Land surface representation for regional rainfall-runoff modelling, upper Blue Nile basin, Ethiopia

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by

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Dedicated to my late sister and brother
Yeshi Desalegn and Hailu Gebremichael
Abstract

The topography based distributed rainfall-runoff model (TOPMODEL) has been applied to the Gilgel Abay Basin which is the biggest of the four main tributaries of the Lake Tana basin. The model is applied to simulate outflows from the catchment and to predict spatial and temporal soil moisture dynamics and variable source areas in space and time. Digital elevation models (DEMs) of grid cell sizes from 30m to 500m have been used to test the effect of DEM grid cell size variation in the derived topographic index and hydrologic simulations. The model is found to perform satisfactorily with Nash-Sutcliff efficiency (Nash and Sutcliffe, 1970) of 0.782 for the 90m resolution DEM. Grid size is found to affect the distribution of topographic index significantly averaging effects on lower values and by increasing mean values as the DEM resolution gets coarser. This effect on the topographic index is also found to propagate in the hydrologic simulations and model efficiency. For the range of DEM grid sizes used in this study (60m to 500m), model efficiency is found to degrade slightly (1.83%) when DEM grid cell size increases from 60m to 500m by using the same calibrated parameters. Internal model predictions, such as overland flow component of the total runoff, the percentage of predicted variable source areas, also have been shown to be affected significantly. Larger grid cell sizes have been found to exaggerate the overland flow component of the total runoff. When the DEM grid cell size increases from 60m to 500m, the percentage of overland flow component of the total runoff increases from 9% to 22.9%. Increase in grid size also has been shown to increase the percentage of predicted variable source areas. While DEM grid size is increased from 60m to 500m the percentage of variable source areas predicted to be fully saturated is found to increase from 5.77% to 27.33% respectively at the peak flow rate.

Key words: Gilgel Abay, TOPMODEL, DEM, Model efficiency, Topographic index, Internal model predictions
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1. Introduction

1.1. General

Water is an essential resource to all mankind. Water is used in irrigation, water supply, transportation, recreation, and power production, etc. There are various processes that keep the global water in a continuous movement called the hydrologic cycle. The major processes are shown in Figure 1.1.

![Figure 1.1 The global water cycle (Fetter 2001)](image)

From the sea, lakes, land surface, or vegetation water evaporates and rises to the atmosphere. The vapor turns into clouds and there it can be transported by the wind to other places. If the condensed water particles become too heavy, the water falls back to the earth again as precipitation. This can be in the form of rainfall, snow, hail, but also as dew. Part of this precipitation may be intercepted and taken up by plants or it infiltrates to the subsurface or becomes runoff and flows to the nearby stream channels. It is also possible that the water can evaporate directly back to the atmosphere. Some of the water flowing through the ground returns to the surface to supply water to springs, lakes, and rivers. The water on the ground surface, in streams in lakes and part of the water used by plants returns back to the atmosphere through the combined action of evaporation and transpiration known as evapotranspiration. This succession of processes keeps the global water in a continuous cycle.
However, the availability of water on the globe is limited and as such Water is a scarce resource. Therefore to keep this scarce resource in a sustainable way, good water management is vital. Understanding hydrological processes and their spatial and temporal pattern over the catchment is basic for a good water management. The application of hydrological models helps in various ways to understand the spatial and temporal variability of processes.

The wide availability of digital elevation models nowadays and the possibility of coupling hydrological models with geographic information systems has made it possible to predict spatial and temporal patterns of hydrological response units in a catchment.

The topography based hydrological model TOPMODEL uses the topographic index of similarity, \( a / \tan \beta \), to define distributions of variable source areas in a catchment. The index is used to predict patterns of saturation over the catchment. All points with the same value of the index are assumed to respond in a hydrologically similar manner, which is advantageous from the fact that calculation of the index can be made only for those areas with different values of the index (Beven 1998).

1.2. Problem statement

A more thoughtful understanding of hydrological processes is vital for sustainable water resources development. There is a high degree of spatial and temporal variability of hydrological processes and in particular the runoff generating mechanisms are highly dynamic in response to rainfall. Therefore it is necessary to understand the major processes that contribute to the runoff at the outlet of the catchment and their spatial and temporal pattern so that a more integrated management at basin scale is possible. The application of hydrologic models helps in this respect. There are various types of models to choose between. Models like TOPMODEL allow spatial and temporal patterns of predictions to be mapped in to catchment space. However the predictive ability of such models has been shown to be affected by the spatial representation of the land surface features. Raster-based digital elevation models (DEMs) are commonly used for representing elevation surface in rainfall-runoff simulations. Topography based runoff models like TOPMODEL use DEMs to represent topography. As the grid size of the digital elevation model decreases, land surface features will be represented in more detail, with the effect propagating in the topographic attributes derived from these DEMs. Therefore, a need is identified to study both the effect of varying the representation of the topography on the derived attributes and the spatial and temporal pattern of these predictions. Although Gilgel-Abay catchment is the largest catchment contributing flow to the Lake Tana basin, only little has been known about the hydrology and runoff behaviour in particular of the catchment owing to the fact that not much study has been conducted in the basin.
1.3. Objectives of the study

The main objective of this study is to apply a physically based rainfall-runoff model which takes into consideration the effect of topography and to perform model sensitivity analysis in the Gilgel-Abay catchment.

Specific objectives are:

- To apply a physically based rainfall-runoff model (TOPMODEL) which explicitly considers the effect of topography so as to better understand the dynamics of the surface and subsurface flow contributing areas in the Gilgel-Abay catchment.
- To predict spatial and temporal soil moisture dynamics and variable flow contributing areas in the Gilgel Abay catchment.
- To analyze the effect of the spatial resolution of the digital elevation model on the distribution of the index and model simulation results.

1.4. Research questions

Based on the research objectives the following research questions are formulated:

- How effective is the application of TOPMODEL to the Gilgel Abay catchment?
- What are the main runoff generating mechanisms of Gilgel-Abay catchment and how are these spatially distributed over the catchment?
- What is the effect of the spatial resolution of a digital elevation model resolution on the topographic index distribution and hydrologic simulation results?

1.5. Thesis outline

This thesis consists of six chapters. The contents of each chapter are described in brief as follows.

The first chapter gives a general introduction about the global water cycle. The problem statement the research objectives and the research questions are also discussed in this chapter.

The second chapter gives a brief introduction of the study area. The location, topography, climate and land cover of the study area are discussed in brief in this chapter.

The third chapter is about the materials and methods used in this study. It gives a review of literatures this study is mainly based on. In this chapter are also discussed the field work activities in the study area.

The fourth chapter details on the data analysis and preparation processes. It explains the various data analysis processes conducted to prepare input data for the final modelling. Extraction and processing of digital elevation models and analysis of hydro meteorological data to prepare areal model input data are discussed in this chapter.

The fifth chapter presents the results obtained in this study and discusses the results. A brief presentation of the various results from the whole study and a discussion is given in this chapter.
The last chapter, chapter six, gives the conclusions drawn from the study and the recommendations based on the study conducted to indicate direction for future studies.
2. study area

2.1. Introduction

The study area is the Gilgel Abay catchment and is located in Lake Tana basin, north western part of Ethiopia. Gilgel-Abay catchment is the largest of the four main sub-basins of Lake Tana Basin. The river originates from the foot of mount Gish at an altitude of 2850m. This river contributes about 60% of the flow to Lake Tana basin and is gauged at two stations: Wetet Abay and Koga gauging stations. The total catchment area up to Wetet Abay gauging station is 1661 km² while the total area of this basin up to Lake Tana is estimated as 5000 km². This study considers part of the basin which is up to Wetet Abay gauging station.

2.1.1. Geographic location and topography

The catchment’s geographic location extends between 10° 56’ to 11°51’ N latitude and 36° 44’ to 37° 23’ E longitude.

Figure 2.1 Location of the study area
Rugged mountainous topography mainly characterizes most part of the catchment, especially the southern part. There is some low land within the basin as well. Elevation ranges from 1805 m a.m.s.l (above mean sea level) to 2811 m a.m.s.l (according to the SRTM DEM as shown in Figure 2.2)

![Figure 2.2 SRTM DEM of Gilgel Abay river basin](image)

**2.1.2. Climate**

The catchment falls within the Cool semi-humid zone that represents altitudes of 1800-2400 m a.m.s.l, with mean annual temperature of 17-20°C and the Cool and humid zones that represent higher altitudes of 2400-3200 m a.m.s.l, with mean annual temperature of 11.5-17°C. The dry season occurs between November and April while the wet season occurs mostly between May and October. Small rains also occur sporadically during April and May.

**2.1.3. Land cover/land use**

Rain fed agriculture is the main economic activity of the basin. Agriculture, predominantly rain-fed, is the mainstay of the whole Tana basin economy. The surrounding flood plains known as Fogera, Dembia, Alefa and Achefer, are intensively cultivated areas for more than centuries. Irrigated land and perennial crops, mainly coffee also cover a great amount of the area. One of the major natural forests covers a small part of the Gilgel Abay catchment at the north, which is known as Zege, the peninsula of the lake. Some marshes and wetlands occur at the floodplains, especially near the lake, with a dominant flora type of Papyrus (Cyperus papyrus). Bushes, shrubs and grassland including grazing land cover the rest of the basin. Many small urban areas are distributed throughout the whole basin of which biggest is Bahir Dar city, which is located on the south corner of the lake where the Blue Nile River starts. (Tessema 2006)
3. Materials and methods

3.1. Methodology

As any modelling, study this study starts with the conceptualization of the hydrology of the basin. General GIS based analysis supplemented by remote sensing imagery is the main methodology used in this study. The study has been conducted in two main phases: data collection phase and data analysis and model development phases.

During the data collection phase, a visit to the study area was made where data was collected by direct field measurements and from the concerned offices. The office data collected include: historical measurements on discharge, precipitation meteorological variables for evapotranspiration calculation and topographic maps of the study area (1:50,000). From the field level, ground control points for georeferencing of satellite images were collected. ASTER images covering the study area which are freely available have been retrieved from WebPages for extraction of digital elevation models (DEM) (http://edcimswww.cr.usgs.gov/pub/imswelcome/)

Model code development is also one of the major tasks performed in this study. The model development is based on the basic assumptions of TOPMODEL which will be explained in section 3.2.3. Basic equations governing flow components in TOPMODEL are organized to develop a code for simulations. The flow chart below shows the sequence of the major steps in the research process and methodology applied in this study.
3.2. Literature review

3.2.1. Previous studies
The hydrology of Gilgel-Abay basin has not been explored in detail. A related and recent study conducted in the basin is the MSC thesis entitled “Catchment Modelling and Preliminary Application of Isotopes for Model Validation in Upper Blue Nile Basin, Lake Tana, Ethiopia (Gragne 2007). In that study the HBV model is applied to the basin.
The main differences with this study are:

- TOPMODEL has been applied in this study that explicitly considers the effect of topography, that is of major influence on the dynamics of the flow contributing areas of the basin.
- Spatial and temporal patterns of soil moisture dynamics and variable source areas (see Figure 3.3) can be mapped in the catchment with TOPMODEL.
- A further objective of evaluating the effect of the spatial resolution of the DEM on simulation results is set under this study which is believed to show how simulation results are affected by the grid cell size of a DEM.

3.2.2. Runoff processes at the hill slope

There are various processes that contribute to the runoff at the outlet of a catchment. The following diagram shows a schematic representation of flow processes which contribute to the catchment runoff.

![Figure 3.2 Runoff processes at the hill slope (After Rientjes, 2007)](image)

Within the regional scale of the catchment, in theory all processes described above can be present. Infiltration excess overland flow (Horton overland flow) is generated by the infiltration excess mechanism when the rainfall intensity exceeds the infiltration capacity and when rainfall duration is longer than the ponding time of small depressions at the land surface. Saturation excess overland flow occurs when the subsurface is completely saturated by the rising groundwater table. This is usually the result of percolated water from upslope areas. A graphical representation of the expansion of such zones due to a rainfall event is shown in Figure 3.3.

On top of these saturated zones, overland flow is encountered in areas lying along the streams and channels. This is caused by the lateral flow contribution generated by exfiltration of subsurface water and by rainfall. These zones are termed as saturation overland flow source areas or variable source areas.
3.2.3. The TOPMODEL concept

Introduction
The more widespread availability of DEMs and the possible integration of the hydrological modelling with GIS has boosted hydrological modelling.

For catchments with moderate to steep slopes and relatively shallow soils overlaying an impermeable bed rock, topography does have an important role in runoff generation, at least under wet conditions arising from the effect of down slope flows (Beven 1998). The index of hydrological similarity, called the topographic index was first introduced by (Kirkby et al. 1975) and is the basis for the rainfall-runoff model, TOPMODEL.

TOPMODEL attempts to define hydrologic similarity of points in a catchment based on the topographic index. Grid cells that have an equal value for the topographic index respond hydrologically in the same way (Wolock 1995). This means that, if all other parameters are assumed to be lumped, grid cells with the same value for topographic index have an equal soil moisture deficit and are therefore assumed to be hydrologically similar. The index helps to understand the dynamics of flow contributing areas in a catchment as the catchment wets and dries.
The original TOPMODEL

TOPMODEL is based on variable contributing area concept. Basically the original TOPMODEL structure consists of the following components (Beven and Kirkby 1978).

(a) A variable contributing area component related to surface soil water storage in which rain falling on the contributing area, $A_c$, will immediately become overland flow.

(b) A surface interception and depression storage, $S_1$, with a maximum value $S_{D_1}$, which must be filled before infiltration can take place. Evapotranspiration is allowed from this store at the estimated potential rate until it is empty.

(c) A near surface infiltration store $S_2$. A storage based approach to infiltration is adopted, with a constant leakage rate, $i_0$, allowed from this store to the exponential subsurface store $S_3$, in the area that is not considered saturated. Input to the store $S_2$ takes place (once the interception store $S_1$ is filled) at the infiltration rate $I$ unless:

$$i > i_{\text{max}} = i_0 + b/S_2$$

(1)

In this case excess rainfall $I - i_{\text{max}}$ is considered to reach the basin outlet by surface route (infiltration excess overland flow). If under extreme conditions a maximum value of near surface storage, $S_c$, is exceeded, then again excess water is considered to reach the sub-basin outlet by a surface route (saturation excess overland flow). Losses due to evaporation are allowed from the store at a decreasing rate depending on the level of $S_2$. Thus

$$e_a = e_x S_2 / S_c$$

(2)

where

$e_x$ = the potential evapotranspiration remaining once the interception store $S_1$ is depleted

$e_a$ = the actual loss from the infiltration store.

Figure 3.4 Schematic representation of the TOPMODEL structure
(d) A nonlinear subsurface saturated soil water store, which provides the delayed response. The simplest form of this store is an exponential store for which

\[ q_b = q_0 \exp(S_3 / m) \]  

where
\[ q_b = \text{the flow reaching the channel from this store and} \]
\[ q_0 = \text{the flow when } S_3 = 0 \]
\[ m = \text{a constant.} \]

(e) A simple non linear routing algorithm based on velocity relationship completes the model.

\[ c(t) = CHA \cdot Q(t)^{CHB} \]  

where:
\[ Q(t) = \text{the discharge at the outflow of the whole basin at time } t \]
\[ c(t) = \text{an average kinematic wave velocity for the network which is assumed to be spatially constant and equal water to water velocity measure by tracer experiments} \]
\[ CHA \text{ and } CHB \text{ are constants.} \]

This original form of TOPMODEL has been used in different catchments by suitting it to specific situations and therefore the model after its inception has many versions. Nevertheless, the basic assumptions and concepts remain the same. More recently the dynamic TOPMODEL concept (Beven and Freer 2001) has been introduced that allows simulation of variable upslope contributing area (dynamic upslope contributing area). The dynamic TOPMODEL was found to have a slightly improved prediction of discharge and the time to peak. But the conclusion TOPMODEL concept since only a slight improvement is achieved. However, spatial predictions were reported to be more consistent, at least qualitatively, with existing conditions of the Slapton Wood catchment.

The topographic index

TOPMODEL uses the topographic index to define hydrologic similarity of different points in a catchment. All points with the same values of the topographic index values are assumed to behave in a hydrologically same manner. The distribution of the index can be calculated for any catchment and used as a basis for the prediction of the source areas, for both saturation excess overland flow and subsurface flow (Quinn et al. 1995). The index has the form:

\[ \ln\left(\frac{a}{\tan \beta}\right) \]  

where:
\[ a = \text{the upslope area per unit contour contributing flow to a pixel} \]
\[ \tan \beta = \text{the local slope angle acting on a cell (approximately taken as the local hydraulic gradient under steady state conditions)} \]

The assumption that all points with the same value of the index are assumed to behave in a hydrologically same manner has advantages from computation point of view in that the computation required to generate spatially distributed local moisture deficit patterns (water table patterns) reduces to one calculation per each class of the index (Beven 2000). The distribution function that can be
derived based on the topographic index is then used to analyse the responses of units at the catchment scale.

![Diagram](image)

**Figure 3.5 Distribution of \( a_i \), \( \tan \beta \), and topographic index across hill slope (Rientjes 2007)**

Fig 3.5 shows the distribution of topographic index in a hill slope where each subsurface element is interpreted as a linear reservoir. Saturated areas are assumed to contribute both to overland flow and subsurface flows while those areas which are not saturated or with moisture deficit only contribute to the subsurface flow during a rainfall event.

**Calculation of the topographic-index**

The calculation of the topographic index can be done by digital terrain analysis. The calculation of upslope contributing area and the local slope angle can be done by analysis of digital elevation models which are nowadays available or can be produced with preferred spatial resolution from available remote sensing imagery.

Various algorithms are available for the calculation of the upslope contributing area. Broadly classified these are single flow and multiple flow direction algorithms. Single flow direction algorithms do not require a contour length term as every pixel has the same contour length. However multiple flow algorithms have variable outflow direction that are dependent on a cells neighbour, and hence contour length must be taken in consideration (Quinn et al. 1995).

The various algorithms used to derive the upslope area calculation have an effect on the resulting topographic index. Wolock and McCabe (1995) have shown the effect of using single flow direction and multiple flow direction algorithms on the derived topographic index. According to this study the \( \ln(\frac{a}{\tan \beta}) \) distributions derived using the single flow direction algorithms have lower mean and higher variance and skew. They have suggested that the choice of algorithms affects spatial distribution of simulated hydrological characteristics such as soil moisture content. In this particular study the D8 (deterministic-8) algorithm with multiple flow directions is implemented.
Basic assumption of top model and governing equation for flow components

Given the original structure and assumptions of the TOPMODEL discussed above, the model has many versions since its inception in 1978. Basic assumptions, however, remained the same. The basic equations used in coding the model for this study are given in the following section. All Equations given in this section are based on paper by Beven (1998) and the book Beven (2000). Only final results of formulas have been given here. Detailed derivation of the formulas can be found in the above mentioned literatures.

The fundamental assumptions underlying the TOPMODEL concept are (Beven 1998)

1. The dynamics of the saturated zone can be approximated by the successive steady state representation of the saturated zone on an area, $a$, draining into a point on a hill slope

2. The hydraulic gradient of the saturated zone can be approximated by local surface topographic slope measured with respect to plan distance, $\tan \beta$

3. To this is added a further assumption about the nature of local transmissivity profile with depth.

The original assumption of TOPMODEL uses an exponential decline of transmissivity with moisture deficit (with depth below the ground) given by:

$$T = T_0 \exp\left(\frac{D}{m}\right).$$  

where

- $T_0 =$ transmissivity at saturation  
  $[L^2T^{-1}]$

- $D_i =$ local storage deficit  
  $[L]$

- $m =$ a scaling parameter  
  $[L]$

Under the above assumptions, at any point $i$ on a hill slope the down slope saturated subsurface flow rate, $q_i$, is given by:

$$q_i = T_0 \tan \beta_i \exp(-D_i / m)$$  

Assuming a spatially homogeneous recharge rate $r$ entering the water table, the subsurface down slope flow per unit length $q_i$ may also be given by:

$$q_i = ra$$

The formula for any point relating the recharge rate $r$, the local topographic index, the scaling parameter and the transmissivity at saturation is

$$D_i = -m \ln\left(\frac{ra}{T_0 \tan \beta_i}\right)$$

Local moisture deficit at any time step is determined based on the catchment average soil moisture deficit. If the catchment average soil topographic index $\gamma$ is given by

$$\gamma = \frac{1}{A} \int_{A} \ln\left(\frac{a_i}{T_0 \tan \beta_i}\right) dA$$
then, the relationship between the local and catchment average storage deficit is given by:

\[
D_i = D - m \left[ \gamma - \ln \frac{a_i}{T_0 \tan \beta_i} \right] \tag{11}
\]

Equation (11) shows that the local moisture deficits at any time step is determined based on catchment average moisture deficit scaled by the parameters \(m\) and \(T_0\). Using equation (11) soil moisture at any time step can be mapped in the catchment (model) space (in a distributed fashion) which helps in understanding the spatial and temporal soil moisture dynamics to be visualized. This is regarded as one of TOPMODEL’s predictive abilities.

If \(Q_0\) is the discharge when the mean deficit is zero, the total drainage from the saturated zone (which is the same as the sum of all the subsurface flows given by (7) as:

\[
Q_b = Q_0 \exp(-D/m) \tag{12}
\]

\[
Q_0 = T_0 \exp(-\gamma) \tag{13}
\]

During model simulations, the catchment average deficit at any time step is determined by subtracting the unsaturated zone recharge from the average deficit at the previous time step and by adding the base flow as calculated for the previous time step

\[
D = D + \left[ \frac{Q^{t-1}_b - Q^{t-1}_U}{A} \right] \tag{14}
\]

where,

- \(Q^{t-1}_U\) = the drainage from the unsaturated zone (gravity) recharge [L^3T^1]
- \(A\) = is the catchment area [L^2]

To initialize the saturated zone of the model, an inversion of equation (12) will be used. If an initial base flow discharge at time zero, \(Q^{t=0}_b\), is known which is assumed to be totally a result of drainage from the saturated zone, equation (12) can be inverted to give an initial value of the catchment average storage deficit \(D\) as

\[
D^{t=0} = -m \ln \left( \frac{Q^{t=0}_b}{Q_0} \right) \tag{15}
\]

Therefore once we have an initial value of \(D\) the local storage deficit \(D_i\) are calculated by using equation (11)

The vertical flux from the unsaturated zone to the saturated zone is given by

\[
Q^{t}_u = \frac{S_{uc}}{S_i * td} \tag{16}
\]

where

- \(Q_u\) = the vertical (gravity) drainage from the unsaturated zone
- \(S_{uc}\) = the unsaturated zone storage
- \(S_i\) = local saturated zone deficit
\[ t_d = \text{time delay constant of the unsaturated zone} \]

Actual evapotranspiration is calculated as a function of potential evapotranspiration and maximum root zone moisture storage deficit. A reduction from the potential evapotranspiration value occurs depending on the moisture status of the root zone.

\[ E_a = E_p \left( 1 - \frac{SRZ}{SRMAX} \right) \]  

(17)

where

- \( E_a \) = actual evapotranspiration
- \( E_p \) = the potential evapotranspiration
- \( SRZ \) = the root zone storage
- \( SRMAX \) = maximum root zone storage deficit

Any remaining water based on the relative values of the unsaturated zone storage and local storage deficit is allowed to evaporate with the maximum limit of \( SRMAX \).

During model, any area without moisture deficit (zero or less local moisture deficit \( D_r \)) or those areas which are saturated during the previous time step will contribute both to the subsurface and surface flow. On the other hand, those unsaturated areas (areas where local moisture deficit greater than zero) will contribute only to the subsurface flow. Therefore once storage deficits are calculated, equation (12) will be used for calculation of base flow from unsaturated areas. From the saturated areas, the total runoff is calculated as the sum of the subsurface flow as given by equation (12) and the overland flow. The overland flow consists of two components: the saturation excess overland flow and the infiltration excess overland flow. Infiltration excess overland flow is calculated based on the Green and Ampt assumptions (see section 3.2.6). The model is then completed by a simple channel network routing algorithm of the form shown in equation (4).

**Choice of transmissivity profile**

In its original form, TOPMODEL assumes an exponential decline of soil transmissivity with depth, a form that is frequently observed in the soil upper layers (Beven, 1984) and that is easy to handle analytically. The transmissivity profile applicable to the particular catchment has impact on the model simulation results because the base flow recession directly depends on the transmissivity profile applied.

![Figure 3.6 Types of transmissivity profiles in TOPMODEL](Ambroise et al. 1996)
Apart from the original form of exponential transmissivity decline with depth, parabolic and linear decline of transmissivity with depth have been identified. Figure 3.6 gives the types of transmissivity profiles and their conceptualization. The transmissivity profile that is applicable to the particular catchment can be determined from base flow recession curve analysis.

The base flow recession curve of a catchment, which expresses the way in which the base flow, $Q_b$, decreases naturally during interstorm and dry periods can take various forms, depending on the nature and complexity of the catchment. Tallaksen (1995) has provided a comprehensive review of recession curve analysis. For many catchments the recession curve could be fitted by very simple functions of time $t$. The two more frequently observed functions are:

1. The exponential function, a form that is typical of little incised streams draining thick aquifers:
   \[ Q_b = Q_s \exp(-t/t_s) \]  
   (18)

2. The second-order hyperbolic function, a form that is typical of shallow aquifers that are well drained by the stream, as is the case for instance in many granitic areas:
   \[ Q_b = Q_s (t/t_s)^{-2} \]  
   (19)

Where $t_s$ is scaling time related to a specified reference discharge of $Q_s$.

And in the original assumption of exponential transmissivity decline of TOPMODEL gives rise to first order hyperbolic function of time:
   \[ Q_b = Q_s (t/t_s)^{-1} \]  
   (20)

These recession curves can be transformed to be linear functions of time as follows:

<table>
<thead>
<tr>
<th>First order hyperbolic</th>
<th>Second order hyperbolic</th>
<th>Exponential</th>
</tr>
</thead>
<tbody>
<tr>
<td>$1/Q_b - 1/Q_s = \tau / Am$</td>
<td>$1/\sqrt{Q_b} - 1/\sqrt{Q_s} = \tau \sqrt{Q_0} / Am$</td>
<td>$\ln Q_b - \ln Q_s = -\tau Q_0 / Am$</td>
</tr>
</tbody>
</table>

These transformed equations are therefore used to determine which transmissivity profile is applicable to the catchment. From the analysis of discharge records of the basin during periods which are not affected by base recharge and large evapotranspiration, the transformed recession curve should plot as a straight line for the assumed transmissivity profile.

### 3.2.4. Deriving model parameters through recession curve analysis

One of the major steps in the modelling process is the parameterization of the model. Basically TOPMODEL requires only a small number of parameters which can be derived from information of the catchment topography (DEM), soils and recession curve analysis.

As discussed in the previous sections the topographic index and the catchment area can be derived from digital terrain analysis process. There are the other parameters that need to be determined will be the scaling parameter and the transmissivity at saturation. These parameters can be derived from analysis of observed flow. The transformed recession curves as given by equation (21) which are linear functions of time have constant slope $\alpha$ for the three cases of transmissivity given by:
Exponential  \[ \alpha = \left( \frac{Q_s}{A_m} \right)^{-1} = \frac{1}{A_m} \]
Parabolic  \[ \alpha = \left( \sqrt{\frac{Q_s}{A_m}} \right)^{-1} = \frac{\sqrt{Q_0}}{A_m} = \left( \sqrt{A y_m} \right)^{-1} \]
Linear  \[ \alpha = t_s^{-1} = \frac{Q_0}{A_m} = \left( \gamma m \right)^{-1} \] (22)

Therefore once the form of transmissivity profile and its transformed slope \( \alpha \) are known from observed discharges the two main parameters \( m \) and \( T_0 \) are estimated by the above conditions (see section 4.2.3).

### 3.2.5. Other TOPMODEL based studies and parameter values used

TOPMODEL has been applied in a number of humid temperate climates. Catchments ranging from less than 1 km\(^2\) up to 500 km\(^2\) have been modelled using TOPMODEL with correspondingly variable parameter values. In nearly all TOPMODEL studies topography related parameters are considered spatially variable while soil related parameters often are considered spatially uniform. Moreover, in most of the modelling studies, the \( T_0 \) parameter as found by calibration and assumed to be spatially constant over the catchment is found to be very large. Some studies also have shown that the DEM grid cell size has an effect on the calibrated \( T_0 \) value , see for example Saulnier et al.(1997). An overview of parameter values used in various TOPMODEL applications is presented in Table 3-1 below.
<table>
<thead>
<tr>
<th>Catchment</th>
<th>Area (Km²)</th>
<th>DEM (m)</th>
<th>γ</th>
<th>m(m)</th>
<th>T₀ (m²/h)</th>
<th>Reference</th>
<th>Comment</th>
</tr>
</thead>
<tbody>
<tr>
<td>Gardsjon, G1, Sweden</td>
<td>0.0063</td>
<td>5</td>
<td>5.1</td>
<td></td>
<td>1.8</td>
<td>Seibert et al., 1977</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>f=13m⁻¹, variable Δθ</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>(f is the equivalent of m)</td>
<td></td>
</tr>
<tr>
<td>Saeternbekken Minifelt, Norway</td>
<td>0.0075</td>
<td>2</td>
<td>5</td>
<td>0.0053</td>
<td>1.31</td>
<td>Lamb, 1996</td>
<td></td>
</tr>
<tr>
<td>Ecerex B, French Guiana</td>
<td>0.015</td>
<td>2.5</td>
<td>5.62</td>
<td>0.0035</td>
<td>7</td>
<td>Molicova et al, 1997</td>
<td></td>
</tr>
<tr>
<td>Ringelbach, France</td>
<td>0.34</td>
<td>5</td>
<td>5.94</td>
<td>0.041</td>
<td>2.75</td>
<td>Ambroise et al., 1996b</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Parameters for exponential transmissivity version</td>
<td></td>
</tr>
<tr>
<td>Slapton wood, UK</td>
<td>1</td>
<td>10</td>
<td>7.87</td>
<td>0.004-0.25</td>
<td>0.01-30</td>
<td>Fisher and Beven, 1996</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Parameter ranges used in Monte Carlo experiments</td>
<td></td>
</tr>
<tr>
<td>White Oak Run, VA USA</td>
<td>5</td>
<td>30</td>
<td>6.04-6.08</td>
<td>0.027</td>
<td>0.0007-0.0012</td>
<td>Wolock and McCabe, 1995</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Different parameters derived from different DEM grid sizes and analysis algorithms</td>
<td></td>
</tr>
<tr>
<td>Maurets, France</td>
<td>8.4</td>
<td>20-120</td>
<td>6.40-6.96</td>
<td>0.025</td>
<td>1.05-1.5</td>
<td>Saulinier et al, 1997</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>For different DEM grid sizes after excluding river pixels from \ln(a/tanβ) distribution</td>
<td></td>
</tr>
<tr>
<td>Wye, UK</td>
<td>10.5</td>
<td>10-100</td>
<td>5-9.8</td>
<td>0.0093</td>
<td>9.223-27.11</td>
<td>Quinn et al., 1995</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td></td>
<td>Different parameters derived from different DEM grid sizes</td>
<td></td>
</tr>
<tr>
<td>North fork Rivanna, VA</td>
<td>456</td>
<td></td>
<td>7.64</td>
<td>0.0092</td>
<td>11.75</td>
<td>Beven and wood, 1983</td>
<td></td>
</tr>
</tbody>
</table>

### 3.2.6. The Green-Ampt infiltration model

Most TOPMODEL studies ignore infiltration excess overland flow. In some of the studies infiltration excess is assumed that it does not contribute to the channel flow, see for example Quinn and Beven (1993). In this study Infiltration excess overland flow is modelled based on the modified Green-Ampt infiltration model which is developed by Beven (1984).
The original Green-Ampt infiltration model
The Green–Ampt model (Green and Ampt 1911) was originally derived to predict the infiltration of ponded water in a uniform soil. In this model, it is assumed that water infiltrates as a piston like sharp wetting front. The soil above the wetting front is assumed to be fully saturated and the soil below the wetting front is at its initial water content, although some studies have indicated that the soil profile above the wetting front may not be fully saturated, see for example Gavin and Xue (2007). According to the model, runoff will not occur until the rainfall intensity is below the saturated hydraulic conductivity $k_s$. Here there cases can be considered for ponding to occur

- The rainfall intensity exceeds the infiltration capacity of the soil throughout the whole time step so that ponding must occur for the entire time.
- The infiltration capacity exceeds the rainfall intensity at the beginning but will later be exceeded by the rain rate. Then ponding starts at later time steps
- The infiltration capacity exceeds the rain rate through out the rainfall event and the rainfall infiltrates for the whole time step

The Figure below illustrates the variation in moisture content $\theta$ with depth $Z$ below the surface, at a point in time when the front has progressed a distance $L$.

![Figure 3.7 Wetting front of the Green-Ampt model](image)

As water advances through the soil the passage of the wetting front increases the moisture content of the soil from its initial value $\theta_i$ to the saturated value $\theta_s$. The difference is the moisture deficit given by:

$$\Delta \theta = \theta_s - \theta_i \quad (23)$$

where:

$\Delta \theta$ = the moisture deficit which is a fraction

The total head that causes infiltration is given by (see Figure 3.7):
\[ h = h_0 + L + \psi \]  \hspace{1cm} (24)

Where

- \( h_0 \) = the ponding at the surface which is usually neglected
- \( L \) = the depth of water that is already infiltrated \([L]\)
- \( \psi \) = a suction head at the wetting front due to capillary attraction in the voids \([L]\)

The cumulative depth of infiltration \( F \) is given by

\[ F = \Delta \theta * L \]  \hspace{1cm} (25)

If the hydraulic conductivity is \( k \) then Darcy’s law can be applied to give the infiltration rate by neglecting the ponding depth as

\[ i = \frac{dF}{dt} = -k \frac{\partial h}{\partial z} = k \left( \frac{\psi + \frac{F}{\Delta \theta}}{\Delta \theta} \right) \]

And the above equation simplifies to

\[ i = k \left( 1 + \frac{\Delta \theta * \psi}{F} \right) \]  \hspace{1cm} (26)

Where \( i \) = the infiltration rate

Equation 26 is called the Green-Ampt infiltration equation

**Beven’s version of the Green-Ampt model**

The modified Green-Ampt model (Beven 1984) assumes, as the original Green-Ampt model, water infiltrates as a piston like wetting front.

In this model, however, the hydraulic conductivity is assumed to be an exponential function of depth, \( z \). The effective hydraulic conductivity behind the wetting front is given in terms of the hydraulic conductivity at the surface, \( k_0 \) as:

\[ k(z) = k_0 e^{fz} \]  \hspace{1cm} (27)

Where

- \( k_0 \) = the hydraulic conductivity at the surface \([LT^{-1}]\)
- \( z \) = depth below soil surface \([L]\)
- \( f \) = constant
If the wetting front has reached a depth $z$, the infiltration rate, $i$, is given by the generalization of the relationship for layered soils (Childs and Bybordi 1969):

$$i = \frac{dI}{dt} = \frac{\Delta \psi + z}{\int_{z=0}^{z=z} [dz / k(z)]}$$  \hspace{1cm} (28)

Where $I$ is the cumulative infiltration [L].

In the approach it is assumed that at the time of ponding $t_p$, the wetting front is at a depth $z_p$ which is given by the infiltration rate $I_p$ divided by the change in moisture content $\Delta \theta$:

$$z_p = \frac{I_p}{\Delta \theta}$$  \hspace{1cm} (29)

The cumulative infiltration at the time of ponding is given by:

$$I_p = r t_p$$  \hspace{1cm} (30)

where

$r$ = the rain rate

At the onset of ponding the rain rate $r$ is assumed to be equal to the infiltration rate $i$ and the infiltration rate is given by

$$r = \frac{k_{0} f (\Delta \psi + I_p / \Delta \theta)}{1 - e^{-b / \Delta \theta}}$$  \hspace{1cm} (31)

At any time after ponding has started, the cumulative infiltration is $I$ and the infiltration rate is given by

$$\frac{dI}{dt} = \frac{k_{0} f^* (\Delta \psi + I / \Delta \theta)}{1 - e^{-b / \Delta \theta}}$$  \hspace{1cm} (32)

The time required for the surface to reach saturation is the time to ponding. Ponding will only occur when the rain rate is greater than or equal to the hydraulic conductivity at the surface. The time to ponding, $t_p$, is calculated as:

$$t - t_p = \frac{1}{k_{0}} \left[ \ln(I + c) - \frac{1}{e^{f^* c}} \ln(I + C) + \sum_{m=1}^{\infty} \frac{\left[ f^* (I_p + C) \right]^m}{m! m} - \lambda \right]$$  \hspace{1cm} (33)

Where

$C = \Delta \psi / \Delta \theta$ is the storage suction factor and

$f^* = -f / \Delta \theta$

$\lambda$ is a constant given by:

$$\lambda = \ln(I_p + c) - \frac{1}{e^{f^* c}} \left[ \ln(I_p + c) + \sum_{m=1}^{\infty} \frac{\left[ f^* (I_p + C) \right]^m}{m! m} \right]$$

Equation (33) is used in Newton-Raphson iterative procedure to calculate the time to ponding. By comparing the infiltration rate with the rain rate the excess rainfall and hence the infiltration excess overland flow is calculated.
3.3. Field work

A visit to the study area was made to collect necessary data for the study and to get visual impression of the landscape and hence the main hydrological features of the catchment. Two types of data were collected that are, office level (archived data) and field level (primary) data.

3.3.1. Field level activities

The field activities include:

- Ground truth collection for land cover classification and geo-referencing of satellite images and also ground control points collection for DEM extraction using hand held GPS

- Soil moisture measurement using soil moisture probe. The main aim of this soil moisture measurement was to gain an insight in the soil moisture distribution in the catchment

- Piezometers installation to measure local hydraulic gradient. The main aim of this installation of the piezometers is to measure the local hydraulic gradient in order to have an idea of the relationship between the topographic slope and water table gradient (hydraulic gradient) since in the TOPMODEL approach it is assumed that local hydraulic gradients are equal to local topographic surface slopes. However, only one piezometer that is installed near the river was able to bear water at a depth of 0.7m and all the other were dry. The holes were dug using auger of 1.5m depth and it was not possible to dig deeper holes then 1.5 meter. Therefore this trial was unsuccessful

Figure 3.8 Piezometers installation to measure local hydraulic gradient
3.3.2. Data collected from offices

Hydro-meteorological data

Hydrological data includes both hydrological data like river discharge and meteorological data such as rainfall, temperature, wind speed, sunshine hours relative humidity and pan evaporation.

Discharge data of Gilgel Abay river basin is collected from Ministry of Water Resources (MoWR). River discharge of Gilgel Abay River measured at Wetet Abay gauging station (see Figure 3.9) from 1973 to 2006 has been collected. A plot of the collected river discharge data is shown in figure 3.10.

Figure 3.9 Wetet Abay gauging station

Figure 3.10 Discharge record of Gilgel Abay River at Wetet Abay gauging station (1973-2004)
Rainfall and other meteorological variables
Rainfall data of 10 meteorological stations for 10 years (1997-2006) have been collected from the National Meteorological Agency of Ethiopia (NMAE). The selected meteorological stations and their location are shown in Figure 3.11. Apart from rainfall data, 5 years of daily data on temperature (max and min) sunshine hours, wind speed, relative humidity and pan evaporation have been collected that are used for the calculation of evapotranspiration. Among the selected stations, only two of the stations, Bahir Dar and Dangilla, are principal stations and it is from these two stations that meteorological variables for evapotranspiration calculations are collected.

Table 3-2 UTM coordinates of selected meteorological stations for the study

<table>
<thead>
<tr>
<th>Station name</th>
<th>UTM X</th>
<th>UTM Y</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abay Sheleko</td>
<td>267107.7</td>
<td>1259013.43</td>
</tr>
<tr>
<td>Bahir Dar</td>
<td>327062.45</td>
<td>1243881.08</td>
</tr>
<tr>
<td>Dangila</td>
<td>262994.68</td>
<td>1244290.59</td>
</tr>
<tr>
<td>Gundil</td>
<td>288627.18</td>
<td>1210920.95</td>
</tr>
<tr>
<td>Kidamaja</td>
<td>259512.91</td>
<td>1216653.11</td>
</tr>
<tr>
<td>Kimbaba</td>
<td>323607.15</td>
<td>1277085.23</td>
</tr>
<tr>
<td>Sekela</td>
<td>304679.72</td>
<td>1214109.59</td>
</tr>
<tr>
<td>Wetet Abay</td>
<td>287108.5</td>
<td>1257029.48</td>
</tr>
</tbody>
</table>

Figure 3.11 Location of meteorological stations selected for the study
Figure 3.12 Annual rainfall of selected meteorological stations for the study (1997-2004)

Topographic maps
Topographic maps of the area are obtained from the Ethiopian Mapping Authority. For this study 13 Topographic maps of scale 1:50,000 were collected to have good coverage.
4. Data analysis and preparation

4.1. DEM extraction and processing

4.1.1. Introduction

Hydrological processes in landscapes are influenced to a great extent by catchment topography. Often, the spatial distribution of topographic attributes serves as a measure of spatial variability of hydrological processes and allows the dynamics of these processes to be mapped over the catchment easily. Nowadays catchment topography is analysed easily with digital elevation models.

A digital elevation model is an ordered array of numbers that represent the spatial distribution of elevations above some arbitrary datum in a landscape. It may consist of elevations sampled at discrete points or the average elevation over a specified segment of the landscape, although in most cases it is the former(Moore et al. 1991). DEMs are a subset of digital terrain models (DTMs) which can be defined as an ordered array of numbers that represent the spatial distribution of terrain attributes

4.1.2. Data networks of digital elevation models

When discussing the use of DEMs it is important to consider the way in which the surface representation is to be used. The ideal structure of a DEM may be different if it is to be used as a structure for a dynamic hydrologic model than if it is used to determine the topographic attributes of the landscape (Maathuis 2007)

Three principal ways of structuring a network of elevation data for its acquisition and analysis are:
1 triangulated irregular networks (TINs)
2 grid based networks
3 contour-based networks

Triangulated irregular networks (TINs)

Triangulated irregular networks form irregular network of points stored as a set of x,y and z coordinates together with pointers to their neighbours in the net(Peucker et al.1978,mark 1975) The elemental area is the plane joining the three adjacent points in the network and is known as a facet(Peucker et al. 1978). The Triangulated irregular network (TIN) is a popular model for representing surface models in geographic information systems (GIS), computer graphics, and virtual reality (VR) because it has a simple data structure and can easily be rendered using common graphics hardware(Yang et al. 2005)

Grid based networks

In grid based networks, the terrain is represented as elevations sampled on a regular grid. The main problem in grid based networks is that elevations are sampled on a regular grid, which may result in under sampling of data points under rugged topography. There are many types of grid based networks. The choice of grid based methods is related primarily to the scale of the area to be examined and the utilization of the DEM. The data can be stored in a variety of ways, but the most efficient is the z
coordinates corresponding to sequential points along a profile with a starting point and grid spacing also specified.

The most widely used data structure consists of square-grid networks because of their ease of computer implementation and computational efficient (Collins and moon 1981). However they do have several disadvantages including:

i. they can not easily handle abrupt changes in elevation, although Frank and Nelson discussed possible ways of modelling these discontinuities;

ii. the size of the grids mesh affects the result obtained and the computational efficiency (Panuska et.al., 1990)

iii. the computed upslope flow paths used in hydrological analysis tend to zig zag and therefore are somewhat unrealistic and

iv. The precision is lacking in the definition of specific catchment areas.

Since regular networks must be adjusted to the roughest terrain, redundancy can be significant in sections with smooth terrain (Peucker et al. 1978), whereas triangular irregular networks are more efficient and flexible in such circumstances.

Contour based networks

Contour based network consist of digitized contour lines that are stored as Digital Line Graphs (DLGs) in the form of x, y coordinate pairs along each contour line of specified elevation. These can be used to subdivide an area in irregular polygons bounded by adjacent contour lines of and adjacent stream lines (Moore, 1988; Moore and Grayson, 1989, 90) and are based on stream path analogy first proposed by Onstad and Braceiokse (1968)

4.1.3. Digital elevation model extraction from ASTER stereo imagery

The Advanced Space borne Thermal Emission and Reflection Radiometer (ASTER) on-board the National Aeronautics and Space Administration’s (NASA’s) Terra spacecraft provides along-track digital stereo image data at 15-m resolution. The sensor is designed to provide image data in 14 visible, near-infrared, short wavelength infrared and thermal infrared spectral bands. Stereo image data are recorded only in Band 3, which is the near-infrared wavelength region from 0.78 to 0.86 Am, using both nadir and backward looking telescopes. The ASTER Earth Observation Satellite offers nearly simultaneous capture of stereo images, minimizing temporal changes and sensor modelling errors. This means that the nadir and backward images are acquired with in a few seconds difference. This provides a stereo images of consistent quality from which a DEM can be automatically extracted.

DEM generation process

Digital elevation model is extracted from the ASTER level1A data of the visible near infrared channel (3N and 3B). The DEM was extracted in ERDAS IMAGINE using LPS Automatic Terrain Extraction module. Some preprocessing of the images was necessary while importing the images to ERDAS IMAGINE to ensure that there is no stripping of lines and to correct panoramic distortion and calibrate the image by applying a polynomial transformation of order1 to the image data. The raw level1A images imported as imagine image format are then used to generate the DEM in LPS (Leica Photogrametric Suit). Photogrametric techniques such as digital image matching are used to extract the images automatically. DEM generation by these methods involves several steps. A block file is
created for the geometric sensor model (orbital push broom) under which the ASTER satellite is fully supported. The reference coordinate systems are then set. The two image pairs are then added to this block file where these are stored for further analysis and generating of DEM. As the raw images do not have pyramid layers the first step in the DEM generation process in LPS is creating pyramid layers for all the images. After the pyramid layers have been created, tie points were collected. In this study all tie points are generated automatically. Moreover, no ground control points (GCPs) were used (which means that the DEM generated is not an absolute DEM but is a relative DEM). 300 tie points per image were generated and 100% Point Success rate for both the 3N and 3B images was obtained. The pattern success rates were 72% for the 3N image and 64% for the 3B image due to the difference in quality of the nadir and back ward looking scene images. The study area is covered by two ASTER images and therefore it was necessary to do two separate block files for the two images. In the second image also the Point Success rate was 100% for both images but the pattern success rate reduces to 57% for the 3B image due to the lower quality of the second image. After the tie points are generated, the coordinates (XYZ) of these points are assigned to the points by the process of triangulation. The simple gross error check after the triangulation shows that the “total image unit weight RMSE” is 0.00179 for the first image and 0.0020606 for the second mage. Once the tie points are assigned the appropriate XYZ coordinates, the next step is the DTM extraction process. The DTM extraction process consists of the following steps (LPS online help)

- **Digital Image Matching for DTM Mass Point Collection.** Ground points appearing within the overlap portion of the left and right images associated with a DTM are identified. This is referred to as digital image matching. The resulting output consists of the image location of ground points appearing within a DTM.
- **Ground Point Coordinate Determination.** The 3D coordinates of a ground point whose image positions have been automatically identified are computed using photogrammetric principles. This is referred to as 3D ground point calculation. The calculated 3D ground points are also referred to as mass points.
- **DTM Construction.** The automatically extracted and calculated mass points are used as a basis for constructing a DTM. DTMs are constructed differently depending on the specified DTM output type.

In this particular study the DTM is extracted in the form of a single mosaic DTM as a DEM. Figure 4.1 show the extracted DEMs of 30, 60 and 90 m resolution. The other DEM grid cell sizes (150, 250, 400 and 500m) are then resampled from the 90m resolution DEM. The resampled DEMs differ from the original extracted DEMs of the same size that means when a 30 m resolution DEM is extracted and resampled to a 60m grid size, the resulting DEM differs from that extracted with 60 m resolution with out resampling.
4.1.4. Effect of DEM grid size and quality on hydrologic prediction and assessment

Effect of grid size on hydrologic simulation results
The accuracy of prediction of most distributed topographic driven hydrological models, such as TOPMODEL depends on the characterization of land surface such as the local topographic slope and contributing areas. Previous studies, for example, by Zhang and Montgomery (1994) have shown the effect of varying the Grid size of DEM on hydrologic simulations by using TOPMODEL and TOPOG. They have shown that increasing the grid size of DEM increases mean values of the topographic index thereby increasing the areas predicted as saturated by the model. This increase in saturated areas was reported to result in the prediction of high overland flow component of the total runoff due to more surface contributing areas.

The accuracy of DEMs partly depends on the accuracy and spatial distribution of the original data from which the DEMs are produced. Grid size represents the horizontal resolution while vertical accuracy is related to the closeness of the height values in the DEM to the height values of points in the real landscape from a certain reference datum. Some studies (Kenward et al. 2000), for example have shown the effect of the vertical accuracy of DEMs on hydrologic simulation. They have shown that vertical accuracy affects simulation results and prediction of mean annual runoff volumes were reported to vary to a great extent when using DEMs of different accuracy. Therefore both the grid size
(horizontal resolution) and the vertical accuracy have effect on the simulated hydrological processes. This study however will concentrate on the effect induced by the change in the grid size of the DEMs.

**Accuracy assessment of extracted ASTER DEM**

In order to assess the accuracy of the extracted DEMs, very accurate height measurements of known points in the field are necessary. The accuracy for example can be determined by comparing the computed Z-coordinate values (extracted height values from the DEM produced) at check points with those collected from the topographic maps or differential GPS (DGPS) surveys.

During the field survey, ordinary hand held GPS is used with a vertical accuracy of ± 10 meters and this is not reliable to be used as a comparison to assess the accuracy of the extracted DEM. Therefore in the absence of a reliable ground control point collected with a differential GPS, the comparison is made with a ground survey made for Koga Irrigation project farm road. These points, however, cover only a small portion of the total image and the elevation range of the control points is not wide. Moreover, the survey area is relatively flat and does not cover a range of elevations. Therefore, only 7 representative points in the region covered by the survey have been taken and compared. Based on the comparison made, the ASTER DEM is found to be closer to the actual surveyed elevations than the SRTM DEM (see Table 4-1). It should, however, be noted that this is not the same for all areas covered by the extracted DEM. The ground survey covered only a relatively flat area which is not covered by dense vegetation. On the other hand, Points collected with the hand held GPS in mountainous areas are closer to the SRTM DEM. Within the range of points used for comparison, the comparison shows there is a good accuracy which is within the design specification of RMSE of 7 to ± 50m see for example Akira et.al (2002)

<table>
<thead>
<tr>
<th>GROUND SURVEY</th>
<th>ASTER</th>
<th>SRTM</th>
</tr>
</thead>
<tbody>
<tr>
<td>2006.073</td>
<td>1990.7</td>
<td>2015</td>
</tr>
<tr>
<td>2007.659</td>
<td>1995.3</td>
<td>2020</td>
</tr>
<tr>
<td>2008.906</td>
<td>1996.4</td>
<td>2020</td>
</tr>
<tr>
<td>2009.358</td>
<td>1995.6</td>
<td>2020</td>
</tr>
<tr>
<td>2010.222</td>
<td>1997</td>
<td>2022</td>
</tr>
<tr>
<td>2011.027</td>
<td>1996.7</td>
<td>2022</td>
</tr>
<tr>
<td>2012.198</td>
<td>1996.7</td>
<td>2023</td>
</tr>
</tbody>
</table>

**Comparison of ASTER and SRTM DEMs**

A comparison of the height values extracted from both the SRTM DEM and the DEM derived from ASTER image is also done by taking vertical horizontal transect lines in both DEMs at same location. The comparison is made using the 90m DEM derived from ASTER image and the 90m resolution SRTM DEM by subtracting height values of ASTER derived DEM from that of SRTM DEM. The result is shown in Figure 4.2 with the statistics. The comparison shows that the height values extracted from the DEM derived from ASTER image in almost all pixels along the transects are lower than that of the SRTM DEM. Relatively similar values are obtained in flatter areas and in areas without dense vegetation and the difference is maximum in rugged and mountainous areas and in areas where vegetation cover is high.
Figure 4.2 Comparison of height values extracted from ASTER and SRTM DEMs (a) along a vertical transect line and (b) along a horizontal transect line
4.1.5. DEM processing

This section is based on the DEM hydro-processing module developed by Maathuis (2006) which supports processing of the raw DEM to obtain relevant hydrological parameters such as drainage network, (sub) catchment boundaries, slope aspect etc. including the various components of topographic index used in TOPMODEL. The flow direction algorithm applied is the Deterministic-8 or (D8) flow direction algorithm with multiple flow direction.

Calculation of Components of the topographic index

The raw DEMs extracted are used to extract a number model input parameters. First local pits are removed from the extracted DEMs and the resulting sink free DEMs are used to derive drainage network of the catchment and the topographic index for all DEMs. The derivation of two main components of the topographic index, the accumulated upslope contributing area per unit contour length and the tangent of local topographic slope, and the drainage networks is discussed in the following section.

Flow determination

Local depressions from the extracted (raw) DEMs are removed by the fill sinks procedure to ensure flow from every pixel in the DEM. The procedure avoids local depressions of single pixels and of multiple pixels from the DEM (Maathuis 2006). In this procedure:

- The height value of a single-pixel depression is raised to smallest value of the 8 neighbors of a single-pixel depression.
- The height values of a local depression consisting of multiple pixels are raised to the smallest value of a pixel that is both adjacent to the outlet for the depression, and that would discharge into the initial depression.

The sink free DEM is then used to derive a flow direction map using the flow direction operation. The flow direction operation determines in which neighboring pixel any water in a central pixel will flow. Flow direction is calculated for every central pixel in input blocks of 3 by 3 pixels, each time comparing the value of the central pixel with the value of its 8 neighbouring pixels. The output map contains flow directions as N (to the North), NE (to the North East), etc.

After determining the direction of flow from each pixel in the DEM, the flow direction map is used to determine the flow accumulation map where a cumulative count of the number of pixels that naturally drain into outlets based on their flow direction is performed. The operation can also be used to find the drainage pattern of a terrain. This operation gives the accumulated count of the number of pixels draining to each cell (pixel). To calculate the upslope contributing area per unit contour length, this count should be multiplied the corresponding DEM pixel size. The calculated flow accumulation maps from the respective DEMs have a minimum value of 1 which implies that the minimum upslope contributing area for each DEM is equal to the corresponding DEM grid cell size. The output of the above operations is shown in Figure 4.3 for the 90m resolution DEM extracted from ASTER image.
Figure 4.3 Result of the DEM hydro-processing of the extracted 90m resolution ASTER DEM for Wetet Abay catchment (a) the original DEM (b) the filled DEM (c) flow direction map and (d) flow accumulation map

DEM optimization

DEM optimization is a process to improve a DEM so that drainage patterns derived from the analysis of DEM better follow the actual drainage networks of the terrain. In flat areas and in areas where rivers meander the drainage network extracted from the DEM do not usually match well with actual drainage network. This can be identified easily by comparing extracted drainage network with actual drainage networks, for example, using other satellite images (like ASTER images) as background.
The DEM optimization operation burns existing (actual) drainage networks of the terrain which can be digitized from satellite images in the DEM so that in the subsequent flow direction operation the existing drainage pattern will be followed better.

In this study the ASTER image has been used to digitize main drainage features of the catchment. These digitized drainage lines are then ‘burnt’ on the original DEM. Subsequent flow direction and flow accumulation operations are then done using the improved DEM. Figure 4.4 shows the optimized DEM with the digitized drainage networks.

![Figure 4.4 Digitized drainage lines from ASTER image (left) and optimized DEM (right) for Wetet Abay catchment](image)

**Drainage networks**

The optimized DEM is used to derive the drainage networks of the catchment by the drainage network extraction operation. Here a threshold value to initiate a stream is required. This operation allows the threshold to be expressed in terms of number of pixels or a stream initiation threshold map based on many factors that influence the drainage density in basin like geology, soil, relief etc. If a geological or a soil map is available the units of this map can be reclassified to represent flow accumulation threshold values where units with coarse grained sandy soils overlaying deeply weathered sandstones can be assigned higher thresholds compared to thin soils occurring over shales reflecting the lower permeability and little resistance to erosion.

In this study, the threshold map is based on internal relief since only the DEM is available which is classified in four threshold upper bound classes. The resulting drainage network is compared with main drainage features in the ASTER satellite image to see if the threshold is assigned properly.

The extracted drainage networks are assigned a Strahler order by the drainage ordering operation. In this operation a minimum drainage length should be specified so that a stream having a length equal to or greater than this specified length remains in the network where as all other streams will be
ignored. In this study, drainage lengths starting from 750 m down have been used as the minimum drainage length. A minimum drainage length of 500m has been found to be proper. Figure 4.5 shows the drainage network of the area by Strahler order. Comparison of the (main) drainage network extracted with that in the aster image shows a good match by using the optimized DEM.

![Drainage network by strahler order of Wetet Abay catchment](image)

**Figure 4.5 Drainage network by strahler order of Wetet Abay catchment**

Calculation of Local topographic slope

The DEMs extracted are further used to derive the local topographic slope map used for the calculation topographic index. In TOPMDEL, the local topographic slope is assumed to be equal to local hydraulic gradient. These maps are derived from the slope map calculated in degrees. Figure 4.6 shows the $\tan \beta$ map calculated for the catchment from the 30, 60, and 90 and resampled 150, 250, 400 and 500m resolution DEM.

It is noted from Figure 4.6 that as the DEM grid cell size increases it tends to decrease the maximum and mean values of the tangent of the local topographic slope. A plot of the cumulative fraction of area against $\tan \beta$ value is shown in Figure 4.7. Minimum values of the $\tan \beta$ for DEM grid cell sizes from 30m to 250m are 0 but become larger than 0 when the DEM resolution decreases (resampled DEMs of 400m and 500m) as shown in Figures 4.6 and 4.7.
Figure 4.6 Map of the $\tan \beta$ for Wetet Abay catchment for different DEM grid cell size (a) 30m (b) 60m (c) 90m and (d) 150m (e) 250m (f) 400m and (g) 500m resolution DEMs
As the grid size of the DEM increases from 30m to 500m the maximum value of $\tan \beta$ reduces from 2.09 to 0.556 and the mean value from 0.547 to 0.159. Table 4.2 shows the statistical summary of calculated $\tan \beta$ maps from different DEM grid sizes and the corresponding plot is shown in Figure 4.8.

Table 4.2 Statistical summary of $\tan \beta$ from different DEM resolution

<table>
<thead>
<tr>
<th>Grid size(m)</th>
<th>30</th>
<th>60</th>
<th>90</th>
<th>160</th>
<th>250</th>
<th>400</th>
<th>500</th>
</tr>
</thead>
<tbody>
<tr>
<td>mean</td>
<td>0.547</td>
<td>0.418</td>
<td>0.368</td>
<td>0.309</td>
<td>0.239</td>
<td>0.182</td>
<td>0.159</td>
</tr>
<tr>
<td>Max.</td>
<td>2.09</td>
<td>1.949</td>
<td>1.172</td>
<td>1.002</td>
<td>0.802</td>
<td>0.65</td>
<td>0.556</td>
</tr>
</tbody>
</table>

Figure 4.7 Cumulative frequency distribution of the $\tan \beta$ values from different DEM grid sizes

Figure 4.8 Variation of mean and maximum values of $\tan \beta$ with DEM grid cell sizes
4.2. Preparation of areal model input data

4.2.1. Areal rainfall

Areal rainfall in this study is estimated by the thiessen polygon method. Seven meteorological stations have been used for averaging of the areal rainfall by the thiessen polygon method. The selected stations and the thiessen polygon map produced is shown below in Figure 4.9

Thiessen weights for each station are calculated by dividing the area of influence of each station by the total area of the catchment and this is used as a weighting factor for these stations. The thiessen weight for each station is given in Table 4-3

![Thiessen polygon method of areal rainfall estimation for Wetet Abay catchment](image)

**Table 4-3 Thiessen weights for areal rainfall estimation**

<table>
<thead>
<tr>
<th>Station</th>
<th>Thiessen weight</th>
</tr>
</thead>
<tbody>
<tr>
<td>Abay_sheleko</td>
<td>0.01424203</td>
</tr>
<tr>
<td>Bahir_Dar</td>
<td>0.10395524</td>
</tr>
<tr>
<td>Dangila</td>
<td>0.14170547</td>
</tr>
<tr>
<td>Gundil</td>
<td>0.21871795</td>
</tr>
<tr>
<td>Kidamaja</td>
<td>0.06244126</td>
</tr>
<tr>
<td>Sekela</td>
<td>0.18082942</td>
</tr>
<tr>
<td>Wetet_Abay</td>
<td>0.27810863</td>
</tr>
</tbody>
</table>

4.2.2. Evapotranspiration

Daily evapotranspiration is estimated by FAO Penman-Monteith method of calculating reference evapotranspiration. For calculation of potential evapotranspiration, time series of data from meteorological stations Dangilla and Bahir Dar are used. Evapotranspiration is calculated using the FAO Penman-Monteith equation.
The FAO Penman-Monteith equation is given by: (Allen et al.)

\[
ET_0 = \frac{0.408\Delta(R_a - G) + \gamma \frac{900}{T + 273}U_2(e_s - e_a)}{\Delta + \gamma(1 + 0.34U_2)}
\] (34)

where:

- \(ET_0\) = reference evapotranspiration [mm.day\(^{-1}\)]
- \(R_a\) = net radiation at the crop surface [MJ m\(^{-2}\) day\(^{-1}\)]
- \(G\) = soil heat flux density [MJ m\(^{-2}\) day\(^{-1}\)]
- \(T\) = air temperature at 2 m height [°C]
- \(U_2\) = wind speed at 2 m height [m s\(^{-1}\)]
- \(e_s\) = saturation vapor pressure [kPa]
- \(e_a\) = actual vapor pressure [kPa]
- \(e_s - e_a\) = saturation vapour pressure deficit [kPa]
- \(\Delta\) = slope of vapour pressure curve [kPa °C\(^{-1}\)]
- \(\gamma\) = psychometric constant [kPa °C\(^{-1}\)]

4.2.3. Recession curve analysis

An estimate on the two main TOPMODEL parameters, the scaling parameter and the transmissivity at saturation can be made by recession curve analysis from historical river discharge data. The base flow recession curve of a catchment shows the way how the base flow decreases naturally in periods which are not affected by large evapotranspiration, snow melt and rainfall (Ambroise et al. 1996). Once the base flow recession curve is calculated it can be transformed in to simple functions of time as shown by equations (21). The constant slope of these curves gives a fixed relation ship between the two parameters depending on the transmissivity profiles applicable to the particular catchment.

In this study the master recession curve is calculated by the RECESS program. The program scans the stream flow dataset, finding periods of continuous recession and when a segment is found the user decides whether to use all or part of the segment. Once a decision is made as to which part of the recession segment is to be used, the program calculates a mathematical expression of the form

\[ t = K_1 \times \log Q + K_2 \]

Where \(t\) is time in days and \(\log Q\) is the logarithm of the flow in ft\(^3\)/s and \(K_1\) and \(K_2\) are coefficients that are determined by linear regression. This expression is used to derive the recession index (days/log cycle), which is the absolute value of \(K_1\). The program similarly proceeds to the next recession period and solves the same regression equation and the process continues until all recession periods are analyzed by selecting recession segments to be analyzed.
The program then performs calculations to determine calculations for the master recession curve by regression equation. Discharge records of 33 years (from 1973 to 2003) are used and the master recession curve is plotted. This master recession curve is then transformed into the three forms explained in equation (21). Figure 4.9 shows the master recession curve and the transformation. From the plots it is clearly noted that the logarithmic or LN (Q) transformation of the discharge axis plots as a straight line. This suggests a linear transmissivity profile! Here in this study, however, an exponential transmissivity based on the original assumption of TOPMODEL has been adopted. The constant slope is calculated which gives a first estimate of the scaling parameter.

![Master recession curve and transformation](image)

**Figure 4.10** Master recession curve and transformation for Wetet Abay catchment (a) master recession curve (b) straight line transformation (c) parabolic straight line transformation and (d) Ln (Q) transformation (which plots as straight line)

### 4.3. TOPMODEL code development

The various relationships discussed in section 3.2 were used to develop The TOPMODEL code used in this study. Due to the relatively large area of the catchment (1659Km²) which, for a 30 m resolution digital elevation model size, has more than 1.8 million pixels in the topographic index file, it was difficult to run the code developed in spreadsheet (Excel). An attempt was also made to integrate the code in to GIS but it was not possible due to the same computing efficiency problems.
Therefore an IDL code has been used. The source code is retrieved from the webpage (http://instaar.colorado.edu/topoflow/Downloads/Code/TopModel_in_IDL/). The first conversion was made by Scott Peckham.

The code is essentially a conversion of Keith Beven’s FORTRAN version of TOPMODEL with some modification and addition of the infiltration excess overland flow component. The IDL code is found to be able to handle the relatively very large size of the topographic index histogram. An increased possibility of simulating in larger number of time steps is also achieved as compared to the FORTRAN version.

Basically the IDL code consists of two of two main subroutines. The first one is to simulate subsurface flow and saturation excess overland flow from the variable contributing areas. The other is an infiltration model to simulate infiltration excess overland flow based on the Green-Ampt model as described in section 3.2.6

The subsurface component simulates subsurface (base flow) based on equation (12). With $T_0$, $\gamma$ and $m$ being constants for the catchment, this equation requires only the average catchment soil moisture deficit. The average moisture deficit $D$ is initialized based on a known base flow at the start of the simulation period according to equation (15). Local storage deficit is calculated based on catchment average moisture deficit according to equation (11).

The other component is the infiltration excess overland flow as discussed in section 3.2.6. In this component the time to ponding, $t_p$ is calculated from equation (34) based on the Newton-Raphson iterative procedure. Excess rainfall and hence infiltration excess overland flow is then calculated based on the assumption that all rain will infiltrate until the time to ponding, $t_p$ has reached.

### 4.3.1. Model inputs

The main inputs of the model are:

- A topographic index file which is a text file that contains two lines. The first one gives the percentage of the basin area with particular values in a histogram bin and the second is the largest value of the topographic index in that histogram bin. Despite the relatively large size of TI files used in this study, no computational efficiency problem is encountered.
- An area distance file which is also a text file that gives the cumulative distribution of area with distance from the main outlet for routing of surface flow. Here the main catchment is subdivided in a number of sub catchments to obtain this distribution.
- A parameter file which is also a text file that contains the values of the different parameters used in this model. In this case there are 12 parameters used. The parameters and their definition is given in Table 4-4
- A file containing observed discharges, rainfall and evapotranspiration with the number and value of time step to be used for computation
### Table 4-4 Parameter values used and their definition

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Description</th>
<th>Unit</th>
</tr>
</thead>
<tbody>
<tr>
<td>$m$</td>
<td>A parameter which controls the rate of exponential decrease of transmissivity with increasing soil moisture</td>
<td>[m]</td>
</tr>
<tr>
<td>LN_T0</td>
<td>The natural logarithm of the transmissivity at saturation</td>
<td>[m²/hr]</td>
</tr>
<tr>
<td>TD</td>
<td>Time delay constant for routing unsaturated zone flow</td>
<td>[-]</td>
</tr>
<tr>
<td>CHV</td>
<td>An effective surface routing velocity</td>
<td>[m/hr]</td>
</tr>
<tr>
<td>RV</td>
<td>Stream velocity</td>
<td>[m/h]</td>
</tr>
<tr>
<td>SRMAX</td>
<td>Maximum root zone storage</td>
<td>[m]</td>
</tr>
<tr>
<td>Q₀</td>
<td>Initial base flow</td>
<td>[m/hr]</td>
</tr>
<tr>
<td>SR₀</td>
<td>Initial value of root zone storage</td>
<td>[m]</td>
</tr>
<tr>
<td>INFEX</td>
<td>An infiltration flag</td>
<td>[-]</td>
</tr>
<tr>
<td>K₀</td>
<td>Hydraulic conductivity at surface</td>
<td>[m/h]</td>
</tr>
<tr>
<td>$\Delta \psi$</td>
<td>Effective suction head</td>
<td>[m]</td>
</tr>
<tr>
<td>$\Delta \theta$</td>
<td>Moisture deficit</td>
<td>[-]</td>
</tr>
</tbody>
</table>
5. Results and discussion

5.1. Distribution of topographic index

The topographic index has been derived from the digital elevation model extracted from ASTER images for DEM grid cell sizes of 30, 60, 90, 150, 250, 400 and 500m.

Figure 5.1 Map of the $\ln(a / \tan \beta)$ index from different DEM grid cell sizes for Wetet Abay catchment (a) 30m (b) 60m (c) 90m (d) 150m (e) 250m (f) 400m and (g) 500m resolution DEMs (all DEMS from ASTER image)
DEM with grid cell size of 150, 250, 400 and 500m have been resampled from the 90 m grid cell size DEM that has extracted from the ASTER image.

The upslope contributing area is calculated per unit contour length from the accumulated number of pixels. The $\tan \beta$ maps calculated are the other inputs for the topographic index map. Some undefined pixels, especially for lower resolution DEMs which have minimum $\tan \beta$ value of zero, appeared in near river pixels in the calculated topographic index maps. These pixels have been filtered out by taking the majority value. This effect decreases as the DEM resolution gets coarser where the minimum values of $\tan \beta$ achieve values higher than zero as can be seen from the $\tan \beta$ maps for different resolution DEMs in Figure 4.6.

It is clearly observed from Figure 5.1 that increasing the DEM grid cell size affects more the lower and mean values of the topographic index than the higher values in lower resolution DEMs. Figure 5.2 shows a plot of the cumulative frequency distribution of the topographic index. It shows that as the grid sizes increase lower and mean values tend to increase while the effect on highest values is minimal on higher resolution DEMs. However, the effect is not the same for the lower resolution resampled DEMs (150m and 250m 400m and 500m) in which case the higher values also tend to decrease significantly. The cumulative frequency distribution curve for lower resolution DEMs crosses the axis before that of higher resolution DEMs.

![Cumulative frequency distribution of the $\ln(a / \tan \beta)$ index for different digital elevation model grid cell sizes](image)

As the grid size of the DEM increases from 30m to 500 m, the minimum value of the index increases from 2.877 to 6.837. The mean value changes only slightly, from 10.49 to 10.88. Although the effect of increasing grid size in the higher values of the index size at the higher resolution DEMs is minimal, it becomes significant for lower(coarser) resolution DEMs (especially the resampled 150,250 400 and
500m resolution DEMs in this case) as can be observed from Figures 5.1 and 5.2. The maximum value of the topographic index as the DEM size increases from 30 to 90 m remains fairly the same (changes from 25.4 to 24.3) while its value decreases significantly when the DEM grid cell size increases from 90m to 500m (from 24.3 for the 90m resolution to 20.057 for the 500m resolution) but mean values still increase slightly. As the $\ln(a / \tan \beta)$ index is an index of wetness and disappearance of lower values and increasing mean values of the index would suggest more areas to be predicted as saturated.

### 5.2. Hydrograph simulation

First calibration was done using the 90m grid size DEM (Figure 5.3). Calibration is done manually each time changing parameter values in the text file containing parameters. The model is able to simulate the observed discharges with satisfactory efficiency. The Nash and Sutcliffe efficiency coefficient (Nash and Sutcliffe 1970), here referred to as $R^2$, is taken as a measure of the model efficiency which is given by:

\[
R^2 = 1 - \frac{\sum_{i=1}^{n}(Q_{\text{sim}} - Q_{\text{obs}})^2}{\sum_{i=1}^{n}(Q_{\text{obs}} - Q_{\text{avg}})^2}
\]

Where

- \( n \) = the number of time steps
- \( Q_{\text{sim}} \) = Simulated flow at each time step
- \( Q_{\text{obs}} \) = Observed flow at each time step
- \( Q_{\text{avg}} \) = average of the flow

A model efficiency ($R^2=0.7823$) has been achieved using the 90m grid size DEM. A relatively bigger misfit is observed in the base flow recession part. The rising limb fits very well to the observed discharge. Peak discharges are also simulated very well except for some locations like the one observed in the year 2003. This can be taken as a good performance of the model since areal rainfall is estimated from seven meteorological stations with only two of the stations at the periphery of the catchment and all the other stations being out of the catchment boundary (As shown in Figure 3.11).
The volumetric assessment of the simulated and observed discharges has been made by plotting the accumulated volume of both the observed and simulated discharges. The curve shows that most of the time the actual volume of water leaving the catchment is comparable with the simulated one except in the year 2003 where a small gap is observed between the simulated and observed volume. Total volume error for the simulation period is 11.2% of the observed discharge.
5.2.1. Model sensitivity

Model sensitivity analysis to parameters has been carried out using the 90m grid cell size DEM. The sensitivity analysis in this case a manual procedure in which parameter values are changed manually each time the model is run in the text file containing the parameters. The model is found to be highly sensitive to 3 out of the 12 parameters used. These are the root zone storage parameter, the scaling parameter, and the transmissivity at saturation. Sensitivity of the model to the three parameters is assessed independently by keeping two parameters unchanged when the model is run to analyse the sensitivity of the third parameter. The results are shown below. In Figure 5.5 sensitivity to \( SRMAX \) is shown with the parameter values with their corresponding model efficiency shown in table 5-1.

![Diagram of model sensitivity](image-url)

Figure 5.5 Model sensitivity to the maximum root zone storage deficit parameter \( SRMAX \)
(a) \( SRMAX = 0.01 \) and (b) \( SRMAX = 0.018 \)
Table 5-1 Model sensitivity to SRMAX

<table>
<thead>
<tr>
<th>Run</th>
<th>SRMAX</th>
<th>$R^2$</th>
<th>hydrograph</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.01</td>
<td>0.782</td>
<td>Figure 5.3</td>
</tr>
<tr>
<td>2</td>
<td>0.018</td>
<td>0.731</td>
<td>Figure 5.4(a)</td>
</tr>
<tr>
<td>3</td>
<td>0.03</td>
<td>0.568</td>
<td>Figure 5.5(b)</td>
</tr>
</tbody>
</table>

So far in most TOPMODEL studies the scaling parameter $m$ and the transmissivity at saturation are found to be most sensitive. The parameter $m$ controls the decline of transmissivity (exponential decline in this case) with increasing moisture deficit. As the transmissivity at any point and any moisture deficit in the catchment depends on the transmissivity at saturation these two parameters, $T_0$ and $m$, are linked to each other. The sensitivity of the model to the scaling parameter is shown in Figure 5.6.
and the corresponding model run with the parameter value and model efficiency are shown in Table 5-2. It is noted from Figures 5.6 (a) and (b) that although the misfit in the recession part of the simulated discharge is improved by optimizing the $m$ parameter (decreases from 0.5 to 0.2) overall efficiency degrades from 0.782 to 0.664 and obviously, the recession part of the hydrograph shows the delayed response of the catchment and the transmissivity has influence in this part of the hydrograph.

<table>
<thead>
<tr>
<th>Run</th>
<th>$m$ (m)</th>
<th>$R^2$</th>
<th>hydrograph</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>0.05</td>
<td>0.782</td>
<td>Figure 5.3</td>
</tr>
<tr>
<td>2</td>
<td>0.03</td>
<td>0.688</td>
<td>Figure 5.6 (a)</td>
</tr>
<tr>
<td>3</td>
<td>0.02</td>
<td>0.664</td>
<td>Figure 5.6 (b)</td>
</tr>
</tbody>
</table>

### 5.3. Spatial and temporal soil moisture dynamics and predicted saturated source areas

One of TOPMODEL’s important capabilities is that it allows mapping of predictions in a distributed fashion the catchment space at any time step in the course of simulation. Spatial and temporal patterns of soil moisture and variable source areas can be mapped back in space at any time step.

Soil moisture at any point in the catchment model space depends on the catchment averaged soil moisture content as given by equation (11). With the average topographic index and the scaling parameters being constants for a catchment, the soil moisture in space entirely depends on the local topographic index values once the average catchment soil moisture deficit is known. Here it has been tried to map soil moisture patterns in the catchment in the course of simulation at five time moments for the hydrograph of the year 2003 of Figure 5.3. This allows mapping of the saturated flow contributing areas at the same time moments. Based on this prediction, the percentage of the (saturated) catchment area that contributes to the saturation excess overland flow have been determined at the corresponding time moments to show the dynamics of surface contributing areas in the catchment during the events. Mapping is started during the dry period and then additional points at the rising limb, at the peak rate of runoff, one at the falling limb and at the end of the simulation period have been used. Figure 5.7 below shows the predicted variable saturated flow contributing areas during the course of simulation.
Figure 5.7 Spatial and temporal prediction of saturated source areas in the Wetet Abay catchment during the course of simulation for the hydrograph of the year 2003 (a) date16/05/2003 (dry period) (b) date26/06/2003 (rising limb) (c) at the time when the runoff rate is at the peak rate (date11/07/2003) (d) date06/10/2003 (falling limb) and (e) at the end of the simulation period (date31/12/2003) [Note: that negative value of moisture deficit means saturation]
### Table 5-3 Percentage of the catchment area predicted as saturated corresponding to Figure 5-7

<table>
<thead>
<tr>
<th>Time from start of simulation(hr/date)</th>
<th>Saturated contributing area (Km²/ %)</th>
</tr>
</thead>
<tbody>
<tr>
<td>20760/(date 16/05/2003)</td>
<td>39.55Km²(2.238%)</td>
</tr>
<tr>
<td>21740/(date 26/06/2003)</td>
<td>128.18Km²(6.8%)</td>
</tr>
<tr>
<td>22100/(date 11/07/2003)</td>
<td>175.09Km²(10.55%)</td>
</tr>
<tr>
<td>24190/(date 06/10/2003)</td>
<td>108.58Km²(6.54%)</td>
</tr>
<tr>
<td>26290/(date 31/12/2003)</td>
<td>55.53Km²(3.35%)</td>
</tr>
</tbody>
</table>

### 5.4. Effect of DEM resolution on the simulation results

#### 5.4.1. Effect on the model efficiency

In order to test the sensitivity of the TOPMODEL approach to grid size the model was run with the calibrated parameters of the 90m resolution DEM using the topographic index from respective DEMs. The size of topographic index file used in the different DEM grid sizes varies significantly. Due to very large size of the topographic index file for the 30m resolution (more than 1,843 744 pixels or 262,202 values in the histogram) simulations are started by the 60 m resolution DEM which has lesser cells (see Table 5.4). The model is then executed using topographic index files from each DEM grid size. The sensitivity of the model prediction to DEM grid size is analysed first using the Nash Sutcliffe model efficiency indicator as criteria. The first calibration with the 90m resolution DEM resulted in a model efficiency of 0.7823. The Model run with the 60m resolution DEM results in a very slight reduction of model efficiency, which is 0.781. For all other DEM grid cell sizes model efficiency degrades slightly with the lowest efficiency of 0.764 when using the 500m grid cell size DEM. This means that the overall model efficiency reduces by only about 1.83% while the DEM grid cell size changes from 60m to 500m (see Figure 5.8).

This result fairly agrees with previous TOPMODEL studies by (Saulnier et al. 1997; Wu et al. 2007). The later have shown that serious efficiency drop was found starting from a DEM grid cell size of 1000m up to which efficiency was fairly constant by using calibrated parameters. The former have shown the dependence of the calibrated $T_0$ value on grid size and how the decrease in model efficiency due to DEM grid cell size can be compensated by calibrating the $T_0$ parameter.

### Table 5-4 Size of topographic index file (number of histogram bins) used for each DEM resolution

<table>
<thead>
<tr>
<th>DEM Grid cell size</th>
<th>30m</th>
<th>60m</th>
<th>90m</th>
<th>150m</th>
<th>250m</th>
<th>400m</th>
<th>500m</th>
</tr>
</thead>
<tbody>
<tr>
<td>size of TI file</td>
<td>262,020</td>
<td>94,373</td>
<td>53,484</td>
<td>23,155</td>
<td>11,579</td>
<td>10,375</td>
<td>4,051</td>
</tr>
<tr>
<td>Total number of grid cells</td>
<td>1,843,744</td>
<td>460,936</td>
<td>204,850</td>
<td>73683</td>
<td>26,456</td>
<td>5,754</td>
<td>6,626</td>
</tr>
</tbody>
</table>
Figure 5.8 Simulated and observed flows for different DEM grid cell sizes with calibrated parameters for the 90m resolution DEM

Figure 5.9 Overland flow component of the total flow hydrograph for different DEM resolution

5.4.2. Effect on the overland flow component

The overland flow component is significantly affected by the change in DEM grid cell size. The percentage of overland flow to the total flow changes from 9% to 22.9% when DEM grid cell size changes from 60m to 500m. This can be explained by the disappearance of the smaller values and
increase of the mean values of the topographic index values when grid cell size is changed from 60m to 500m. The overland flow component of the total flow hydrograph is shown in Figure 5.9 and the percentage of overland flow to the total flow is shown in table 5.5 when simulated using respective DEM grid cell sizes.

Table 5-5 Model efficiency and percentage of overland flow predicted using respective DEM resolution

<table>
<thead>
<tr>
<th>DEM resolution</th>
<th>60m</th>
<th>90m</th>
<th>150m</th>
<th>250m</th>
<th>400m</th>
<th>500m</th>
</tr>
</thead>
<tbody>
<tr>
<td>% overland flow</td>
<td>9%</td>
<td>11.6%</td>
<td>12.9%</td>
<td>16%</td>
<td>20.2%</td>
<td>22.9%</td>
</tr>
<tr>
<td>Model efficiency</td>
<td>78.1%</td>
<td>78.23%</td>
<td>78.04%</td>
<td>77.70%</td>
<td>77%</td>
<td>76.4%</td>
</tr>
</tbody>
</table>

5.4.3. Effect on the spatial and temporal soil moisture dynamics and variable source areas

The spatial and temporal distribution of soil moisture distribution and the extent of the saturated source areas are also, to a great extent, affected by the change in DEM grid cell size. Saturated source areas are predicted at the peak runoff rate for all grid sizes and a high variation is observed as the DEM resolution changes. When the DEM resolution changes from 60m to 500m, the corresponding predicted percentage of saturated overland flow source area changes from 6% (95.73Km²) to 27.3% (453.5Km²) respectively. This can be explained by considering equation (11). The spatial distribution of soil moisture deficit depends on the scaling parameter, the difference of average and local value of the topographic index and the average soil moisture deficit. The scaling parameter being kept constant for all grid sizes, the local soil moisture deficit depends on the average moisture deficit and the difference between local and average topographic index values, both of which depend on DEM resolution. The percentage of area predicted as saturated at the peak rate of runoff for the hydrograph of the year 2003 is given in Table 5-6 and the variable source areas mapped back in catchment space at the same time moments for the 60m and 500m resolution DEMs is shown in Figure 5.10.

Table 5-6 Percentage of saturated contributing area at the peak flow rate from different DEM resolution

<table>
<thead>
<tr>
<th>DEM resolution</th>
<th>60m</th>
<th>90m</th>
<th>150m</th>
<th>250m</th>
<th>400m</th>
<th>500m</th>
</tr>
</thead>
<tbody>
<tr>
<td>Saturated area %</td>
<td>5.77</td>
<td>7.4</td>
<td>11.09</td>
<td>16.58</td>
<td>23.26</td>
<td>27.33%</td>
</tr>
<tr>
<td>Saturated area Km²</td>
<td>95.734</td>
<td>124.424</td>
<td>184.02</td>
<td>275.125</td>
<td>385.92</td>
<td>453.5</td>
</tr>
</tbody>
</table>
Figure 5.10 Predicted saturated contributing area at the peak flow rate for the 60m resolution (left) and 500m resolution DEMs (right) (negative value of soil moisture deficit indicates saturation)
6. Conclusions and recommendations

6.1. Conclusions

This study has applied TOPMODEL to test the applicability and efficiency of the model in the Gilgel Abay basin and to assess the effect of DEM grid cell size on the topographic attributes (topographic index) and simulation results. Predicting spatial and temporal soil moisture dynamics and variable source areas is also the aim of this study. Based on the study conducted the following conclusions are drawn:

- Although four out of the six rainfall measuring stations used for areal rainfall estimation are out of the catchment boundary and despite the high inconsistency that resulted from errors in observation observed in the rainfall and discharge data used for modelling and relatively large size of the catchment area, TOPMODEL is proven to perform with an efficiency of 78.23% in the Gilgel Abay basin using a 90m resolution DEM which is considered as satisfactory efficiency.

- The capability of TOPMODL in terms of its spatial and temporal prediction abilities has been investigated. The approach is able to model variable source areas in the course of simulation and spatial and temporal soil moisture dynamics can be predicted. A high variation of saturation excess overland flow has been observed in the course of simulation which reaches the maximum at the peak flow rate. These results, however, must be verified with actual distributed data from the field.

- DEM grid cell size has been found to affect the distribution of topographic index significantly. As the DEM grid size becomes larger it tends to increase the minimum values for all DEM grid sizes. Minimum value of the index increases from 2.87 to 6.84 when the DEM grid cell size increases from 30m to 500m. Mean values also have been observed to increase slightly. The effect on the maximum values, however, has been found to be minimal up to some size (150m in this case) and has been shown to be significant on the resampled DEMs of 150m, 250m, 400m, and 500m grid cell size.

- The effect of DEM grid size on the topographic index values has been observed to propagate in the TOPMODEL predictions. For the range of DEM grid cell sizes used in this study (60m to 500m) model efficiency has been found to degrade slightly. However, the effect on the other model predictions like overland flow component and percentage of saturated contributing areas is found to be very significant. Overland flow component of the total runoff changes significantly (from 9% to 22.9%) as DEM grid cell size changes from 60m to 500m. The percentage of area predicted as saturated also changes significantly (from 5.77% to 27.33% for the 60m and 500m grid cell size DEMs respectively). Therefore for the range of DEM grid sizes used in this study, the change in DEM grid cell size affects the model efficiency in terms of the Nash and Sutcliffe efficiency coefficient only slightly. However, the amount of saturated contributing area and overland flow components of the total runoff have been affected significantly.
6.2. Recommendations

Although TOPMODEL is found to perform with a fairly very good efficiency in the Gilgel Abay catchment, the model efficiency and predictions can be improved considering the following:

- So far in TOPMODEL studies the transmissivity profile assumed to model a particular catchment has been shown to have impact on TOPMODEL predictions, see for example (Ambroise et al. 1996b). Although, a quick analysis of the recession curve using the RECESS program has shown applicability of a linear transmissivity profile where the transformed master recession curve plots as a straight line for the LN(Q) transformation as shown in Figure 3.9, this study has applied an exponential transmissivity profile with depth by retaining the original assumption of transmissivity profile. Therefore it is recommended to test the applicability of linear profile for the catchment with a more detailed analysis of the recession characteristics of the base flow.

- The range of DEM grid cell sizes applied in this study varies between 30m and 500m due to constraints in the availability of sources for finer resolution DEMs than 30m. Previous TOPMODEL based studies, for example, Zhang and Montgomery (1994) have suggested more finer resolution DEM as an appropriate size for the TOPMODEL simulations. Therefore the model performance can be checked for those finer resolution DEMs in the future.

- TOPMODEL has been shown, in this study, to be sensitive for a set of parameters like the scaling parameter, the transmissivity at saturation and the maximum root zone moisture deficit. The model sensitivity analysis for these parameters is performed manually. This can, however, be done automatically by Monte Carlo simulations which should give a better estimate of parameter space.

- Digital elevation model for this study is extracted from ASTER image without using any ground control points. The quality of the extracted DEM could, however, be improved by using reliable ground control points from other sources like differential GPS or a ground survey which are fairly distributed over the entire catchment area.
References


Maathuis, B. (2007). digital elevation model extraction processing and parameterization for hydrology, ITC Educational material.


## Appendices

### Appendix 1: acronyms and abbreviations

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>ASTER</td>
<td>Advanced Space born Thermal Emission Radiometer</td>
</tr>
<tr>
<td>DEM</td>
<td>Digital Elevation Model</td>
</tr>
<tr>
<td>DTM</td>
<td>Digital Terrain Model</td>
</tr>
<tr>
<td>GCP</td>
<td>Ground Control Points</td>
</tr>
<tr>
<td>GIS</td>
<td>Geographic Information System</td>
</tr>
<tr>
<td>GPS</td>
<td>Global Positioning System</td>
</tr>
<tr>
<td>IDL</td>
<td>Interface Definition Language</td>
</tr>
<tr>
<td>LPS</td>
<td>Leica Photogrametric Suit</td>
</tr>
<tr>
<td>MoWR</td>
<td>Ministry of Water Resources</td>
</tr>
<tr>
<td>NASA</td>
<td>National Aeronautics and Space Administration</td>
</tr>
<tr>
<td>NMSA</td>
<td>National Meteorological Services Agency</td>
</tr>
<tr>
<td>RMSE</td>
<td>Root Mean Square Error</td>
</tr>
<tr>
<td>SRTM</td>
<td>Shuttle Radar Topographic Mission</td>
</tr>
<tr>
<td>USGS</td>
<td>United States Geological Survey</td>
</tr>
<tr>
<td>UTM</td>
<td>Universal Transverse Mercator</td>
</tr>
</tbody>
</table>
Appendix 2: Spatial and temporal predictions of soil moisture corresponding Figure 5.7

Figure A-2 Spatial and temporal prediction of soil moisture dynamics in Wetet Abay catchment during the course of simulation for the hydrograph of the year 2003 (a) date 16/05/2003 (dry period) (b) date 26/06/2003 (rising limb) (c) at the time when the runoff rate is at the peak rate (date 11/07/2003) (d) date 06/10/2003 (falling limb) and (e) at the end of the simulation period (date 31/12/2003)

[Note: that negative value of moisture deficit means saturation]
Appendix 3: topographic index maps from SRTM DEM

Figure A-3 Topographic index maps from SRTM DEM (a) 90m resolution (b) Resampled 150m resolution DEMs
Appendix 4: flow components of the total runoff for the 90m resolution DEM

Figure A-4 base flow and overland flow contributions to the total runoff in Wetet Abay catchment