OVERLAND FLOW AND TUNNEL FLOW GENERATION ON A SEMI-ARID CATCHMENT: FROM FIELD AND EXPERIMENTAL STUDIES TO SIMULATION MODELING

by

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A thesis submitted in conformity with the requirements for the degree of Doctor of Philosophy
Graduate Department of Geography
University of Toronto

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Overland Flow and Tunnelflow Generation on a Semi-arid Catchment: From Field and Experimental Studies to Simulation Modeling

Ph.D. 1998, Tongxin Zhu
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Abstract

The major barriers to modeling hydrological processes in semi-arid and arid areas are a lack of understanding and model representations of some of the distinctive features and processes associated with runoff generation in those regions and a paucity of field data. This study incorporates field and experimental investigations into simulation modeling to explore stormflow generation processes on a semi-arid agricultural catchment of the Loess Plateau in China.

Overland flow and tunnelflow processes were first investigated in the field. The mechanisms responsible for runoff generation processes and their spatial variation within the catchment were further explored using field rainfall experiments. TOPOG, developed by CSIRO in Australia, was then modified by adding model representations of some of the predominant features and processes identified by field and experimental investigations. The modified models were used to continuously simulate both slowly changing hydrologic states during interstorm periods and fast-responding overland and tunnel flows during stormflow periods. Finally, model simulations, under a wide range of rainfall and spatially-temporally varied land cover conditions, were validated by a comparison of observed and simulated stormflow discharges from both catchment outlet and internal plots.
Field investigations indicated that stormflow generation processes in this catchment are dominated by an infiltration-excess mechanism. Tunnel flows are almost entirely derived from the overland flow entering via inlets, but tunnel flow processes may be significantly disturbed by tunnel instability. Tunnel systems are likely to be initiated in catastrophic storm events, although the subsequent storms do expand these systems. Tunnel conduits are developed above materials of low permeability. The spatial variation in soil infiltration capacity is largely responsible for the spatial variability in runoff generation. The spatial variation in soil infiltration capacity is mainly caused by the spatial variation in crusting, which results from varied land management activities. The simulations showed that the model represents reasonably well stormflows generated by rainfall events with recurrence intervals > 2 years, which account more than 60 per cent of runoff and 70 per cent of sediment leaving this area. The simulations are extended by the inclusion of some features associated with land management, such as terracing and the existence of a plow pan. Considerable variability in simulation accuracy was found among storm events and within the catchment.
Acknowledgments

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# Table of Contents

<table>
<thead>
<tr>
<th>Section</th>
<th>Page</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abstract</td>
<td>i</td>
</tr>
<tr>
<td>Acknowledgments</td>
<td>iii</td>
</tr>
<tr>
<td>Table of Contents</td>
<td>iv</td>
</tr>
<tr>
<td>List of Figures</td>
<td>v</td>
</tr>
<tr>
<td>List of Tables</td>
<td>viii</td>
</tr>
<tr>
<td>1. Introduction</td>
<td>1</td>
</tr>
<tr>
<td>2. Overland Flow: Field and Experimental Investigations</td>
<td>11</td>
</tr>
<tr>
<td>3. Tunnel Flow: Field Study</td>
<td>40</td>
</tr>
<tr>
<td>Continuous modeling of intermittent stormflows on a semi-arid agricultural catchment. J. Hydrol. (to be submitted).</td>
<td></td>
</tr>
<tr>
<td>5. Summary and Conclusions</td>
<td>105</td>
</tr>
<tr>
<td>6. Bibliography</td>
<td>108</td>
</tr>
</tbody>
</table>
LIST OF FIGURES

2.1. The Yangdaogou catchment, plot locations, measurement sites and topography 14
2.2. Depth-duration-frequency of rainfall events in the study area. 15
2.3. Annual rainfall, rainfall during rainy seasons, runoff-generation rainfall and runoff yield over the monitoring periods, 1956-70 and 1987-90. 21
2.4. The rainfall intensities to initiate runoff for the runoff-generation storms observed in the periods, 1956-1970 and 1987-1990. 24
2.5. Plot of the recurrence interval against cumulative percentage of total runoff discharge and sediment load. 25
2.6. Runoff generation source areas on the hillslopes in the runoff-generation storms over the period from 1963 to 1968. 28
2.7. Infiltration processes on different land use in the portable-sprinkler experiments with low kinetic energy of simulated rainfall. 33
2.8. Infiltration processes on different land use in the downspraying-sprinkler experiments with high kinetic energy of simulated rainfall. 34
2.9. Effects of the crusts formed during the previous storms and the present storm on infiltration processes. 36
3.1. The Yangdaogou experimental catchment, tunnel connectivities and measurement sites. 43
3.2. Frequency distribution of tunnel inlet size. 45
3.3a. Cumulative rainfall in the Yangdaogou in 1989 and 1990. 49
3.3b. Frequency distribution of maximum 30-min rainfall intensity for runoff-generation storms in a long-term and the monitored period.

3.4. Rainfall, tunnel and basin outflow hydrographs of the August 11, 1990 storm.

3.5. Water discharges at tunnels 1, 3, 4 and 6 in 1989 and 1990.

3.6. Rainfall, tunnel and basin outflow hydrographs of the July 26, 1990 storm.

3.7. Temporal variations in runoff coefficients of tunnels and experimental catchment at the beginning of the 1990 rainy season.

4.1. Element network for the upper Yangdaogou catchment.

4.2. Frequency distribution of runoff-generation rainfall events over the periods of 1956-70 and 1989-90.

4.3a. Hyetograph and hydrograph for a short-duration storm event in the YDG catchment.

4.3b. Hyetograph and hydrograph for a long-duration storm event in the YDG catchment.

4.4 Simulation of LAI, interception loss, soil evaporation, vegetation transpiration and soil water storage (4m) over the period of 1957-70.

4.5a. Comparison of simulated and observed discharges over the periods of 1957-70 and 1989-90 under the condition of crusted on the croplands.

4.5b. Comparison of simulated and observed discharges over the periods of 1957-70 and 1989-90 under the condition of uncrusted on the croplands.

4.6. Comparison of simulated and observed tunnelflow discharges over the period of 1957-70.
4.7. Simulated contributions of tunnel flows to catchment outflows over the period of 1957-70.

4.8a. Comparison of simulated and observed stormflow discharges on the upper gentle cultivated slopel and over the period of 1963-68.

4.8b. Comparison of simulated and observed stormflow discharges on the barren gully slope over the period of 1963-68.

4.8c. Comparison of simulated and observed stormflow discharges on the whole sideslope over the period of 1963-68.

4.9. Comparison of simulated stormflow discharges with and without terrace lands over the period of 1957-70.

4.10. Comparison of simulated stormflow discharges with and without terrace lands over the period of 1989-90.

4.11. Comparison of simulated stormflow discharges with and without considering the effect of plow pan.
LIST OF TABLES

2.1. Physical and chemical properties of red earth and loess. 16

2.2. Major characteristics of vertical zones in the Yangdaogou catchment 17

2.3. Correlation matrix between annual rainfall, rainfall in the rainy season, runoff-generation rainfall and runoff yield over the periods of 1956 to 1970 and 1987 to 1990. 22

2.4. Rainfall features of different event rain amount categories derived from rainfall data, 1956 to 1970 and 1987 to 1990. 22

2.5. Correlation between rainfall and hydrologic variables. 27

2.6. Runoff yield, runoff coefficient and peak discharge of basin outflow with the difference in runoff source areas on the hillslope over the period of 1963 to 1968. 29

2.7. Description of field monitoring plots and runoff yields. 29

2.8. Correlation analyses between runoff yields from various zones and rainfall variables over the period of 1963 to 1968. 31

2.9. Description of experimental conditions at the plots 32

2.10. Runoff depth from the plots or basin in the first major runoff-generation storm of rainy season from 1963 to 1968. 37

3.1. Major characteristics of the tunnel systems in Yangdaogou subbasin. 47

3.2. Annual precipitation with various return periods. 48

3.3. Tunnel and overland flow start time. 52

3.4. Tunnel and overland flow duration. 55
3.5. $R^2$ values in linear regression analyses between runoff discharge and effective rainfall ($I > 0.20$ mm/min).

3.6. Contribution of tunnel flow to the experimental catchment outflow.

4.1. Input parameters to Topog_IRM and STORM models for the YDG watershed.

4.2. Description of the surface conditions of the plot.
1. INTRODUCTION

In semi-arid areas, rainfall is one of the most limiting factors to vegetation growth and crop yield. On the other hand, high-intensity storms may cause severe flooding and erosion. Appropriate landuse management can increase rainfall infiltration into soils and thereby reduce runoff and erosion. Understanding of the hydrological processes is the prerequisite to implement landuse management properly. The aim of this study is to explore runoff generation processes and their spatial variation within a semi-arid agricultural catchment of the Loess Plateau in China through field, experimental and modeling approaches.

A large number of field studies have been conducted in humid environments to investigate the source areas of streamflow and runoff pathways on the hillslopes since the early work done by Hewlett (1961) and Betson (1964) (e.g. Hewlett and Hibbert, 1967; Dunne and Black, 1970; Anderson and Burt, 1978). These studies demonstrated that spatial variation in soil moisture leads to the non-uniformity of hydrologic responses and the most effective areas in terms of contribution to stormflow are the channel itself and a belt of varying width extending on both sides of the channel, in which soil is saturated or nearly saturated. Stream baseflow might be mainly or completely contributed by subsurface flows. All these findings question Horton's (1933) theory which had dominated hydrology over several decades. With the understanding of physical hydrologic processes and the availability of increasingly powerful computers, physically-based models have been developed (e.g. Beven and Kirkby, 1979; Abbott et al., 1986, etc.). More
recently, interest has arisen in integrating such models into ecological models to simulate
the dynamic interactions between soil-water-vegetation-climate systems at various scales
(Band et al., 1991, 1993; Band, 1993; Moore, et al., 1993; Dawes et al., 1997).

In semi-arid and arid areas, evidence from field and experimental studies re-emphasizes the importance of Hortonian overland flow (e.g. Yair and Lavee, 1985; Dunne
and Aubry, 1986; Abrahams et al., 1994). The infiltration rate is controlled by many
variables: rainfall (Sharon, 1980); vegetation (Nicolau, et al., 1996; Parsons, et al., 1996);
topographic features (Muhs, 1982; Poesen, 1984); antecedent soil moisture (Dunne and
Dietrich, 1980); soil properties (Schumm and Lusby, 1963; Edwards and Larson, 1969;
Bryan et al., 1978; Morin et al., 1981; Yair and Lavee, 1985); and stone cover (Poesen, et

Non-uniformity of runoff generation has also been found but the major controlling
factors seem to be more diverse than in humid areas. Bryan et al. (1978) experimentally
indicated that the spatial variation in runoff generation patterns is mainly the result of
heterogeneity in lithology. Yair (1974) found that the non-uniformity in runoff generation
is controlled by two topographic variables, slope length and slope angle which are largely
responsible for the variation in soil depth. He and his associates (1980) also conducted
rainfall simulation experiments in the Zin Valley and found that aspect differences are
reflected in variation of surface properties despite homogenous bed-rock, which cause
marked differences in hydrological response. Many workers (e.g. Osborn and Lane, 1972;
Renard, 1977; Goodrich, 1990) noted that spatial rainfall heterogeneity has a significant
impact on runoff generation. Foley et al. (1991) and McFarlane et al. (1992), among
others, indicated that the soil hydraulic properties can be highly dependent on land use and land management practices.

Saturation overland flow and lateral subsurface flow are not commonly considered important agents of runoff generation in semi-arid and arid environments, although recent evidence indicates they do occur. For example, Lopes and Ffolliot (1993) found that prolonged frontal rainfall or snowmelt can saturate the shallow, low-permeability soils, causing overland runoff to be generated in the pinyon-juniper and ponderosa pine woodlands of Arizona. Wilcox et al. (1997) found that lateral subsurface flow is a major mechanism of runoff generation in a semiarid ponderosa pine hillslope in northern New Mexico.

A considerable number of studies have recently been conducted to simulate the non-uniformity of runoff generation using distributed models in semi-arid and arid areas (e.g. Moore and Foster, 1990; Goodrich, 1990; Grayson et al., 1992a; El-hames and Richards, 1994.). Due to the absence of baseflow and short-duration of stormflows, those models are often discrete-event models with short-time intervals which are able to capture the temporal variation in rainfall intensity. The models have often been demonstrated to satisfactorily reproduce the observed stormflow hydrographs, but the stormflow data for model validation were usually collected only from one point (catchment outlet) in a few individual storm events. Grayson et al. (1992a) demonstrated that different representations of spatially variable parameters (e.g. soil moisture and hydraulic conductivity) and processes can produce outflow hydrographs which agree well with observed flows. As a result, fitted outflow hydrographs may not represent real hydrologic variability and runoff
generation processes within the catchment. Moreover, effectiveness of model simulation in a few events may not be applicable to a wide range of runoff events. Michaud and Sorochian (1994) tested two distributed models using 24 large storm events collected by USDA-ARS from the Walnut experimental watershed and found neither is able to simulate storm runoff volumes or peak discharges accurately.

Since Carson and Kirkby (1972) stated that piping is very important for subsurface flow and debris removal, but its quantitative contribution cannot yet be estimated, considerable advances have been made in our understanding of nature and function of pipes. However, most of the quantitative studies of pipeflows were conducted in humid environments, especially in Britain (Gilman and Newson, 1980; Jones, 1981,1982,1987; Jones and Crane, 1984). Those pipes are usually characterized by shallow depth (less than 1 m) and small size (less than 30 cm). Though deep-seated tunnel systems were reported from many places in semi-arid areas (Heede, 1971; Hughes, 1972; Drew, 1982; Imeson et al., 1982; Bryan and Harvey, 1985; Baillie et al., 1986), very few studies have been conducted to investigate tunnel flow processes and their hydrological roles in catchment outflow routing (Heede, 1971; Bryan and Harvey, 1985). No attempt has been made to model tunnel flow processes.

Overall, the major barriers to model improvement in semi-arid and arid areas are (1) lack of understanding and model representation of the distinctive features associated with runoff generation in those areas; and (2) lack of field data for model input and validation.
A great deal of research has been carried out in the Yangdaogou catchment, where the present study area is located. As it is an experimental catchment of Shanxi Institute of Soil and Water Conservation (SISWC), detailed field data on runoff and soil loss both at the plot scale and the watershed scale were collected during the period 1956-1970. Such data have been used to analyze the variation in annual sediment yield and establish the nitrogen mass balance (Wang, 1991; Davis et al., 1992). Since 1987, great efforts have been made to develop a Soil Erosion Management Geographic Information System (SEMGIS) for Decision Support System, in which physically based simulation models of crop production and soil erosion are linked with an economic accounting procedure and a geographic information system (GIS) (Band and Fu, 1992). Among people working on this project, Hamilton (1990) quantified the sediment source areas and the losses of dissolved and sediment-associated N from the basin. Wang (1991) developed a hillslope erosion model to predict soil losses caused by raindrop detachment and rill erosion from a hillslope land element. Li (1991) experimentally investigated the spatial variation in infiltration and developed an infiltration model. Luk et al. (1992) developed a soil erosion model, ERODE, which combines a hillslope submodel with a valley slope submodel and a channel submodel. Tague (1994) incorporated a hydrological model, WATER, into a crop productivity model, YIELD, to estimate the potential crop yields and the effects of land management practices on them. Mitchell (1994) examined the effects of vegetation and cropping practices on erosion processes. In the current version of ERODE, the hydrology component is a semi-empirical model. The effect of land use and land management on runoff generation is accounted for by a number of fitted empirical coefficients. In fact,
how various landuse management practices affect runoff generation processes is still not well known. Tunnel flow is not simulated in the model. Antecedent soil moisture is estimated from rainfall data and no consideration is given to the effect of topographic features (slope, aspects etc.). As a result, it seems that the hydrological component in ERODE needs to be redeveloped in order to apply ERODE widely in the Loess Plateau. To redevelop the stormflow hydrological model, we must advance our understanding of the mechanisms responsible for runoff generation processes and their spatial variation in this area.

In this study, overland flow and tunnelflow processes were first investigated in the field. The mechanisms responsible for runoff generation processes and their spatial variation within the catchment were further explored using field rainfall experiments. TOPOG, developed by CSIRO in Australia, was then modified by adding model representations of some predominant features and processes identified by field and experimental investigations. The modified models were used to continuously simulate both slowly changing hydrologic states during interstorm periods and fast-responding overland and tunnel flows during stormflow periods. Finally, the model simulations under a wide range of rainfalls and spatially-temporally varied land cover conditions were validated by comparison of observed and simulated stormflow discharges from both catchment outlet and internal plots. The effectiveness of the model simulations was discussed as it pertains to the linkage of the runoff generation processes identified by field and experimental investigations to their representation by the models.
Three papers were produced from this study: (1) Runoff Generation on a Semi-Arid Agricultural Catchment: From Field and Experimental Studies; (2) Deep-Seated, Complex Tunnel Systems: A Hydrological Study of a Semi-Arid Catchment, Loess Plateau, China; and (3) Continuous Modeling of Intermittent Stormflows in a Semi-Arid Agricultural Catchment.

The objective of the first paper was to investigate overland flow generation processes and the mechanisms responsible for them through field and experimental approaches. Analysis of 19-year matched rainfall-runoff data collected from the study catchment indicated that only a small proportion of rainstorms generated runoff. Data collected from the surface plots shows that both runoff occurrence and yields were highly variable within the catchment. To explore further the mechanisms responsible for runoff generation processes and their spatial variation, rainfall simulation experiments were conducted in the field. It was found that soil crusts play a critical role in runoff generation processes and their spatial variation. Soil crusting can be significantly affected by land use and land management activities.

The objective of the second paper was to investigate tunnel flow processes in the field. Tunnel networks were first identified in the field using smoke bombs. Tunnel flows from six tunnel systems were then monitored in the study catchment over two years. The difference and similarity between tunnel flow and overland flow and the major factors affecting tunnel flow processes were identified. Finally, the hydrological role of tunnel flow in the catchment was evaluated.
The objective of the third paper was to simulate both overland and tunnel flow processes using "TOPOG", developed by CSIRO in Australia. The TOPOG model was modified by adding model representations of some features and processes identified from field and experimental investigations. The modified model was applied to simulate stormflow generation processes over 16 years divided into two periods (1957-1970 and 1989-1990). The effectiveness of model simulation was validated by the comparison of observed and simulated stormflow discharges from the catchment outlet, tunnel systems and internal surface plots.
2. Overland flow: Field and Experimental Investigations

The following paper is intended to explore overland flow generation processes and the mechanisms responsible for them through field and experimental approaches. Kirkby (1987) stated that the generation of overland flow occurs under at least three sets of conditions. When rainfall falls faster than it can be taken in by the soil, and infiltration capacity is exceeded, then the excess flows over the surface. This is the infiltration process described by Horton, and the surface runoff has been termed Hortonian or infiltration-excess overland flow. Overland flow also occurs where the soil is saturated, when any further rainfall, even at low intensities, is able to generate overland flow. This type of overland flow is then termed saturation-excess overland flow. A third type of overland flow is caused by return flow, which can occur even after rainfall has ceased where subsurface flow is forced up to the surface by the soil or slope configuration. In most semi-arid environments, runoff generation is dominated by rainfall-excess or infiltration-excess overland flow mechanism.

The scarcity of long-term matched rainfall-runoff records for small watersheds in semi-arid areas has left the relationship between them still unclear over a wide range of rainfall conditions. Basic questions such as how much runoff occurs, at what frequency it occurs, and under what conditions it occurs remain largely unanswered on the hillslope scale (Thornes, 1994; Wilcox et al., 1997).

In the following paper, 19-year matched rainfall-runoff data were analyzed to identify the relationship between runoff generation and rainfall characteristics. Data collected from the surface
plots were used to identify the spatial variation in runoff generation. Furthermore, the mechanisms responsible for runoff generation processes and their spatial variation were examined by rainfall simulation experiments.
Runoff Generation on a Semi-arid Agricultural Catchment: Field and Experimental Studies

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2.1. Introduction

In most semi-arid and arid environments, runoff generation is dominated by rainfall-excess overland flow, the process whereby the rainfall rate exceeds the infiltration capacity of the soil (Horton, 1933; Abrahams et al., 1994). The infiltration rate is controlled by many variables: rainfall (Sharon, 1980); vegetation (Nicolau, et al., 1996; Parsons, et al., 1996); topographic features (Muhs, 1982; Poesen, 1984); antecedent soil moisture (Dunne and Dietrich, 1980); soil properties (Schumm and Lusby, 1963; Edwards and Larson, 1969; Bryan et al., 1978; Morin et al., 1981; Yair and Lavee, 1985); and stone cover (Poesen et al., 1990; Abrahams and Parsons, 1991). Runoff generation is also highly spatially variable and the major controlling factors seem to be more diverse than in humid areas. Bryan et al. (1978) experimentally indicated that the spatial variation in runoff generation patterns is mainly caused by heterogeneity in lithology. Yair (1974) found that the non-uniformity in runoff generation is controlled by two topographic variables, slope length and slope angle, which are largely responsible for the variation in soil depth. He and his associates (1980) also conducted rainfall simulation experiments in the Zin Valley and found that aspect differences are reflected in variation of surface properties despite homogenous bed-rock, which cause marked differences in hydrological response. Many workers (e.g. Osborn and Lane, 1972; Renard, 1977; Goodrich, 1990)

noted that spatial rainfall heterogeneity has a significant impact on runoff generation. Foley et al. (1991) and McFarlane et al. (1992), among others, indicated that the soil hydraulic properties can be highly dependent on land use and land management practices.

Because runoff-producing events are often highly variable and of short duration in semi-arid and arid areas, the time required to adequately characterize runoff is relatively long (Wilcox et al., 1997). The scarcity of long-term matched rainfall-runoff records for small watersheds in semi-arid areas has left the relationship between them still unclear over a wide range of rainfall conditions. Basic questions such as how much runoff occurs, at what frequency it occurs, and under what conditions it occurs remain largely unanswered on the hillslope scale (Thornes, 1994; Wilcox et al., 1997).

The present study was carried out in the Loess Plateau of China. The Loess Plateau, located in the middle reaches of the Yellow River of China, has an area of 380,000 km² and has been cultivated over thousands of years. Currently, three severe problems exist in this region and its neighboring area. First, water deficit in the soil is the crucial factor hampering agricultural development (Zhao and Zhao, 1991). Second, erosion caused by storm runoff is one of the most severe environmental problems in China. The average and maximum erosion rates are 150 mg ha⁻¹ yr⁻¹ and 390 mg ha⁻¹ yr⁻¹ respectively (Chen and Luk, 1989). Finally, flood risks in the lower reaches cause severe social problems. The storm runoff and eroded soil eventually find their way into the Yellow River and the alluviation in the lower reaches has raised the beds of river channels above the surrounding flood plain by more than 10 m. As a result, the lives and property of millions of inhabitants are in a hazardous flood zone (Whitney and Chen, 1992). These problems can be attributed to the undesirable movement of water. One of the
fundamental solutions to these problems is to change current inappropriate landuse and improve land management so as to increase infiltration and thereby reduce runoff and erosion. Clearly, understanding of runoff generation processes is a prerequisite to implementing these changes.

The objectives of this study are (1) to understand runoff generation and its spatial variations under a wide range of rainfall conditions through field monitoring; (2) to explore the mechanisms responsible for them using the experiments.

2.2. Study site

The study site, Yangdaogou catchment, has an area of 20.3 ha and is located about 4 km north of Lishi town, Shanxi Province (Fig. 2.1). It is a typical first-order drainage subbasin in the Loess Plateau. The climate is semi-arid warm temperate with an annual precipitation of approximately 500 mm, over 70% of which is received during June through September. Fig.2.2 shows the depth-duration-frequency relationship of rainfall events.

Local deposits consist mainly of thick silty loess, which is believed to be windborne dust derived from central Asia in the Quaternary (Liu, 1964). The Tertiary clayey red earth is widely exposed at the lower slope section near the basin outlet. The major physical and chemical properties of red earth and loess are shown in Table 2.1. As an unmanaged experimental basin, the land use had undergone limited changes before 1970. However, after then, a check dam with a sedimentation pond was built at the basin outlet and upper slope has been partially terraced.
Figure 2.1. The Yangdaogou catchment, plot locations, measurement sites and topography
Fig. 2.2. Depth-duration-frequency of rainfall events in the study area.
Table 2.1. Physical and chemical properties of Red earth and Loess

<table>
<thead>
<tr>
<th>Materials</th>
<th>Particle size (%)</th>
<th>Bulk density (g/cm³)</th>
<th>CaCO₃ (%)</th>
<th>Organic Matter (%)</th>
<th>Al₂O₃ (%)</th>
<th>Fe₂O₃ (%)</th>
<th>MnO (%)</th>
<th>MgO (%)</th>
<th>CaO (%)</th>
<th>Na₂O (%)</th>
<th>K₂O (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Loess</td>
<td>13.5</td>
<td>58.1</td>
<td>28.4</td>
<td>1.13-1.19</td>
<td>12.73</td>
<td>1.029</td>
<td>1.7</td>
<td>4.3</td>
<td>0.08</td>
<td>2.23</td>
<td>7.23</td>
</tr>
<tr>
<td>Red earth</td>
<td>6.4</td>
<td>38.1</td>
<td>56.5</td>
<td>1.27-1.40</td>
<td>8.85</td>
<td>0.737</td>
<td>6.6</td>
<td>5.52</td>
<td>0.10</td>
<td>2.28</td>
<td>4.98</td>
</tr>
</tbody>
</table>
Like the other first-order subbasins in the Loess Plateau hilly region, the hillslopes in the Yangdaogou can be divided into four vertical zones from the divide to gully bottom.

The major characteristics of each zone are summarized in Table 2.2

**Table 2.2.** Major characteristics of vertical zones in the Yangdaogou catchment

<table>
<thead>
<tr>
<th>Zone</th>
<th>Soil</th>
<th>Slope (°)</th>
<th>Area (percentage of the catchment)</th>
<th>Landuse and topographic feature Before 1970</th>
<th>Landuse and topographic feature After 1970</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>Loess</td>
<td>&lt; 5</td>
<td>4.3</td>
<td>gentle cultivated slopeland, vegetation land</td>
<td>terraceland, vegetation land</td>
</tr>
<tr>
<td>2</td>
<td>Loess</td>
<td>10-25</td>
<td>46.0</td>
<td>steep cultivated slopeland</td>
<td>terraceland, slopeland, vegetation land</td>
</tr>
<tr>
<td>3</td>
<td>Loess, Red earth</td>
<td>&gt; 30</td>
<td>38.9-48.5</td>
<td>steep gullyslope with sparse shrubs</td>
<td>vegetation land, steep gullyslope with sparse shrubs</td>
</tr>
<tr>
<td>4</td>
<td>Loess, Red earth</td>
<td>&lt; 5</td>
<td>1.2-10.8</td>
<td>narrow gully bottom</td>
<td>narrow gully bottom, sedimentation pond</td>
</tr>
</tbody>
</table>

The soils in the Yangdaogou were developed from Quaternary loess and Tertiary clay and are classified as Typic Ustorthents and Lithic Ustorthents, respectively (Soil Survey Staff, 1975). They occupy about 80 and 20 % of the total basin area (Hamilton and Luk, 1993). Cultivation takes place only at Zone 1 and Zone 2. The crops in the cultivated lands include maize, beans, wheat, sunflower and millet. On the lower slope (Zone 3), shrubs including *Caragana korshinskii*, *Abrotanum lavandulaefolia*, and *Periploca sepium* are present.
2.3. Field setup and experimental design

Rainfall and runoff data were collected in two periods (1956 to 1970 and 1987 to 1990). During the first period (1956 to 1970), a Parshall flume was constructed at the basin outlet to monitor the hydrological processes on the storm event basis. During each storm, the water level readings and sediment samples were taken manually with a varied time interval from 0.5 to 20 minutes. The time interval decreases with the runoff discharge and its changes over time. Based on the calibration formulae, the water level readings were first converted into discharge, and the sediment discharge was then calculated. In order to identify the spatial variation of runoff generation on the hillslope, runoff data were collected from the plots located in different parts of the catchment from 1963 to 1968 (Fig.2.1). During the second period (1987 to 1990), although the focus of the monitoring program had shifted to tunnel flow, the hydrological processes of intermittent stream flow were still monitored from the upper catchment and from the plots with different landuse. The catchment area that contributes runoff to the flume was reduced from 20.3 ha to 12.6 ha. Basic sampling procedures were the same as in the first period except for a shorter time interval (1 to 3 minutes). In both periods, rainfall data were recorded in detail.

In order to explore the mechanisms responsible for runoff generation and its spatial variation, two kinds of rainfall simulators, portable and downspraying sprinklers, were employed to conduct field experiments. The portable rainfall sprinkler described in detail
by Li (1991) ejects water drops from a height of 1.5 m. These drops hit a screen with a net size of 1 mm² at a height of 0.5 m to create a spatially uniform rainfall. The simulated rainfall intensity is constant during any one experiment but can be adjusted from 54 mm/h to 108 mm/h. The plot size is 0.5x0.5 m. Owing to the small drop sizes and low fall-height, water drops have very low kinetic energy and are thereby unable to breakdown aggregates and form crusting. Details of the downspraying sprinkler can be found in Luk et al. (1986). Briefly, simulated rainfall was sprinkled from Spraco Full Cone nozzles, pointing downwards from a height of 4.57 m. The kinetic energy of simulated rainfall is about 90% of natural rainfall and rainfall intensity is similar to that of the portable sprinkler. The experimental plots were laid out with a standard dimension of 1 m cross-slope and 5 m downslope (measured horizontally).

Rainfall simulation experiments were carried out on forest land, terrace land, cultivated slope land and barren gully slope with both portable and downspraying sprinklers.

To evaluate the impacts of crusts formed during both previous storms and the present storm, a pair of cultivated slopeland plots (Plots A and B) with identical gradients (approximately 11 degrees) were selected for downspraying-sprinkler experiments only. Both plots were plowed first and subjected to rainfall simulation experiments for 60 minutes, which is long enough to develop a crust. Three days later, the crust on Plot A was carefully broken to a depth of 1-2 cm and Plot B was left as it was. Subsequently, both plots were exposed to the simulated rainfall for another 60-minute.
During all the experiments, runoff discharges were measured with a time interval of 1 to 3 minutes.

2.4. Results and discussion

2.4.1 Rainfall and basin runoff

Scarcity of matched rainfall-runoff records for small watersheds in semi-arid areas has left the rainfall-runoff relationship between them still unclear over a wide range of rainfall conditions. Fairly long-term and detailed records of rainfall and runoff processes in the Yangdaogou basin have enabled this problem to be tackled. Figure 2.3 shows annual rainfall, rainfall during the rainy season (June to September), runoff-generating rainfall and runoff yield in the basin over a 19-year monitoring period. Average annual precipitation is 500 mm or so with extreme values of 243.5 and 756.3 mm. Over 70% of rainfall and almost all runoff events occur within the rainy season. The runoff-generating storms produce an average of 155.4 mm of rain yearly with a range of 15.3 to 416.3 mm. The mean annual runoff yield is 29.6 mm with a minimum and maximum of 0.2 and 94.5 mm respectively.

Two very characteristic features can be found from the comparison of these four sets of data. First, the coefficients of variation of annual precipitation, rainfall at the rainy season, runoff-generating rainfall and runoff yield (24%, 29%, 63% and 93% respectively) show that runoff-generating rainfall and runoff yield have a much higher inter-annual variability. Second, good correlation exists between rainfall variables but not between rainfall variables and runoff yield (Table 2.3). In the latter correlations, the weakest is
Figure 2.2. Annual rainfall, rainfall during rainy seasons, runoff-generation rainfall and runoff yield over the monitoring periods, 1956-70 and 1987-90.
between annual rainfall and runoff yield \( (R^2=0.25) \). As an example, the highest recorded annual precipitation was 756.3 mm in 1964 but it yielded only 20.6 mm of runoff, which was even below the average of 29.6 mm.

**Table 2.3.** Correlation matrix between annual rainfall, rainfall in the rainy season, runoff-generating rainfall and runoff yield over the periods of 1956 to 1970 and 1987 to 1990 \((n=19)\).

<table>
<thead>
<tr>
<th></th>
<th>Annual rainfall</th>
<th>Rainfall in rainy season</th>
<th>Runoff-generating rainfall</th>
<th>Runoff yield</th>
</tr>
</thead>
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<tr>
<td>Rainfall in rainy season</td>
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<td>Runoff-generating rainfall</td>
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<tr>
<td>Runoff yield</td>
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</table>

From the perspective of individual events, the number of rainfall events sharply decreases with event rain amount and over 80% events have rain amount less than 10 mm \( (Table 2.4)\). The percentage of runoff-generating storms shows a reverse order. Overall, 

**Table 2.4.** Rainfall features of different event rain amount categories derived from rainfall data, 1956 to 1970 and 1987 to 1990.

<table>
<thead>
<tr>
<th>Event rainfall (mm)</th>
<th>&lt; 10</th>
<th>10 - 20</th>
<th>20 - 30</th>
<th>30 - 40</th>
<th>40 - 70</th>
<th>&gt; 70</th>
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</thead>
<tbody>
<tr>
<td>Total events</td>
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<td>170</td>
<td>63</td>
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<td>12</td>
<td>5</td>
</tr>
<tr>
<td>Total precipitation (mm)</td>
<td>3475.8</td>
<td>2438</td>
<td>1306.8</td>
<td>1025.7</td>
<td>732.2</td>
<td>459.1</td>
</tr>
<tr>
<td>Runoff-generating events</td>
<td>34</td>
<td>44</td>
<td>23</td>
<td>11</td>
<td>12</td>
<td>5</td>
</tr>
<tr>
<td>Total runoff-generating precipitation (mm)</td>
<td>215.5</td>
<td>656.7</td>
<td>533.2</td>
<td>356.6</td>
<td>732.2</td>
<td>459.1</td>
</tr>
<tr>
<td>Percentage of runoff-generating events</td>
<td>2.7</td>
<td>25.9</td>
<td>36.5</td>
<td>35.5</td>
<td>100</td>
<td>100</td>
</tr>
</tbody>
</table>
only 8% of the total rainfall events generate basin outflow. However, runoff-generating and non runoff-generating storms share a wide range of event rain amount. The lowest recorded rain amount of a runoff-generating storm was 1.9 mm while the highest non runoff-generating was 39.8 mm. Hence, rainfall amount can not be used as a criterion to differentiate non runoff-generating storms from runoff-generating storms in this area. Instead, rainfall intensity is a more appropriate measure. Owing to the small catchment area and very quick response of stream flow to rainfall, the rainfall intensity which leads to runoff initiation for each storm can be identified by comparing the hyetograph and the streamflow hydrograph. Figure 2.4 shows the rainfall intensity to initiate runoff for all runoff-generating storms observed during the monitoring periods. It can be seen that the minimum rainfall intensity to generate runoff was around 5.5 mm/h.

Over the 19-year periods (1956-1970 and 1989-1990), the runoff-generating storms have recurrence interval ranging from 0.3 to 35 years. Figure 2.5 shows the recurrence interval against cumulative per cent of runoff and sediment load. It can be seen that only 18 per cent of total runoff discharge was generated by the storms of recurrence interval less than 1 year, which removed about 14 per cent of the total sediment load. 63 per cent of runoff and 72 per cent of sediment were contributed by those less frequent to rare storms with recurrence intervals > 2 years.

In order to understand the between-event variations in rainfall and basin runoff, 15 rainfall variables and 6 basinflow hydrological variables were selected for correlation analysis using the rainfall-runoff data of the first period (1956-1970). The aim was to identify those relationships that would merit further study with experiments. The results
**Figure 2.4.** The rainfall intensities to initiate runoff for the runoff-generating storms observed in the periods, 1956-1970 and 1987-1990.
Figure 2.5. Plot of the recurrence interval against cumulative percentage of total runoff discharge and sediment load.
are given in Table 2.5. Three points deserve to be noticed. First, even for the runoff-generating storms only, the best rainfall indicator (highest $r^2$) to predict runoff yields is not the total rainfall amount $P$ but the rainfall amount with an intensity of over 12 mm/h $P_2$. Second, as expected, runoff peak time shows strong positive relation with rainfall peak time. However, the relationship between peak intensity and peak runoff discharge is not as high as expected. Instead, once again, rainfall amounts with an intensity of over 12 mm/h $P_2$ is the better indicator to estimate peak discharge $Q_p$. Third, weak or no correlation exists between times to stream flow initiation $T_{ni}$ and antecedent rainfall $P_a$ or rainfall intensity at the beginning of the storm $I_{ph}$.

2.4.2. Vertical zonation in runoff generation

Monitoring the various plots enabled the spatial variation in runoff generation on the hillslope to be determined. It was assumed that runoff was only generated in Zone 4 (gully bottom) if runoff was recorded at the basin outflow flume but not on any of the hillslope plots. Figure 2.6 shows the runoff generation source areas. Among the total of 52 runoff-generating storms over the period from 1963 to 1968, 11 storms generated runoff on Zone 4 only; 11 storms on Zones 3 and 4, 4 storms on Zones 2, 3 and 4, and 26 storms on all zones. Runoff generation usually required a rainfall intensity in excess of 6 mm/h on Zone 4, 12 mm/h on Zone 3 and 18-24 mm/h on Zones 1 and 2 respectively. With the expansion of runoff source areas, hydrological variables of the basin outflow such as runoff coefficient, runoff yield and peak discharge show an overall increase as well although a wide overlap still exists between them (Table 2.6).
Table 2.5. Correlation between rainfall and hydrologic variables

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<th></th>
<th>Qₐ</th>
<th>Qₚ</th>
<th>Tₚₚ</th>
<th>Tᵣᵢ</th>
<th>Dᵣ</th>
<th>Rₑ</th>
<th>P</th>
<th>Pₐ</th>
<th>Dₚ</th>
<th>P₀₁</th>
<th>P₀₂</th>
<th>P₀₃</th>
<th>P₀₄</th>
<th>P₀₅</th>
<th>Iₑᵥₑ</th>
<th>Max I₁₀</th>
<th>Max I₃₀</th>
<th>Iᵣ</th>
<th>Iᵣᵢ</th>
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<td>0.11</td>
<td>0.09</td>
<td>0.34</td>
<td>1</td>
<td></td>
</tr>
</tbody>
</table>

* Antecedent rainfall is calculated by the equation: \( P_a = \sum (Pt - Rt)^k \), where \( P_a \) is antecedent rainfall; \( Pt \) and \( Rt \) are daily precipitation and runoff depth \( t \) day prior to present day; \( K = 0.83 \) and \( n = 10 \) days.

Qₐ, Storm flow per unit area (mm); Qₚ, Peak discharge (m³/s); Tₚₚ, Peak discharge time (min); Tᵣᵢ, Runoff initiation time (min); Dᵣ, Duration of storm runoff (hour); Rₑ, Runoff coefficients (%); P, Storm rainfall (mm); Pₐ, Antecedent rainfall (mm); Dₚ, Duration of storm (hours); P₀₁, P₀₂, P₀₃, P₀₄, P₀₅, Precipitation with intensity over 0.1, 0.2, 0.3, 0.4, 0.5 mm/min respectively; Iₑᵥₑ, Average rainfall intensity of the storm; Max I₁₀, Max I₃₀, Intensity for the most intense 10 and 30 minutes during the storm respectively (mm/min); Iᵣᵢ, Intensity to runoff generation (mm/min); Iᵣᵢ, Intensity at the beginning of storm (mm/min); Pₘ, Initial losses of rainfall before runoff generation (mm); T₁₀, Time of peak rainfall intensity (min).
Figure 2.6. Runoff generation source areas on the hillslopes in the runoff-generation storms over the period from 1963-1968. Zone 1: hilltop (gentle cultivated land); Zone 2: upper slope (moderate to steep cultivated land); Zone 3: lower slope (barren gully slope); Zone 4: gully bottom.
Table 2.6. Runoff yield, runoff coefficient and peak discharge of basin outflow with the difference in runoff source areas on the hillslope over the period of 1963 to 1968. Values in the parenthesis are the average.

<table>
<thead>
<tr>
<th>Runoff Source Area</th>
<th>Runoff Yield (mm)</th>
<th>Runoff Coefficients (%)</th>
<th>Peak Discharge (m³/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zone 4</td>
<td>0.01-0.5 (0.18)</td>
<td>0.1-8.3 (2.4)</td>
<td>0.009-0.147 (0.051)</td>
</tr>
<tr>
<td>Zones 3 and 4</td>
<td>0.1-4.8 (1.32)</td>
<td>1.3-27.5 (9.2)</td>
<td>0.027-1.003 (0.374)</td>
</tr>
<tr>
<td>Zones 2, 3 and 4</td>
<td>0.3-4.2 (1.52)</td>
<td>1.3-22.5 (12.3)</td>
<td>0.110-3.087 (0.996)</td>
</tr>
<tr>
<td>Zones 1, 2, 3 and 4</td>
<td>0.2-26.9 (4.49)</td>
<td>2.2-43.4 (13.6)</td>
<td>0.049-6.678 (1.188)</td>
</tr>
</tbody>
</table>

Runoff yields also show great variations on the hillslope (Table 2.7). Table 2.7 should be read in conjunction with Figure 2.1. The total runoff yield on the whole hillslope (Plot 4), 144.3 mm, was comparable with that of the basin, 148.8 mm, and also with the weighted average of upper and gully slope sections (Plots 3 and 5), 151.5 mm. This may suggest that the overall run-on losses over the slope length were limited because of the relatively steep terrain and the crusted surface. Plots 1, 2, and 3 have identical land use (cultivated lands) and surface material (loess), differing mainly in slope. The greater runoff yield from Plot 3 than from Plot 2, is consistent with the field
experimental result that runoff yield increases with slope gradient (Luk et al., 1992). However, Plot 1 has gentler slope but produced greater runoff yields than Plots 2 and 3 despite the smaller number of runoff events, which will be discussed later. The greater runoff yield from Plot 5 compared to Plots 1, 2, and 3 is a result of the steeper gradient and different land use, although the data can not be used to evaluate the relative importance of these two factors. Plots 5 and 6 differ in slope gradient, slope aspect, and surface material. The steeper gradient of Plot 5 should result in greater runoff but the higher runoff yield from Plot 6 indicates that the difference in surface material appears to be the most important factor. Red earth has higher clay content and bulk density than loess (Table 2.1) and thereby is more favorable to runoff generation than loess.

Simple regression analysis was conducted between runoff yields on different zones and total rainfall amount or rainfall amount above a certain intensity (Table 2.8). The results once again clearly show that event rainfall amount is not the best index to predict spatial variation in runoff yields on the hillslope. Instead, the rainfall amount above a certain intensity is the better one.

Table 2.8. Correlation analyses between runoff yields from various zones and rainfall variables over the period of 1963 to 1968

<table>
<thead>
<tr>
<th>Plots or Subbasin</th>
<th>Total rain amount (mm)</th>
<th>Rain amount with I &gt; 6 mm/h</th>
<th>Rain amount with I &gt; 12 mm/h</th>
<th>Rain amount with I &gt; 18 mm/h</th>
<th>Rain amount with I &gt; 24 mm/h</th>
<th>Rain amount with I &gt; 30 mm/h</th>
</tr>
</thead>
<tbody>
<tr>
<td>Zone 1</td>
<td>0.622</td>
<td>0.774</td>
<td>0.642</td>
<td>0.468</td>
<td>0.176</td>
<td>0.129</td>
</tr>
<tr>
<td>Zone 2</td>
<td>0.288</td>
<td>0.368</td>
<td>0.42</td>
<td>0.488</td>
<td>0.466</td>
<td>0.286</td>
</tr>
<tr>
<td>Zone 3 (Red earth)</td>
<td>0.381</td>
<td>0.355</td>
<td>0.587</td>
<td>0.659</td>
<td>0.660</td>
<td>0.531</td>
</tr>
<tr>
<td>Zone 3 (Loess)</td>
<td>0.278</td>
<td>0.282</td>
<td>0.485</td>
<td>0.615</td>
<td>0.634</td>
<td>0.505</td>
</tr>
<tr>
<td>Zones 1 and 2</td>
<td>0.261</td>
<td>0.332</td>
<td>0.535</td>
<td>0.610</td>
<td>0.477</td>
<td>0.191</td>
</tr>
<tr>
<td>Zones 1, 2 and 3</td>
<td>0.473</td>
<td>0.539</td>
<td>0.617</td>
<td>0.641</td>
<td>0.548</td>
<td>0.347</td>
</tr>
<tr>
<td>Subbasin</td>
<td>0.457</td>
<td>0.510</td>
<td>0.704</td>
<td>0.730</td>
<td>0.599</td>
<td>0.491</td>
</tr>
</tbody>
</table>
2.4.3. *Sprinkler experiments*

The field observation reveals the fundamental characteristics of rainstorms and the spatial variation in runoff generation on hillslopes in this area. The simple regression analyses also suggest some possible cause and effect relationships between rainfall and runoff variables. However, owing to the widely varied antecedent conditions and uncontrollable rainfall conditions, it is often difficult to assess the relative significance of individual factors on runoff generation and its spatial variation. Moreover, regression analysis cannot indicate the actual mechanisms responsible for them. For this reason, it is important to get assistance from field experiments.

Rainfall simulation experiments were conducted on forest land, terrace land, cultivated land and barren gully slope using portable and downspraying sprinklers respectively. The experimental conditions are given in Table 2.9 and the results are shown in Figures 2.7 and 2.8. It can be seen that in both sets of experiments, runoff initiation time decreased in the order of forest land, terrace land, slope land and barren gully slope with a range from more than 20 minutes to less than 1 minute. Final infiltration capacity was highest on forest lands and the difference between the two types of rainfall simulation experiments was very limited: 60-66 mm/h for the portable-sprinkler method and 48-56 mm/h for the downspraying method. For each rainfall simulator method, terrace land and cultivated slope land had similar final infiltration capacity but the final infiltration capacity for the downspraying sprinkler (12-22 mm/h) was much lower than for the portable-sprinkler (36-42 mm/h). In both sets of
Table 2.9. Description of experimental conditions at the plots

<table>
<thead>
<tr>
<th>Plots</th>
<th>Sprinkler</th>
<th>Landuse</th>
<th>Sprinkler</th>
<th>Tillage</th>
<th>Materials</th>
<th>Slope (°)</th>
<th>Initial Moisture (%)</th>
<th>Rainfall Intensity (mm/h)</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>portable</td>
<td>forest land</td>
<td>portable</td>
<td>no tillage</td>
<td>Loess</td>
<td>10</td>
<td>8</td>
<td>83.4</td>
</tr>
<tr>
<td>2</td>
<td>portable</td>
<td>terrace land</td>
<td>portable</td>
<td>tilled</td>
<td>Loess</td>
<td>&lt;5</td>
<td>10.2</td>
<td>83.4</td>
</tr>
<tr>
<td>3</td>
<td>portable</td>
<td>cultivated slope</td>
<td>portable</td>
<td>tilled</td>
<td>Loess</td>
<td>25</td>
<td>N.A.</td>
<td>84</td>
</tr>
<tr>
<td>4</td>
<td>portable</td>
<td>barren slope</td>
<td>portable</td>
<td>no tillage</td>
<td>Loess</td>
<td>31</td>
<td>9.15</td>
<td>84</td>
</tr>
<tr>
<td>5</td>
<td>downspraying</td>
<td>forest land</td>
<td>downspraying</td>
<td>no tillage</td>
<td>Loess</td>
<td>&lt;5</td>
<td>N.A.</td>
<td>54</td>
</tr>
<tr>
<td>6</td>
<td>downspraying</td>
<td>terrace land</td>
<td>downspraying</td>
<td>tilled</td>
<td>Loess</td>
<td>3</td>
<td>14.5</td>
<td>51</td>
</tr>
<tr>
<td>7</td>
<td>downspraying</td>
<td>cultivated slope</td>
<td>downspraying</td>
<td>tilled</td>
<td>Loess</td>
<td>12</td>
<td>13.5</td>
<td>72</td>
</tr>
<tr>
<td>8</td>
<td>downspraying</td>
<td>barren slope</td>
<td>downspraying</td>
<td>no tillage</td>
<td>Loess</td>
<td>27</td>
<td>12.5</td>
<td>67.8</td>
</tr>
</tbody>
</table>
Figure 2.7. Infiltration processes on different landuse in the portable-sprinkler experiments with low kinetic energy of simulated rainfall.
Figure 2.8. Infiltration processes on different landuse in the downspraying-sprinkler experiments with high kinetic energy of simulated rainfall.
experiments, final infiltration capacity on barren gully slope was low and there was little
difference between the rainfall simulator methods.

The major difference in the simulated rainfall between these two kinds of sprinkler is
rainfall kinetic energy. The very low kinetic energy raindrops from the portable sprinkler do
not breakdown soil aggregates. Consequently, no crusts or very weak crusts develop on the
surface even after large-amount of simulated rainfall with long durations. In contrast, the
downspraying sprinkler produces raindrops compatible with natural rainstorms and for terrace
land and cultivated slopeland. The large difference in final infiltration capacity between the two
methods is due to the formation of crusts when using the downspraying-sprinkler method.
However, on forest lands, a thin layer of litter on the surface largely dissipates the kinetic
energy of raindrops and thereby prohibits the development of crusts, and final infiltration
capacity is very high using both rainfall simulators. Low final infiltration capacity for both
methods on barren gully slopes is ascribed to the crusts existing before the experiments.

The infiltration capacities for the experiments to determine the effects of crusts are
shown in Figure 2.9. At Plot B, runoff was initiated very quickly and the infiltration capacity
decreased very sharply with time. In contrast, Plot A had a longer runoff initiation time and
gentler decrease in the infiltration capacity over time. The infiltration capacity became
comparable near the end of the experiment on both plots, which suggests that new crusts were
gradually formed on the Plot A. With differing infiltration processes, the runoff yields from the
60-minute experiments are 30.7, and 48.8 mm for Plots A and B, respectively.
Figure 2.9. Effects of the crusts formed during the previous storms and the present storm on infiltration processes. Both plots have same area (1x5 m) and identical gradients (approximately 11 degrees). The crust on Plot A was carefully broken to a depth of 1-2 cm and the crust on Plot B was left as it was prior to the experiment.
2.4.4. Effects of cultivation and ploughing: evidence from field observation

Cultivation and ploughing activities occur only on Zone 1 and Zone 2 in Yangdaogou and break up pre-existing crusts. The cultivated land is ploughed with animal-driven ploughs, mainly in May, before seeding. During the crop growth season, occasional hoeing may occur from time to time. However, because the owners of small land parcels have different habits, such activities are more or less unpredictable in both space and time. On gully slopes (Zone 3), crusts exist all the time since no cultivation activities occur there.

Table 2.10 lists the runoff yields of the plots on different zones in the first runoff-generating storm in each of the 6 years of the monitoring period (1963 to 1968). For these

Table 2.10. Runoff depth from the plots or basin in the first major runoff-generating storm of rainy season from 1963 to 1968.

<table>
<thead>
<tr>
<th>Year</th>
<th>Date</th>
<th>Rainfall (mm)</th>
<th>Runoff Depth (mm)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>Plot 1</td>
</tr>
<tr>
<td>1963</td>
<td>5/23</td>
<td>62.3</td>
<td>1.1</td>
</tr>
<tr>
<td>1964</td>
<td>5/20</td>
<td>23.7</td>
<td>0.2</td>
</tr>
<tr>
<td>1965</td>
<td>7/19</td>
<td>15.3</td>
<td>0</td>
</tr>
<tr>
<td>1966</td>
<td>6/29</td>
<td>11.0</td>
<td>0</td>
</tr>
<tr>
<td>1967</td>
<td>7/17</td>
<td>37.5</td>
<td>/</td>
</tr>
<tr>
<td>1968</td>
<td>5/22</td>
<td>25.3</td>
<td>0</td>
</tr>
</tbody>
</table>

storms, the runoff yields from the cultivated lands (Plots 1, 3) were much smaller than those from the gully slope (Plot 5), typically by a factor of 10 or less, compared to a factor of 2 for
the average over the 6 years. This can be attributed to the ploughing on the cultivated lands. After several rainfall events, the development of crusts on the cultivated lands increases the runoff yield and the difference in runoff yields between cultivated lands and gully slope decreases. Thereafter, occasional hoeing may cause small-scale and random effects on runoff generation.

Another remarkable effect of cultivation is the formation of a plough pan on the cultivated lands in Zone 1 owing to the long history of cultivation. Although cultivated lands are also extensively distributed in Zone 2, no obvious plough pan is found in steep slopelands (Cai, et al., 1990). Field investigation indicates that the plough pans are located at 15-30 cm below the surface and have a thickness of 10-20 cm. The bulk densities of the surface soil layer, plough pan and sub-soil layer are 1.13-1.19, 1.23-1.35, and 1.19 -1.25 g/cm³ (Cai et al., 1990) respectively. Portable sprinkler experiments indicate that the infiltration capacity for plough pans is less than 6 mm/h (Li, 1991). This is even lower than the infiltration capacity of crusted surface. During rainstorms, when the infiltration wetting front reaches the plough pan, rainfall may gradually accumulate above it owing to the difference in infiltration capacity between surface layer and plough pan. Finally, the soils above the plough pans become saturated, which needs a rainfall amount of 80 mm or so under dry antecedent conditions (Li, 1991). Thus, the effects of plough pans on runoff generation are important only in storms with extremely large amounts of rainfall or large antecedent rainfall.

2.5. Conclusions
(1) In this area, runoff generation is predominantly caused by infiltration-excess processes, although some of the runoff may reach the channels via deep-seated tunnel systems (Zhu, 1997). Only a small proportion of rainfall events is able to generate runoff. More than 60 per cent of runoff and 70 per cent of sediment load are contributed by the storms of recurrence interval > 2 years.

(2) Both runoff source areas and runoff yields show spatial variations over the hillslope. The spatial variation in runoff generation on the hillslope is mainly ascribed to the difference in soil infiltration capacity.

(3) Of all the factors affecting soil infiltration capacity in this area, crusting is the most important. Formation of crusts can reduce the final infiltration capacity of loess soils from 35-45 mm/h to 10-20 mm/h or so. Soil crusting can be significantly affected by landuse management.
3. Tunnelflow Processes: Field Study

In the previous paper, overland flow processes and their spatial variation within the catchment were studied through field and experimental studies. The aim of the following paper is to investigate tunnel flow processes in the field.

Tunnels or soil pipes are reported from a wide range of environments. Since Carson and Kirkby (1972) stated that piping is very important for sub-surface flow and debris removal, but its quantitative contribution cannot yet be estimated, considerable advances have been made in our understanding of their nature and function. However, almost all the quantitative studies of pipeflows were conducted in humid environments, especially in Britain. Those pipes are usually characterized by shallow depth (less than 1 m) and small size (less than 30 cm). We have very limited knowledge of large-sized, deep-seated tunnel systems which are typically distributed in semi-arid areas. How do tunnel networks drain water in the catchment? What are the major factors affecting tunnel flow processes? What is the difference between tunnel flow processes and overland flow processes? How important is tunnel flow in catchment hydrology?

In the following paper, the hydrological characteristics of tunnel flows are analyzed based on the data collected from the study catchment over a period of two years. The major factors affecting tunnel flow processes are then discussed. Finally, the hydrologic role of tunnel flows in the catchment is evaluated.
Deep-seated, Complex Tunnel Systems - A Hydrological Study in a Semi-arid Catchment, Loess Plateau, China*  

T.X.Zhu  

(Department of Geography, University of Toronto, Toronto, Ont. M5S 1A1)  

3.1. Introduction  

The hydrological roles of pipe and tunnel flow in streamflow generation have attracted increasing attention over the past two decades. However, most research has been conducted in humid areas (Jones, 1981, 1982, 1987; Jones and Crane, 1984; Gilman and Newson, 1980; McCaig, 1983; Morgan, 1972; Walsh and Howell, 1988), and comparatively few studies have been conducted in semi-arid or arid areas. Even fewer attempts have been made to monitor the hydrologic significance of deep-seated tunnel systems which refer to those located at depths of over 1 m below the slope surface. Two exceptions exist to this general scarcity. One is the early work conducted by Heede (1971) in Colorado. Despite prolonged monitoring, few records were obtained as flow occurred only during snowmelt. The other is more recent work conducted by Bryan and Harvey (1985) in the Alberta badlands. Two tunnel systems were monitored through three storms. It was observed that about 10 per cent of the total water flow in the study catchment may be routed through tunnel systems.  

Deep-seated tunnel systems have been reported from many places in semi-arid areas, especially in badlands in Morocco (Imeson et al., 1982), Tunisia (Baillie et al.,  

1986), Alberta and Southern Saskatchewan, Canada (Bryan and Harvey, 1985; Drew, 1982), Southeast Spain (Harvey, 1982), Colorado (Heede, 1971), and in loess areas such as the Loess Plateau of China (Fuller, 1922; Thorp, 1936; Chen, 1958) and parts of New Zealand (Cumberland, 1944; Hughes, 1972). The reasons for the comparative lack of detailed field monitoring may be ascribed to the difficulties in the delimitation of individual tunnel systems, the installation of monitoring equipment in the often steep landscapes, as well as the observational problems caused by the short-duration tunnel flow. Nevertheless, to evaluate the hydrological roles of pipe and tunnel flow in a wide area, deep-seated tunnel systems deserve more attention. To serve this purpose, a monitoring program was conducted during 1989 and 1990 by the author and during the subsequent years by DeCenzo (1997) in the Yangdaogou, a small subbasin of Loess Plateau in North China and all of the tunnel outlets observed to be active were investigated. The preliminary results on tunnel flow hydrology are reported in this paper.

3.2. The Yangdaogou catchment

The Yangdaogou catchment is located about 4 km north of Lishi town, Shanxi Province, China (Figure 3.1). It is a first order drainage basin with an area of 0.203 km². The climate is semi-arid warm temperate. Long term rainfall records at the Shanxi Institute of Soil and Water Conservation, 3 km west of the catchment, indicate that the mean annual rainfall is about 500 mm, over 70% of which falls within the summer from June to September. Local deposits mainly comprise Quaternary loess and Tertiary clayey
Figure 3.1. The Yangdaogou experimental catchment, tunnel connectivities and measurement sites.
red earth. The loess is believed to be wind-borne dust derived from central Asia during the Quaternary (Liu, 1964).

3.3. Description of tunnel systems and tunnel flow monitoring methods

The field monitoring area covers the upper and middle parts of the Yangdaogou catchment, with an area of 0.122 km². A total of 77 tunnel inlets were found in the Yangdaogou subbasin and 75 are located in the experimental catchment. Both diameter and depth of those tunnel inlets range from less than half a meter to more than 20 meters, with a mean of 4.8 and 5.0 m respectively (Figure 3.2). In order to delimit the catchment area of the tunnel systems, tunnel networks were first traced with smoke bombs at the beginning of monitoring period. During the monitoring period, these tracing experiments were also repeated many times in order to detect their temporal changes. The pressure differential between the tunnel inlet and outlet facilitates this method of evaluation. If connectivity is established during any one of the experiments, it is indicated in the topographic map. It was observed that 45 tunnel inlets were connected to 6 outlets which drained most of the tunnel catchment area (Figure 3.1). The tunnel catchments of those connected tunnel systems were then delimited in great detail using a 1:1000 topographic map, supplemented by field survey (Table 3.1).

The outlets of tunnels 1, 3, 4 and 6, being located at relatively flat sites, were suitable for the installation of weirs to monitor flow processes. However, the outlets of
Figure 3.2. Frequency distribution of tunnel inlet size.
tunnel 2 and 5 are located on the cliffs, so that two- and three-order flow dividers were installed there respectively. Although two tubes with a diameter of 10 cm were used to connect the tunnel outlets with flow dividers, tunnel flow mixed with trapped air still readily caused the bricks, mortar seal and tubes to burst a few meters away. After frequent repair, we finally had to give up the process of monitoring those two tunnel systems, although the start and end time of tunnel flow were still recorded during some events. At the exit of the experimental catchment, a concrete flume was constructed by Hamilton (1990) to check the outflow. Additionally, for the purpose of comparison between overland flow and tunnel flow, five surface plots, with areas from 8 to 21500 m², were established in the Yangdaogou subbasin, but only two of the plots were monitored for processes. Owing to the high sediment concentration in the flow, automation of runoff and sediment monitoring was difficult. Thus, stage readings were manually taken every minute throughout the runoff event and sediment samples were taken every three minutes during the first half hour and every six minutes during the second half hour and every twelve minutes thereafter. The stage readings were first converted into discharge using the formulae developed and tested by Zeng (1983) and then sediment discharges in the flow were further calculated.

The field monitoring program was conducted from late July of 1989 and throughout the rainy season of 1990. The annual precipitation was 340 mm in 1989 and 623 mm in 1990 (Figure 3.3a), which represents dry and wet years with recurrence intervals of 10 and 5 years, respectively (Table 3.2). A total of 15 storms were monitored. Those storms were characterized by a short duration and an early peak. By comparing the
Table 3.1. Major characteristics of the tunnel systems in Yangdaogou subbasin

<table>
<thead>
<tr>
<th>Tunnel System</th>
<th>No. of Inlets</th>
<th>Minimum Tunnel Length (m)</th>
<th>Mean Slope</th>
<th>Outlet Area (m²)</th>
<th>Minimum Catchment Area (m²)</th>
<th>Physical Characteristics</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>4-5</td>
<td>2</td>
<td>0.35</td>
<td>0.13</td>
<td>2410</td>
<td>Developed on an old landslide and the bottom of tunnel is the contact between loess and red earth. It drains water from the gentle slopes covered with dense shrub (Caragana).</td>
</tr>
<tr>
<td>2</td>
<td>1</td>
<td>6</td>
<td>1.08</td>
<td>0.68</td>
<td>2050</td>
<td>Formed within loess deposits. Two layers of tunnel conduits are connecting to the outlet but the upper one is abandoned. Tunnel drains water from cultivated slopelands.</td>
</tr>
<tr>
<td>3</td>
<td>8-12</td>
<td>132</td>
<td>0.57</td>
<td>0.13</td>
<td>12700</td>
<td>Developed within the loess and the outlet is located just above the gully bottom land. Three series of inlets ranked by size up the valley slope. Tunnel catchment consists of a large proportion of terrace lands.</td>
</tr>
<tr>
<td>4</td>
<td>5-9</td>
<td>45</td>
<td>0.72</td>
<td>0.36</td>
<td>3285</td>
<td>Developed within the loess and the outlet is also located just above the gully bottom land. Tunnel catchment underwent a great change over the monitoring period.</td>
</tr>
<tr>
<td>5</td>
<td>3</td>
<td>44</td>
<td>0.46</td>
<td>0.3</td>
<td>3210</td>
<td>The outlet is hanging on the cliff. Though the uppermost tunnel inlet is adjacent to the above terraceland there is no indication overland flow enters the tunnel system. Thus the main drainage area is confined to the valley slope.</td>
</tr>
<tr>
<td>6</td>
<td>24-25</td>
<td>254</td>
<td>0.36</td>
<td>3.10</td>
<td>25600</td>
<td>The largest tunnel in the basin. Red earth is scattered along the gulley bottom of the catchment. In 1990 rainy season, a new inlet leading to 40 m or so tunnel conduit was formed during one single storm.</td>
</tr>
</tbody>
</table>
frequency of maximum 30-min rainfall intensity between the monitored storms and the long-term rainfall records (1955-90) of the Shanxi Institute of Soil and Water Institute, we found the monitored storms covered all but the largest runoff-generating storms in this area except the very heavy ones (Figure 3.3b).

Table 3.2. Annual precipitation with various recurrence intervals

<table>
<thead>
<tr>
<th>Category</th>
<th>Wet Year</th>
<th>Average Year</th>
<th>Dry Year</th>
</tr>
</thead>
<tbody>
<tr>
<td>Recurrence intervals</td>
<td>200 100 50 20 10 5</td>
<td>2 5 10 20 50</td>
<td></td>
</tr>
<tr>
<td>Annual Precipitation (mm)</td>
<td>908 865 807 733 671 601</td>
<td>483 399 334 299 240</td>
<td></td>
</tr>
</tbody>
</table>

After Zhang et al. (1992).

3.4. Results and discussion

3.4.1 Tunnel flow hydrological responses

a) Tunnel flow start time Tunnel flow start time is here defined as the lag time between the onset of rainfall and the emergence of flow at the tunnel outlet. Tunnel flows occurred in 12 of 15 monitored storms. Overall, their response to rainfall was very fast, with the start times ranging from 1 to 67 min (Table 3.3). No distinctive difference exists between tunnel flow start times and overland flow start times, which implies that tunnel flow velocity is very fast. An estimate of tunnel flow velocities was made for tunnel 6. The shortest start time for this tunnel was 5 minutes. Even if we assume that overland
Figure 3.3a. Cumulative rainfall in the Yangdaogou in 1989 and 1990.
Figure 3.3b. Frequency distribution of maximum 30-min rainfall intensity for runoff-generating storms in a long-term and the monitored period.
flow was initiated immediately and entered the last tunnel inlet after the rainfall start, the tunnel flow still had an average velocity of 0.26 m s\(^{-1}\). It is noted that the tunnel length of this section was actually measured by the author passing through it.

b) *Tunnel flow peaks* A total of 42 discharge peaks were recorded during 35 complete flow events at four tunnels (numbers 1, 3, 4, 6). Most tunnel hydrographs only have a single peak, which occurs within half an hour of the start of tunnel flow. The number of peaks seems to be determined by rainfall characteristics and is not affected by tunnel complexity. But exceptions did exist. For example, the storm of 7/11/90 was characterized by a single rainfall peak. Likewise, one discharge peak occurred in tunnels 1, 4 and 6. However, two flow peaks were produced in tunnel 3 (Figure 3.4). The reason is unclear but may be related to the runoff generation zonation. In small or medium storms, runoff in the tunnel catchment is mainly generated from the cultivated slopes and the steeper valley side slopes and can be quickly directed into the tunnel system. However, in large storms such as the above-mentioned ones, a large amount of runoff can also be generated from the extensive terrace lands and takes considerably longer time to reach tunnel inlets owing to the relatively remote locations and gentle slopes. Therefore, this runoff formed a second discharge peak for the same rainfall peak. The uneven peaks in the flume hydrograph in Figure 3.4 imply the slight difference in peak time among different tunnels.

(c) *Tunnel flow duration* In small and medium storms, tunnel flow duration shows only limited differences among the tunnels, and is more or less comparable to the effective rainfall duration (Table 3.4). Here, effective rainfall is defined as the rainfall with
Table 3.3. Tunnel and overland flow start time

<table>
<thead>
<tr>
<th>Date</th>
<th>Precipitation (mm)</th>
<th>Antecedent Rainfall (mm) (5 days)</th>
<th>First 10-min Intensity (mm/min)</th>
<th>Runoff start time (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td>T1</td>
<td>T2</td>
</tr>
<tr>
<td>8/6/89</td>
<td>24.8</td>
<td>0</td>
<td>0.99</td>
<td>10</td>
</tr>
<tr>
<td>8/10/89</td>
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<td>24.8</td>
<td>0.37</td>
<td>N</td>
</tr>
<tr>
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<td>28.7</td>
<td>13.1</td>
<td>0.65</td>
<td>21.4</td>
</tr>
<tr>
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<td>28.7</td>
<td>0.15</td>
<td>N</td>
</tr>
<tr>
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<td>21</td>
<td>8.4</td>
<td>0.04</td>
<td>N</td>
</tr>
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<td>29.5</td>
<td>1.19</td>
<td>12</td>
</tr>
<tr>
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<td>39.7</td>
<td>30.5</td>
<td>1.8</td>
<td>15</td>
</tr>
<tr>
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<td>0.16</td>
<td>49</td>
</tr>
<tr>
<td>7/22/90</td>
<td>18.2</td>
<td>1.6</td>
<td>0.42</td>
<td>67</td>
</tr>
<tr>
<td>7/26A/90</td>
<td>33.3</td>
<td>46.9</td>
<td>1.05</td>
<td>19</td>
</tr>
<tr>
<td>7/26B/90</td>
<td>19.8</td>
<td>33.3</td>
<td>0.09</td>
<td>M</td>
</tr>
<tr>
<td>7/30/90</td>
<td>15.8</td>
<td>53.1</td>
<td>0.4</td>
<td>M</td>
</tr>
<tr>
<td>8/11/90</td>
<td>20.2</td>
<td>3.9</td>
<td>0.63</td>
<td>20</td>
</tr>
<tr>
<td>8/13/90</td>
<td>35.4</td>
<td>24.1</td>
<td>0.9</td>
<td>M</td>
</tr>
<tr>
<td>8/28/90</td>
<td>53</td>
<td>0</td>
<td>1.4</td>
<td>13</td>
</tr>
</tbody>
</table>

T: Tunnel System
S: Surface Plot
F: Experimental basin Flume
M: Missed events
N: No runoff occurs
Figure 3.4. Rainfall, tunnel and basin outflow hydrographs of the August 11, 1990 storm.
an intensity of over 12 mm/h (Zhu, et al., 1997). However, in the large storms, such as those on 7/11/90, 7/26/90 and 8/28/90, tunnels 3 and 4 had considerably longer flow duration than the other tunnels. This is because the runoff could be generated from the terrace lands in large storms and it took quite long time to enter the tunnel networks for tunnel 3 and 4, whereas the tunnel flow duration for the remaining tunnels was still quite similar to that in small and medium storms, owing either to relatively shorter distances between terrace land and nearest inlets or to higher slope gradients in the tunnel catchments.

3.4.2 Impacts of instability within tunnel systems on tunnel flow hydrology

Deep-seated tunnel systems in this area are characterized by great instability. Collapses within tunnel systems are very common. Small scale collapses may cause fluctuations of sediment concentration but have no effect on tunnel flow discharges. However, large collapses may exert profound impacts on tunnel flow hydrology. If the collapses are extremely large or associated with surface depression or sediment deposition, tunnel systems could be totally blocked. Rapid tunnel flow could, in turn, reopen the blocked tunnel systems. Such temporal shifts of tunnel systems can be detected using smoke bombs before and after storm events. In the 1989 rainy season, the outlet of tunnel 1 was connected to two series of inlets, but the major branch was blocked in the 1990 rainy season. This led to a great disparity in tunnel flow discharge between 1989 and 1990 (Figure 3.5). In tunnel 3, the southern branch, consisting of four tunnel inlets, was blocked throughout the 1989 and 1990 rainy seasons. But it was reopened in
Table 3.4. Tunnel and overland flow duration

<table>
<thead>
<tr>
<th>Date</th>
<th>Precipitation (mm)</th>
<th>Antecedent Rainfall (mm) (5 days)</th>
<th>Rainfall Duration (min)</th>
<th>Rainfall with 1 &gt; 0.15 mm/min Duration (min)</th>
<th>Runoff Duration (min)</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>T1</td>
<td>T2</td>
</tr>
<tr>
<td>8/6/89</td>
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<td>0</td>
<td>45</td>
<td>45</td>
<td>25</td>
</tr>
<tr>
<td>8/10/89</td>
<td>13.1</td>
<td>24.8</td>
<td>35</td>
<td>35</td>
<td>N</td>
</tr>
<tr>
<td>8/15/89</td>
<td>28.7</td>
<td>13.1</td>
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<td>21</td>
</tr>
<tr>
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<td>28.7</td>
<td>535</td>
<td>10</td>
<td>N</td>
</tr>
<tr>
<td>7/6/90</td>
<td>21</td>
<td>8.4</td>
<td>450</td>
<td>0</td>
<td>N</td>
</tr>
<tr>
<td>7/7/90</td>
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<td>29.5</td>
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<td>8</td>
<td>11</td>
</tr>
<tr>
<td>7/11/90</td>
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<td>30.5</td>
<td>132</td>
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<td>37</td>
</tr>
<tr>
<td>7/13/90</td>
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<td>110</td>
<td>110</td>
<td>51</td>
</tr>
<tr>
<td>7/22/90</td>
<td>18.2</td>
<td>1.6</td>
<td>67</td>
<td>30</td>
<td>31</td>
</tr>
<tr>
<td>7/26A/90</td>
<td>33.3</td>
<td>46.9</td>
<td>185</td>
<td>70</td>
<td>60</td>
</tr>
<tr>
<td>7/26B/90</td>
<td>19.8</td>
<td>33.3</td>
<td>220</td>
<td>60</td>
<td>M</td>
</tr>
<tr>
<td>7/30/90</td>
<td>15.8</td>
<td>53.1</td>
<td>85</td>
<td>50</td>
<td>M</td>
</tr>
<tr>
<td>8/11/90</td>
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<td>3.9</td>
<td>83</td>
<td>30</td>
<td>24</td>
</tr>
<tr>
<td>8/13/90</td>
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<td>24.1</td>
<td>914</td>
<td>54</td>
<td>M</td>
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<tr>
<td>8/28/90</td>
<td>53</td>
<td>0</td>
<td>255</td>
<td>140</td>
<td>62</td>
</tr>
</tbody>
</table>

T: Tunnel System  
S: Surface Plot  
F: Experimental Basin Outflow Flume  
M: Missed events  
N: No runoff occurs
The middle branch was also blocked during the 8/11/90 storm and reopened by the storm of 8/28/90, which caused the tunnel flow discharge of the former storm to be disproportionally low. In tunnel 4, three connecting tunnel inlets, about 30 m south of the tunnel outlet, were abruptly joined into tunnel 4 during the 8/13/90. Those newly joined tunnel inlets subsequently added a large amount of runoff from slope land and terrace land to the tunnel system.

The most significant event we observed during the two consecutive rainy seasons was the abrupt initiation of one tunnel inlet on 8/13/90. The inlet, with a diameter of 1.5 m and depth of 1.9 m, was developed in the middle of a road, located on the upper drainage boundary of tunnel 6. Smoke bomb tests indicated that it was connected to an inlet of tunnel 6 about 40 m away! The runoff from the village and the neighboring subbasin, which used to flow into another basin via the excavated road, was redirected into tunnel 6 through the newly developed tunnel inlet and conduit. This led to the discharge of tunnel 6 being unusually high during the 8/13/90 storm (Figure 3.5). After that storm, the inlet was filled in by the villagers, since it hindered traffic. As a result, discharge at tunnel 6 returned to normal in the subsequent storm of 8/28/90.

Totally blocked tunnel branches can be readily detected with smoke bombs and thereby their impacts on tunnel flow hydrology can be explicitly evaluated. However, in most cases, the tunnel systems may not be totally blocked but partially dammed or blocked first and reopened later during the same storm. In these situations, smoke bombs are useless and it is extremely dangerous to crawl into the tunnel systems after storms to investigate. Thus normally no direct evidence is available. However, in July of 1989, I did manage to pass through the last section of tunnel 6, a 76 m long tunnel conduit, and
Figure 3.5. Water discharges at tunnels 1, 3, 4 and 6 in 1989 and 1990.
found a bridge-like constriction inside. Apparently, it used to be a dam caused by a collapse and sometime later tunnel flow penetrated the dam and formed the opening under the “bridge”. Here, in contrast to the intended objective of this section, we use the monitored tunnel flow hydrologic processes to identify the possible partial damming during the event. Owing to the lack of direct evidence, the results presented here must be considered tentative and examined further in the future. In the 7/26/90 storm (Figure 3.6), tunnel flows in all tunnels except tunnel 4 were characterized by an early discharge peak, which was caused by a peak in rain intensity immediately after rainfall onset. However, discharge at tunnel 4 was very low in the first hour and peak discharge did not occur until 77 minutes after rainfall onset. After the peak, the discharge sharply dropped to a very low level and lasted for another 40 minutes or so. The total discharge appeared to be normal and the absence of the first peak probably resulted from partial damming, which was flushed away later by accumulated water inside the tunnel. In contrast, in tunnel 6, after the first discharge peak, tunnel flow simply stopped. It was unlikely that no runoff had been generated by the second rainfall peak from the tunnel catchment, the largest one in the basin. The total discharge from tunnel 6 during this storm was also quite low. Twelve hours later, another storm occurred, with a rainfall of 19.8 mm. Flow in tunnel 6 started 18 minutes after rainfall onset and the discharge was so high that the trapezoidal weir shifted. Accordingly, no water discharge data were collected, although sediment samples were still taken throughout the event. Actually, the mean rainfall intensity for the first 20 minutes (5.4 mm/h) was very low in this storm. Though the antecedent soil moisture was very high, it is unlikely to have produced such a high flow if no water had
Figure 3.6. Rainfall, tunnel and basin outflow hydrographs of the July 26, 1990 storm
been trapped by damming during the previous storm. Two rainfall peaks produced four discharge peaks in tunnel 3 in Figure 3.6, similar to that in Figure 3.4, may have been caused by runoff generation zonation as well.

To further evaluate the impact of instability on tunnel flow discharge, we compared it with overland flow discharge. Owing to the limited number of events that were monitored on the surface plots during our monitoring periods, we have used the data collected by Shanxi Institute of Soil and Water Conservation during the period of 1963-68 from the Yangdaogou subbasin (SISWC, 1981). Three surface plots with areas of 400, 1855 and 4167 m² were selected in the subbasin comprising upper slope, lower slope, and combined slope, respectively. It was found that a good correlation exists between runoff discharge and rainfall with an intensity of more than 12 mm/h for all three surface plots (Table 3.5). For tunnel systems, such a good correlation could only be found for tunnel 3. However, if we discard those events affected by tunnel instability, identified above, correlation coefficients are improved for all tunnels, especially for tunnel 1 and 4. The still poor correlation for tunnel 6 may be ascribed to unidentified tunnel instabilities within this large and complex system.

Table 3.5. \( R^2 \) values in linear regression analyses between runoff discharge and effective rainfall (I > 12 mm/h)

<table>
<thead>
<tr>
<th>Plot No.</th>
<th>Overland Flow</th>
<th>Tunnel Flow</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>( R^2 ) (including all events)</td>
<td>Tunnel No.</td>
</tr>
<tr>
<td>S-1 (Zone 1)</td>
<td>0.642 (30)</td>
<td>T1</td>
</tr>
<tr>
<td>S-2 (Zones 1 and 2)</td>
<td>0.535 (34)</td>
<td>T3</td>
</tr>
<tr>
<td>S-3 (Zones 1, 2 and 3)</td>
<td>0.617 (40)</td>
<td>T4</td>
</tr>
<tr>
<td></td>
<td></td>
<td>T6</td>
</tr>
</tbody>
</table>

Note: Values in parentheses represent the number of events.
3.4.3 Hydrological significance of tunnel flow

Tunnel flow discharge data are only available from tunnels 1, 3, 4 and 6, but they account for about 90% of the catchment area of all tunnel systems in the basin. Here, two methods are used to evaluate the contribution of tunnel flow to the basin outflow. In Method 1, only those monitored events and tunnel systems which could represent the minimum contribution of tunnel flow are considered. In Method 2, catchment area and effective rainfall are used to interpolate the discharges of tunnels 2 and 5, as well as those missed events for the other four tunnels. The results are listed in Table 3.6.

Overall, the total tunnel flow contribution to basin outflow is 42.9 - 52.5%, which is slightly higher than the proportion of tunnel catchments to total experimental area, 40% or so. Great interstorm variations exist. In the small storms with low intensities such as those of 8/10/89, 8/16/90 and 7/6/90, no tunnel flow was observed though overland flow was generated. Field investigations showed that overland flow generated within the tunnel catchment area was reduced by collapse debris or the surrounding materials of the tunnel conduits while flowing into tunnel systems via inlets. Such reduction was maximized at the beginning of the rainy season and increased with tunnel length. Because we did not monitor the beginning of the 1989 rainy season, we analyze here the temporal variation of tunnel flow discharge only for the 1990 rainy season. The first runoff-generating storm occurred on July 6. A total of 21 mm of precipitation fell and 120 m³ of runoff passed through the low flume, but no tunnel flow emerged from any tunnel outlets. On the following day, a total of 9.5 mm of rain fell within 8 minutes and tunnel flow occurred in
Table 3.6. Contribution of tunnel flow to the experimental catchment outflow

<table>
<thead>
<tr>
<th>Date</th>
<th>Rainfall (mm)</th>
<th>Effective Rainfall (mm)</th>
<th>Tunnel Flow&lt;sup&gt;a&lt;/sup&gt; (m³)</th>
<th>Tunnel Flow&lt;sup&gt;b&lt;/sup&gt; (m³)</th>
<th>Outflow (m³)</th>
<th>Contributions&lt;sup&gt;a&lt;/sup&gt; (%)</th>
<th>Contributions&lt;sup&gt;b&lt;/sup&gt; (%)</th>
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<tbody>
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<td>8/6/89</td>
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<td>24.8</td>
<td>142.2</td>
<td>224.0</td>
<td>280.7</td>
<td>50.7</td>
<td>79.8</td>
</tr>
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<td>8/10/89</td>
<td>13.1</td>
<td>13.1</td>
<td>0</td>
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<td>36.6</td>
<td>0</td>
<td>0</td>
</tr>
<tr>
<td>8/15/89</td>
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<td>22</td>
<td>194.7</td>
<td>221.8</td>
<td>330.7</td>
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</tr>
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<td>360.4</td>
<td>40.9</td>
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<td>82.5</td>
<td>332.8</td>
<td>0</td>
<td>25.0</td>
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<td>7/30/90</td>
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<td>14.2</td>
<td>80.6</td>
<td>85.8</td>
<td>104.5</td>
<td>77.1</td>
<td>82.1</td>
</tr>
<tr>
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<td>18.1</td>
<td>20.0</td>
<td>23.9</td>
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<td>22.9</td>
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<td>486.6</td>
<td>/</td>
<td>/</td>
<td>/</td>
</tr>
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<td>307.8</td>
<td>1530.2</td>
<td>1871.6</td>
<td>3566.7</td>
<td>42.9</td>
<td>52.5</td>
</tr>
</tbody>
</table>

Notes: <sup>a</sup> Estimated from Method 1;  
<sup>b</sup> Estimated from Method 2.
all tunnel systems except tunnel 6. Tunnel 1, with the blockage of a longer branch (70 m or so) and the opening of a shorter one (2 m or so), had an even higher runoff coefficient than the whole experimental catchment (Figure 3.7). However, for tunnel 3, which had at least a 45-m-long conduit from the last inlet to the outlet, the runoff coefficient was still very low. No tunnel flow was observed from the outlet of tunnel 6, which had a 76-m-long conduit from the last inlet to the outlet. Even in the July 11 storm, the third one of the rainy season, tunnel 6 contributed only 62.3 m$^3$ or 7.3 % of runoff at the experimental subbasin outflow. This runoff coefficient was still smaller than the proportion of the basin occupied by the tunnel catchment. In contrast, the runoff coefficients for the other tunnels were more or less comparable to the proportions of the whole experiments. Hamilton (1990) has also stated “for some tunnel outlets there was no evidence of flow during the first two events of the year (1988) whereas major discharges were directly observed from these tunnels during the third events of the rainy season.”

From Table 3.6, it can be seen that the tunnel flow contributions to the subbasin outflow are often significant from medium storms such as 8/6/89, 8/15/89, 7/22/90 and 7/30/90. This may be due to several reasons. First, after the tunnel was wetted at the beginning of the rainy season, the thick tunnel roof prevents evaporation from wetted tunnel conduits. Second, after sheet or rill flows enter tunnel systems, infiltration losses should be reduced owing to the small contact area. Third, particle size analyses indicated that all six tunnel floors were composed of weak permeable layers with higher clay contents than the surface, which could further reduce infiltration losses. In large storms, such as 7/11/90, 7/26/90 and 8/28/90, tunnel flow contributions to the outflow fall off
Figure 3.7. Temporal variations in runoff coefficients of tunnels and experimental catchment at the beginning of the 1990 rainy season.
again and are more or less comparable to the proportions of the basin occupied by the tunnel catchment areas. This is because the overland flow is generated throughout the basin and the difference in infiltration losses between tunnel systems and untunnelled surface areas is considerably reduced.

3.5. Conclusion and implications

Deep-seated tunnel systems in this area act as conduits for overland flow. Soil throughflow is too shallow to reach tunnel systems during storm events. Field experiments and investigations indicate that the wetting depth is usually less than 30 cm and rarely exceeds 50 cm during the storms (Li, 1991), whereas almost all the tunnel systems are located well below this level. Ground water is too deep to recharge tunnel flow, although some tunnel inlets are 20 m deep they are still not deep enough to be close to the ground water table.

Tunnel flow does not simply mirror overland flow. First, high levels of absorption of water at the beginning of the rainy season and the high efficiency of flow delivery later create significant differences in the temporal variation of flow discharge over the season. Second, instability within tunnel systems could make tunnel flow discharge or processes highly erratic within storms. Third, some inlets not connected to any outlets still drain water from the catchment. The trapped water is not only absorbed by surrounding materials but may also recharge the ground water.
Field rainfall simulation experiments conducted in this area showed that micropipes could also be developed close to the surface. Initiation of those shallow pipes is closely related to rill development (Bryan et al., 1978; Yair et al., 1980; Wang, 1991). However, in this area, they seem unlikely to develop into deep-seated tunnels. Instead, the deep tunnel systems are more likely to be initiated in catastrophic storms, although the subsequent storms do expand them. One tunnel inlet leading to about 40 m of tunnel conduits, was developed in a single storm on the 7/13/90. Another example was reported from Wudu, Gansu province in the western Loess Plateau. A total of 123 tunnel inlets were developed within an area of 0.2 km² during a single storm with a precipitation of 37 mm and duration of 115 minutes in 1962, and the volume of developed tunnel totaled around 1000 m³ (Wang, 1989).
4. Overland and Tunnel Flows: Simulation Modeling

In the previous two papers, both overland flow and tunnel flow processes were investigated in the field. In the following paper, TOPOG is modified to simulate both overland flow and tunnelflow discharges. It is noted that stormflow velocity is, however, not simulated.

Daily precipitation, frequently used as a rainfall index in hydrological modeling with GIS, is not suitable in this area since it largely masks the temporal variations in rainfall intensities which can be crucial in determining stormflow discharges and hydrological processes. To shorten time steps may achieve better results but will substantially increase data requirements and computation in a year-round continuous-time hydrological model and thereby cripple its applicability to a wide region. This study incorporates the Topog_IRM model, a terrain-based ecohydrological model, into a runoff generative processes model called STORM. Topog_IRM simulates slowly changing hydrologic states and plant growth during interstorm periods at daily time steps and STORM simulates fast-responding runoff during storm periods with time steps of minutes.

The modified models are applied to the study catchment over the periods of 1956-1970 and 1989-1990. The model simulation under a wide range of rainfall and spatially-temporally varied land cover conditions is tested by comparison of simulated and observed stormflow discharges from both the catchment outlet and internal plots. The effectiveness of model simulation is discussed as it pertains to the linkage of the runoff generative
processes identified by field and experimental investigations to the model representations of them.
4.1. Introduction

In semi-arid areas, precipitation is one of the most limiting factors for vegetation growth and crop yield. On the other hand, short-duration and high intensity rainstorms may cause severe flooding and erosion. Appropriate landuse and land management can increase rainfall infiltration into soils and reduce surface runoff. The evaluation of the hydrologic responses of landuse management necessitates the use of distributed models. Due to the dominance of infiltration-excess flow by short-duration, highly intensity-varied storms, models require very short time intervals to capture the temporal change of rainfall intensity and, therefore, discrete event models are appropriate.

A considerable number of studies have recently been conducted to simulate discrete stormflow events using distributed models (e.g. Moore and Foster, 1990; Goodrich, 1990; Grayson et al., 1992a; El-hames and Richards, 1994). The models have often satisfactorily reproduced the observed stormflow hydrographs, but the stormflow data for model validation was usually collected only from one point (catchment outlet) in a few individual storm events. Grayson et al. (1992a) demonstrated that different representations of spatially variant parameters (e.g. soil moisture and hydraulic conductivity) and processes can produce outflow hydrographs which agree with observed flows. As a result, fitted outflow hydrographs may not reflect differences in hydrologic

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conditions and runoff generation processes within the catchment. To overcome such problems, we need, first of all, a clear understanding of runoff generation processes and their spatial variation within the catchment. Moreover, the field runoff data for model evaluation needs to be collected not only from the catchment outlet, but from the internal areas as well (Grayson et al., 1992b). These data should cover a range of rainfall conditions and a variety of runoff events (Goodrich et al., 1997; Smith et al., 1997).

This paper is an extension of detailed field and experimental studies on the processes of surface and subsurface flowpaths in Loess Plateau watersheds (Zhu, et al., 1997; Zhu, 1997). We adapted TopogIRM and STORM models, developed by CSIRO in Australia (Hatton et al., 1992; Vertessy et al., 1993; Dawes et al., 1997), to continuously simulate the slowly changing hydrologic states (e.g. soil moisture) and vegetation cover during the interstorm periods and the fast-responding runoff generative processes during the storm periods. The models are applied to a semi-arid agricultural catchment with complex-terrain and mixed landuse in the Loess Plateau over the periods of 1956 to 1970 and 1989 to 1990. Landuse experienced a considerable change between these two periods. The model simulation under a wide range of rainfalls and spatially-temporally varied land cover conditions is validated by the comparison of observed and simulated stormflow discharges from both the catchment outlet and internal plots. The effectiveness of model simulation is discussed as it pertains to the linkage of the runoff generative processes identified by field and experimental investigations to the model representation of them.

4.2. Study Catchment
The study catchment, Yangdaogou, is located in the Loess Plateau of China. The Loess Plateau experiences some of the highest rates of soil erosion in the world. The control of soil and water losses by various conservation methods, while maximizing food production, has become a high priority (Huang, 1983; Chen et al., 1988). The study catchment has a drainage area of 20.3 ha. The climate is semi-arid warm temperate. The annual precipitation is approximately 500 mm and over 70% falls in the period from June to September. The average annual temperature is around 9°C with the lowest monthly temperature of -6.2°C in January and highest monthly temperature of 24.6°C in July. The catchment has a maximum relief of 170 m and an average gradient of 12.7%.

One of the unusual geomorphic features of this catchment is the deep-seated, complex tunnel systems. Bryan and Yair (1982) stated that a tunnel has larger and more permanent forms than a soil pipe. A total of 77 tunnel inlets and six tunnel outlets were found in the catchment. Both the diameter and depth of those tunnel inlets range from less than 0.5 m to more than 20 m. The development of the inlets appears to be associated with the cleavage in the loess and the sloping tunnel conduits are developed above materials of lower permeability (e.g. clayey red earth). To investigate the tunnel networks, smoke bombs were dropped in the various inlets to trace their connectivity to outlets. The locations of inlets and outlets, together with their connectivity were recorded on a topographic map with a scale of 1:1000.

About 80% of the catchment is covered with loess and the other 20% is covered with Tertiary clayey red earth which is located in the lower slope sections near the
catchment outlet. As an experimental catchment run by the Shanxi Institute of Soil and Water Conservation (SISWC), the landuse was intentionally kept unchanged from 1956-70 and was characterized by cultivated slopelands on the upper slopes, abandoned lands with sparse shrubs on the lower slopes, and a narrow gully bottom at the foot of the slopes. Since 1970, the upper slopes have been partially terraced, and the lower slopes have been planted with patchy shrubs (Caragana korshinskii). In 1980, a check dam was built at the catchment outlet and a sedimentation pond has developed behind it. The stormflow processes were monitored over the period from 1956 to 70 from the catchment outlet and several internal plots by SISWC. During 1989 and 1990, stormflows were monitored from the upper catchment and the tunnel outlets within the catchment. For more information on the physical properties and details of the field measurements the reader is referred to Hamilton and Luk (1992), Zhu et al. (1997) and Zhu (1997).

4.3. Adaptation of the TOPOG model

TOPOG is a terrain analysis-based, distributed-parameter hydrologic modeling package. The terrain analysis procedures used in TOPOG enable the partitioning of a catchment into a series of interconnected elements (Vertessy, et al., 1993). Each element is bounded with a pair of flow trajectories and upper and lower contours. A suite of topographic attributes (e.g. slope and upper contributing area) can be calculated for each element. Fig. 4.1 shows a network of elements for the Yangdaogou catchment derived from a 1:1000 contour map. Tunnel inlets and outlets are co-registered on the element
network, and the upper drainage area for each tunnel inlet is defined by the element network.

Two hydrologic modules in TOPOG, Topog_IRM, and STORM, are adapted to the study catchment. The Topog_IRM model simulates slowly changing hydrologic states and plant growth over long-term sequences at daily time steps. Soil evaporation is computed by the method of Choudhury and Monteith (1988). Transpiration is computed using the Penman-Monteith equation and the canopy resistance term in the equation is calculated using the stomatal resistance equation of Ball et al. (1987), scaled by the leaf area index. A plant growth module computes daily rates of carbon assimilation, allocation and respiration. The computations were described by Hatton and Dawes (1991) and Dawes et al. (1997). The infiltration and vertical percolation of water through the unsaturated zone are handled by the Richards’ equation. To solve this equation, a finite difference Newton-Raphson scheme is applied. The model developed by Broadbridge and White (1988) is used to estimate the K-Ψ-Θ relationship. It is assumed no distinctive difference between crusted surface and subsoil in the K-Ψ-Θ relationship. A full description is given by Vertessy et al. (1993). The lateral water flux between elements in the saturated zone is computed with a Darcy equation. Dawes et al. (1997) applied the Topog_IRM model to a cropping rotation catchment over a period of one and a half years and found simulated evapotranspiration, soil moisture content at selected depths and leaf area index were all in agreement with published and field measurements.
The STORM model simulates runoff generation processes on each element during a discrete storm event. The major time series data input to the model is rainfall intensity with time interval varied from minutes to tens of minutes. Evapotranspiration is neglected during the storm period. Infiltration is handled by the solution of Richards’ Equation, and subsurface flow is calculated with the Darcy equation.

We modified the STORM model to allow surface water detention to occur anywhere. A simple tunnel flow submodel has also been added. Overland flow generated from the upper contributing area of the element where a tunnel inlet is located is referred as tunnel inflow. Transmission losses within the tunnel system are estimated by comparing tunnel outflows with overland flow from non-tunnelled plots. Those surface plots have similar characteristics to the neighboring tunnel systems in terms of topographic features, vegetation cover and landuse. The estimated transmission losses are subtracted from the inflow before coming out of the tunnel outlet as tunnel outflow. Tunnel outflow is added to the overland flow at the element where the tunnel outlet is located.

In the coupled Topog_IRM and STORM models, stormflow periods are simulated with STORM, and interstorm periods are simulated with Topog_IRM. Distributed soil moisture on the uppermost layers output from Topog_IRM is input to the STORM model as the initial soil moisture for stormflow generation, and the soil moisture status at the end of a stormflow event simulated by STORM is updated for Topog_IRM simulation. The integration of the two models with different time resolution enables the capture of the temporal changes of hydrologic processes in storm periods and interstorm periods with reasonable input data requirements and computation time. The continuous simulation by
TopogIRM model can overcome the subjectivity in estimating the spatial pattern of initial soil moisture prior to discrete storm events.

4.4. Input Data and Parameter Estimation

Daily temperature (Maximum & Minimum) data were obtained from a Lishi County meteorological recording station located about 4 km south of Yangdaogou catchment and were scaled to Yangdaogou catchment using MTCLIM (Running et al., 1987) based on the elevation difference. Precipitation data were collected from the tipping bucket rain gauge in Yangdaogou. We assume that rainfall is spatially uniform within this small catchment. Rainfall events are differentiated as non-runoff generating and runoff generating ones by the maximum 10-minute intensity of 6 mm/hour (Zhu et al., 1997). On non-runoff generating days, only daily precipitation is input to the TopogIRM model. Vapor pressure deficit and radiation are estimated from temperature and rainfall data with MTCLIM. On runoff-generating storm days, rainfall data with time intervals ranging from 1 to tens of minutes are input to the STORM model to simulate the stormflow period and the interstorm period is still simulated with the TopogIRM.

Saturated hydraulic conductivity is estimated by the experimental infiltration process data (Hillel, 1971; Helalia et al., 1988). Li (1991) compared the saturated hydraulic conductivities estimated by rainfall simulator method with those measured by Guelph permeameter and indicated that both are comparable in this area.
It is difficult to directly measure water detention capacity in the field. Consequently, data collected from rainfall simulation experiments are used to estimate water detention capacity. The water detention capacity is estimated by

\[ \text{Wd} = (I-f)dt \]

Where \( \text{Wd} \): water detention capacity;

\( I \): rainfall intensity (mm/h);

\( f \): final infiltration capacity (mm/h);

\( dt \): time for the detention stores to be filled after runoff initiation (h)

It is noted that since the infiltration capacity during the water detention period is higher than the final infiltration capacity, \( \text{Wd} \) is overestimated. The potential error is high on uncrusted surfaces and low on crusted surfaces.

The vegetation growing on the catchment can be divided into three groups: deciduous trees, shrubs, and crops. The root distribution for all of the vegetation groups is assumed to be an exponential decrease from the ground surface to the estimated maximum rooting depth. The effect of the plow pan on root growth is, however, not considered in the model. The input parameters for vegetation growth simulation for each group are derived from the recommended data in the TOPOG user manual. Mean sowing and harvest dates were obtained by visiting local farmers (Mitchell, 1994). A summary of input parameters to the TopogIRM and STORM models is given in Table 4.1.
Table 4.1. Input parameters to TopogIRM and STORM models for the YDG watershed

<table>
<thead>
<tr>
<th>Soil parameters</th>
<th>Terraceland</th>
<th>Slopeland</th>
<th>Vegetationland</th>
<th>Gullyslope</th>
<th>Gullybottom</th>
</tr>
</thead>
<tbody>
<tr>
<td>Saturated hydraulic conductivity (m/d)</td>
<td>1.15 (uncrusted)</td>
<td>1.01 (uncrusted)</td>
<td>1.44</td>
<td>0.43</td>
<td>0.14</td>
</tr>
<tr>
<td></td>
<td>0.58 (crusted)</td>
<td>0.58 (crusted)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td>0.10 (plow pan)</td>
<td>0.10 (plow pan)</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>Volumetric soil moisture content at saturated state (cm³/cm³)</td>
<td>0.5</td>
<td>0.5</td>
<td>0.55</td>
<td>0.35</td>
<td>0.4</td>
</tr>
<tr>
<td>Volumetric soil moisture content at air dry state</td>
<td>0.07</td>
<td>0.07</td>
<td>0.07</td>
<td>0.07</td>
<td>0.07</td>
</tr>
<tr>
<td>Soil structure parameter</td>
<td>1.2</td>
<td>1.2</td>
<td>1.3</td>
<td>1.2</td>
<td>1.2</td>
</tr>
<tr>
<td>Surface detention capacity (m)</td>
<td>0.0096</td>
<td>0.001</td>
<td>0.00151</td>
<td>0</td>
<td>0.010</td>
</tr>
<tr>
<td>Vegetation parameters</td>
<td>Crops</td>
<td>Shrubs</td>
<td>Deciduous trees</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Rainfall interception coefficient (mday⁻¹ per unit LAI)</td>
<td>0.0003</td>
<td>0.0003</td>
<td>0.0003</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Canopy albedo</td>
<td>0.15</td>
<td>0.15</td>
<td>0.15</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Light extinction coefficient</td>
<td>-0.65</td>
<td>-0.60</td>
<td>-0.42</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Maximum assimilation rate (umoles/m²/s)</td>
<td>40</td>
<td>35</td>
<td>35</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Leaf on date (year day)</td>
<td>150</td>
<td>120</td>
<td>120</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Maximum rooting depth (m)</td>
<td>1.5</td>
<td>2.5</td>
<td>4.0</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Leaf respiration coefficient (kg/kg/C/day)</td>
<td>0.002</td>
<td>0.002</td>
<td>0.002</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Stem respiration coefficient (kg/kg/C/day)</td>
<td>/</td>
<td>0.0004</td>
<td>0.0008</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Root respiration coefficient (kg/kg/C/day)</td>
<td>0.0002</td>
<td>0.0002</td>
<td>0.0002</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Optimal temperature (C) for assimilation</td>
<td>22</td>
<td>20</td>
<td>20</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

Other variables

| Tunnel transmission loss rates at the beginning of rainy season (m³/m) | 0.00125 |
| Plow pan depth (m) | 0.20 |
| Sowing date | May 15 |
| Harvest date | September 20 |
4.5. Model Application

The model was calibrated for the first year (1956) and some input variables were adjusted within the range of known variability. The calibrated model was then applied to two periods (1957-70 and 1989-90) with fixed input parameters. During these two periods, yearly precipitation ranged from 243.3 in 1965 to 756.3 mm in 1964. The runoff-generating storms accounted for an average of 155.4 mm of rainfall per year with a range of 15.3 to 416.3 mm, yielding a mean annual runoff of 29.4 mm and minimum and maximum runoffs of 0.2 and 94.5 mm, respectively. A total of 129 storms generated runoff. Event rainfall ranged from 1.9 mm to 104 mm with rainfall duration from 0.12 to 20.5 h, and maximum 30 min rainfall intensity from 3.4 to 70.8 mm/h (Fig. 4.2).

Fig. 4.3a represents a typical stormflow hydrograph at the catchment outlet. Due to an initial high intensity, runoff was generated quickly after the onset of rainfall. The runoff peaks corresponded well with rainfall intensity peaks with only a very short time delay, and runoff ceased even before the end of rainfall, which indicates the dominance of infiltration excess flow mechanisms in runoff generation. In some large storms, the extensive recession period with low discharge rates may imply the contributions of saturation excess flow and shallow subsurface flows (Fig. 4.3b).

4.5.1. Slowly changing hydrologic states and vegetation covers

Fig. 4.4 shows the simulated LAI, interception loss, soil evaporation, transpiration and storage in the top 4 m of soil over the period of 1957-1970. It can be seen that the difference in vegetation cover between years was closely related to the variability in yearly
Figure 4.2  Frequency distribution of runoff-generation rainfall events over the periods of 1956–70 and 1989–90: (a) event rainfall amount; (b) maximum 30-min rainfall intensity; and (c) rainfall duration.
Figure 4.3a. HYetograph and hydrograph for a short-duration storm event in the Yangdaogou catchment.
Figure 4.3b. Hyetograph and hydrograph for a long-duration storm event in the Yangdaogou catchment.
(to be continued)
Figure 4.4. Simulation of LAI, interception loss, soil evaporation, vegetation transpiration and soil water storage (4 m) over the period of 1957-1970.
precipitation. Vegetation had negligible canopy cover during the winter months, and the cover increased in spring during the spring and summer. Interception loss was maximized in the late summer and minimized in the winter when crops were harvested and all of vegetation died back. The interception loss in the winter was caused by litter on the vegetation areas. There was no transpiration from cultivated land and vegetation lands and soil evaporation was very low during the winter. With the growth of vegetation and crops in spring and summer, transpiration rapidly increased over time and finally exceeded soil evaporation in crop lands. However, since a large proportion of the catchment is gully slopes with sparse vegetation, the overall transpiration is still relatively low in comparison to soil evaporation on the catchment. Temporal changes of the ratio between evaporation and transpiration had a considerable effect on soil moisture distribution in the soil profile as soil evaporation mainly derives water from the uppermost soil layers and transpiration can derive water from deeper soil through root up-take. Since the initiation of runoff is mainly controlled by soil water in the upper layers, soil evaporation and vegetation transpiration may impose differing impacts on stormflow generation. Summer water storage peaks existed in wet years but were absent in dry years.

4.5.2. Stormflow: effect of crusting

Field experiments demonstrated that crusts vary greatly among different landuses. No crusts or very weak crusts exist on shrub and deciduous tree lands because a litter layer on the surface dissipates rainfall kinetic energy. On the other hands, barren gully slopes are largely crusted throughout all the storm events. On the crop lands, plowing and
cultivation activities break pre-existing crusts, and the subsequent storms can develop new crusts. However, because the owners of small land parcels have different habits, hoeing is not predictable, and it is difficult to keep a complete record of such an activity in such long periods. Moreover, it is very difficult to incorporate the crusting development processes into the infiltration model at the watershed level. Consequently, we run the models under two extreme conditions: fully crusted and uncrusted on crop lands. Fig. 4.5 (a) and (b) compare observed and simulated stormflow discharges over the first period (1957-1970) under the two conditions (crusted and uncrusted on crop lands). The total observed stormflow discharges over the periods were 406.9 mm, and total simulated stormflow discharges were 340.6 mm under uncrusted conditions, 476.4 mm under crusted conditions. As expected, the total stormflow discharge is overpredicted under fully crusted conditions and underpredicted under uncrusted condition. From Fig. 4.5, it can be seen that the effectiveness of model simulation varies significantly among different storms. It is clear that stormflow is better simulated in large events with recurrence intervals over two years than in small and medium events (e.g. recurrence interval < 2 years) under crusting. The poorer simulation of stormflow discharges in small and medium storms may be due to the effect of an inaccurate simulation of antecedent soil moisture, water detention storage capacity, transient crusting processes, vegetation interception, and/or the disturbance of tunnel flows. With the increase of storm severity, the effects of those factors are reduced. For large storms (recurrence interval > 2 years), simulated stormflow discharges are significantly below the observed ones under uncrusted conditions (i.e., simulated discharge = 0.54 x observed discharge, r² = 0.76). However, stormflow
Figure 4.5a. Comparison of simulated and observed stormflow discharges over the periods of 1957-1970 under the condition of fully crusted on the croplands.
Figure 4.5b. Comparison of simulated and observed discharges over the period of 1957-1970 under the condition of uncrusted on the cropland.
discharges are not overestimated under fully crusted conditions (i.e., simulated discharge = 0.96 x observed discharge, r² = 0.86). In other words, treating the soil surface on the cultivated lands as fully crusted throughout a large storm does not cause significant errors in predicting stormflow discharge. This may imply that crop lands are predominately crusted throughout a large storm event, irrespective of whether the pre-existing crusts are broken or not by hoeing prior to the event. This assumption is also supported by the experimental observations on the surface of disturbed loess. Examination of the soil micromorphology in the lab rainfall simulation experiments indicated that initial crusts developed within five minutes of the onset of rain (Luk, et al., 1990). To capture crusting processes in an infiltration-process model is difficult and often requires too many input parameters; some of which are difficult to derive in the field (e.g. Assouline and Mualem, 1997). The above treatment largely simplifies our approach to meet one of the major goals of hydrologic modeling in this area: predicting flood hazards during large storms. The implication of the above simulations for land management is that mechanical cultivation activities (e.g. plowing, hoeing) can be effective in increasing soil moisture and hence crop yields (in small to medium storms) but may not be very effective in relieving flood hazards.

4.5.3. Tunnel discharges

Fig.4.6 compares the simulated and observed total event tunnelflow discharges over the period of 1989-90. It can be seen that tunnel flows are relatively poorly simulated by the model. The poor simulation seems to be related to tunnel instabilities, the size of simulated storm and the method of subtracting transmission losses. Field investigations
Figure 4.6. Comparison of simulated and observed tunnelflow discharges over the period of 1989-1990.
showed that frequent blockages of tunnels were caused by collapses inside the tunnels, which could be reopened later in the event or during a subsequent event (Zhu, 1997). However, in the tunnel flow submodel, only transmission losses at the beginning of rainy season are considered and the effect of tunnel instabilities is disregarded. Figure 4.7 shows the contribution of tunnel flows to catchment outflows during 1957-70 simulated by the model. The total simulated tunnel flow discharges are 129.7 mm, which contributed about 27.3% of the total catchment outflow. From Fig 4.7, we can see that in the small and medium runoff events, the contributions of tunnel flows are widely varied. With the increase of storm severity, tunnel flow contribution tends to be stable, around 25%, which is comparable to the proportion of tunnel drainage area to the entire catchment. The highly variable contributions in small and medium events simulated by the model suggest great spatial variability in runoff generation among different events. The very good agreement between simulated and observed catchment outflow in the few extremely large storm events (see Fig. 4.5a) suggests that the effect of instability on tunnel flow discharges is likely to be limited. This can be explained by the fact that partial damming or blockage caused by the collapses during the very large storm events may be flushed away during the same event as a large amount of tunnel flow with high velocities is involved. As a result, the hydrologic processes of tunnel flows may be significantly disturbed, but the total discharge of tunnel flows may be less affected by the instabilities during large events than small events.
Figure 4.7. Simulated contributions of tunnel flows to catchment outflows over the period of 1957-1970.
4.5.4. Effect of land cover

Three internal plots of various sizes are selected to validate model simulation within the catchment. These three plots cover upper gentle cultivated slope land, barren red-earth gully slope and an entire slope, respectively (Table 4.2). Stormflows were monitored on the plots from 1963 to 1968. Fig.4.8 compares the simulated and observed stormflow discharges among the plots. It is noted that the simulation of each plot is pulled out of the full basin simulation. Again, the simulation accuracy increases with storm size. The significant variability in the accuracy of model simulation among the internal plots may be attributed to the complexity of runoff generation processes and to the closeness of model representation of initial conditions and plot characteristics. On the barren gully slope, no cultivation activities (e.g. hoeing) disturbed the crusted surface, and surface detention capacity is also negligible because of the steep slope and the dominance of concentrated flows. Accordingly, stormflow discharges are predicted with relatively good

Table 4.2. Description of the surface conditions of the plots

<table>
<thead>
<tr>
<th>Plot</th>
<th>Landuse</th>
<th>Surface materials</th>
<th>Mean Slope (°)</th>
<th>Slope Aspect</th>
<th>Slope Length (m)</th>
<th>Area (m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Upper slope</td>
<td>crop land</td>
<td>Loess</td>
<td>4</td>
<td>NE</td>
<td>40</td>
<td>400</td>
</tr>
<tr>
<td>Gully slope</td>
<td>barren slope</td>
<td>Red earth</td>
<td>29</td>
<td>NW</td>
<td>50</td>
<td>630</td>
</tr>
<tr>
<td>Entire slope</td>
<td>cropland +</td>
<td>Loess + red earth</td>
<td>29</td>
<td>NE</td>
<td>185</td>
<td>4167</td>
</tr>
</tbody>
</table>
Figure 4.8a. Comparison of simulated and observed stormflow discharges on the upper gentle cultivated slopeland over the period of 1963-1968.
Figure 4.8b. Comparison of simulated and observed stormflow discharges on the barren gully slope over the period of 1963-1968.
Figure 4.8c. Comparison of simulated and observed stormflow discharges on the whole sideslope over the period of 1963-1968.
accuracy. On the upper gentle cultivated slopeland, irregular cultivation activities disturbed the surface from time to time during the growth season; as discussed above, these activities are not captured by the model. In addition, surface detention capacity, as one of the most important parameters to estimate stormflow generation on the plot, is assumed to be constant throughout all storm events in the model. In fact, irregular hoeing and changes in micromorphology during the storm led to changes in surface detention capacity from event to event.

Landuse changed a considerably change between the first and second periods. In the first period the upper slope was dominated by cultivated slope lands, while in the second period it was partially converted to terrace lands. As indicated by the field experiments, the major difference between slopeland and terrace land in terms of stormflow generation is that terrace lands increase surface water detention capacity. To evaluate the effect of water detention capacity on stormflow discharges under a wide range of rainfall conditions, we simulate catchment outflow under non-terraced and partially terraced conditions over the two periods. Fig. 4.9 shows the comparison of simulated stormflow discharges over the 1957-70 period, both with and without considering water detention capacity on the cultivated lands. Simulation results indicate that if the same area of terrace land had been built in the first period as was built in the second period, the total stormflow discharge over the period would have decreased by 32.9%. Similarly, stormflow would have increased by 19.1% over the second period if no terrace land had been built over the second period (Fig.4.10). The rates of decreased runoff vary with storm sizes. For some of the small runoff-generating storm events,
Figure 4.9. Comparison of simulated stormflow discharges with and without terrace lands over the period of 1957-1970.
water detention capacity made little or no difference to the total runoff discharge. This was because stormflow was generated only on the barren gully slope and gully bottom and little or no runoff was generated from the cultivated land. It seems that the largest reduction rates occurred in the medium-sized storm events. In very large storm events, the relative significance of water detention in reducing runoff discharge seems to diminish.

4.5.5. Effect of plow pans

Plow pans are the result of the compaction by plows and are very common in the cultivated lands. A plow pan is typically located at 15-30 cm below the surface and has a thickness of 10-20 cm. The bulk densities of the surface soil layer, plow pan and subsoil layer are 1.13-1.19, 1.23-1.35 and 1.19-1.25 g/cm³ in the Yangdaogou (Cai, et al., 1990). Li (1991) indicated that the infiltration capacity of a plow pan is less than 6 mm/h, which is much lower than the infiltration capacity of crusted surface. Fig. 4.11 shows the comparison of the simulated storm discharges with and without considering the presence of plowpans on the catchment. The plow pan increased the stormflow discharges in a few extremely large storm events with recurrence intervals greater than 10 years but had no effect on other storm events. The biggest difference occurred in the storms with large amounts of rainfall but relatively low rainfall intensities. For example, for the event of 8/20/59, which had a total rainfall of 104 mm and the maximum 30-min rainfall intensity of 26 mm/h, the simulated stormflow discharges are 37.8 mm and 14.02 mm with and without considering the plow pan’s effect, respectively. The increased runoff was
Figure 4.10. Comparison of simulated stormflow discharges with and without terrace lands over the period of 1989-1990.
Figure 4.11. Comparison of simulated stormflow discharges with and without considering the effect of plow pan.
contributed by saturation overland flow caused by a perched water table above the plow pan. Increased stormflow discharges in the extremely large events, though very few, could cause disastrous flooding and erosion.

4.6. Summary and Conclusions

This study adapted the TopogIRM and STORM models to a semi-arid agricultural catchment. The different time intervals used in the two models enabled the simulation to capture the slowly changing hydrologic states and vegetation cover during interstorm periods as well as the fast-responding runoff generation processes during the storm periods with reasonable input data requirements and computation time. Continuous simulation over long periods avoids substantial uncertainty in estimating the antecedent conditions (e.g. soil moisture, vegetation cover, etc.) prior to each storm event.

The simulations indicated that the model represents reasonably well storms with recurrence intervals > 2 year, which amount for more than 60 per cent of the runoff and 70 per cent of the sediment from this area.

The model has shown that the inclusion of some features associated with land management has improved the simulation of stormflow generation. The results have highlighted the critical importance of crusting in increasing stormflow runoff. Simplifying the crusting formation processes as the fully stable crusts throughout the storm event does not cause significant errors in estimation of stormflow discharges during large storm events in this area. Surface detention on terrace lands is very effective in medium events,
but its effectiveness becomes increasingly insignificant in large storms. Plow pans can dramatically increase stormflow in extremely large storm events but have no effects on smaller events.

The results have highlighted the considerable variability in simulation accuracy. Generally, the model simulations yield better results for large storm events than for small events. This is probably because of the inaccurate simulation of water detention storage, antecedent soil moisture, and of reinfiltration on the surface and disturbance of instability in tunnel systems during small stormflow events. The influence of these factors seems to be reduced with the increasing severity of storms. The instability of tunnel systems may also significantly disturb the total event stormflow discharges during such storm events. However, for extremely large events, the effect of such instability on the total tunnel flow discharges seems not very remarkable. This may be because the partial damming or blockage of tunnel systems can be flushed away during the storm event. The accuracy of the model simulations varies considerably among the entire catchment and the internal areas. The accuracy of the model simulations seems to be related to the complexity of the runoff generation processes, and the ability of the model to represent these processes, the scale of simulation, and the inherent variability and uncertainty in parameterization.

Finally, some recommendations can be drawn from this study for land management and landuse changes. Currently, to reduce the effects of crusting on runoff generation, farmers are advised to destroy the crust by hoeing when a storm is expected. Our simulation results indicate that this is effective for small to medium events, but this effectiveness seems to be dramatically reduced in very large events. The alternative
measure is to put crop residues on the surface to dissipate rainfall kinetic energy. To reduce the effect of the plow pan, the farmers should deepen the plow depth, thereby increasing the water retention capacity of the soil layers above the plow pan. In general, the simulation results support the overall arrangement implemented in the Yangdaogou catchment since the 1980’s. The conversion of the upper gentle slopes into terrace lands increases crop yields and reduces the runoff. As tunnel systems virtually acted as the conduits of overland flows, terracing the upper slope can reduce the water entering tunnel systems via tunnel inlets and thereby reduce tunnel erosion. However, terracing should be combined with the planting of vegetation to stabilize terrace banks. With the development of bottomland which can be used as croplands, all the slope lands should be replaced with vegetation lands in the future.
5. SUMMARY AND CONCLUSIONS

This study incorporates field and experimental investigations into distributed modeling to explore overland flow and tunnelflow generation processes and their spatial variation within a semi-arid agricultural catchment of complex-terrain with mixed landuses in the Loess Plateau of China.

Field investigations demonstrated that infiltration-excess mechanisms dominate in most events but on occasion shallow subsurface flows and saturation overland flows might be also involved. Analysis of 19 years of matched rainfall and runoff data indicated that only a small proportion of rainfall events generated runoff. Both runoff occurrence and yields were found to be highly variable within the catchment. The spatial variation in runoff generation is mainly ascribed to differences in soil infiltration and water detention capacity. Comparison of two sets of rainfall simulation experiments indicated that differences in infiltration are mainly caused by the spatial variation in crusting. On vegetation lands, either no crusts or weak crusts are developed owing to the existence of a litter layer on the surface. On the cultivated lands, plowing and hoeing activities can break pre-existing crusts and subsequent rainfall events can develop new crusts. On the barren gully slopes, crusts exist throughout the year.

Detailed monitoring of deep-seated, complex tunnel systems in the upper catchment showed that all tunnel flow is derived from overland flow generated entering via inlets and no soil throughflow and ground water is involved. But tunnel flow processes do not simply mirror overland flow processes. They can be significantly disturbed by frequent blockages of tunnels, which can be reopened in subsequent events, and by the
occasional abrupt opening of new inlets. In addition, tunnel flows are significantly reduced by the collapse of dry debris and materials surrounding the tunnel conduits at the beginning of rainy season. During the monitoring period, tunnel flows contributed at least 43% of the total outflow from the upper catchment of the study basin, with the contributions ranging from 0 to 78%, from event to event.

The Topog_IRM and STORM models were modified to continuously simulate the slowly changing hydrologic states and vegetation cover during the interstorm periods and the fast-responding runoff during the storm periods. The simulated runoffs were compared with the observed stormflow discharges from both the catchment outlet and internal plots. The simulations indicated that the model can represent reasonably well storms with recurrence intervals > 2 year, which account for more than 60 per cent of runoff and 70 per cent of sediment leaving this area. The results have highlighted the critical importance of crusting in increasing stormflow runoff. Simplifying the crusting formation processes as fully stable crusts throughout the storm events does not cause significant errors in estimation of stormflow discharges during large storm events. Surface detention on terrace lands is most effective in medium-sized events, and least effective in large storms. Plow pans can dramatically increase stormflow in the extremely large storm events but have no effect on other events. Large storm events are better simulated than small events, and tunnel flows are overall more poorly simulated than catchment outflows. This is probably due to the sensitive impacts of inaccurate simulation of water detention storage, antecedent soil moisture, reinfiltration on the surface, and disturbances in tunnel systems during small and medium stormflow events. The influence of these factors seems to
diminish with the increasing severity of storms. The accuracy of model simulation is considerably varies among the entire catchment and the internal areas. The accuracy of the model simulations seems to be related to the complexity of runoff generation processes, ability of the model to represent these processes, scale of simulation, and the inherit variability and uncertainty in parameterization.

Some recommendations can be drawn from this study for land management and landuse changes. Currently, to reduce the effects of crusting on runoff generation, farmers are advised to destroy the crust by hoeing when a storm is expected. The simulation results in this study indicated that this is effective for a small to medium events but not for very large events. The alternative measure is to put crop residues on the surface to dissipate rainfall kinetic energy. To reduce the effect of the plow pan, the farmers should deepen the plow depth, thereby increasing the water retention capacity in the soil layers above the plow pan. In general, the simulation results support the management changes implemented in the Yangdaogou catchment since 1980’s. Conversion of the upper gentle slopes into terrace lands increases crop yields and reduces the runoff entering gully and tunnel systems. However, terracing should be combined with the planting of vegetation to stabilize terrace banks. With the development of bottomland which can be used as croplands, all the slope lands will be replaced with vegetation land.
6. Bibliography


IMAGE EVALUATION
TEST TARGET (QA-3)

1.0  1.25  1.4  1.6
1.1  1.8
1.25  1.4  1.6

1.0  1.28  2.5
1.1  1.22
1.25  1.4  1.6

150mm

6"