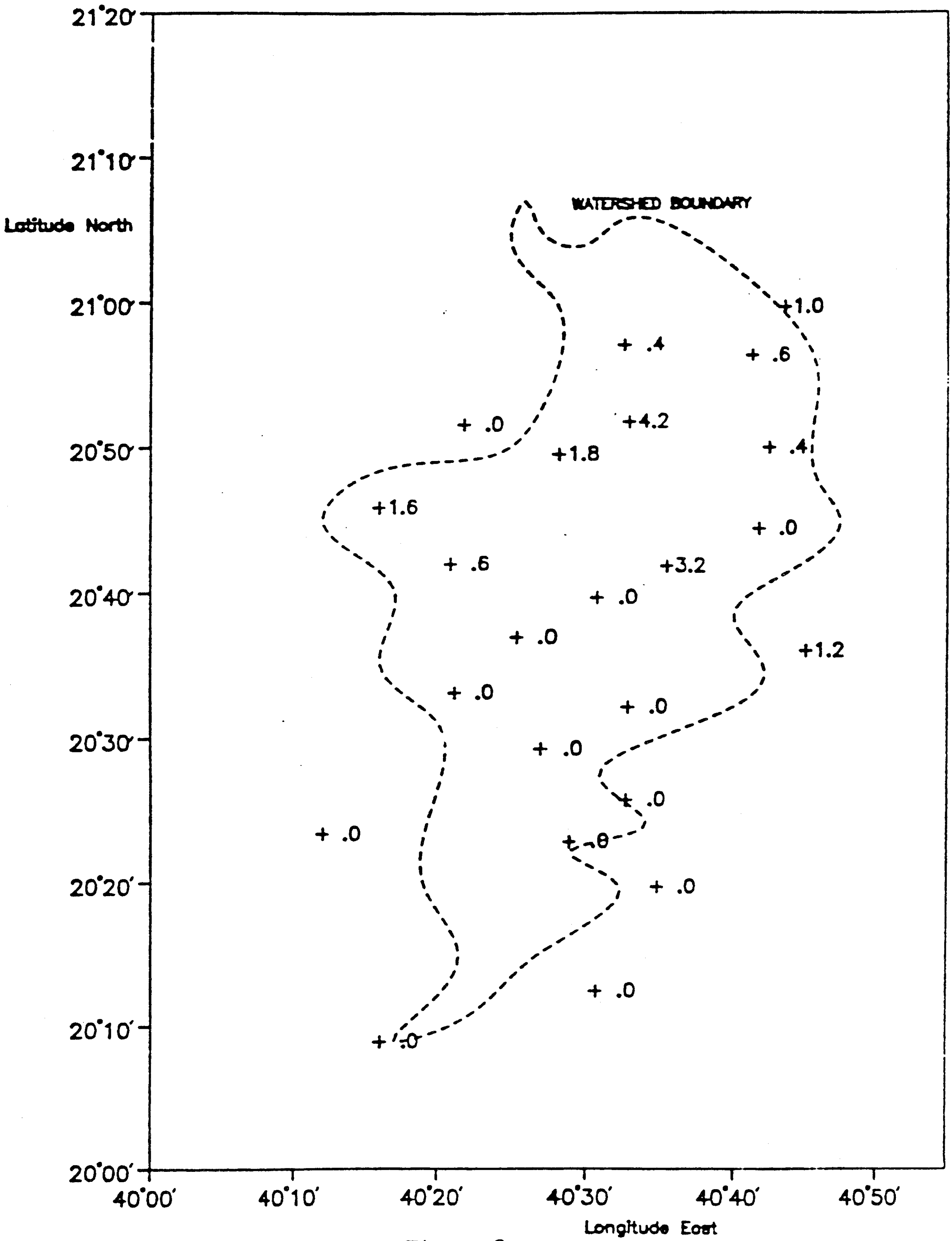


Figure 2c

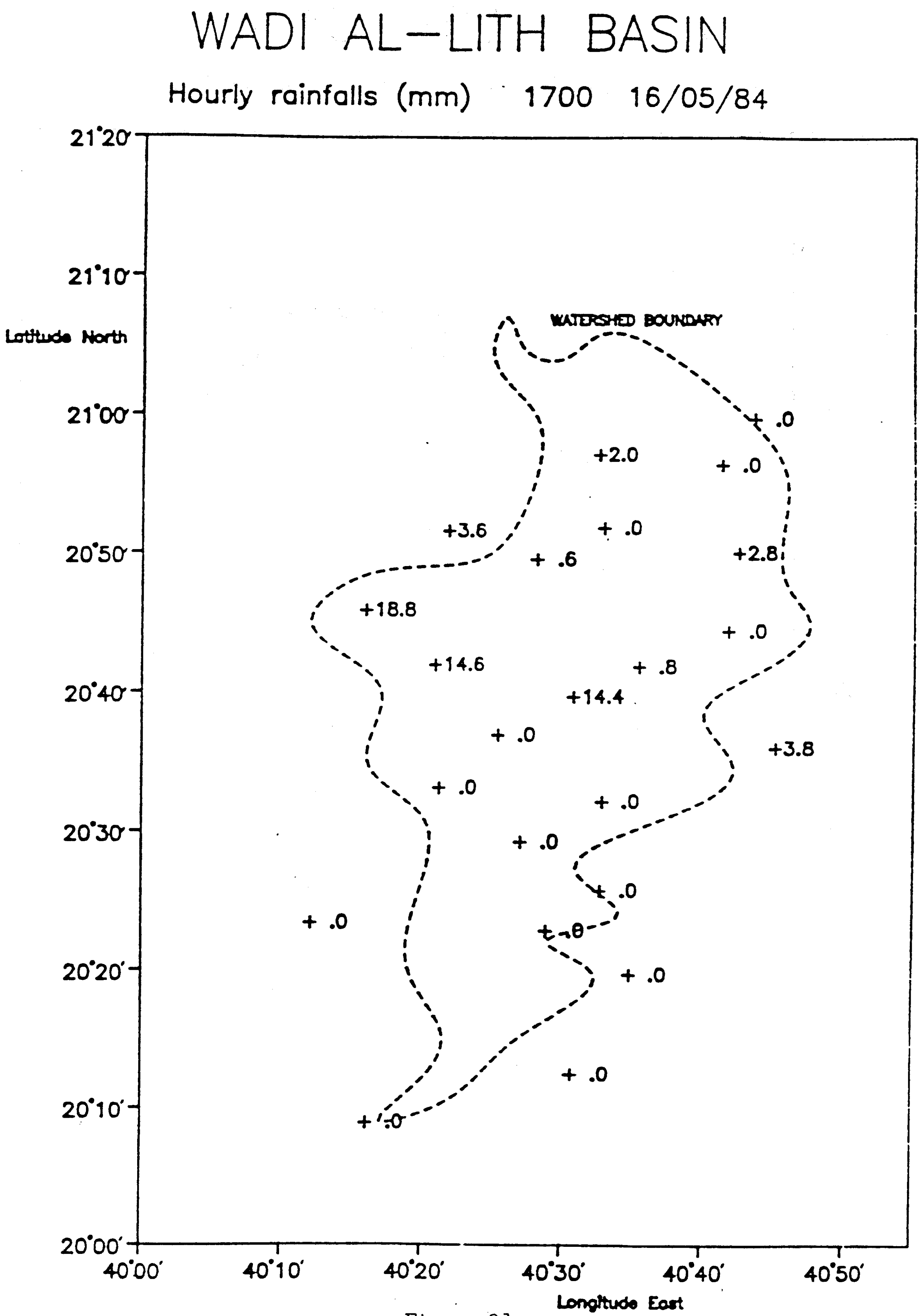

WADI AL-LITH BASIN
Hourly rainfalls (mm) 1700 16/05/84
WATERSHED BOUNDARY
Latitude North
Longitude East
21°20'
21°10'
21°00'
20°50'
20°40'
20°30'
20°20'
20°10'
20°00'
40°00'
40°10'
40°20'
40°30'
40°40'
40°50'
+ .0
+2.0
+ .0
+3.6
+ .0
+ .6
+2.8
+18.8
+ .0
+14.6
+ .8
+14.4
+ .0
+3.8
+ .0
+ .0
+ .0
+ .0
+ .0
+ .8
+ .0
+ .0
+ .0

Figure 2d

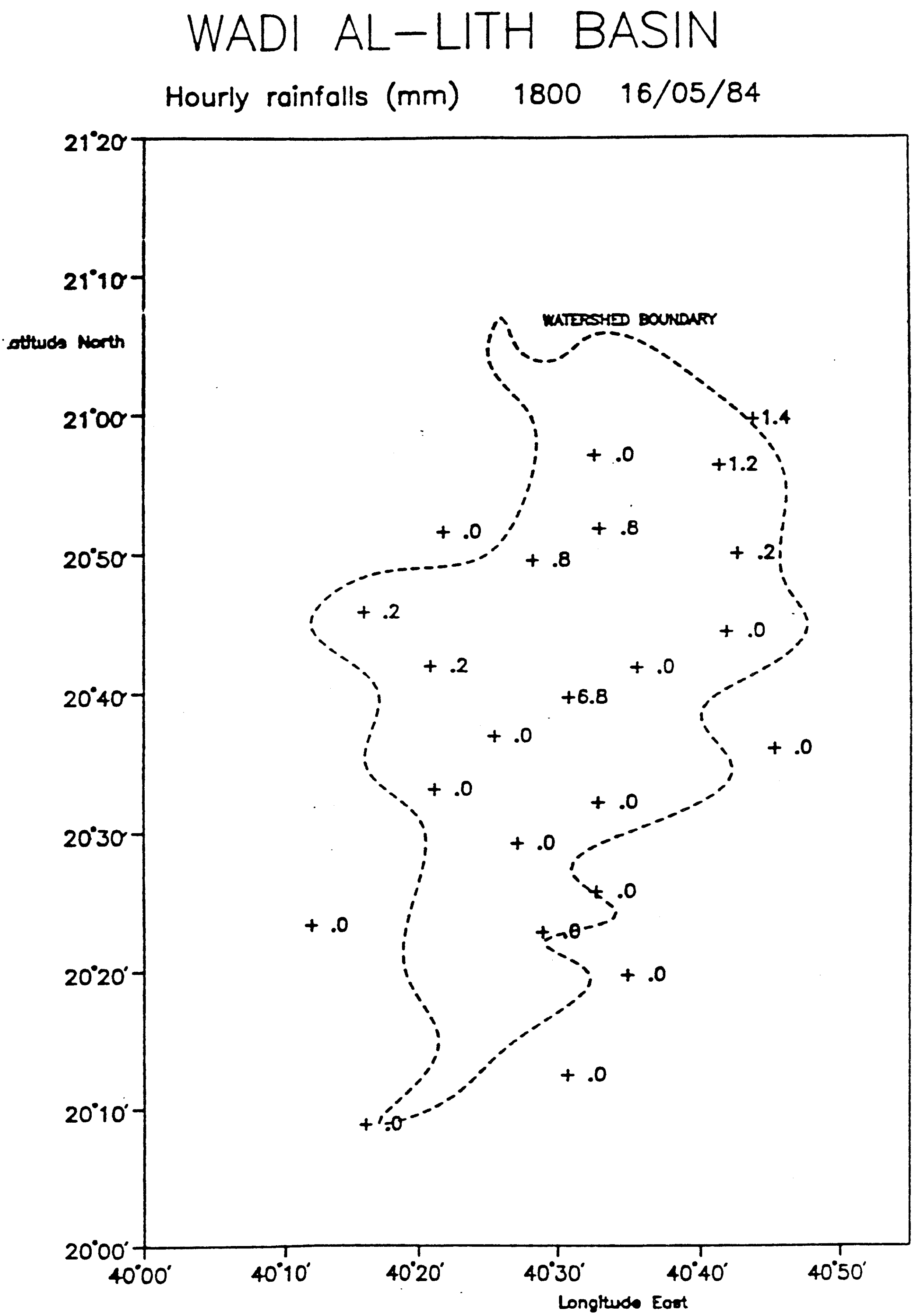

WADI AL-LITH BASIN
Hourly rainfalls (mm) 1800 16/05/84
WATERSHED BOUNDARY
Latitude North
Longitude East
21°20′
21°10′
21°00′
20°50′
20°40′
20°30′
20°20′
20°10′
20°00′
40°00′
40°10′
40°20′
40°30′
40°40′
40°50′
+1.4
+ .0
+1.2
+ .0
+ .8
+ .8
+ .2
+ .2
+ .0
+ .2
+ .0
+6.8
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .0
+ .8
+ .0
+ .0
+ .0

Figure 2e

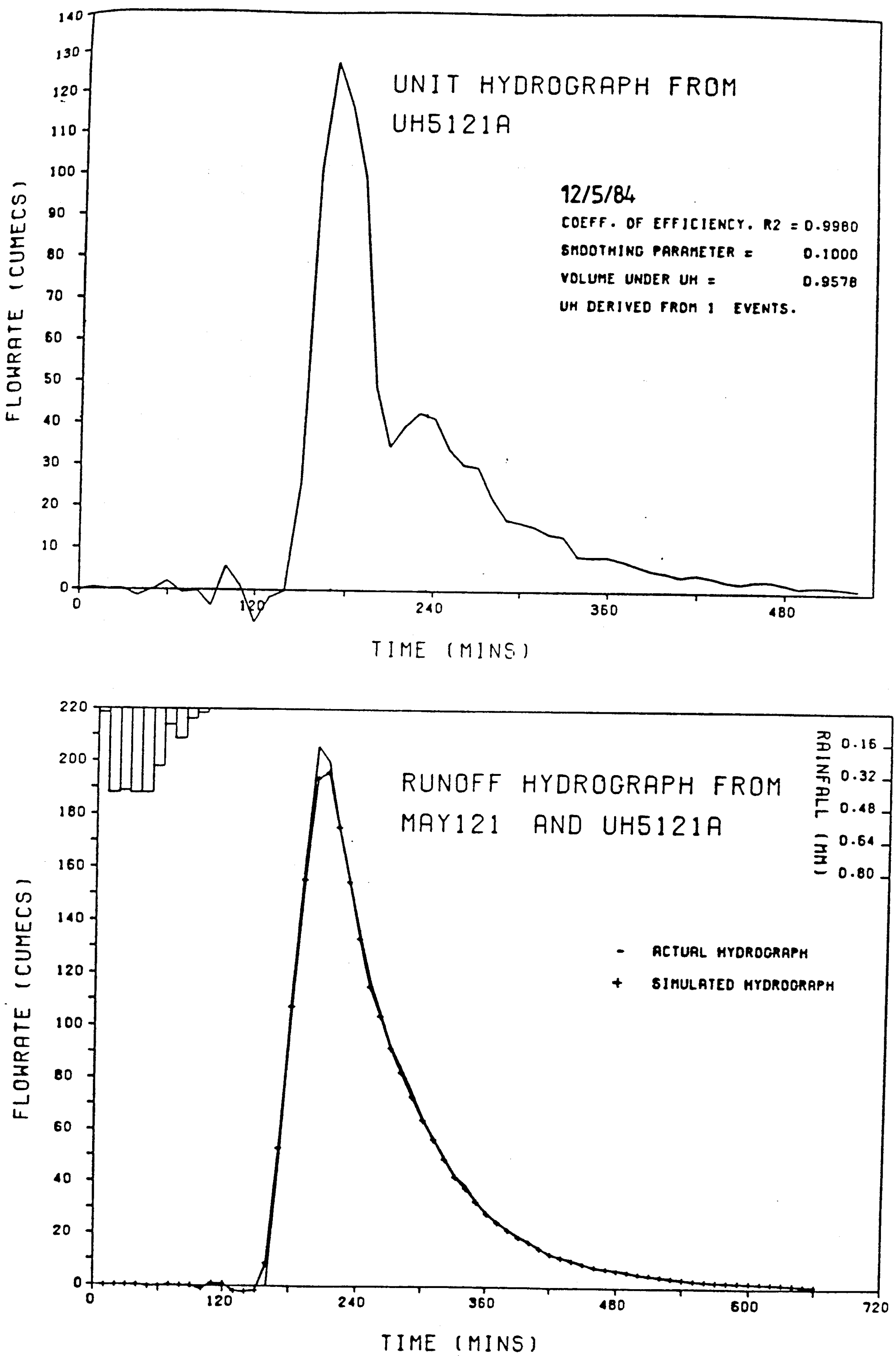


Figure 3

Surface water hydrographs, Wadi Ghat 12 May 1984 - observations and unit hydrograph simulation

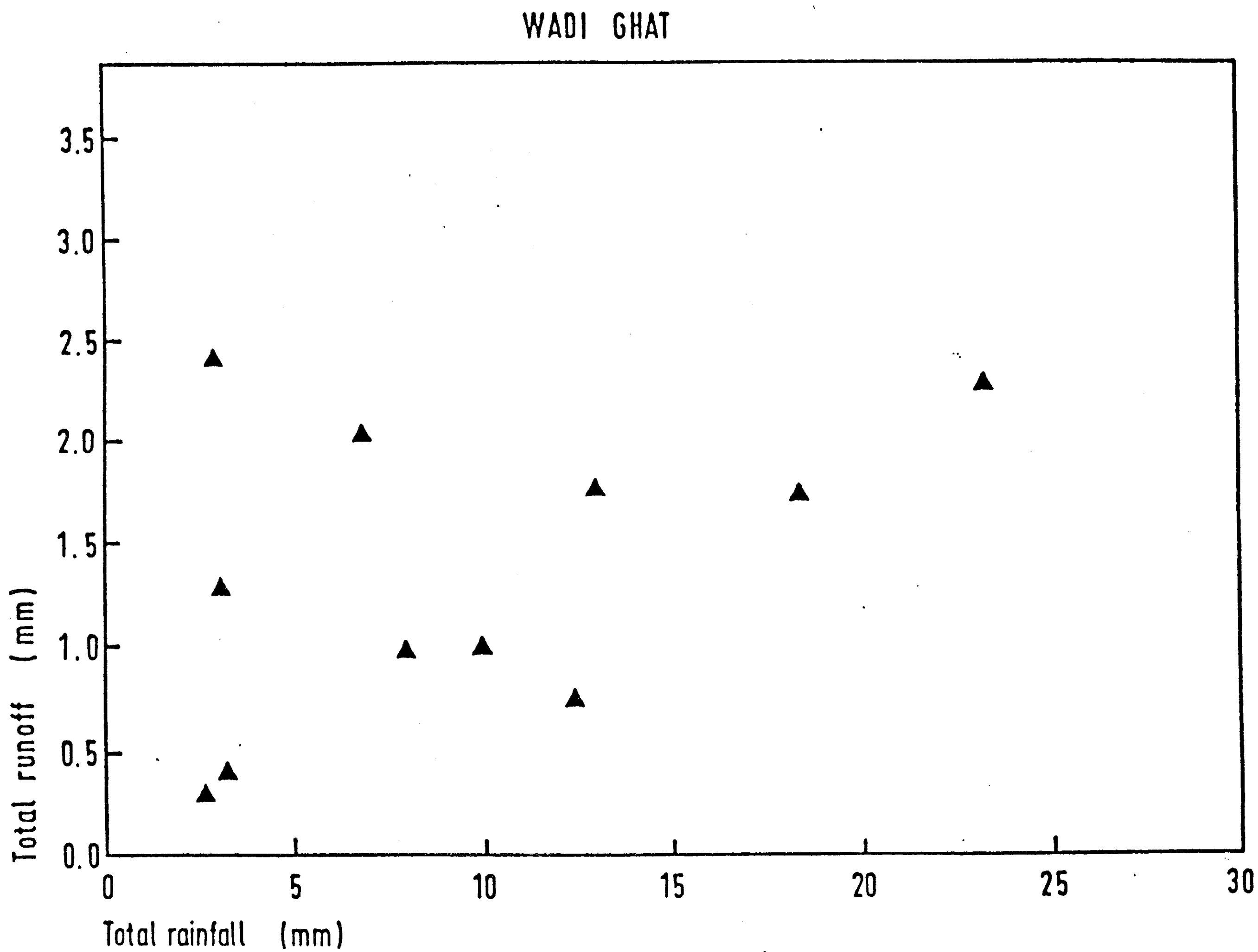


Figure 4

Storm runoff as a function of rainfall, Wadi Ghat

Figure 5

Frequency distribution of spatial coverage of Walnut Gulch rainfall. Observed vs. alternative smulations.

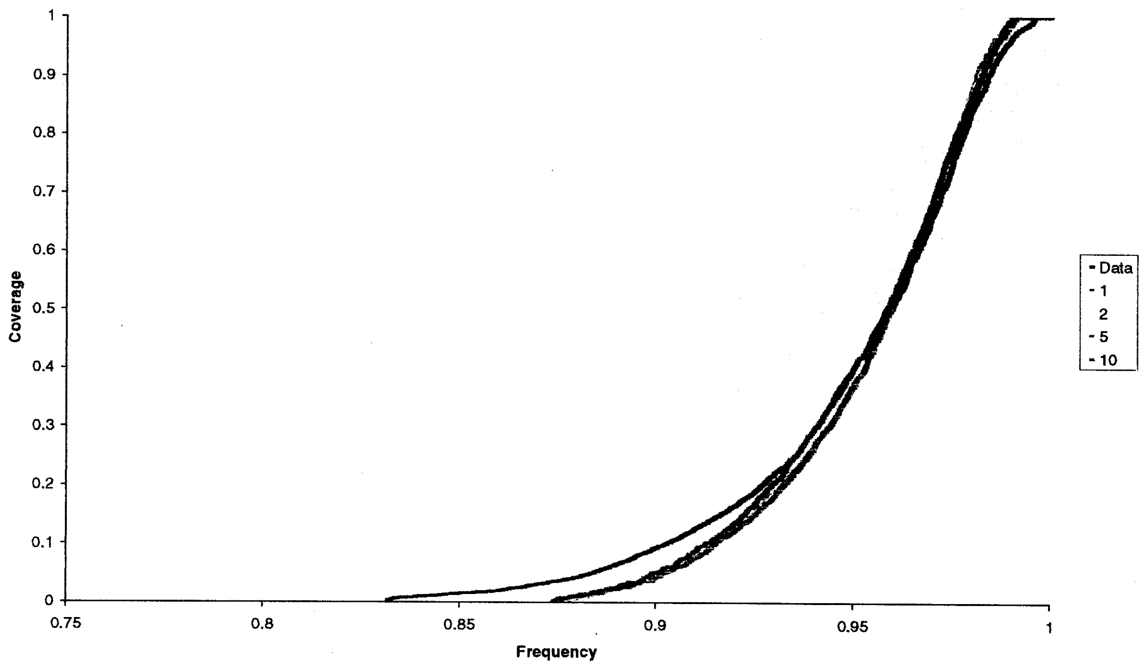
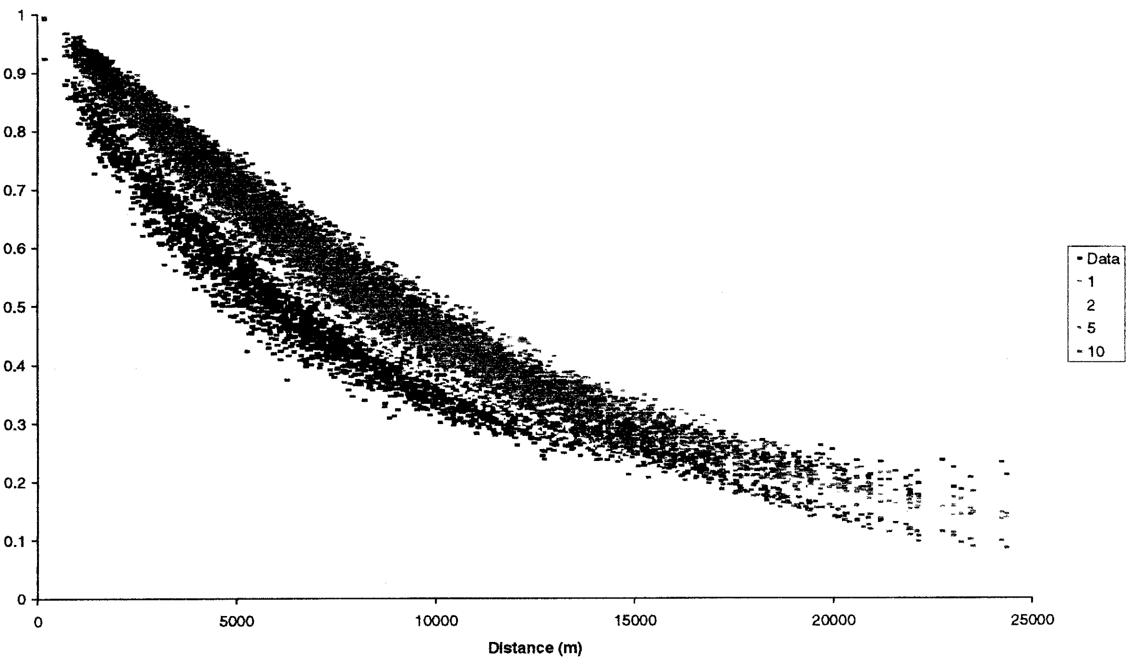


Figure 6

Spatial correlation of Walnut Gulch rainfall.
Observed vs. alternative simulations.



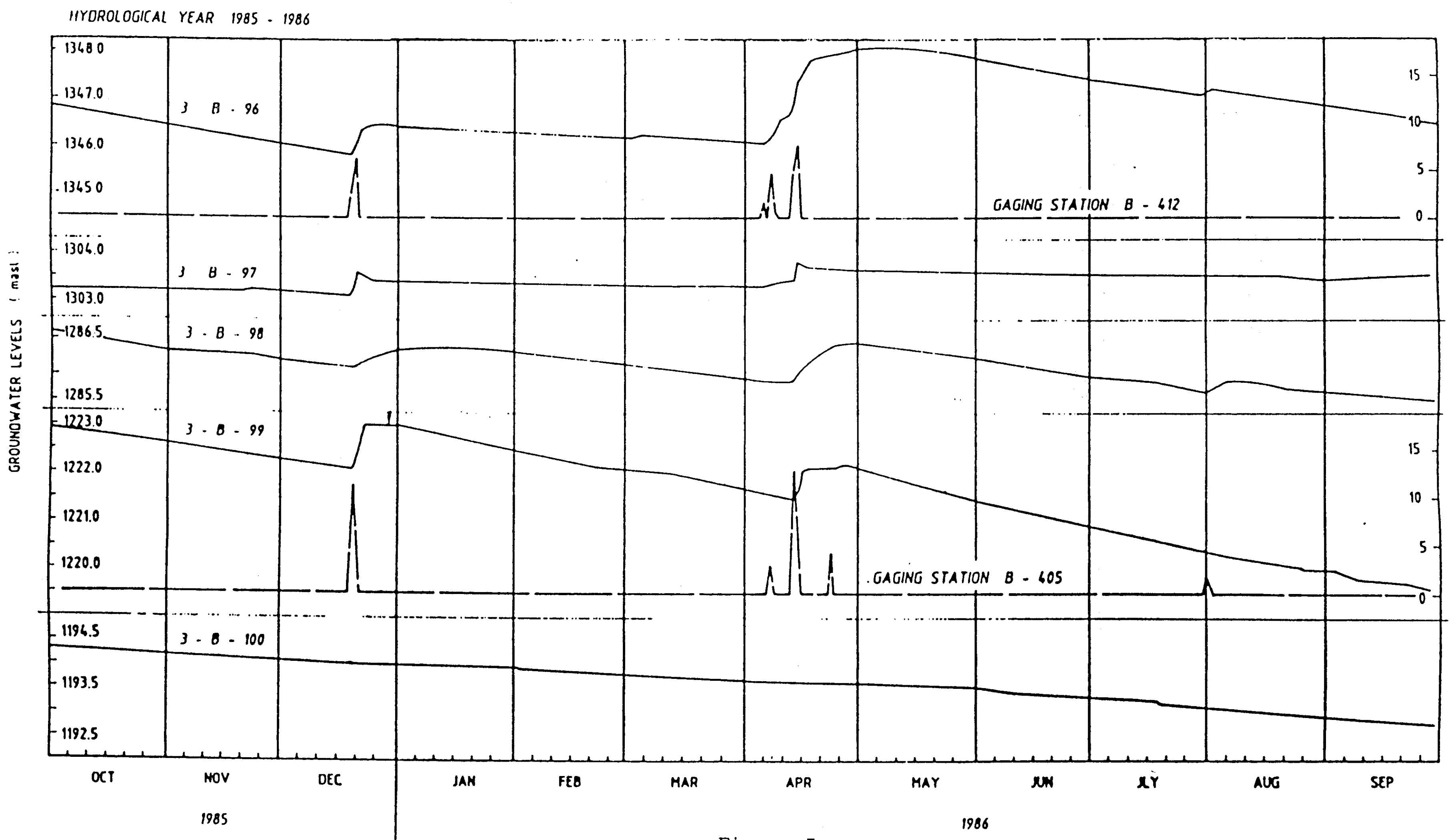


Figure 7

Longitudinal sequence of wadi alluvium well hydrographs and associated surface flows, Wadi Tabalah, 1985/86

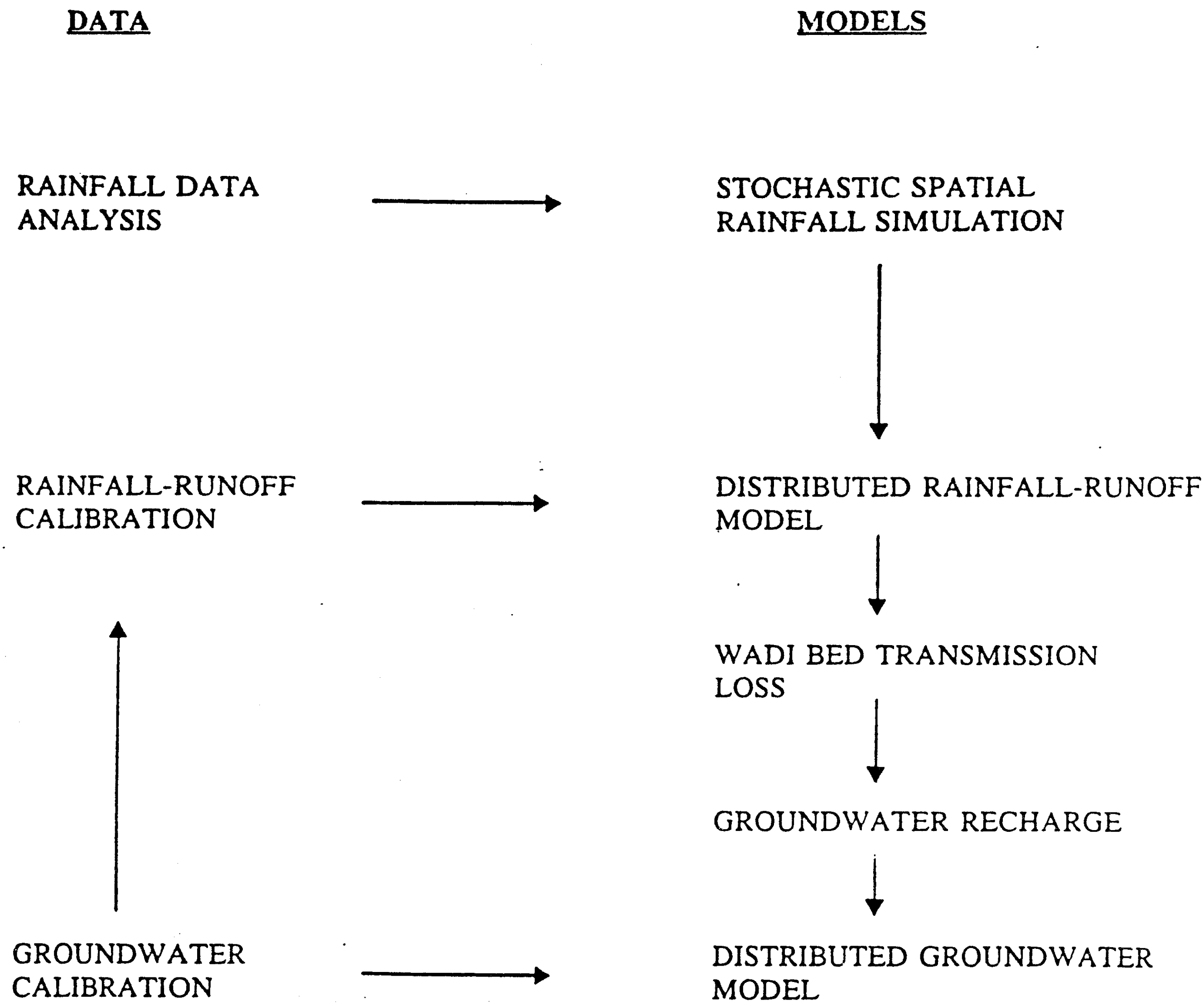


Figure 8

Integrated modelling strategy for water resource evaluation.

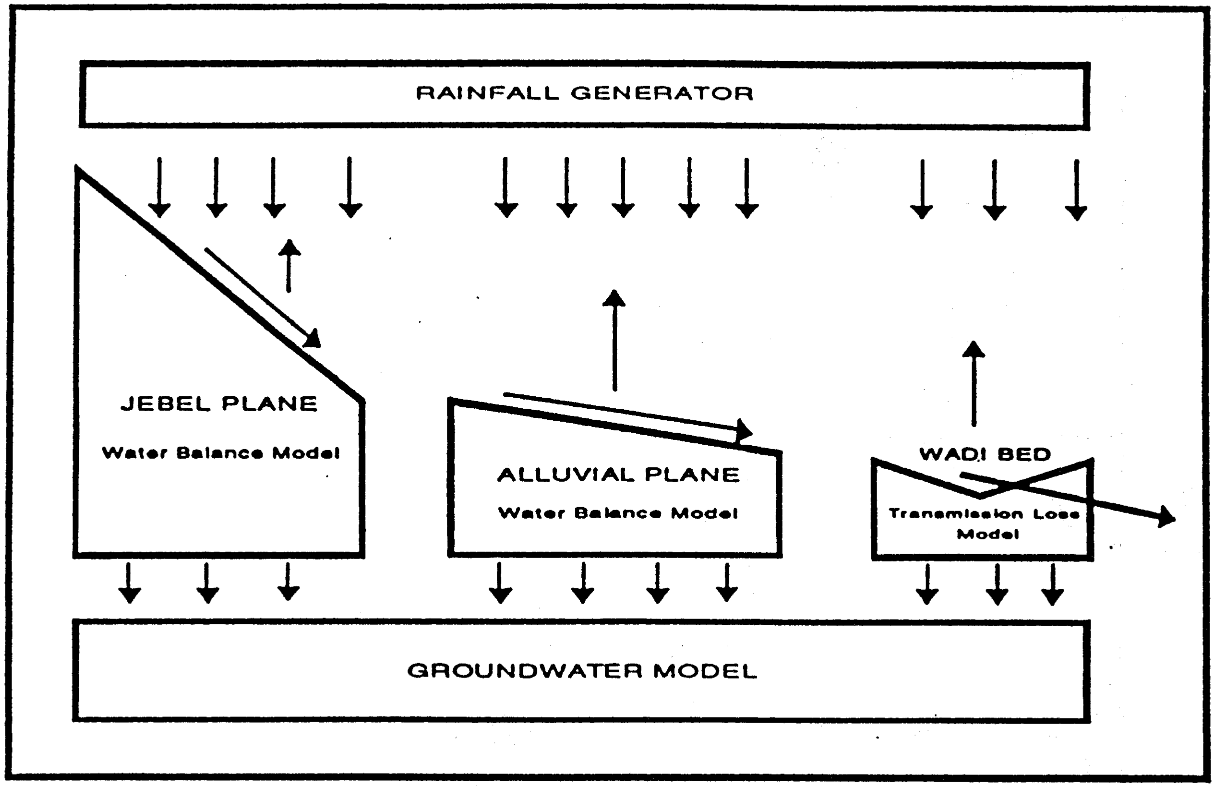


Figure 9

Schematic of the distributed water resource model

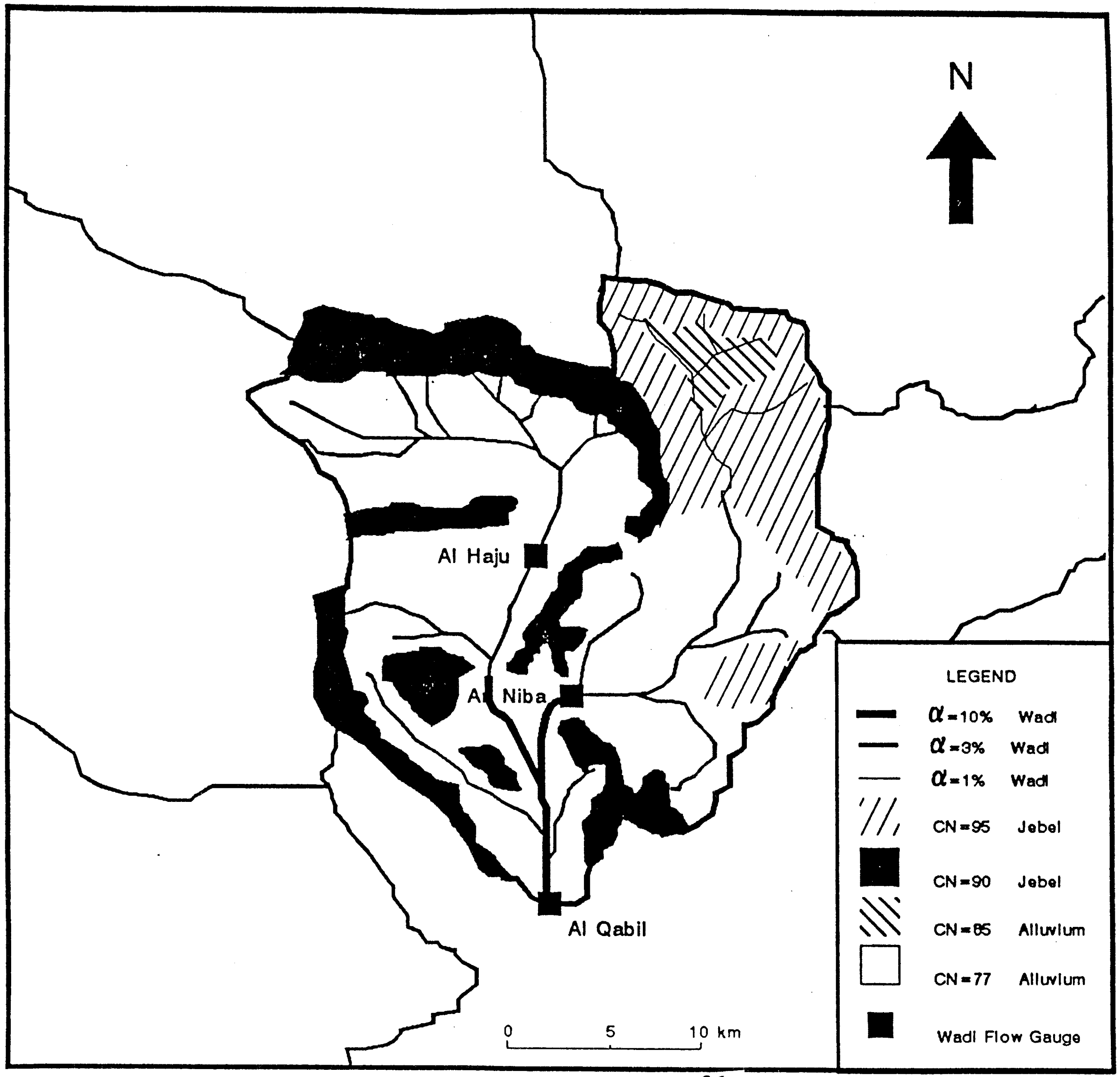


Figure 10

Distributed calibration parameters, water balance model.